The role of metamorphic fluids in the formation of ore deposits

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Abstract: Many ore deposits are hosted by metamorphic rocks, and metamorphic fluids have been invoked as a source for various deposits, especially gold deposits. Metamorphic fluid compositions reflect original sedimentary environment: continental shelf sequences yield saline metamorphic fluids with little dissolved gas while metasediments from accretionary and oceanic settings host less saline fluids with significant CO2 contents.

The principal difficulty in reconciling ore deposits with a metamorphic origin is that many form quickly (c. 1 Ma), whereas metamorphic heating is slow (c. 10–20⁰/Ma). Gravitational instability means that fluid cannot be retained. Metamorphic ores may nevertheless form by: (a) segregation leading to enrichment of pre-existing concentrations; (b) infiltration of water-rich fluids from schists into marbles at high temperature overstepping decarbonation reactions and allowing fast reaction that locally draws down temperature; and (c) rapid uplift driving dehydration reactions owing to pressure drop.

Some orogenic lode gold deposits fit well with a purely metamorphic origin during rapid uplift, but others are problematic. At Sunrise Dam, Western Australia, anomalies in Sr-isotope ratios and in apatite compositions indicate a partial mantle/magmatic source. Low salinity, H2O–CO2 fluids commonly associated with hydrothermal gold reflect the effect of salt on gas solubility, not the origin of the fluid.

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Despite some clear evidence to the contrary (e.g. Roedder 1972), until recently crustal fluids were generally assumed to carry only small amounts of potential ore metal in solution. Thus the problem of understanding ore deposits was seen as one of accounting for extensive focussing of fluid flow. With the recognition in recent years that crustal fluids can carry very large dissolved loads (e.g. Yardley & Bottrell 1992; Heinrich et al. 1992; Campbell et al. 1995; Audetat et al. 2000; Yardley 2005; Heijlen et al. 2008; Newton & Manning 2008), this consideration is no longer such a dominant constraint for many metals, although some are always relatively insoluble. While attention has rightly focussed on mass balance considerations, the development of refined methods of geochronology has also highlighted the duration of ore-forming episodes, and in particular has demonstrated that the duration of ore-forming events is often very short in geological terms. Timing provides a further constraint for models of ore genesis, since the requisite volumes of fluid must flow or circulate for relatively short periods of time compared with the metamorphic evolution of the host rocks. This constraint is hardly novel to field geologists used to determining the relative,
rather than absolute, duration of ore formation within a complex regional terrane, but is now quantified for a range of deposit types.

**When and where do fluids occur in metamorphic rocks?**

Metamorphic rocks in a classic orogenic belt evolve from relatively porous sediments, with correspondingly high water content, to crystalline rocks with very little porosity. Once they have recrystallized to a low-porosity metamorphic rock they can accommodate significant (percentage) levels of fluid only if they are fractured or as a result of specific metamorphic reactions which generate secondary porosity. Such reactions are known from certain rock types, but they are not widespread and the enhanced porosity is very transient. Furthermore, far from being a ubiquitous source of fluid, metamorphic rocks have the potential to act as a sink, rather than a source, for fluid, absorbing water or other fluid species in retrograde reactions as they cool.

It summary, metamorphic rocks only release fluids during specific parts of their metamorphic history (Fig. 1), normally when they are being heated so that volatile-bearing minerals break down and are replaced by mineral assemblages with a lower volatile content. These reactions are strongly endothermic. During progressive heating, fluid is released and fluid pressure rises until the rock acquires sufficient permeability for the fluid to escape (Yardley 1986; Ingebritsen & Manning 2010). Conventionally, this is assumed to be at a value close to lithostatic pressure, and there is extensive evidence from both veins and from phase equilibrium calculations to indicate that this condition is widespread during prograde metamorphism. In contrast, on cooling retrograde reactions reverse the devolatilization and rapidly consume any remaining pore fluid so that the rock becomes essentially dry (Frost & Bucher 1996; Yardley & Valley 1997). After the onset of retrogression, the chemical potential of water or other volatile species in a rock is defined by the coexistence of peak metamorphic minerals and their retrograde products. Typically, the values are at least an order of magnitude lower than would be expected if a free fluid was present.

To summarize the fluid history of a reactive metamorphic rock, we can identify three distinct types of fluid regime. During initial prograde metamorphism, fluid is pervasively released by reactions into the grain boundary network. Further fluid loss may arise from reduction in porosity. The composition of the pore fluid evolves progressively as fluid is released from minerals and is lost by flow owing to its low density. Fluid pressures are very high, indicating that the rock is very impermeable. Except for very low-grade rocks then, irrespective of the temperature reached, the onset of cooling is generally accompanied by incipient retrogression, lowering fluid pressure, unless lithostatic pressure also drops markedly (see below). This effectively consumes all free fluid from the pore space. The rock is now dry and a potential sink for fluid, and any fluid that infiltrates along fractures or elsewhere will tend to react to form retrograde assemblages. Many high-temperature minerals are prone to rapid retrograde reaction with infiltrated water (e.g. olivine, orthopyroxene, Al-silicates, biotite), whereas others are much more sluggish to respond.

In the event of a subsequent orogenic episode, the response of a pre-existing metamorphic rock depends on the relative conditions attained in the two metamorphic episodes. Often, a later orogeny is marked by pervasive retrogression, deformation and shearing of earlier rocks and the rock continues to act as a fluid sink until it has been fully rehydrated under the new metamorphic conditions. The classic Laxfordian shear zones of the NW Highlands are an example of this pattern (Beach 1973). It is also possible for a later metamorphism to attain more extreme conditions and to drive further volatile loss.

This pattern of fluid availability during the metamorphic cycle (Fig. 1) provides the fundamental framework for understanding the role that

**Fig. 1.** Schematic representation of the changes in the water contents of a sequence of supracrustal rocks during an idealized metamorphic cycle. The horizontal axis represents time and temperature, with the ends corresponding to surface conditions and the middle to amphibolites facies metamorphism at c. 650 °C. Separate trends are shown for pore water and water combined in minerals, for pelite, psammite and basalt lithologies. Modified from Yardley (1996).
metamorphic fluids can play in ore genesis. During heating there is a possibility of ore-forming fluids being generated in metamorphism, but for much of their crustal residence history, metamorphic rocks are a potential sink for introduced fluids, and may host ores formed from external fluids that become sucked into them.

