GEOSEA V Proceedings Vol. II, Geol. Soc. Malaysia, Bulletin 20, August 1986; pp. 1–19

Global tectonics and resources

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Abstract: The formation of most metallic resources involves the focussed flow of very large (km^3) fluid volumes. Different types of fluids, sea water, metamorphic fluids, meteoric fluids, with or without biomolecules, may be specific for certain metals or suites of metals. Whenever large fluid volumes are involved in a geologic process there is a potential for local metal enrichment. During weathering processes, short-term leaching produces fertile soils, long-term leaching produces laterite.

INTRODUCTION

Virtually all our resources are the result of processes that occur in the top 10 km or so of the earth's crust. Almost all our resources are the result of interactions between surface rocks and the atmosphere, hydrosphere and frequently the biosphere. Plate tectonics involves processes where new crust is formed mainly at ocean ridges and in volcano-plutonic arcs above subduction zones while some older crust is recycled to the deep mantle at subduction zones. New primary crust is dominantly igneous and dominantly formed of extrusive basalt or andesite. This primary crust rarely produces a useful resource until it is changed by some form of interaction with the hydrosphere. There are a few important exceptions, the formation of a chromite or platinum deposit in a layered ultramafic body being possible examples.

In this paper, I wish to focus mainly on resources where formation involves interactions between a moving fluid and rock. Such resources include everything from fertile soil to a quartz vein with gold. In general, all the processes involve similar factors of chemistry and flow dynamics.

An ore deposit is simply a volume of crust where a given element occurs at a higher concentration than in common primary rocks. The necessary concentration factor varies with the primary abundance of the element (e.g. 5x for iron, 200x for copper, 10,000x for gold) and in a general way the value of an ore is related to abundance and the concentration factor.

To produce an ore there must be mass motion, a given element may be transported from a rock and deposited in a more concentrated form or the bulk of the rock may be transported leaving a more concentrated residue. The former is the case for gold and copper, the latter is common for aluminium. We can consider all such processes as ***chemical transport processes.*** All such processes require a transport medium and a driving force for flow. For gold we normally consider the medium to be an aqueous fluid driven by thermal energy; for a bauxite again we assume aqueous fluid with flow driven by gravitational energy.

My thinking about this process was greatly influenced by work in the chemical industry on chemical transport processes (see Schafer, 1964; Fyfe, 1977). We consider a simple example of transport as shown in Figure 1. Two chambers which are held at different constant temperatures are connected by a path which allows diffusion between the chambers. One chamber contains an impure solid and they both contain a gas or fluid phase. The solids will have a finite vapour pressure so that components will enter the vapor phase. But if the vapour

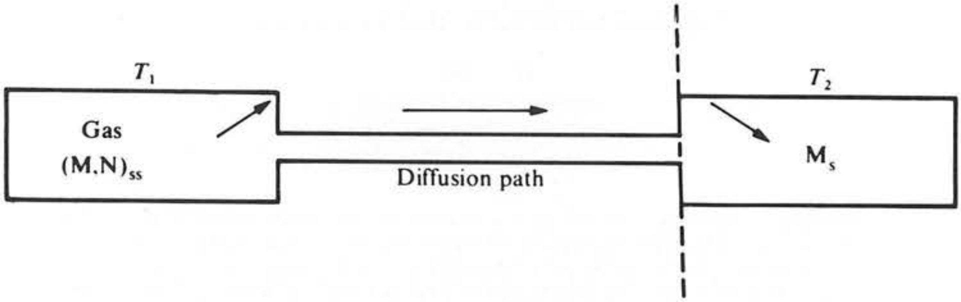


Fig. 1. A typical system involved in a chemical transport purification process. Transport rates for all chemical species and even isotopes must lead to purification.

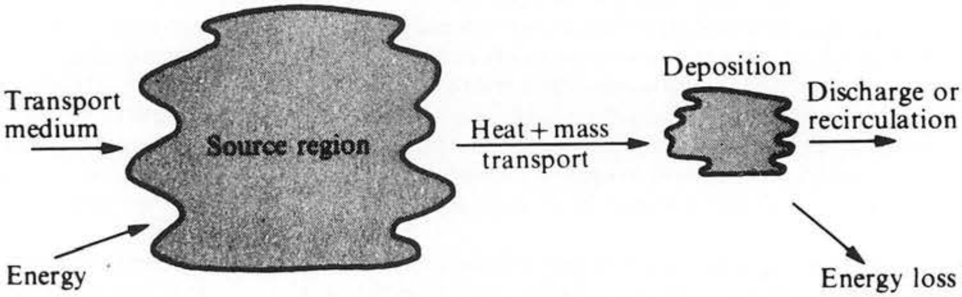


Fig. 2. The typical transport proceses associated with ore formation. The geologic case can be considered as an array of systems as in Fig. 1.

is not chemically inert, the concentrations of the solid components in the vapour will normally be enhanced. Depending on the thermal regime in the entire system and the molecular chemistry in the fluid phase, certain components will tend to diffuse away from the source region and may precipitate in the hotter or colder chamber. Consider a simple example: chamber A at T_1 and chamber B at T_2 with $T_1 > T_2$. Let the solid be a mixture of SiO_2 and Au. If the fluid is gaseous chlorine, we know that gold forms volatile molecules $(AuCl_3)_n$ while SiO_2 is unreactive. Gold ($AuCl_3$) will diffuse away from T_1 and precipitate at T_2. The precipitated gold will be of higher purity than the original but the residue at T_1 will be nearer pure SiO_2. If we replaced the chlorine gas phase with H_2O, it would be silica that would move to produce a higher purity SiO_2 in the container at T_2. Two important factors appear from this simple case. In any case where a fluid moves through a complex solid, there will be separation and purification. It is inconceivable that all components would have the same concentrations and diffusion properties in the fluid phase. Further, the chemical nature of the fluid may be highly specific for the transport of given components and a chemist would use his knowledge of molecular chemistry to choose the fluid to selectively transport a chosen element. Such specificity may play a very large part in ore-forming processes. Such processes are so specific that they can be used to separate even heavy isotopes.

It is not difficult to write down the terms which will control the ***quantitative success*** of the experiment (see Schafer, 1964). Clearly transport will be proportional to:

— the gas ⇔ solid chemistry
— concentrations in the gas phase as a fraction of T.
— diffusion velocities and dimensions of the diffusion path
— time
— energy parameters which maintain the system

The geologic case is more complex (Figure 2) but one can consider a very large array of chambers, each at a different T. To extract an element from a given rock we must consider:

(a) the extractability of a given rock (i.e. the nature of its permeability to a fluid, via cracks, porosity, etc.)
(b) the nature of the energy sources which drives flow (gravity and thermal energy)
(c) the nature and mass of the fluid, the degree of its chemical specificity
(d) the nature of flow, single pass and multipass recirculation
(e) concentrations in the source rock
(f) the geologic structure which controls the flow regime and flow focussing
(g) the nature of thermal gradients which will influence concentration gradients
(h) lithology along the flow path, the influence of wall rock in modifying the fluid chemistry (e.g. a change in local PO_2)
(i) the possibility of mixing of fluids in different flow cells
(j) the problems of preservation of an ore once it forms
(k) the precipitation mechanism

But I would again emphasize conclusions from the simple case of Figure 1. For this example a steep thermal gradient and a fluid which is chemically specific will be important. It is no accident that so many ore deposits are formed in regions with igneous activity and high heat flow and thermal gradients.

In what follows I wish to look at tectonic processes and environments where ores might form. Certain fundamental questions must be kept in mind. Are the source rocks sufficiently permeable to allow efficient extraction? Is the necessary solvent volume available and what is its chemistry? Can there be locally enriched source rocks? What is the scale of local energy sources?

