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ORIGINAL ARTICLE

An overview of hydrodynamic studies of mineralization

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Abstract Fluid flow is an integral part of hydrothermal mineralization, and its analysis and characterization constitute an important part of a mineralization model. The hydrodynamic study of mineralization deals with analyzing the driving forces, fluid pressure regimes, fluid flow rate and direction, and their relationships with localization of mineralization. This paper reviews the principles and methods of hydrodynamic studies of mineralization, and discusses their significance and limitations for ore deposit studies and mineral exploration. The driving forces of fluid flow may be related to fluid overpressure, topographic relief, tectonic deformation, and fluid density change due to heating or salinity variation, depending on specific geologic environments and mineralization processes. The study methods may be classified into three types, megascopic (field) observations, microscopic analyses, and numerical modeling. Megascopic features indicative of significantly overpressured (especially lithostatic or supralithostatic) fluid systems include horizontal veins, sand injection dikes, and hydraulic breccias. Microscopic studies, especially microthermometry of fluid inclusions and combined stress analysis and microthermometry of fluid inclusion planes (FIPs) can provide important information about fluid temperature, pressure, and fluid-structural relationships, thus constraining fluid flow models. Numerical modeling can be carried out to solve partial differential equations governing fluid flow, heat transfer, rock deformation and chemical reactions, in order to simulate the distribution of fluid pressure, temperature, fluid flow rate and direction, and mineral precipitation or dissolution in 2D or 3D space and through time. The results of hydrodynamic studies of mineralization can enhance our understanding of the formation processes of hydrothermal deposits, and can be used directly or indirectly in mineral exploration.

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1. Introduction

Most ore deposits formed from circulation of geologic fluids, and the mineralization processes generally involve the extraction, transport, and deposition of the ore-forming materials (Guilbert and Park, 1986). Although these processes are both chemical and physical, there are far more studies of the chemical processes (e.g., complexing and solubilities of metals in hydrothermal fluids, and deposition mechanisms of ores) than the physical processes (e.g., fluid flow scale, direction, speed and duration), as reflected by the contents of textbooks of ore deposit geology and monographs in the general field of economic geology (e.g., Barnes, 1979, 1997; Guilbert and Park, 1986; Evans, 1993; Hedenquist et al., 2005; Robb, 2005).

Nevertheless, fluid flow related to mineralization has been the subject of a number of studies, as summarized in many review papers (e.g., Cathles, 1981, 1997; Garven and Raffensperger, 1997; Cathles and Adams, 2005; Cox, 2005; Ingebritsen et al., 2006). These studies have demonstrated the driving forces, scales, and patterns of fluid flow in various hydrothermal mineralization systems, including notably mineral deposits associated with magmatic intrusions (e.g., White et al., 1971; Norton and Cathles, 1973, 1979; Whitney, 1975; Cathles, 1977, 1981, 1997; Norton and Knight, 1977; Norton, 1978; Henley and McNabb, 1978; Hayba and Ingebritsen, 1997; Driesner and Geiger, 2007; Fu et al., 2010), ocean floor hydrothermal venting systems (e.g., Lowell, 1975; Ribando et al., 1976; Wolery and Sleep, 1976; Parmentier and Spooner, 1978; Fehn and Cathles, 1979; Cathles, 1983; Fisher and Becker, 1995; Yang et al., 2006), mineral deposits formed in sedimentary basins (e.g., Cathles and Smith, 1983; Garven and Freeze, 1984a, 1984b; Bethke, 1986; Bethke and Marshak, 1990; Sanford, 1994; Garven, 1995; Raffensperger and Garven, 1995a, 1995b; Garven and Raffensperger, 1997; Chi and Savard, 1998; Cathles and Adams, 2005; Xue et al., 2010, 2011), and mineralizing systems associated with structures (e.g., Phillips, 1972; Fyfe et al., 1978; Cox and Etheridge, 1983a, 1983b; Etheridge et al., 1983; Sibson et al., 1988; Sibson, 1994; Valenta et al., 1994; Oliver, 1996; Cox, 1999, 2005; Hobbs et al., 2000; Schaubs et al., 2006; Zhang et al., 2006, 2011).

This paper is not a review of research results related to fluid flow in various hydrothermal systems, nor a summary of hydrodynamic models for various types of mineral deposits. Instead, it intends to summarize the various methods or approaches that can be used to study the hydrodynamic characteristics of mineral deposits, or to solve problems related to hydrodynamic processes in the formation of mineral deposits. Although much emphasis has been placed on mathematical analysis and numerical modeling in previous studies, as was done in most of the papers cited above, there are many other methods that can provide useful information about the hydrodynamic regimes of the hydrothermal systems, which are reviewed in this paper.