In more detail, there are differences arising directly or indirectly from the grade of metamorphism. During low-temperature but prograde metamorphism, commonly up to the garnet zone, rocks are still relatively fine grained and therefore have potential to carry relatively large amounts of fluid in grain boundaries. The amount of pore fluid may decline as rocks become coarser, but will also be influenced by reaction rate: the faster the rate of fluid production, the larger the porosity that may persist. Where salt is present, even in modest amounts, water and non-polar (CO\textsubscript{2}–CH\textsubscript{4}) fluids are immiscible, and relative permeability effects are likely to inhibit fluid loss. At higher metamorphic grades water is still produced but will often, although not invariably, become less saline (Yardley & Graham 2002), with less likelihood of two-phase behaviour. Metal concentrations may, however, be enhanced through the effect of temperature. At the highest grades, melt takes over the role of metamorphic fluid and so different patterns of behaviour result.

**Rheological implications of the metamorphic fluid cycle**

The formation of ore deposits requires fluid flow through rocks and along focussed pathways: the nature of such pathways is dependent on rock strength, which in turn is strongly influenced by water. The effect of high fluid pressures at metamorphic temperatures is to make rocks undergoing metamorphism very weak. They are unable to sustain significant fluid-filled porosity or cracks over vertical extents of more than a few tens of metres at most (Yardley 1986), because the density difference between fluid and rock will drive the fluid to migrate upwards as a 'porosity wave' (Connolly & Podladchikov 2004) through the weak host. Nevertheless, the low effective stresses that result from the high fluid pressure mean that rocks undergoing devolatilization readily crack in response to deviatoric stress, hence the abundance of syn-metamorphic veins in regionally metamorphosed, but not in contact metamorphosed, rocks. Despite their potential for brittle behaviour, rocks undergoing prograde metamorphism are most extensively deformed in a ductile manner, with pervasive strain accompanying folding and producing metamorphic fabrics. Deformation mechanisms are dominated by pressure solution (Rutter 1983) and by grain boundary processes, although minor intracrystalline plasticity is also widespread, especially in quartz.

The transition from fluid-saturated to dry conditions with the onset of cooling causes a marked change in the rheology of metamorphic rocks (Yardley & Baumgartner 2007). The water-weakening processes that affect crustal rocks undergoing prograde metamorphism require the presence of a fluid. Once the fluid has been consumed by retrograde reactions, such mechanisms are not possible, hence the rocks are much stronger and deformation tends to be localized rather than pervasive. At all but the lowest temperatures, once cooled crystalline rocks are fractured, infiltration of water from an external source can cause transient weakening and ductile behaviour where it penetrates, resulting in retrograde shear zones (Mancktelow & Pennacchioni 2005, Van Diggelen et al. 2010).

**The chemical composition and evolution of metamorphic fluids**

Classically, metamorphic fluids have been treated as mixtures of H\textsubscript{2}O and CO\textsubscript{2} with scant regard for their dissolved load. A few studies have used fluid inclusions to explore metamorphic fluid solute composition (e.g. Crawford & Hollister 1986; Yardley & Graham 2002), and experiments have investigated cation exchange between minerals and inclusions to explore metamorphic fluid solute composition (e.g. Crawford & Hollister 1986; Yardley & Graham 2002), and experiments have investigated cation exchange between minerals and fluids under metamorphic conditions (Orville 1972; Eugster & Gunter 1981; Shmulovich & Graham 2008). The dissolved salt load is particularly significant for its effect on CO\textsubscript{2}–H\textsubscript{2}O phase relationships (Liebscher 2007), potentially resulting in fluid immiscibility over a wide range of metamorphic conditions (Yardley & Bottrell 1988; Heinrich 2007). While there are circumstances where metamorphic fluids are dilute H\textsubscript{2}O–CO\textsubscript{2} fluids, this is far from being the universal situation, and metamorphic fluids can vary between very dilute and salt-saturated end members.

Aluminosilicates are not very soluble in pure water under most crustal metamorphic conditions, irrespective of salinity, but just as surface waters and shallow formation waters can carry very varying amounts of dissolved solids, so too can metamorphic fluids. Like shallow waters, however, the make-up of the dissolved load is not particularly 'rock-like'. Experimental studies indicate that levels of Si in silicate-saturated waters are strongly dependent on temperature and moderately dependent on pressure (Newton & Manning 2000), so under hot lower crustal conditions, concentrations can rise to over 1% (Shmulovich et al. 2006), although even here the total dissolved load
is likely to be much greater than this (Yardley & Graham 2002). Even under modest conditions, sufficient Si enters solution for silicate minerals to be able to segregate readily into fractures to form veins. Al concentrations in metamorphic fluids have until recently been thought to be extremely low. However recent experimental work (Newton & Manning 2008) has demonstrated that, while corundum solubility in pure water is low, Al is able to form aqueous complexes with Si and Na so that concentrations in brines equilibrated with Al-silicate assemblages may be much higher. In contrast, Si solubility is not greatly affected by the presence of Na salts (Shmulovich et al. 2006), although it is significantly reduced in the presence of CO₂.

While the aluminosilicate contents of metamorphic fluids are at most a few thousand ppm, metal concentrations may be much higher, because of their ability to form soluble salts. In all but the most dilute natural fluids, chloride is the dominant anion and the metal budget is dominated by metals that form soluble chlorides. Metal concentrations of metamorphic fluids are known from analyses of geothermal waters (Williams & McKibben 1989) and from some fluid inclusion studies (e.g. Banks et al. 2000). They are primarily dictated by interactions with minerals in the host rock, themselves dependent on temperature, and by the chlorinity of the fluid (Yardley 2005). Na, K and Ca can all be important, with Ca concentrations being greater in high-salinity fluids (e.g. Houston et al. 2011). Base metal concentrations increase with temperature and Fe reaches percentage levels in hot salty fluids. In some brines, sulphate can also be abundant and this has an impact on the oxidation state of the fluid, retaining a capacity to oxidize host rocks to elevated temperatures and in environments where most rocks are strongly reduced.

Progressive dewatering of rocks undergoing metamorphism should lead to low-salinity metamorphic fluids with only modest salt contents. However the evidence from fluid inclusions is that metamorphic fluids are often quite saline. Yardley and Graham (2002) compiled evidence of fluid inclusion salinity from a wide range of the literature and showed that metasediments derived from continental shelf sequences commonly retained fluids with salinities greater than seawater, while in evaporite-bearing successions, pore fluids could still be salt-saturated at amphibolite facies temperatures, amounting in effect to wet salt melts. In contrast, sediments with an oceanic or accretionary provenance did not exhibit such extreme salinities, although salt concentrations greater than seawater were not uncommon. This study indicates that metamorphic dewatering is not as efficient at washing salt out of sedimentary sequences, as might be supposed, and for some types of sedimentary sequences it is possible to have extremely salt-rich fluids at moderate to high metamorphic grade, which will have correspondingly high base metal concentrations. On the other hand, in slate belts derived from deep water settings without evaporites, low-salinity fluids (i.e. comparable salinity to seawater) are likely to be widespread.