Studies show that many of the metals of great economic interest are present in natural hot fluids at or below the ppm level. Thus if we are considering the formation of ore deposits with a million tons of mental 10^{12}g) the fluid volume required will be of the order of 10^{18} g (10^3 km^3). Note that this is about the annual flow of the Yangtze River and this is the mass which must pass the ore zone. We are thus searching for very large focussed flow paths in the crust.

If the flow is gravity driven, it will obey some form of Darcy's law when Q(flux) = KIA with K the permeability, I the hydraulic gradient and A the cross section of the flow path.

If the flow is thermally driven we will need to consider factors which influence the Rayleigh number of the system and the adiabatic gradient. The Rayleigh number tells us about the vigour of convection while the adiabatic gradient tells us whether or not convection will occur, it must be exceeded for convective flow.

Straus and Schubert (1977) have shown that the adiabatic gradient for porous medium convection of water is greatly exceeded when gradients exceed 20°C km^{-1}. For a porous or cracked medium, the Rayleigh number for convection is, $Ra = k\beta\, \Delta T\, g\, H/K_m\, \upsilon$ where K is the permeability, β the coefficient of expansion of the fluid, ΔT the temperature gradient, g the gravitational acceleration, H the thickness of the permeable layer, K_m the thermal diffusivity and υ the kinematic viscosity. Large Ra's and vigorous convection require high permeability, large T gradients, thick units, while convection is opposed by a thermally conductive high viscosity fluid. For most volcanic situations, large Ra's are found (Fyfe and Lonsdale, 1981). The same is true when igneous intrusion into any wet, porous material occurs.

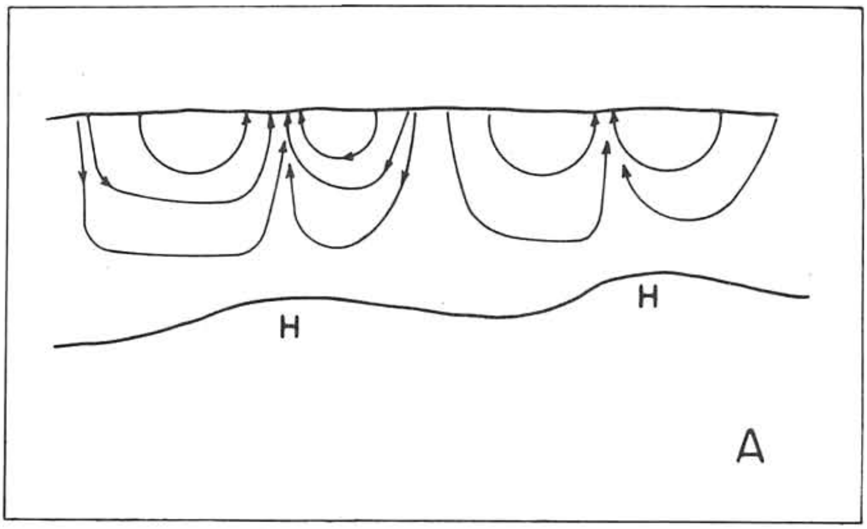
In what follows I would like to examine some types of geologic situations where fluid movement involves potentially focussed flow of fluid volumes in the multi-km^3 range. In all such cases there must be resource potential. Some of these situations are rather well understood but some are not. In each case I also wish to discuss what features in a rock may tell us that such a process has occurred and assist in finding the much smaller enriched volume.

OCEAN RIDGE SYSTEMS

Convective processes in the modern earth sweep almost 50 per cent of the planet's internal energy to ocean ridge sites where new crust is formed at a rate of about 12 $km^3\ a^{-1}$. This new crust is basaltic and made up of the typical ophiolite structure of lavas, dykes and intrusive gabbro bodies (Emiliani, 1981). The crust is formed in a dominantly submarine environment by cooling material introduced at about 1200°C. When any melt cools it must contract (ρ basalt lava $\cong 2.7\ g\ cm^{-3}$, ρ basalt $\cong 3.0\ g\ cm^{-3}$) and for basalt the cooling can produce pore space of about 10%. Most of this space will be in the forms of cracks or joints. The permeability of the freshly cooled crust will be large near the surface but will be reduced at deeper levels in the intrusive zones by plastic deformation. The cooling and crack propagation process has been discussed by Lister (1972).

In the near ridge environment consideration of Rayleigh numbers shows that convective cooling must occur. Studies of heat flow near ridges (Wolery and Sleep, 1976; Ribando *et al.*, 1976) have clearly shown that large multi-km convection cells are set up with massive input of sea water which carries away almost 50% of the heat of the igneous bodies, about 4×10^{19} cal a^{-1} on average. It is now also well established that sea water penetrates several km into the ocean crust and may even reach the Moho (Lewis and Snydsman, 1977, McCulloch *et al.*, 1980).

The scale of the process is impressive. Given that the ocean mass is 1.4×10^{24} g, and given that on average sea water involved in cooling is heated to 100°C, then it takes only 3 million years to process the entire ocean mass through ridges. As with most geothermal systems (see Elder, 1976) while recharge of water into the crust involves large areas and volumes, the discharge may be highly focussed and now, we are all familiar with the famous black smokers that have been found at ridges (Edmond and Von Damm, 1983). Large magma chambers



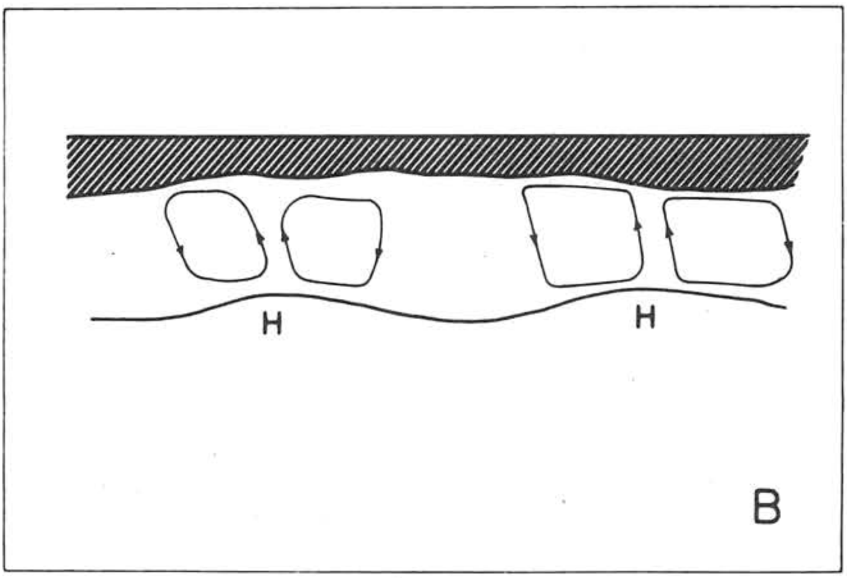


Fig. 3. Styles of convection cells involved in cooling young oceanic crust. In A, sediments are not present and the discharge vents to the sea floor perhaps producing a black smoker. Cells may have dimensions of 10 km or so. In B, an impermeable sediment blanket confines major circulation to the igneous layer. In this case, stratabound deposition may occur where cells rise. It is possible that to form major ore deposits, some type of cover is required to prevent dispersion of metal-bearing fluids.

beneath ridges have been found by seismic methods and may have dimensions of 50 km^3 or so. The energy from cooling such a body to say 300°C ($\cong 6 \times 10^{19}$ cal) could heat 200 km^3 of sea water to 300°C. At such a temperature the copper or zinc content of discharged fluids will be at the ppm level, so given appropriate precipitation mechanisms during cooling, there is potential for metal transport at the million ton level. Without doubt, the typical ophiolite copper-zinc deposits, and Mn-Fe oxide deposits, have formed in such a process.

But I would stress that the cooling igneous body derives its permeability mainly from cooling cracks. If the crack separation is large, the ability of the circulating fluid to extract metal may be limited to small volumes near crack walls. This process may be reasonably efficient for metals like Cu or Zn present in the rock at the 100 ppm level, but may be of limited efficiency for metals like gold at the ppb level. Ophiolite alteration can be highly variable and is not pervasive in the host rock.