The purpose of this paper is twofold. On one hand, it aims to provide an easy-to-understand account of hydrodynamics of mineralization for geologists, especially those studying mineral deposits, explaining the basis of geologic fluid flow and how it can be used in understanding ore genesis and helping mineral exploration. On the other hand, it aims to provide a summary of various methods related to the hydrodynamics studies of mineral deposits, other than numerical modeling, for geoscientists focusing on mathematical analysis of fluid flow, emphasizing the importance of constraints obtained from these methods on numerical models.

The analysis of fluid flow in mineralizing systems inevitably involves various mathematic equations, especially partial differential equations. However, it is not the purpose of this paper to demonstrate how these equations are derived or how they are solved. Rather, we intend to minimize the use of mathematical equations and try to illustrate the principles of hydrodynamics of mineralization with a plain language.

As indicated above, the formation of mineral deposits involves both physical (hydrodynamic) and chemical (reactive) processes, and it would be desirable to study both of them together. Indeed, over the last 15 years there have been increasing attempts to combine hydrodynamic studies with geochemical reactions related to ore transport and deposition (e.g., Garven and Raffensperger, 1997; Appold and Garven, 2000; Zhao et al., 2000; Geiger et al., 2002; Xu et al., 2006). Nevertheless, we choose to focus on the hydrodynamic aspect to keep with the main topic of this paper, although we do briefly discuss about reactive transport models in the section of numerical modeling.

We will first analyze the various driving forces of geologic fluid flow, and then describe various methods that can be used to study the hydrodynamic regimes and processes of mineralization, including filed observations, petrographic, microstructural, and fluid inclusion microthermometric analyses, and numerical simulations. Finally we will discuss the usefulness and limitation of hydrodynamic studies of mineralization for understanding mineralization processes and helping mineral exploration.

2. Driving forces of geologic fluid flow

Because the solubilities of most metals in hydrothermal fluids are generally low (Seward and Barnes, 1997), the formation of a mineral deposit requires that a large amount of fluid flows through the site of mineralization. Whether or not a mineral deposit can be formed and how large it is depend on if there is a driving force of fluid flow and how long this force exists, in addition to other conditions such as sources of ore-forming materials and ore deposition mechanisms. Therefore, understanding the driving forces of fluid flow is a fundamental part of the hydrodynamics of mineralization. Although various driving forces differ greatly in detail, they can be generally related to Darcy's law, which describes fluid flow in porous media. In the following Darcy's law is described first, and then the various driving forces and their relationships with Darcy's law are discussed. As shown below, fluid flow rate is directly related to rock permeability; therefore, although rock permeability is not part of the driving force of fluid flow, it is briefly discussed in this section.

2.1. Hydraulic potential gradient as the general driving force of fluid flow

Darcy's law can be described as

$$q = -K\frac{dh}{dl} \tag{1}$$

where q is flow rate, K is hydraulic conductivity, h is hydraulic head, and l is distance. Fluid flows from high hydraulic head toward low hydraulic head, therefore hydraulic head gradient can be considered as the general driving force of fluid flow.

Although hydraulic head, which is measured in length (e.g., meters), is commonly used in studies of modern groundwater systems, hydraulic potential, which has the same unit as fluid pressure, is easier to understand as a driving force (Hubbert, 1940; Bethke, 1985). The relationship between hydraulic head and hydraulic potential is deduced as follows.

In Darcy's experiment, hydraulic head is made of two components, elevation head (z) and pressure head (h_p):

$$h = z + h_p \tag{2}$$

The pressure head (h_p) is related to fluid pressure by:

$$h_p = \frac{P}{\rho g} \tag{3}$$

Substituting (3) into (2) we get:

$$h = z + \frac{P}{\rho g} \tag{4}$$

Re-arranging (4) leads to:

$$\rho gh = \rho gz + P \tag{5}$$

ρgh is defined by Hubbert (1940) as hydraulic potential (Φ), so we have

$$\Phi = \rho gz + P \tag{6}$$

The hydraulic conductivity (K) in Eq. (1) is related to rock permeability (k), fluid density (ρ), and fluid dynamic viscosity (μ) as follows:

$$K = \frac{\rho gk}{\mu} \tag{7}$$

Substituting (7), (4) and (6) into (1), we get:

$$q = -\frac{k}{\mu}\frac{\mathrm{d}\Phi}{\mathrm{d}L} \tag{8}$$

So, flow rate (q) is related to rock permeability (k), fluid viscosity (μ), and the gradient of hydraulic potential (Φ). Fluid flows from high hydraulic potential toward low hydraulic potential, therefore hydraulic potential gradient is the general driving force of fluid flow.