Gas species are less soluble in chloride-rich fluids than in pure water (‘salting-out’ effect), and as a result ligands such as bicarbonate and bisulphide, whose concentrations are related to the solubility of CO₂ and H₂S, respectively, are likely to be more abundant in both absolute and relative terms in low salinity fluids. The lack of fluids rich in both CO₂ and salt, except under extreme high-pressure conditions, is well known from theory, experiment and fluid inclusion observations (Liebscher 2007). For sulphide there is little analytical data from fluid inclusions, but the high concentrations of Fe and other chalcophile elements with insoluble sulphides in Cl-rich brines (Yardley 2005) strongly suggest low sulphide concentrations. In contrast, sulphide levels of the order of 10⁻³ M were reported from low salinity CO₂-bearing fluids from the Brusson gold-bearing quartz veins by Yardley et al. (1993), and these are comparable to the values commonly assumed in calculations involving geothermal field fluids.

It is clear from the foregoing that the composition of metamorphic fluids depends not just on the environment of metamorphism, but on the nature and quantities of fluid present in a sedimentary sequence at the onset of metamorphism. Only where previously metamorphosed rocks, already recrystallized and with very little porosity, are subject to a later event is it likely that a ‘pure’ metamorphic fluid, lacking inherited solutes, can be generated. During initial progressive metamorphism, formation waters present in the original sediments remain in the pore spaces and, while they are doubtless progressively displaced as porosity is reduced, they also interact to varying degrees with the fluids released from mineral structures. Sedimentary basin brines show large variations in salinity and can range from bittern brines, the residua of halite precipitation, with high Br/Cl and high levels of a wide range of cations relative to Na, to redisolved halite brines with similar chloride levels, but dominated by NaCl. Houston et al. (2011) have argued that concentrated brines in porous sediments can contain a sufficiently high dissolved load that mineral–fluid interactions are sometimes unable to buffer the fluid composition because there is insufficient reactive mineral material present (mass-limited fluids). If sediments containing such concentrated brines begin to undergo metamorphism, it seems likely that the brines will migrate and progressively become equilibrated with new
host rocks, perhaps causing albitization or other metasomatic changes in the process. Figure 2 is from Yardley (2012) and shows the relationship between the proportion of Ca to Na in sedimentary brines and their Cl/Br ratios. Redissolved evaporite brines are very low in both Br and Ca, but low-grade metamorphic brines associated with emerald deposits from Columbia (Banks et al. 2000), although clearly derived by halite dissolution (very high Cl/Br) are much richer in Ca than similar sedimentary brines. This is taken to indicate that there has been enhanced mineral fluid interaction during metamorphism, so that the Ca content has moved closer to a rock-buffered value. The concentrations of base metals in the Columbian emerald brines are also significantly higher than in sedimentary brines, and it seems reasonable to assume that, if even very concentrated brines like these are able to equilibrate with their host rocks during low-grade metamorphism, then metamorphic fluids in general are likely to have compositions buffered by the rocks that host them, except for the most conservative components such as Cl and Br, which are concentrated in the fluid phase from the outset (see also Yardley 2009). Metamorphic fluids, as defined at the outset, do not have a clear beginning, any more than there is clear distinction between diagenesis and metamorphism, but evolve continuously through interactions with their hosts.

Yardley (2005) found that the base metal concentrations of crustal brines, ranging from shallow shield brines to magmatic brines, vary systematically with temperature and salinity. As an example, results for Fe are shown in Figure 3. Individual data sets of results for fluids formed over a narrow temperature range define parallel trends on the plot of Fe v. Cl (Fig. 3a), but if Fe is normalized to Cl the plot of log (Fe/Cl) v. 1/T is a straight line fit over the entire range of crustal temperatures (Fig. 3b). Yardley (2005) interpreted this behaviour as indicating that base metal concentrations in crustal brines are determined by mineral fluid interactions at the prevailing temperature, so that the fluid is normally a saturated solution of its host. Differences in mineral assemblages must have some influence on the results, especially for a metal like Fe whose solubility is also dependent on redox, but the effect is not a first-order one, although it doubtless accounts for much of the scatter in the data about the trend. The Columbian brines discussed previously fit on the same trend as the other data collated by Yardley (2005), yielding the same range of Fe/Cl ratios as brines from the Salton Sea geothermal system (Williams & McKibben 1989), interpreted to have formed at very similar temperatures, confirming that by this stage of the metamorphic history they are equilibrated sufficiently that their cation composition does not reflect their origins.

**Mineralizing potential 1: chemical considerations**

Based on chemistry alone, it is possible to argue that in some circumstances metamorphic fluids can contain high concentrations of metals and may
therefore be potential ore fluids. Brines, especially those associated with halite deposits where salinity can continue to rise as more salt is able to dissolve with increasing temperature, may carry large loads of a range of metals under metamorphic conditions. In addition to base metals, many other metals can be complexed by Cl, including the rare earth elements (Banks et al. 1994).

Dilute fluids might appear to have less ore-forming potential, simply because of their lower dissolved load, but they have been widely implicated in the formation of Au ore bodies (Phillips 1993; McCuag & Kerrich 1998; Groves et al. 2003; Phillips & Powell 2010) because they can carry relatively high levels of sulphide, and gold, unlike most other metals, is readily transported in aqueous solution as sulphide complexes (Seward 1973). Thus dilute gas-rich fluids have potential to precipitate aluminosilicate and carbonate minerals, and Au, but relatively little capacity to exchange metals with wall rocks, and this provides a mechanism to generate Au deposits with only minor amounts of base metals. The levels of chalcophile elements are already suppressed by the relatively high levels of S needed to mobilize Au. Although dilute ore fluids are often ascribed to a metamorphic origin, it is worth noting that dilute H₂O–CO₂ fluids can also evolve from magmas, and are particularly commonly associated with Li and Be pegmatites (Cameron et al. 1953; London 1986).