Energy balance shows that on average, each gram of the rocks of the new ocean floor crust is influenced by something like 10 g of sea water during the cooling process. This flow must produce alteration which can be tracked (see Figure 3).

When oxygenated sea water enters hot basalt, during the recharge cycle certain dominant processes occur (Fyfe and Lonsdale, 1981):

(a) the basalt is hydrated forming minerals like clays, chlorites, amphiboles, epidote, serpentine depending on the local temperature
(b) oxygen is fixed in phases like magnetite and hematite
(c) sulphate and carbonate are precipitated as anhydrite or pyrite and calcite respectively
(d) the basalt is enriched in metals like Na, K, Mg, U, Rb
(e) silica is leached during recharge and precipitated during discharge
(f) transition metals (Fe, Mn, Cu, Zn, Ag, etc.) are leached during recharge and may be precipitated during discharge as sulphides or oxides
(g) there are important isotope exchange reactions e.g. ^{18}O is normally enriched during alteration while $^{87/86}Sr$ ratio tends to change from its original value of 0.702 towards sea water values of 0.709.

The extent of all these changes will indicate the overall water/rock ratio in a given region and hence the chance of there being a major metal deposit.

Finding recharge zones is generally rather simple as in such regions there will be significant oxidation, pyrite-calcite precipitation, K-Na enrichment, etc. Regions of discharge will be shown by high degrees of reduction (the hot fluids carry excess hydrogen) and silica veining (Figure 3).

In sea floor systems of type A (Figure 3) where the discharge vents directly to the sea floor, the black smoker system , there may be problems of preservation. The black smoke made up of sulphide particles, may readily back-oxidize to sulphates, unless sedimentation rates are high, a situation not common in ridge environments far removed from a continental margin. A more interesting situation is that of B (Figure 3) where discharge is throttled by a sediment cover (see Edmond and Von Damm, 1983). The rising metal-rich and reduced hot waters may interact with carbonate sediments and sulphate-bearing waters to precipitate sulphides which are protected from dispersion. In a sense this same process has occurred in the Red Sea where dispersion is retarded by the high density of the deep discharge brines. Barriga (1983) has demonstrated that in the case of the giant massive sulphides of Aljustrel, Portugal, that precipitation occurred under a thin veneer of siliceous sediments.

Edmond and Von Damm (1983) stress that dispersion of elements enriched in vent fluids may explain the patterns of metal rich sediments seen in the South Pacific. As I will discuss below, considerable volumes of these deep sea sediments may later be subducted and the sediments may become involved in subduction zone magmas and ore-forming processes.

While there has been great recent attention to ridge processes and in ophiolitic rocks, the same cooling processes must influence magma chambers in "hot spot" ocean islands, particularly when they are at a stage of evolution where their magma chambers are submarine. But such islands tend to erode fast as the moving plates pass over the mantle plumes and the root zones of the island sink. But as there may be a tendency for islands to "scrape off" during subduction, some so called ophiolite fragments with their deposits, may represent such ocean islands and not normal ocean floor.

THE CONVEYOR BELT

Conductive cooling theory would require that the thermal anomaly at a ridge would decay in proportion to $(age)^{1/2}$. While it is clear that in older ocean crust the heat flow approaches that expected for conductive cooling, there is still evidence (Anderson *et al.*, 1979) for mild convection beneath the sediment that blankets older crust (Figure 3B). It is also possible that additional heat sources may be involved and in particular highly exothermic serpentinization of deep ultramafics (Fyfe and Lonsdale, 1981). Study of the nature of the Moho in very old crust may indicate that deep levels are serpentinized. It is interesting to note that if serpentinization is slow and fluid penetration slow, the process peridotite + sea water $\rightarrow$ serpentine + salt could lead to the formation of brines with extreme salinity In general, metal solubility increases with salinitiy (Andrews and Fyfe, 1976). It is thus possible that very slow circulation under sediments may lead to "stratabound" mineralization at the sediment-basalt interface, a possibility which would be strongest where sedimentation rates are high. Ideal situations may be provided by the Juan de Fuca Ridge system (Davis and Lister, 1977) where very steep thermal gradients are found under a thin sediment cover or in the Guyamas Basin (Edmond and Von Damm, 1983). But such sediments may be subducted (see below). But where one finds examples of sea flow basalts with pelagic sediment cover of similar age, there could be good reason to search for metal-enriched sediments near the interface.

SUBDUCTION

Present plate tectonic models assume that virtually all ocean floor crust created at the ridge systems is eventually subducted (Hallam, 1976). Except for a few obducted ophiolite complexes, there is little ocean floor crust older than 200 million years. This means that the altered "spilitic" crust is almost quantitatively returned to the mantle. In this process volatile phases such as H_2O, CO_2 ,S are recycled into the mantle along with crust-hydrosphere phases such as K, Rb, Na, U and the like.

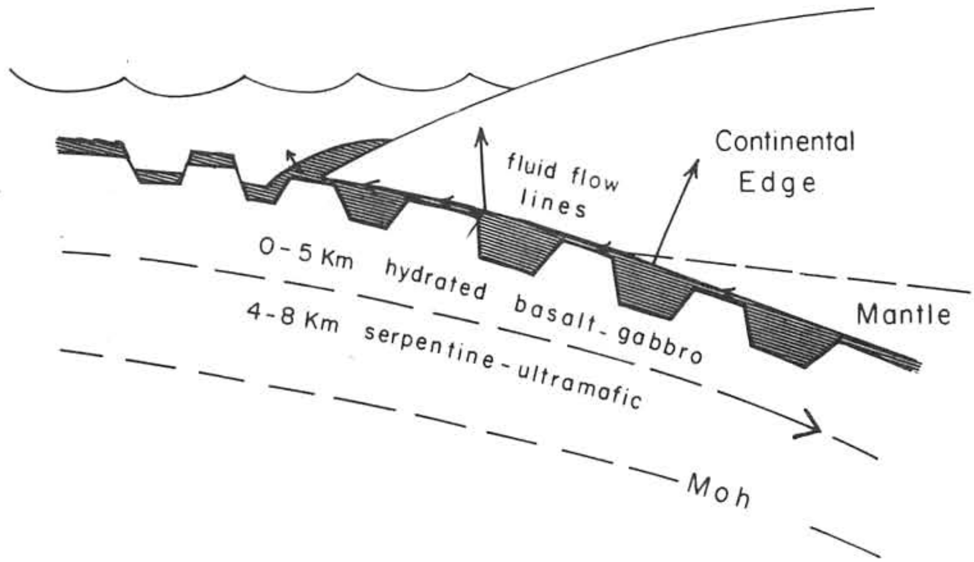


Fig. 4. Structure of the oceanic crust in a subduction zone. The bending lithosphere forms horst and graben structures which trap sediments. As the ophiolites and sediments are compacted, fluids flow out as shown by arrows. High water/rock ratios are most likely to occur in horst tops (A).

Recent studies in the environment of ocean trenches (Hilde and Uyeda, 1984; Uyeda, 1983; Fyfe, 1984) shows that as the oceanic lithosphere bends in the trench environment it cracks and develops a horst and graben structure with graben depths of a km or so. These grabens may contain rather undisturbed pelagic sediments as do the horst tops (Figure 4). During the subduction motion, as the plate underthrusts, grabens must fill with scraped-off sediments or, if there is not enough sediment to fill the tectonic roughness, the overplate may be eroded, a process Uyeda terms tectonic erosion. The extent to which a concretionary sediment wedge forms may depend on the balance of sediment input and plate roughness. The process has been called by Hilde, the "buzz-saw" model of subduction.