From the geological point of view, there are many different driving forces of fluid flow, including fluid overpressure, topographic relief, rock deformation, and fluid density variation. The relationships between these geologic forces and the hydraulic potential are discussed below.

2.2. Fluid overpressure

Fluid overpressure (OP) or excess pressure refers to the difference between fluid pressure (P) and hydrostatic pressure at a given depth (Bethke, 1985; Zhao et al., 1998a) (Fig. 1a):

$$OP = P - \rho gd \tag{9}$$

where ρ is fluid density and d is depth from the surface. If we use the earth surface as the datum for elevation, the elevation (z) at any depth beneath the surface is negative, and Eq. (6) becomes

$$\Phi = P - |\rho gz| \tag{10}$$

where $|\rho gz|$ is equal to the hydrostatic pressure at that depth. It becomes clear from Eq. (10) that hydraulic potential (Φ) is actually the difference between fluid pressure (P) and hydrostatic pressure, or hydraulic potential is equal to fluid overpressure. Because hydraulic potential gradient is the driving force of fluid flow as explained above, we can say that fluid overpressure gradient is the driving force of fluid flow. Thus, although fluid does not always flow from high pressure areas to low pressure areas, it generally flows from high overpressure areas to low overpressure areas. When fluid pressure is at hydrostatic value everywhere, fluid overpressure is equal to zero everywhere, so there is no fluid flow.

Fluid overpressure is produced when the pore fluid supports part or all of the weights of the overlying rocks (lithostatic) rather than those of the overlying pore water (hydrostatic). This can be caused by a number of different geologic processes, which can be divided into two types, i.e., a reduction in pore space or an increase in fluid volume. In both cases, fluid overpressure is produced because the rate of change in pore space or fluid volume is higher than what can be compensated by fluid flow, which is limited by low-permeability rocks or seals (e.g., shales). In