Hydrothermal ores formed from brines may also contain Au, which can form aqueous Cl complexes (Loucks & Mavrogenes 1999), but they are more likely to be dominated by base metals, since these dominate the dissolved load of brines, and so ores contain only low concentrations of Au.

In summary, equilibrium geochemistry and fluid inclusion studies can take the case for the involvement of metamorphic fluids in ore formation to the point of demonstrating that there are metamorphic fluids with significant dissolved metal loads. What they cannot do is show that such fluids are capable of delivering the scale of metal concentration needed to form ore bodies.

Mineralizing potential 2: metamorphic segregation or infiltration?

Veins are almost ubiquitous in regionally metamorphosed rocks, and can develop under the entire range of temperatures for which fluid may be present. Quartz veins in schists normally have no economic potential, and normally lack metasomatic wall rock alteration, apart from depletion in quartz in some instances. For this reason we believe that they arise mainly by segregation of material from nearby host rocks rather than by infiltration of silica-rich fluids from elsewhere. In this context, segregation implies redistribution of material within a volume of rocks that is subject to essentially uniform P and T, so that it occurs under the influence of rather small, local differences in chemical potential. As a result, minerals that occur in segregation veins are also normally stable in the host rocks; there are no intervening metasomatic zones. Yardley & Bottrell (1992) have argued that the movement of fluid between evolving fractures as the rock mass deforms provides reversible flow and allows segregation over distances of perhaps tens of metres without ever developing the major disequilibrium and resultant wall rock alteration that accompany irreversible flow.

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Fig. 3. Plots of Fe–Cl relationships for a range of sedimentary (circles), metamorphic (triangles) and igneous brines (diamonds), after Yardley (2005). (a) The general correlation between Fe and Cl can be seen to be made up of parallel clouds of data points from the three different settings, with the highest Fe-levels for igneous fluids. (b) The plot of Fe/Cl against 1/T demonstrates that temperature plays a major and systematic role in determining the Fe content of a brine of any particular chlorinity.
Segregation can also give rise to concentrations of other rock forming minerals, such as andalusite (Cesare 1994), with material being extracted from a significant volume of the surrounding rocks. It is evident that minerals with quite low solubility can be concentrated in this way provided there is a significant chemical potential gradient to focus precipitation at a specific site. What is less clear is whether ore minerals can be concentrated by segregation in the same way. Metal concentrations in some metamorphic fluids can be quite high, and it may be that there are settings in which a specific nucleus for growth allows concentration by segregation without long-range, irreversible flow.

A number of ore deposits have been interpreted as having been enriched in grade by redistribution during metamorphism, for example Frost et al. (2011) suggest that the Broken Hill orebody arose through melting of earlier-formed sulphides at regional peak metamorphic conditions, something that is possible because of the relatively low eutectic temperatures of many sulphide systems. For Au mineralization in metamorphic sequences, Pitcairn et al. (2006) reported that Au and related elements found in ore bodies hosted by greenschist facies rocks are themselves depleted in higher-grade (amphibolite facies) schists. This suggests transfer downgrade by metamorphic fluid and so does not strictly correspond to segregation as defined here, although it does amount to redistribution on a large scale.

**Fluid production in metamorphism**

With the advances in geochronology in recent years, dating of some individual ore bodies has become sufficiently precise that the duration of the ore-forming event is constrained by dates. In many cases it is now clear that hydrothermal ore bodies can form over relatively short time spans in geological terms. Goldfarb et al. (1991) showed that the time available for the formation of the entire Juneau gold belt was no more than 1.5 Ma. Likewise there are examples of similar short time spans needed for the formation of magmatic hydrothermal systems. Thus, while there are also ore systems which have been active over extended periods of time, any model for ore genesis must be able to account for the formation of major ore bodies on time scales of hundreds of thousands of years.

In contrast, regional metamorphism is normally considered to be a slow process because it requires input of heat through insulating materials (rocks). For slow burial with normal crustal heat flow, this constrains the rate at which temperature can rise to a few tens of degrees per million years at most. England & Thompson (1984) estimated that regional heating rates are in the range 5–20 °C/Ma, while Carlson et al. (1995) estimated a heating rate for Barrovian regional metamorphism of 10 °C/Ma. For the central Alps, Janots et al. (2009) reported an estimate of 8–15 °C/Ma from geochronology. This raises the question as to whether it is possible for regional metamorphism to give rise to the relatively rapid fluid release events needed to form hydrothermal ore bodies. There are three distinct issues here: how fast can fluids be generated during regional metamorphism, how effectively can such fluids be retained for subsequent release and how effectively can metamorphic rocks be drained to contribute to focussed flow?

Metamorphic devolatilization reactions are strongly endothermic, and so their rates must in general be limited by heat supply (Yardley 1977; Fisher 1978; Walther & Orville 1982; Carlson et al. 1995). To a first approximation, the heat required to release the water from 1% muscovite in a rock undergoing metamorphism is comparable to the heat required to raise the temperature by 1 °C. The latent heat requirement is coupled closely to the amount of water released, so breakdown of an equivalent amount of chlorite requires more heat. Under amphibolites facies conditions, pelitic rocks commonly lose large quantities of muscovite or chlorite by continuous reactions taking place over a temperature range of a few tens of degrees, and so we can conclude that even the isograd reactions that accompany the appearance of garnet or the breakdown of staurolite over a 20 °C range have a heat requirement, equivalent to raising the temperature by c. 40 °C and therefore requiring between 4 and 8 Ma to pass through a single regional isograd if metamorphism is the result of heating with a normal crustal heat flux. Clearly the details of the time needed for a particular metamorphic belt to go through dehydration will depend on the mix of rock types present and their geometry (thin reactive beds exchange heat more efficiently with unreactive, and therefore faster-heating, interbeds), but in general, there does seem to be a mismatch between the duration of ore-forming events hosted by metamorphic rocks and the rate at which metamorphic fluids can be released during thermally limited progressive metamorphism. Faster dehydration during burial is difficult to achieve without significant magmatism to introduce additional heat. Fast dehydration owing to rapid uplift and erosion is however a real possibility, explored below.