The implications of such observations from trench topography, seismology and trench wall drilling are profound. The combination of ridge faults and subduction horst-graben topography, provides a mechanism for locking sediment into the oceanic plate and providing the mechanics for their deep subduction. There is increasing evidence for recycling of sediments into the deep mantle (Turcotte, 1982; White and Hoffman, 1982) and as stated by Allegre *et al.* (1984) sediment loss may be of the order of 1-2 km^3 yr. At this present rate the continental crust (1.6×10^{25}g) will be recycled through the mantle in less than 4 billion years, a continent the size of Australia in 350 Ma (Fyfe, 1983).

In the context of this paper it must be stressed that large volumes of metal-enriched sediments or even sea floor massive sulphides will be subducted along with wet ocean crust.

Both the metal anomalies and rocks with high fluid content are associated. Further, if a metal anomaly exists on a continent which is shedding debris into a trench environment, the same anomaly may be recycled in the same geographic position with repetition of a certain type of geochemical anomaly. This could be of great interest to patterns of tin mineralization which often appear to be localized through time (Fyfe, 1983).

The fluids which are carried down with the subducting material must be recycled. For example, given the present subduction rate of spilitic crust ($\cong 4 \times 10^{16}$g a^{-1}) containing water in clays, chlorites, epidote, amphibole, serpentine, etc., if we assume an average bound water content of 5%, the water subduction rate is 2×10^{15} g a^{-1} (2 km^3). As the ocean mass is 1.4 $\times 10^{24}$ g, this implies recycling the ocean mass in about 700 Ma. Clearly there must be a return flow or the oceans would vanish. But such considerations do raise the interesting question as to whether oceans will grow or shrink if mixing continues in a planet which is cooling. It is virtually certain that given cooling, more water will remain in the upper mantle and oceans may slowly diminish in mass (Fyfe, 1984).

As the lithosphere descends, pore fluids will be rapidly expelled and pulses of fluids will be produced as the wet zone undergoes progressive prograde metamorphism (see Fyfe and Kerrich, 1984). Expulsion of such high pressure fluids has been directly observed during trench wall drilling (Anderson, 1981). The same fluids may cause the spectacular metasomatism seen in blueschist terraines. But ore deposits do not seem to be common in regions near the trench axis. The reason for this may be simple. Normally, thermal gradients near the trench are low, 10°C km^{-1} or less, as shown by blueschist mineralogy. Thus fluids in the upper part of the subducting system are cold. While flow may be focussed along the top of the descending plate and may lubricate the thrusts, the low thermal gradients would lead to highly dispersed precipitation and only slight mental enrichments.

A most interesting situation must arise if, for some reason, subduction ceases leaving a cold slab hanging at high levels, a situation which could result in the case of continental collision (see Figure 5). In this case a wet slab with ophiolites and sediments would appear underthrust say 50 km. The temperature at depth A (Figure 5) might be about 300°C in the blueschist facies while the upper parts would be in the prehnite-pumpellyite and zeolite facies. If underthrusting ceases, the entire block will tend to reach normal thermal gradients (20°C/km) and high grade amphibolites or even granulites will develop in much of the deeper parts.

Consider processes in the slab along A-B in a 10 km thick block of ophiolites and sediments, along a strike length of 1 km. The deeper 20 km or so of the slab may reach the amphibolite facies during thermal equilibration and may lose about 3% H_20. Thus the mass of water evolved from this region, $20 \times 10 \times 1$ km, will be of the order of 2×10^{16}g (20 km^3) per km of strike length along the suture. Some of the fluid may move by hydrofracture mechanisms into the overplate, while some may move up the thrust plane A-B. The metamorphic fluid evolved at temperatures between 300-600°C, will carry silica and metals like gold and silver (see Fyfe and Kerrich, 1984-b). Silica precipitated along a flow path like A-B in response to falling temperatures could approach 1% of the solvent mass ($\cong 2 \times 10^{16}$g) and could form a vein or sheet 5 m thick. The high temperature fluids would also carry large volumes of CO_2 from silicate-carbonate reactions in the prograded ophiolites and sediments.

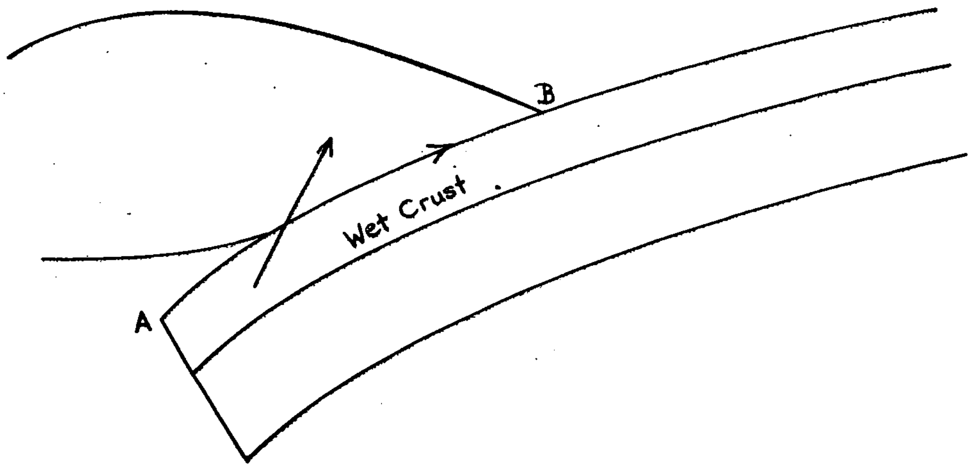


Fig. 5. A section of wet ophiolitic or sedimentary crust is left underthrust when subduction ceases. As this region warms up to normal thermal gradients, fluids will be released along tracks such as A-B.

Is this the type of phenomena that created the famous Mother Lode gold deposits of California? Recent seismic studies in this region (Nelson *et al.*, 1984) and geologic evidence of ophiolites and blueschists, suggest that a suture zone is involved. But what appears clear from the above analysis is that all suture zones deserve careful study for they may be the sites of massive hot fluid release with structures to focus flow.

Given continuous subduction,as the wet ophiolitic regions of the descending material eventually pass into dry eclogite facies rocks, fluids must flow either along the thrusts or into the overplate. Phlogopite may carry water to depths of several hundred km and must be a common mineral where K-enriched materials are sandwiched between ultramafic rocks of the oceanic lithosphere and overlying mantle (Wyllie and Sekine, 1982).

Given the thermal structures near the cold descending lithosphere, it is not yet certain whether the wet slab eventually melts or whether fluid release into the hotter over-lying mantle causes its partial fusion. But either mechanism will lead to the extraction of subducted soluble or easily fusible subducted materials. The ultra-high pressure fluids at temperatures like 600-1000°C, will transport large quantities of solute species like SiO_2, K, and trace metals. Such species will become enriched in basaltic melts which result from the water fluxed melting process.

As shown in Herzberg *et al.* (1983), the density of common mantle melts at Moho processes ($\cong$ 10 kbar) is greater than the density of continental crust. There is thus a tendency for mantle magmas to pond beneath, and underplate continental crust. There is ample evidence for the complexity of the continental Moho (Oliver, 1982; Meissner, 1973). The great root zones beneath the Andes may be caused by such underplate mechanisms. Given widespread development of mantle underplates at temperatures exceeding 1000°C, it follows that the crustal basement will suffer large contact aureole effects leading to ultra high grade metamorphism and partial melting producing melts of the granite family (Brown and Fyfe, 1970; Wyllie, 1977). Eichelberger (1978) has shown how basaltic-andesite and granite melts may mix at this interface and produce hybrid magma types. The same mixing process, and fractional crystallization of underplate magmas can provide a mechanism for transport of metal-enriched components to higher crustal levels and in turn produce volcanic-plutonic sources of anomalous metal content (Figure 6).