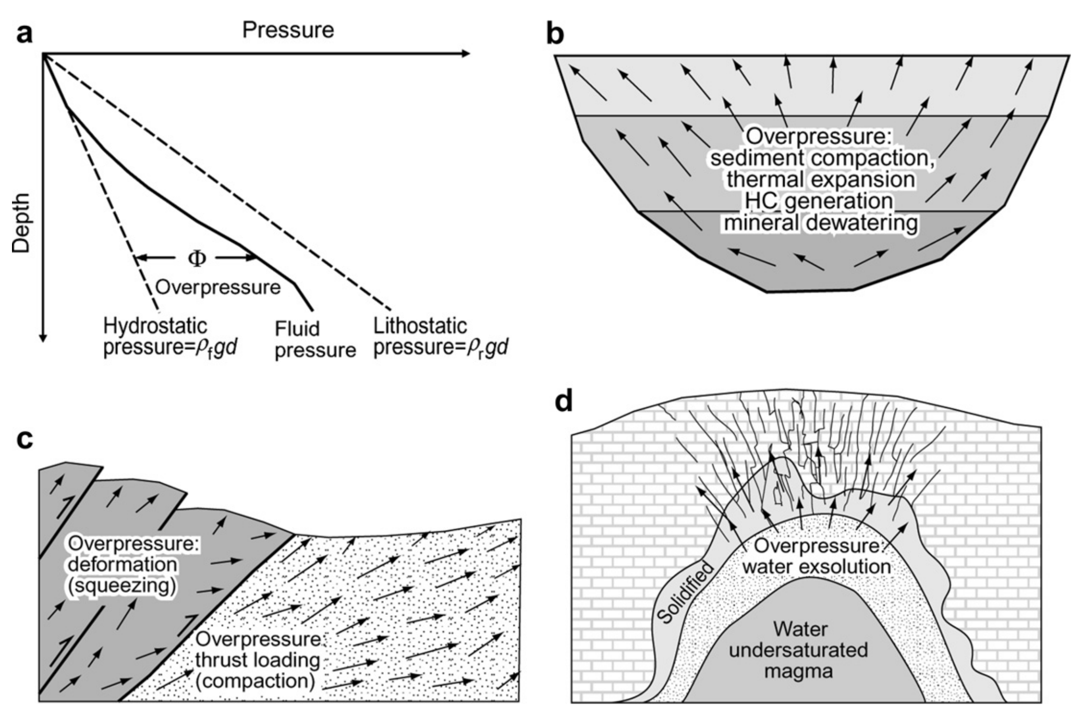

Figure 1 Fluid overpressure as a major driving force of fluid flow. a) definition of fluid overpressure; b) fluid overpressure generated within sedimentary basins; c) fluid overpressure caused by compressional deformation and thrusting; d) fluid overpressure generated in magmatic intrusions by fluid exsolution.

sedimentary basins (Fig. 1b), pore space reduction can result from compaction, and fluid volume increase may be caused by thermal expansion of fluid, hydrocarbon generation, and mineral dewatering (Bethke, 1985; Swarbrick et al., 2002). In metamorphic and deformation belts as well as in the adjacent foreland basins (Fig. 1c), both the reduction of pore space (due to compressive deformation or squeezing and compaction due to thrust loading) and metamorphic dewatering contribute to the development of fluid overpressure (Fyfe et al., 1978; Oliver, 1986; Ge and Garven, 1992). In magmatic intrusions (Fig. 1d), the exsolution of magmatic fluids from the magmas in the outer part of the intrusions (between the solidified shell and the water-undersaturated inner part of the intrusion) results in an increase in the total volume, thus causing fluid overpressure (Burnham, 1997). In all these cases, fluids are expelled from high overpressure areas to low overpressure areas, as illustrated by the flow vectors in Fig. 1.

2.3. Topographic relief

It is known that topographic relief is a major driving force of modern groundwater flow systems, with water generally flowing downward in recharge areas, and upward in discharge areas (Hubbert, 1940; Freeze and Cherry, 1979; Domenico and Schwartz, 1998; Fetter, 2001). Although elevation difference on the surface (topographic relief) is the ultimate driving force of the whole fluid flow system, elevation alone cannot determine the flow direction, as indicated by development of both downward and upward flows in the system (Fig. 2a). Therefore, the driving force is still hydraulic potential gradient, not elevation gradient.

At a given elevation, fluid flows from high pressure to low pressure areas based on Eq. (6). The rightward flow vectors shown in Fig. 2a reflects a rightward decrease in fluid pressure, which is related to the rightward decrease in thickness of the water column at any given elevation. In the vertical orientation, the effect of downward decrease in elevation is more than that of downward increase in pressure in the recharge area, such that there is a net decrease in hydraulic potential downward, whereas in the discharge area, the effect of downward decrease in elevation is less than that of downward increase in pressure, so there is a net increase in hydraulic potential downward. As a result, water flows downward in the recharge area and upward in the discharge area.

2.4. Rock deformation

Rock deformation has been known as a major mechanism responsible for fluid flow in the crust (Sibson, 1994; Oliver, 1996; Cox, 2005). Although much has been focused on the effect of fracturing on enhancing permeability and fluid flow (e.g., Zhang and Cox, 2000; Cox, 2005; Zhang et al., 2008), deformation is also a major driving force of fluid flow as it induces gradients of hydraulic potential both locally and regionally. Two types of effect of deformation on hydraulic potential may be envisaged: one is development of fluid overpressure due to compression, which is often accompanied with mineral dewatering in metamorphic conditions, and the other is development of fluid underpressure (pressures $<$ hydrostatic values) due to creation of open space during fracturing. In addition to fluid overpressure developed in deformation belts and foreland basins as discussed in 2.2 above (Fig. 1c), fluid overpressure is commonly developed at depth of the crust (aseismic crust) due to ductile deformation and devolatilization, and fluid flow is pervasive (Fig. 2b). The large difference in hydraulic potential or fluid overpressure between the aseismic and the brittle crust provides a major driving force for upward focused fluid flow in the upper part of the crust (Fig. 2b), as demonstrated by the fault-valve model (Sibson et al., 1988). In the upper part of the crust, dilation associated with brittle deformation can create large hydraulic potential gradient between the fracture zones (where fluid pressure may be close to or even lower than hydrostatic values) and the surrounding rocks (where fluid pressure are likely higher than hydrostatic values), such that fluid is drawn into the fracture zones (Fig. 2b), as illustrated in the suction pump models (Sibson, 1987, 1994).