Whether or not the rate at which metamorphic fluids might form ore deposits is coupled to the rate of metamorphic fluid production is related to the storativity of rocks undergoing metamorphism. If metamorphic rocks have sufficient porosity, or
include porous lithologies, they might be able to accumulate the fluid generated over millions of years and release it over a much shorter time span. The exact porosity of a rock undergoing metamorphic recrystallization is very hard to observe or estimate, but there are a number of reasons to suggest that it is small, in addition to the buoyancy arguments discussed above. Textural observations in the laboratory today cannot necessarily give a perfect insight into the state of the grain boundary structure at the peak of metamorphism, but they can show that such rocks have tightly interlocking grains with little obvious porosity. Most of the porosity of shallow crystalline rocks is in cracks, and observations of the distribution of retrograde alteration in metamorphic rocks suggests that they retain effectively no pore water at the peak of metamorphism (the freshest rocks are virtually unaltered, meaning that only very small traces of water can have been present when they began to cool), and where water is present it has been introduced, normally along fractures, after the peak of metamorphism. Where there is evidence for the development of porosity in metamorphic rocks it is clearly related to exceptional events, such as the formation of skarn assemblages through decarbonation reactions (Yardley 2009). There is clear textural evidence in such situations to show that minerals have grown into open space and this makes it very unlikely that other rocks might have remained porous for long periods but left no trace.

There is a further theoretical argument to suggest that metamorphic rocks have very low storativity, which derives from the widespread evidence in metamorphic rocks to suggest that fluid pressures approach lithostatic pressure during prograde metamorphism. Rocks undergoing metamorphism differ from overpressured sedimentary reservoirs in that they have recrystallized to have minimal porosity. Thus the high fluid pressure is retained by a low permeability throughout the rock mass, rather than by an impermeable cap. High fluid pressures coupled with high permeability are only possible for brief periods before all the available fluid has drained away.

It has been argued by several authors (Yardley 1986; Ingebritsen & Manning 2010) that permeability in metamorphism is not an intrinsic property of a rock as in the shallow crust, but is a dependent variable dictated by the rate at which fluid is being generated by thermally driven reactions. Yardley (1986) calculated the fluid flux likely to arise in a low-\(P\), high-\(T\) metamorphic setting, and combined it with the assumption that the driving force for fluid flow was equivalent to a lithostatic gradient in fluid pressure, to obtain a value for the permeability of rocks undergoing metamorphism. The resulting value of \(10^{-19}\) m\(^2\) is probably a high estimate for regional metamorphism accompanying burial and is consistent with extremely low porosity. It appears likely that metamorphic recrystallization reduces porosity and permeability to the minimum required to allow fluid to escape, and so these will be even lower in non-reactive lithologies.

The low permeability and porosity of rocks undergoing metamorphism also has a bearing on the ability of fluid to drain into channels that might form ore bodies. If upward flow of metamorphic fluid is driven by a lithostatic gradient in fluid pressure, a still steeper gradient must be established to divert that flow towards a more permeable zone, should one develop. It is difficult to see how it is possible for metamorphic fluid to be diverted into a permeable channel from more than the immediately surrounding rocks, given their very low permeability, whereas large focussed fluxes of metamorphic fluid require the draining of large volumes of metamorphosing rocks. Effective draining requires either pervasive enhanced porosity or pervasive fracturing, and either is likely to leave permanent, visible traces.

**Mineralizing potential 3: flow considerations**

On the face of it, despite their sometimes favourable chemical compositions, regional metamorphic fluids will be unable to form ore deposits because they are not available in large quantities to flow through focussed flow channels over geologically short periods of time. Instead they are produced only very slowly, and the evidence for near-lithostatic fluid pressures during metamorphism demonstrates the very low permeabilities of metamorphosing rocks. Nevertheless, there are circumstances in which the overall limitation that endothermic reactions must be constrained by heat supply will not apply, and these may provide opportunities for fast metamorphism leading to ore formation from metamorphic devolatilization reactions.

The main consequence of a reaction proceeding faster than latent heat is supplied is that the temperature of the reacting rock mass will drop. In some circumstances, this will not bring the reaction to a halt, and these are the situations in which metamorphism may be accelerated and metamorphic ore bodies might develop. The first situation is where a reaction is overstepped sufficiently that the drop in temperature fails to quench it. This might conceivably arise simply as a result of delayed porphyroblast nucleation in pelites. It is however a common phenomenon where carbonate rocks are interbedded with pelitic rocks. Decarbonation reactions such as the breakdown of calcite + quartz
to wollastonite, or dolomite + quartz to diopside, take place at higher temperatures in the absence of water. However they result in a reduction in solid volume which translates into the creation of a transient secondary porosity and a concomitant increase in permeability (Balashov & Yardley 1998). It follows that if water is present in lithologies interbedded with a quartz-bearing marble then, when one of these reactions is initiated in the marble, the influx of water from the interbedded schists will lower the equilibrium temperature, so that reaction can continue unconstrained by heat supply (Fig. 4). Vuggy growth textures in skarns and marbles document the transient porosity (Yardley 2009). Adjacent unreactive lithologies will thermally equilibrate with the reacting marble or skarn when one of these reactions is initiated in the marble.

Fig. 4. $T–X_{\text{CO}_2}$ diagram to demonstrate the effect of water infiltration on a porosity-generating, decarbonation reaction. In the absence of water, quartz-bearing marble must be heated to point A before reaction commences. As a result of the enhanced porosity, water infiltration shifts the fluid composition to point B, at which point the original reaction is now strongly overstepped. Provided water continues to infiltrate, reaction can proceed faster than heat is supplied, drawing down the local temperature towards point C, which in practice could be as much as 100 °C lower than point A.

Proximities of water. However they result in a reduction in solid volume which translates into the creation of a transient secondary porosity and a concomitant increase in permeability (Balashov & Yardley 1998). It follows that if water is present in lithologies interbedded with a quartz-bearing marble then, when one of these reactions is initiated in the marble, the influx of water from the interbedded schists will lower the equilibrium temperature, so that reaction can continue unconstrained by heat supply (Fig. 4). Vuggy growth textures in skarns and marbles document the transient porosity (Yardley 2009). Adjacent unreactive lithologies will thermally equilibrate with the reacting marble or skarn once the rapid reaction has ceased so that the temperature drop is both local and transient.

Focussed flow through carbonate (Oliver & Wall 1987) to produce skarn is of course an important setting for mineralization associated with igneous intrusions, where hot magmatic fluids may introduce significant quantities of ore metals. The skarn-forming process provides a drain for overpressured fluids in the vicinity, whether of metamorphic or magmatic origin. Despite the large fluxes that may take place through purely regional metamorphic skarns (Yardley et al. 1991), we are not aware of examples of mineralization in purely regional skarns without magmatic input. Recent work on Au mineralization at Navachab in the Damara Belt, Namibia, has demonstrated an association with a metamorphosed carbonate sequence including skarns (Kisters 2005), but although Wulff et al. (2010) have argued for a metamorphic origin for the ore fluid, the evidence appears inconclusive while the deposit and nearby prospects show a close association with late orogenic granites and pegmatites.