Thus, subduction zone magmatism may involve a complex array of events leading to a locally metal-rich magma series:

(a) formation of metal enriched ocean floor ophiolites and/or sediments
(b) their subduction

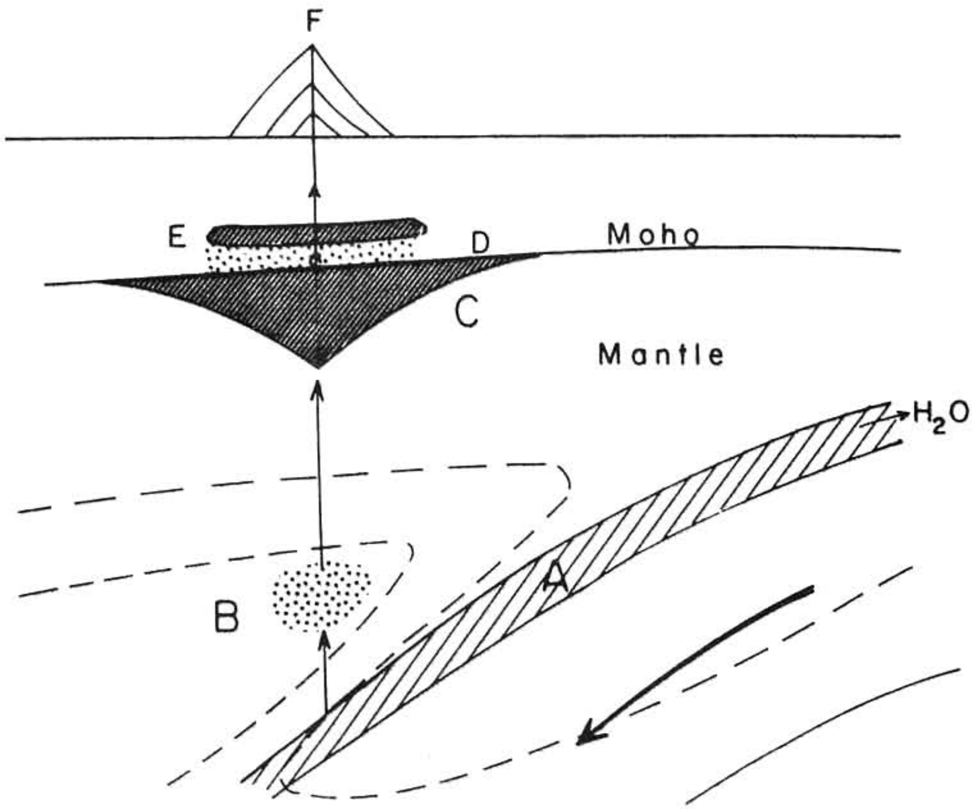


Fig. 6. Possible situation above descending oceanic lithosphere. Fluids released from the ophiolite zone A flux melting in the overlying mantle at B. Dense mantle magmas underplate the continental crust or intrude and extrude. The base of crust in region E-D melts producing granites or mixed-magma types.

(c) volatile-assisted transport to overlying mantle
(d) rise of volatile-fluxed mantle magmas to the continental Moho
(e) transfer of enriched late magma fractions to crustal melts
(f) high level intrusion and extrusion of anomalous magma batches

What I wish to stress, is that such sequences of events are the result of volatile fixation in oceanic lithosphere by near-ridge processes. They would not occur on a dry planet.

In the high-level crust above subduction zones where plutonism and volcanism occur, again high heat flow, and permeable regimes will be created by updoming tectonics and magma contraction. If plutonic bodies move into the zones of groundwater flow ($\cong$ 5 km), just as for ocean ridges, cooling will involve ground water circulation. The potential volumes are impressive. A typical granite pluton (Fyfe, 1973) has a volume of the order of the 600 km^3 In cooling from 900 to 300°C, the energy available for water heating is about 5×10^{20} cal. This could heat 1000 km^3 of ground water to 300°C. The potential for mineralization and focussed flow is enormous as is well shown by continental hot spring environments and geothermal power potential of such regions. Exactly what may result will depend on the level of emplacement, the magma compositions, the composition of the host rocks and the fluid composition, in particular its salinity, which may be highly variable.

The energy of a simple granitic pluton is large and the possibility of finding giant ore deposits is much larger than in present ridge environments where gabbroic plutons are probably an order of magnitude smaller. The large Cu-Mo-W-Sn-Ag deposits associated with acid magmatism are as might be expected.

But all high-level continental ore deposits formed in the thickened crust and high topography above subduction zones will be subject to rapid erosion and such deposits should hence be most common in young regions with limited removal of cover. I would note here that there is increasing evidence to show that in Archaean times, many, if not most, granitic plutons were emplaced in submarine environments. This would account for the development of the much larger "exhalative" sulphide deposits. Exactly the same processes would occur as at a modern ocean ridge, but the granitic batholithic heaters would focus more energy at a localized site at a given time. It is interesting to note that if we take typical plutonic energy, and metals like iron at a level as for the Salton Sea brines of California, >2000 ppm (Barnes, 1979), a single large pluton could mobilize 10 billion tons of iron.

HIGH GRADE METAMORPHISM

The development of extensive terrains of high grade metamorphic rocks of the amphibolite-granulite facies requires abnormal geothermal gradients, crustal thickness or heat flow into the base of the crust (Fyfe *et al.*, 1958). Clearly, magma underplating may provide a common condition to promote such processes. In subduction zone regions above the zones of mantle melting, abnormal heat flow is normally associated with partial fusion at the base of the crust. Given that a mantle magma underplates the crust, structures so brilliantly described by Ramberg (1967) will develop by the rise of plutons. Consider for example, a situation as shown in Figure 7 (c.f. Ramberg, Figure 67, p. 115). Consider a normal section of crust about 30 km thick. The metamorphic grades before underplating may be amphibolite near the base grading into a large mid-crust zone of greenschists. A dense basic magma

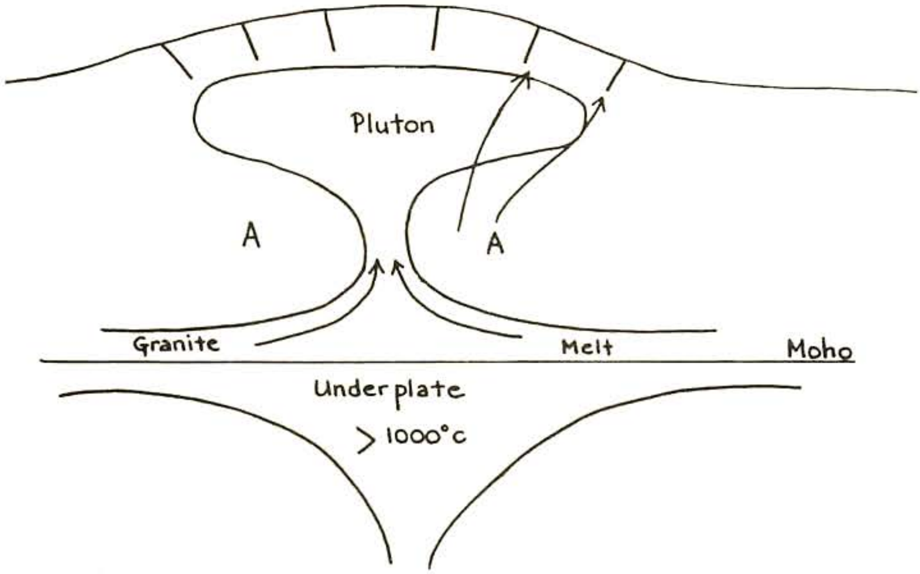


Fig. 7. Crustal section above a region of magma underplating. First, a pluton rises and later the large mid-crust region A progrades to the amphibolite facies.

underplates the base of the crust and a large thermal aureole is developed eventually leading to melting and development of a granulite residue. Light, viscous acid magmas rise ($\rho \cong 2.4$ g cm^{-3}) and form a high level pluton ($\rho \cong 2.65$ g cm^{-3}) at a depth of say 5 km. Extensional cracking in the roof leads to water cooling by near surface waters and hydrothermal deposits result.

In the large zone A, the greenschist zone (T 300-400°C) is now heated from below and above by the melting process and at a time following intrusion, is upgraded to amphibolite facies at temperatures near 500°C. In zone A, a huge degassing event is initiated.

The typical volume of zone A may be of the order of 1000 km^3. The initial greenschist facies rocks might have a typical water content of 4% while the product amphibolites 1%. The water mass evolved during this metamorphic event could reach 10^{17} g, or 100 km^3. The low salinity metamorphic fluid would be dominated by H_2O with an appreciable quantity of CO_2. It is exactly the type of fluid found in most lode gold deposits (Fyfe and Kerrich, 1984-b).

But this metamorphic fluid has been evolved on almost every grain boundary in the primary greenschist. Every epidote crystal, every chlorite crystal, has lost water. If the fluid has solvent capacity for a trace metal like Au or W, the mechanism provides an extremely efficient extraction process. The entire rock mass is pervasively influenced by the evolving fluid which creates transient high porosity (Fyfe *et al.*, 1978).

The crustal structure is also ideal for focussing of the flow of the metamorphic fluids. Sub-vertical structures will be created in zone A by the downflow of the crust around the rising plutons. As the pluton itself cools and contracts, permeable pathways may be created near the pluton borders or even in the pluton itself. Given that the solubility of a species like gold is a function of T as with silica, gold-quartz vein deposits may form. As discussed in detail by Fyfe and Kerrich (1984-b) many of the current facts about gold deposits fit a model of metamorphic degassing, often associated with granitic plutonism. Precipitation temperatures in the greenschist facies range and oxygen isotope systematics of fluids confirm the model. The frequent occurrence of gold vein deposits in Archaean and younger rocks of granite-greenstone terraines appears related to this type of process and the gold deposits are often within the granites. But I would stress, that any major prograde metamorphic events likely to be highly efficient in mobilizing elements which have low background concentration levels. While such processes will be most frequent above subduction zones, they may also occur where any "hot spot" attacks continental crust.

CONTINENTAL THRUSTS

At the present time there is a great renewed scientific effort to describe the deep structure of continents by high resolution seismic techniques. Groups like COCORP in the U.S.A. (see Cook *et al.*, 1980) and BIRPS in the UK (see Matthews, 1982) are producing impressive data. But one dominant result coming from such studies is that at deep crustal levels, low angle thrust structures are ubiquitous. Given the classic work from the Alps (Heim, 1921; Trumpy, 1969) this should come as no surprise. Processes which build continents (and destroy them) are often related to and associated with thrust structures (Fyfe and Kerrich, 1984). Hubbert and Rubey (1959) discussed the mechanism of large low-angle thrust structures. They concluded that for such structures to develop, fluid pressures in the thrust plane must equal or exceed the lithostatic pressure; thrusts float on the thrust plane. Thus low angle thrusts require the generation of fluids for their formation.

What I wish to stress here is that whenever major thrusts occur, there is almost certainly release of large masses of fluids and hence the potential for equally large mineralization processes. I would like to use one example from our planet, possibly the greatest present thrust event.

The January 5-11, 1984, issue of *Nature* (*307*, 17-36) describes the impressive recent work from a French-Chinese cooperative program on the structure of the Himalayan thrust belt. This great thrust belt with dimensions E-W of almost 3000 km, and N-S 1500 km, involves an area of 4.5×10^6 km^2 where continental crust of almost double thickness has been created by the thrusting of parts of the Indian and Lhasa continental blocks beneath Asia. Seismic studies reveal a complex stepped Moho related to separate thrust structures.

The average thrust velocity is about 5 cm a^{-1}. This implies that a thrust zone 100 km in dimensions could form in 2 million years. Given the time constants of thermal equilibration, thrusting at such velocities can be considered isothermal.

Consider the situation of Figure 8 where a 30 km block is thrust over another 30 km block in a very short time. Immediately following thrusting, an inverted thermal structure will result with repeated metamorphic isograds (Figure 9). As thermal equilibrium commences, the

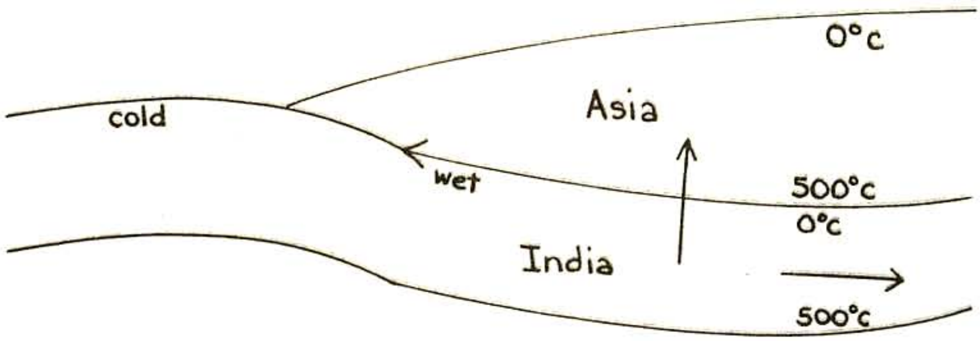


Fig. 8. Possible local Himalayan structure when double thickness continental crust is created by thrusting. The thermal structure is irregular. Fluids from the underthrust block must flow into the overthrust and lubricate the thrust plane.

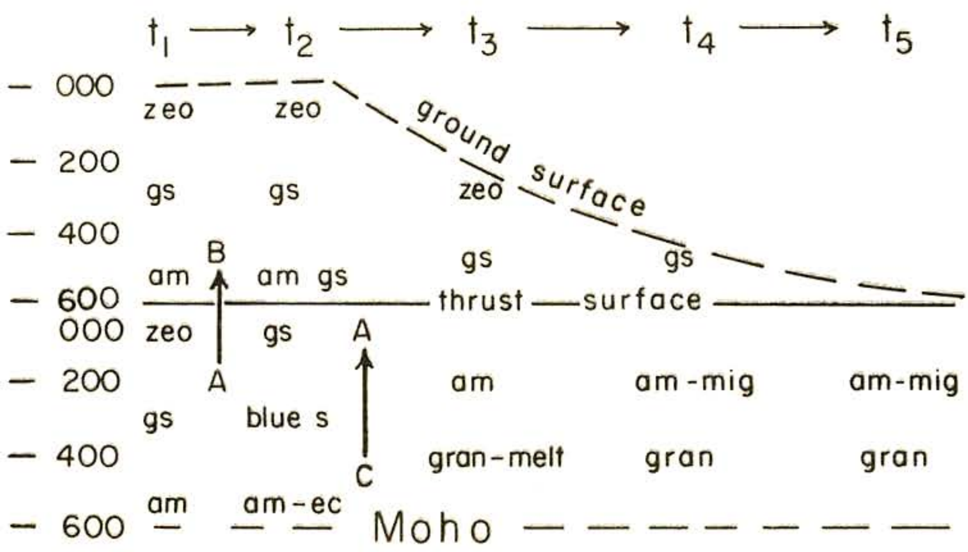


Fig. 9. Possible metamorphic patterns which develop following a large continental thrust event. At the time of thrusting (t_1), isograds are repeated. Facies changes at later times t_2, t_3 etc. are shown as the blocks approach thermal and erosional equilibrium. (zeo = zeolite facies; gs = greenschist; am = amphibolite; blue s = blue schist; ec = eclogite; gran = granulite; mig = melting).

entire underthrust plate will undergo prograde metamorphism with melting at the base. The young granites of the Himalayas with initial $^{87}Sr/^{86}Sr$ ratios up to 0.75 can be no surprise. These are truly granites formed by crustal fusion (Ferrara *et al.*, 1983).