A second setting in which reactions that are normally thought of as prograde can proceed despite decreasing temperature is one of rapid uplift and erosion. The positive slope of most dehydration reactions on a $P–T$ diagram means that they can be crossed as a result of drop in pressure as well as when temperature rises (Fig. 5). Once reaction is initiated, it will consume heat, bringing the temperature of the reacting unit down towards the equilibrium curve. Path A–C on Figure 5 illustrates the change in $P–T$ conditions in an unreactive rock mass uplifted at 2.5 mm a$^{-1}$ for 4 Ma with no reaction, nor conductive heat losses, whereas path A–C′ shows a trend to lower temperatures buffered along a dehydration equilibrium curve applicable to reactive lithologies. Conduction of heat from unreactive interbeds will tend to average the temperature between reactive and unreactive layers. The significance of rapid uplift as a means of driving reactions is that it can allow much more rapid reaction, with a pressure drop of a kilobar possible in just a few hundred thousand years, based on modern uplift rates of the order of 1 cm per year. Geochronology yields cooling rates between 30 and 85 °C/Ma for the central Alps (Janots et al. 2009).

In this setting, the key to focussed flow probably lies in the changes in rheology that accompany uplift. Rocks that experience devolatilization reactions during rapid uplift will retain a pervasive fluid phase at elevated pressure, and so will be particularly prone to hydrofracturing as exceptionally rapid fluid production is coupled to falling overburden pressure. The resulting pervasive hydrofractures create permeability and allow rapid draining of metamorphic fluid. As the rocks rise further, however, the steepening temperature gradient will mean that devolatilization reactions are crossed in a retrograde sense (Fig. 5, path D–E). At this point, the loss of the pervasive fluid as it is consumed in retrograde reactions will result in rocks becoming stronger; they may however continue to crack and provide focussed pathways for the release of metamorphic fluid being generated at deeper levels (cf. Sibson 2004).

Gold-bearing quartz (carbonate) veins hosted by metamorphic rocks and formed at a late stage in the orogenic cycle have been widely interpreted to arise from the passage of metamorphic fluids. Detailed
Studies have demonstrated that the ore fluid composition is compatible with a metamorphic origin (Phillips 1993; Phillips & Powell 2010), but the rapid timescale for the formation of many such deposits appears to preclude an origin by progressive metamorphism during burial and heating. On the other hand, an origin during uplift at the end of the orogenic cycle (Pettke et al. 2000) not only fits better with field relationships in many cases, but it is also compatible with a short duration for the ore-forming event. Craw et al. (2010) have presented detailed petrology and geochronology for the formation and subsequent uplift of young orogenic gold veins in Taiwan. They argue that the metamorphic fluid release that gave rise to Au mineralization took place in a short-lived event bracketed by greenschist facies metamorphism below 10 km depth at 5 Ma and late veining at a depth of c. 4 km, dated to 2 Ma.

Studies such as the work of Craw et al. (2010) appear to demonstrate beyond reasonable doubt that some vein deposits of Au hosted by metamorphic rocks do indeed derive from the passage of fluids released during metamorphism. This is not to say that all ‘orogenic gold deposits’ have formed in the same way, however. The signature fluid, rich in CO₂ and of relatively low salinity, is often considered typical of metamorphism, but we have already noted that many metamorphic fluids are quite unlike this, whereas similar fluids can be exsolved from magmas, depending on the relative proportions of H₂O, Cl and CO₂ in the magma. As noted earlier, low-salinity, gas-rich fluids are particularly common in certain types of pegmatite. Since salts reduce the solubility of gas species in water, it is not surprising that all crustal fluids, irrespective of origin, appear to divide into a low-salinity, gas-rich type and a high-salinity, gas-poor type. This is a distinction that reflects phase relations in the H₂O–CO₂–NaCl system; it is not a reflection of the fluid’s ultimate origin.

Furthermore, although the importance of declining pressure in the late metamorphic cycle is well known (England & Richardson 1977), most orogenic belts do not show evidence for prograde, purely uplift driven metamorphism. Secondly, late orogenic magmatism is widespread in many orogenic belts and provides a vector for deep crust and mantle fluids to be introduced into the upper crust. Perhaps late orogenic magmatism can be driven by fluids from late orogenic magmatism in belts where this is present, and in the second part of this paper we explore a possible example of an orogenic gold deposit for which some of the ore fluid has characteristics that do not fit well with a metamorphic origin.
A case study of a major orogenic gold deposit: evidence for an external fluid source at the Sunrise Dam Gold Mine

Late-orogenic gold-only deposits whose characteristics fit with an origin from metamorphic fluids generated by rapid uplift are normally found in low-grade metamorphosed hosts and are typically rather small with a single cycle of mineralization. Many are Phanerozoic in age. At present, however, the term ‘Orogenic Gold Deposit’ is used to include a much wider range of gold deposits, which, although hosted by rocks that had experienced metamorphism before mineralization, have more complex histories of ore formation, sometimes including multiple events spread over a sustained period of time, and/or evidence of a more diverse range of fluids. Many recent studies assume that Au ores formed in cooled metamorphic rocks must be generated through metamorphic processes continuing at depth. There is, however, widespread evidence for late-orogenic magmatic activity in many orogenic belts, and in this case study we explore the evidence for a contribution from magmatic fluids to the formation of a major orogenic gold deposit. Our purpose in doing this is to examine whether the simple metamorphic model for ‘orogenic gold’ is equally applicable to all deposits commonly included in this category, with the emphasis on those of particular economic importance.

Gold mineralization at the Sunrise Dam deposit (Western Australia) provides an example of the issues and problems in assessing the origin of fluids in a large, complex, high-grade orogenic gold system. The Sunrise Dam Gold Mine (AngloGold Ashanti) contains c. 14 Moz gold and is the largest known deposit in the Laverton Tectonic Zone of the Eastern Goldfields province, Western Australia (Nugus pers. comm.). The Sunrise Dam mineral system is structurally complex and experienced a protracted deformation history, with at least five recognized events (Blenkinsop et al. 2007; Table 1). Gold is emplaced into a complex volcanic succession dominated by rhyolite–basaltic andesite volcanic rocks, Fe-rich sediments, and some ultramafic units. Unlike many Yilgarn gold systems there is little evidence for thick sequences of metamorphosed basalts or ultramafic rocks. The complexity of the local volcanic lithologies and the interplay between structural and lithological localization of fluid flow makes this system complicated to unravel geometrically. This older (c. 2814 Ma, AGA internal report) sequence was intruded by a series of felsic and mafic dykes and sills (2674 ± 3 Ma, Brown et al. 2002), some of which have complicated intrusion–deformation relationships within the large shear zones. Timing relationships for mineralization are poorly constrained but it has been suggested that gold was introduced between 2665 and 2650 Ma (Brown et al. 2002; Mueller, 2002).

Gold at Sunrise Dam is hosted in shear-parallel flat veins, steep stockwork veins and hydrothermal breccia bodies. A common feature in most of these is that gold is hosted within the veins with very little evidence for gold within the wall rock. Gold-related alteration does occur, however, around these veins and breccias and is variably sericite–carbonate–pyrite, typical for vein hosted systems.

Metamorphism and architecture

The metamorphic and architectural evolution of the Eastern Goldfields and specifically the Laverton tectonic zone, which hosts Sunrise Dam, has been summarized in Korsch & Blewett (2010); Henson et al. (2010) Czarnota et al. (2010) and Goscombe (2009). This regional-scale information is summarized in Table 1. Generally, gold is believed

| Event | Geology | Age (Ma) | T (°C) | P (kbar) | Deformation | Gold? |
|-------|---------|----------|--------|----------|-------------|-------|
| M1    | Localized high PT | 2700–2685 | 550–650 | 6–8 | D2 East–west compression | ?? |
| M2    | Regional | 2685–2665 | 350–400 | 3–4 | | D3 lithospheric extension, doming and late basin formation |
| M3a   | Late basin formation and doming | 2665–2650 | 450–550 | 3.5–5 | | Minor |
| M3b   | Rapid exhumation | 2650–2630 | 550 to >300 | 4–5 to <1 | D4a and b, transpressional with stress switching | High-grade |
|       | Late strike-slip | 2630–2625 | 300 to <1 | | D5 strike-slip | |
to form at the transition from regional compression to extension, including doming with an elevated thermal gradient, leading to exhumation taking rocks from \( c. 15 \) km depth to \( <3 \) km.

Magmatism occurred in three distinct phases during the metamorphic evolution of the Eastern Goldfields, summarized from Goscombe et al. (2007) and references therein: (a) M2-related high-Catonalite–trondhjemite–granodiorite (TTG) granitoid bloom; (2) mafic and syenitic magmatism (including carbonatites) synchronous with lithospheric extension (M3a); and (3) late tectonic low-Ca granite bloom across the Yilgarn craton (M3b). The onset of major gold deposition is broadly synchronous with lithospheric extension and the mantle-related M3a magmatism.

**Fluids and structures**

There have been a number of previous studies of the Sunrise Dam system with a focus on both structural aspects (see references in Blenkinsop et al. 2007) and ore fluid chemistry (Brown et al. 2003; Blenkinsop et al. 2007; Baker et al. 2010). Using a summary of previous and new fluid inclusion data integrated with structural models, Baker et al. (2010) concluded that gold at Sunrise Dam was precipitated as a result of un-mixing of \( \text{CO}_2–\text{H}_2\text{O}–\text{Cl} \) fluids during deformation-driven depressurization from supralithostatic to hydrostatic pressures by failure across shear zones. They conclude that this led to gold deposition in steep cross-cutting vein and breccia systems, also arguing that this process was overprinted by mineralization from a later, \( \text{CO}_2 \)-rich fluid that was driven along preferential flow paths such as porphyry dyke margins, giving rise to later high-grade gold. While Baker et al. (2010) allude to magmatic or mantle fluid sources based on the fluid compositions (\( X_{\text{CO}_2} > 0.8 \)), it is difficult to determine direct evidence in the fluid inclusion data to demonstrate the magmatic fluid source.

An interesting feature of the Sunrise Dam gold system highlighted in recent work by Oliver et al. (2012) is the fact that high-grade gold (tens to thousands of ppm) occurs in veins that are hosted in, but not symmetrical to, the alteration halos. In many instances the alteration halos are much bigger than the gold-bearing veins. These alteration halos represent sites of metasomatism of the metamorphic rocks, typically with volume-increasing reactions such as carbonation or sericitization. As discussed earlier in this paper, retrogressive hydration of metamorphic rocks is associated with infiltration of cooled, and therefore dry, rocks by externally derived fluids penetrating along cracks, rather than by release of fluids from the rocks themselves. The concentration of gold in thin, discontinuous veins developed within the broad metasomatic halos causes problems for predictive mining and is common in high-grade Au systems in metamorphic rocks.

**Gold associations**

The gold at Sunrise Dam is dominantly hosted in quartz-carbonate veins within zones of sericitic–carbonate alteration. There are two principal vein assemblages that host the gold: (a) high-fineness gold (\( \text{Au}–\text{Ag} \) amalgams with \( >0.9 \) Au) in carbonate–pyrite–apatite seams; and (b) \( \text{Au}–\text{Ag}–(\text{Te}) \) phases with complex sulfosalts and base metal sulphides. Details of these are reported in Cleverley et al. (2006, 2009) and Sung et al. (2007). The base metal-rich assemblages have been observed to include rarer phases such as Kolarite (\( \text{PbTeCl}_2 \)).

**Fluid sources: apatite and gold-related fluids**

Gold shows a variety of textural relationships with other phases in the veins, ranging from gold

![Image](https://via.placeholder.com/150)

**Fig. 6.** ‘Thick laminated vein’ from the 2077 ore drive within the Sunrise Shear Zone. This vein and ore style are in contrast to the thinner laminated (Te-bearing) veins (Fsz). (a) Exposure of typical vein with thick quartz and included dark laminations, often containing a breccia at the margins. (b) Thin section sample showing the horizontal laminations and steeper cross-cutting structures (Te–base metal–Au).
nano-inclusions within pyrite, to coarse gold cross-cutting all phases, to replacement of pyrite (Cleverley et al. 2011). Of particular interest here is gold associated with apatite, since this mineral can be used to track fluid chemistry owing to its anion substitutions (OH–F–Cl). Apatite occurs with gold in small dark tourmaline–dolomite–apatite–pyrite–gold seams, hosted within gently dipping quartz veins (Fig. 6). The seams are akin to micro-breccias within the quartz veins and they contain gold intergrown with or replacing pyrite (Fig. 7a, b). Quartz from the surrounding vein occurs as clasts within dolomite, showing that the seams post-date the quartz vein. The apatite is strongly zoned and the rim region is intergrown with the tourmaline–dolomite–pyrite–gold assemblages (Fig. 7c). The core of the apatite, bright in backscatter electron images, contains numerous magnetite and rare earth element-rich inclusions. Magnetite is only found in the core of apatite and not within the bulk vein assemblage.