If we consider the total Himalayan event with 4.5×10^6 km^2 of thrust surface, the prograded underthrust plate will have a volume of about 1.35×10^8 km^3. Given an average loss of H_2O of 4% from this volume corresponding to zeolite $\rightarrow$ greenschist, and greenschist $\rightarrow$ amphibolite transitions in the underplate, the fluid mass released is of the order of 1.6×10^{22} g or 1.6×10^7km^3 (Note, the world's ice caps contain 3×10^7 km^3 of water). This means that about 4 km^3 of fluid will pass through each km^2 of thrust surface. The potential for ore deposit formation and for geothermal power must be enormous (see Fyfe and Kerrich, 1984).

While such giant thrust processes are impressive, smaller events are common as revealed by recent seismic studies. A particularly interesting model has been described by Allis (1981) in the Southern Alps of New Zealand. It is proposed that relatively small degrees of thrusting ($\cong$ 100 km) may accompany a dominantly transcurrent or transform motion. Fluid evolution along large transform faults is well known and in fact may be required to lubricate such faults (Fyfe *et al.*, 1978). Any curvature on a transform must lead to local zones of periodic over or under thrusting and again potential for mineralization must exist.

Recently we have described a thrust structure in Brazil (Lobato *et al.*, 1983) where massive uranium mineralization is associated with a thrust which moved Archaean basement over Proterozoic sediments. Fluid motion caused the formation of a major uranium deposit of at least 100,000 tons.

FLUIDS, EROSION AND RESOURCES

Our present planet can be divided into the highly seismic regions which form the plate boundaries and the much larger regions between the boundaries. High elevations are created at the boundaries while regions of low elevation have been produced by long periods of erosion and sea level fluctuations.

Erosion occurs by removal of surface both by solution and transport of particulate matter. Given that average runoff contains about 100 ppm of dissolved mineral matter and that the total global runoff is 3.6×10^{19} g a^{-1}, erosion by solution alone moves about 3.6×10^{15} g of continental material to the oceans each year. A recent study of river sediment load (Millman and Meade, 1983) shows that suspended matter transport is of the order of 13.5×10^{15} g a^{-1}. Thus the total mass of continental material which is transported is about 1.7×10^{16} g a^{-1}. Given that the mass of continental crust is 1.6×10^{25} g, the present rates of erosion would rework this mass in one billion years. Erosion is a major tectonic process!

If we consider a low-lying stable area with no transport by particle transport but erosion by solution processes alone, and give 1 m of rain per year, land surface will be removed at a rate of 50 m per million years; if tectonic forces cease, the Himalayas would be dissolved to sea level in 100 million years!

It is this slow solution process with minimal particle transport that produces first fertile soil and eventually infertile laterites and bauxites (Fyfe *et al.*, 1983). In this case the development of resources is related to the accumulation of a residue of the least soluble phases.

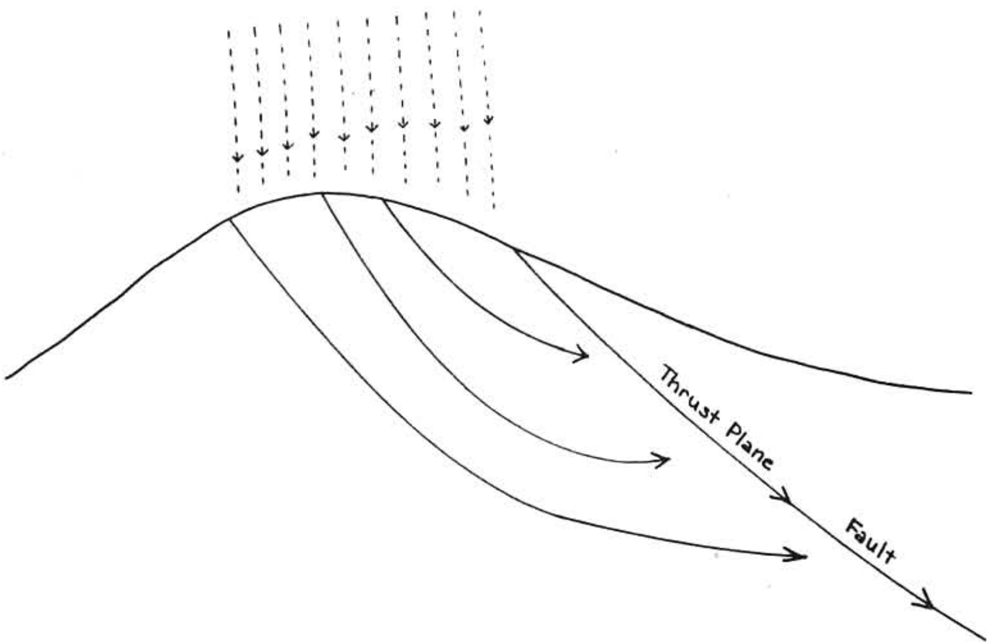


Fig. 10. Focussed motion of deep meteoric fluids caused by faults or thrusts may lead to zones of metal enrichment.

In a general way solution weathering proceeds via a series of steps:

primary rock → rock + smectite clays → smectite + kaolin → kaolin + gibbsite → gibbsite

Forming a fertile soil (rock + smectite) on a primary igneous rock may take a thousand years or more. To obtain a bauxite residue may take millions of years. Fertile soils always tend to occur in regions with plentiful fresh rock debris which contain the supply of essential plant nutrients.

This incongruent leaching process provides the global resources for aluminium. But many other materials also occur, from rare earth elements and phosphates from carbonatites, nickel or titanium rich laterites, gem-bearing laterites. In this near surface environment, transport of interesting elements often occurs by complexing molecules derived from biomolecules and a wide range of exotic materials may be so formed. Elements such as vanadium, molybdenum, copper, uranium, gold, platinum may accumulate in peats and coals. We know remarkably little about such phenomena but there is increasing evidence to show that gold nuggets in weathered terrains and some uranium accumulation may be associated with bacterial and algal processes (Beveridge *et al.*, 1983; Mann and Fyfe, 1984). There is little doubt that a host of potential resources may yet be discovered in this environment.

Another scenario which must be of interest involves the introduction of meteoric water into faults and thrust surfaces in high topography. Consider the scenario shown in Figure 10. Fluids penetrating the ground surface may be focussed to flow out on a thrust surface. Input fluids will be oxidized while at depth reducing conditions will attain. Elements more soluble in the oxidized regime (U, Au, Cu) will tend to precipitate along the major deep aquifer. Again, given a metre of rain per year, a km^2 catchment area, the fluid volumes so focussed may reach a km^3 in a thousand years or so. While there may be good examples of uranium deposits formed by such processes, I doubt if we have explored all such possibilities.

CONCLUDING STATEMENT

In this paper I have stressed that wherever large (km^3) volumes of fluid flow in focussed structures, whether the fluids are hot or cold, there is potential for resource formation. Fertile soils result from limited flow, bauxites from extensive flow. Probably most gold deposits result from the flow of metamorphic fluids, massive sulphide deposits from thermally driven circulation of sea water, etc.

The search for ore deposits follows fashions. Most economic geologists like the security of saying my deposit is of this or that type. Perhaps this attitude restricts our imaginations in exploration. The situation is, that given fluids and structure, there is always potential for new discoveries. I might predict that in the next decades we will see a multitude of new ore types associated with thrusts and deep weathering processes.

REFERENCES

ALLEGRE, C.J. *et al.*, 1984. Structure and evolution of the Himalaya-Tibet orogenic belt. *Nature, 307*, 17-22.

ALLIS, R.G.,1981. Continental underthrusting beneath the Southern Alps of New Zealand. *Geology, 9*, 303-307.

ANDERSON, R.N., 1981. Surprises from the Glomer Challenger. *Nature, 293*, 261-262.

ANDERSON, R.N., Hobart, M.A. and Langseth, M.C., 1979. Geothermal convection through oceanic crust and sediments in the Indian ocean. *Science, 204*, 828-832.

ANDREWS, A.J. and FYFE, W.S., 1976. Metamorphism and massive sulphide generation in oceanic crust. *Geosci. Canada,3*, 84-94.