Microprobe analyses of the apatite show that they are F-apatite with cores enriched in Sr, reaching concentrations only previously recorded from carbonatite apatites (>0.9 wt% SrO; Fig. 8a). The enriched Sr contents of the apatite allowed the use of Laser Ablation Inductively Coupled Mass Spectrometry (LA-ICP-MS) to measure the $^{86}\text{Sr}/^{87}\text{Sr}$ ratios and Cleverley et al. (2006, internal Cooperative Research Centre report) reported that an age corrected $^{86}\text{Sr}/^{87}\text{Sr}$ ratio of 0.7016 is very close to the calculated mantle value at 2.66 Ga (0.70135) (Fig. 8b). This data is similar to that reported by Clout (1989) from anhydrite in the Kalgoorlie superpit gold deposit. The apatite rims, which are intimately intergrown with pyrite–gold assemblages, exhibit a decrease in Sr abundance and lack magnetite inclusions. This may indicate that the system evolved or mixed with a locally derived fluid, which is consistent with data from Brown et al. (2002) showing Sr-isotope values of alteration carbonate closer to crustal values (Fig. 8b).

**Fig. 7.** Backscattered electron images of two areas from dark, shallow dipping laminations in Figure 6b. (a) Apatite (ap)–dolomite (dol) with brecciated quartz fragments (qz) from surrounding vein and course pyrite, and (b) the same field of view with the contrast set to show the gold–pyrite relationship. (c) Comparable laminations with tourmaline (to) also present; the gold is distinguished from pyrite in (d).
Despite the preponderance of evidence for Au mineralization by CO$_2$–H$_2$O fluids, these results indicate that within small high-grade, gold-bearing seams there is evidence for a mantle, rather than metamorphic, fluid component, and for either evolution of a single fluid by interaction with wall rocks or mixing of fluids from two different sources. It appears that a significant component of the fluid in the system is not derived by metamorphism of a crustal sequence. In view of the evident high Sr content of the fluid, it is likely to have had a significant Cl content, consistent with what is known of mantle mixed-volatile fluids from inclusions in diamond (Izraeli et al. 2001). Similarly, the second, more complex, base metal-rich assemblage has been shown above to include Cl-rich phases. Despite this, no hypersaline brines have been observed despite extensive fluid inclusion studies, with Brown et al. (2003) recording ranges of 3–5.5 wt% NaCl (equiv.) This could be a function of the preservation of fluid inclusions from different stages of the system’s development or
lack of quartz precipitation during the critical gold event. This relationship between inferred Cl and CO$_2$–H$_2$O fluids, especially with the Au–As–Sb–Te association, is similar to the model proposed by Cameron & Hattori (1987) to explain the oxidized Au-rich systems in the Archean rocks of the Canadian shield.

The evidence from fluid inclusions, mineral chemistry and Sr-isotopes indicates that there was an essential fluid component in the Sunrise Dam mineral system that was derived from a mantle or mantle-like source. Micro-textures of the gold itself, as discussed in Cleverley et al. (2009, 2011) and Sung et al. (2007), show that gold has a complex history in these systems and may be introduced in one or multiple fluid events. The evolution between gold–phosphate–pyrite and amalgam–telluride–base metal sulphasalts may actually relate to two distinctive sources or a time evolution during uplift and the orogenic cycle.

Sunrise Dam gold mine is a large, high-grade system and the observations made here may not be applicable to smaller deposits. We believe that, while there is a good observational and theoretical basis for the generation of gold–quartz vein systems through purely metamorphic processes, other processes also contribute to late-orogenic gold mineralization, most notably late-orogenic magmatism. It is likely that major deposits such as Sunrise Dam have a more complex origin involving magmatic as well as metamorphic processes. An evolution from magmatic to metamorphic gold systems has been proposed for the nearby Carosue Basin, where Witt & Hammond (2008) suggest that a magmatic-dominant system was superimposed by a metamorphic system in its distal parts. It may be that large systems like Sunrise Dam result from superposition rather than spatial zonation of gold systems.

The Sunrise Dam deposit illustrates the problem of type classification in mineral deposits. This large, high-grade system does not fit the accepted metamorphic model for ‘orogenic gold deposits’, which is mainly based on detailed studies of smaller deposits in simpler metamorphic hosts. We believe, however, that Sunrise Dam is representative of many high-grade, large-ounce systems and an external fluid source may be a critical ingredient in their development. Arguably these deposits are the most important exploration targets for the future.

Summary and conclusions

Metamorphic fluids can undoubtedly carry significant metal contents, although for chloride-complexed metals maximum concentrations are generally lower than for sedimentary brines. While some metamorphic fluids are low salinity, H$_2$O–CO$_2$ fluids, most are salt solutions ranging in strength from around seawater concentrations to what are effectively wet salt melts. The extended times required for metamorphism, especially on a regional scale, also militate against metamorphic events in general giving rise to ore deposits. Nevertheless, we suggest that there are three situations in which ore deposits may arise as a result of metamorphic fluid processes: (a) where relatively metal-rich metamorphic fluids provide a medium for segregation; (b) where decarbonation reactions lead to focussed fluid flow and skarn formation; and (c) where rapid uplift drives dehydration reactions despite falling temperature, so that the rate of fluid production is not limited by heat flow. Magmatic activity is common in certain metamorphic settings, such as low $P$–high $T$ metamorphism, and it is likely that some ore deposits hosted by metamorphic rocks form as a result of a combination of metamorphic processes and magmatic inputs.

Different orogenic gold deposits may have different origins. Late-orogenic uplift metamorphism provides a viable mechanism for the origin of many small deposits, but magmatic fluids may also play an important role and some deposits have a composite origin. What they all have in common, the high abundance of Au relative to sulphides, arises because the ore fluids are gas-rich but generally of relatively low salinity, so that their capacity to transport Au is enhanced by the relatively high levels of H$_2$S, whereas their ability to transport base metals is limited by the low concentration of chloride. It is primarily the phase relationships in CO$_2$–H$_2$O–NaCl–H$_2$S fluids that lead to Au mineralization and the origin of such fluids may be metamorphic or igneous.

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