BARNES, H.L., 1979. *Geochemistry of hydrothermal ore deposits*. 2nd edit., Wiley-Interscience, 798 pp.

BARRIGA, F.J.A.S., 1983. *Hydrothermal metamorphism and ore genesis at Aljustrel*, Portugal. Ph.D. thesis, University of Western Ontario, 368 pp.

BEVERIDGE, T.J., MELOCHE, J.D., FYFE, W.S. and MURRAY, R.G.E., 1983. Diagenesis of metals chemically complexed to bacteria. *Applied and Environmental Microbiology, 45*, 1094-1108.

BROWN, G.C. and FYFE, W.S., 1970. The production of granite melts during ultra-metamorphism. *Contrib. Mineral. Petrol., 28*, 310-318.

COOK, F., BROWN, L. and OLIVER, J., 1980. The Southern Appalachians and the growth of continents. *Scientific American, 243*, 156-168.

DAVIS, E.E. and LISTER, C.R.B., 1977. Heat flow measured over the Juan de Fuca ridge: evidence for widespread hydrothermal circulation in a highly heat transportative crust. *J. Geophys. Res., 82*, 4845-4860.

EDMOND, J.M. and VON DAMM, K., 1983. Hot springs on the ocean floor. *Scientific American, 248*, No. 4, 78-93.

EICHELBERGER, J.C., 1978. Andesitic volcanism and crustal evolution. *Nature, 275*, 21-27.

ELDER, J., 1976. *The bowels of the earth*. Oxford University Press, 222 pp.

EMILIANI, C., 1981. *The Oceanic Lithosphere*. Wiley-Interscience, 1738 pp.

FERRARA, G., LOMBARDO, B. and TONARINI, S., 1983. Rb/Sr geochronology of granites and gneisses from the Mount Everest region, Nepal Himalaya. *Geologische Rundschau, 72*, 119-136.

FYFE, W.S., 1984. Subduction and the geochemical cycle.*Tectonophysics, 99* (in press).

FYFE, W.S., 1983. *Advances in understanding subduction and continental structure above subduction zones*. Proceedings conference on geology and mineral resources of Thailand. Dept. Mineral Resources, Thailand. 1-7.

FYFE, W.S., 1973. The generation of batholiths. *Tectonophysics*, 17, 273-283.

FYFE, W.S., 1977. Introductory remarks on the transport problem. *Spec. Pub. 7, Geol. Soc. London*, 1-3.

FYFE, W.S. and KERRICH, R., 1984. Fluids and thrusting. *Chem. Geol.* (in press).

FYFE, W.S. and KERRICH, R., 1984-b. Gold: natural concentrations processes. *In*: R.P. Foster (Ed.), *Gold '82: The geology, geochemistry and genesis of gold deposits*, A.A. Balkema, Rotterdam, 99-128.

FYFE, W.S., KRONBERG, B.I., LEONARDOS, O.H. and OLORUNFEMI, N., 1983. Global tectonics and agriculture: a geochemical perspective. *Agriculture, Ecosystems and Environment, 9*, 383-399.

FYFE, W.S. and LONSDALE, P., 1981. Ocean floor hydrothermal activity. *In* Emiliani, C. (Ed.), *The oceanic lithosphere*, Wiley-Interscience, 589-638.

FYFE, W.S., PRICE, N.J. and THOMPSON, A.B., 1978. *Fluids in the Earth's crust*, Elsevier, 383 pp.

FYFE, W.S., TURNER, F.J. and VERHOOGEN, J., 1958. Metamorphic reactions and metamorphic facies. *Geol. Soc. Am. Memoir 73*, 259 pp.

HALLAM, A., 1976. How clearly did the continents fit together? *Nature, 262*, 94-95.

HEIM, A., 1921. *Geologie der Schweiz*, Tauchnitz, Leipzig, 476 pp.

HERZBERG, C.T., FYFE, W.S. and CARR, M.J., 1983. Density constraints on the formation of the continental Moho and crust. *Contrib. Mineral. Petrol., 84*, 1-5.

HILDE, T.W.C., and UYEDA, S., 1984. Convergence and subduction. *Tectonophysics, vol. 99* (a special issue, in press).

HUBBERT, M.K. and RUBEY, W.W., 1959. Role of fluid pressure in mechanics of over-thrust faulting. *Geol. Soc. Am. Bull., 70*, 115-166.

LEWIS, B.R.T. and SNYDSMAN, W.E., 1977. Evidence for a low velocity layer at the base of oceanic crust. *Science, 266*, 340-344.

LISTER, C.R.B., 1972. On the thermal balance of a mid-ocean ridge. *Geophys. J. R. Astron. Soc., 26*, 515-535.

LOBATO, L.M., FORMAN, J.M.A., FYFE, W.S., KERRICH, R. and BARNETT, R.L., 1983. Uranium enrichment in Archaean crustal basement associated with overthrusting. *Nature, 303*, 235-237.

MANN, H. and FYFE, W.S., 1984. Selective uptake of barium and uranium by algae. *Chem. Geol.* (in press).

MATTHEWS, D.H., 1982. BIRPS: deep seismic reflection profiling around the British Isles. *Nature, 298*, 709-710.

MCCULLOCH, M.T., GREGORY, R.T., WASSERBURG, C.J. and TAYLOR, H.P., 1980. A neodymium, strontium and oxygen isotope study of the Cretaceous Samail ophiolite and implications for the petrogenesis and seawater and hydrothermal alteration of oceanic crust. *Earth Planet.Sci. Lett., 46*, 201-211.

MEISSNER, R., 1973. The "moho" as a transition zone. *Geophys. Surveys, 1*, 195-216.

MILLMAN, J.D., and MEADE, R.H., 1983. World-wide delivery of river sediments to the oceans. *J. Geol., 91*, 1-21.

NELSON, K.D., ZHU, T.F., GIBBS, A., HARRIS, R., OLIVER, J.E.,KAUFMAN, S. and BROWN, L., 1984. *Cocorp deep seismic reflection profiling in the Northern Sierra Nevada Mountains, California* (in preparation).

OLIVER, J., 1982. Changes at the crust-mantle boundary. *Nature, 299*, 398-399.

RAMBERG, H., 1967 *Gravity, deformation and the Earth's crust*. Academic Press, 214 pp.

RIBANDO, R.J., TORRANCE, K.E. and TURCOTTE, D.L., 1976. Numerical models for hydrothermal circulation in the oceanic crust. *J. Geophys. Res., 81*, 3007-3012.

SCHAFER, H., 1964. *Chemical transport reactions*. Academic Press, New York, 161 pp.

STRAUS, J.M. AND SCHUBERT, G., 1977. Thermal convection of water in a porous medium: effect of temperature and pressure-dependent thermodynamic and transport properties. *J. Geophys. Res., 82*, 325-333.

TRUMPY, R., 1969. Die helvetischen decken der Ostschweiz. *Eclogae geol. Helv., 62*, 105-142.

TURCOTTE, D.L., 1982. Geochemical light on mantle convection. *Nature, 296*, 487-488.

UYEDA, S., 1983. Comparative subductology. *Episodes, 1983*, 19-24.

WHITE, W.M. and HOFFMAN, A.W., 1982. Sr and Nd isotope geochemistry of oceanic basalts and mantle evolution. *Nature, 296*, 821-825.

WOLERY, T.J. and SLEEP, N.H., 1976. Hydrothermal circulation and geochemical flux at mid-ocean ridges. *J. Geol.*, 249-275.

WYLLIE, P.J., 1977. Crustal anatexis: an experimental review. *Tectonophysics, 43*, 41-71.

WYLLIE, P.J. and SEKINE, T., 1982. The formation of mantle phlogopite in subduction zone hybridization. *Contrib. Mineral. Petrol., 79*, 375-380.

Manuscript received 17th Sept. 1984