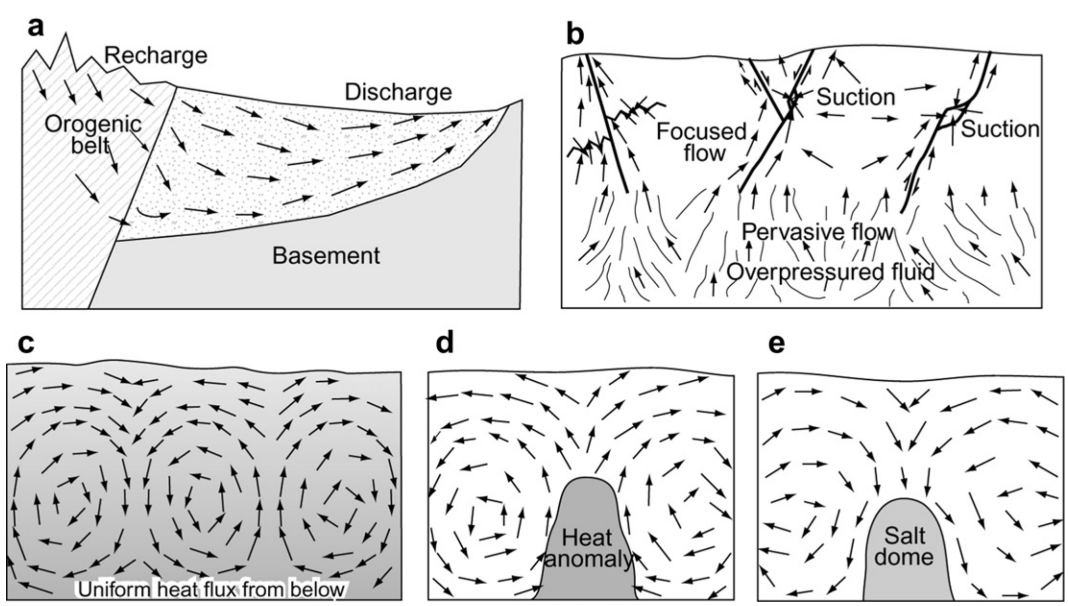

Figure 2 Fluid flow systems driven by various forces. a) topographic relief-driven fluid flow system; b) fluid flow systems driven by fluid overpressure at great depths and those driven by local pressure reduction (suction); c) fluid convection due to uniform geothermal heating; d) fluid convection around magmatic intrusion driven by heat anomaly; e) fluid convection around salt dome driven by density variation.

2.5. Fluid density change

As indicated by Eq. (6), hydraulic potential is related to fluid density, so a change in fluid density can induce a gradient in hydraulic potential. In addition, a change in fluid density can also cause buoyancy, which provides another driving force of fluid flow. There are two major factors affecting fluid density in geological conditions: one is temperature, and the other is salinity. Fluid density decreases with increasing temperature, and increases with increasing salinity.

In the case of temperature-related density change, two situations can be envisaged: one is related to normal temperature increase with depth, and the other is related to local heat anomaly (e.g., magmatic intrusions). In the first situation, fluid density decreases with increasing depth, and when the combination of geothermal gradient, rock permeability, thickness and other thermodynamic parameters is such that the Rayleigh number of the system exceeds its corresponding critical value, fluid convection can take place (Zhao et al., 1997, 1998b, 2001; Turcotte and Schubert, 2002). In some areas fluid flows upward, whereas in other areas it flows downward (Fig. 2c). This may be understood as such that the combination of gravity and buoyancy produces a net upward driving force in certain areas and downward driving force in other areas, forming convection cells (Fig. 2c).

If the increase of temperature with depth is not uniform, i.e., there is a temperature gradient in the horizontal orientation such as in the case of a magmatic intrusion, fluid convection is inevitable (Norton and Cathles, 1979; Cathles, 1997). The fluid surrounding the intrusion tends to move upward due to the upward buoyancy force being stronger than the downward hydraulic potential gradient. In the horizontal direction, the hydraulic potential decreases toward the intrusion due to decreased density, which induces a horizontal component of fluid flow toward the intrusion. As a result, fluid circulation cells are established around the intrusion, with the fluid rising above the intrusion (Fig. 2d).

The fluid flow pattern around a salt dome is opposite to that surrounding an intrusion (Ranganathan and Hanor, 1988). In this case, the density of the fluid near the dome is elevated due to salt dissolution, and the hydraulic potential decreases away from the dome in the horizontal direction. As a result, fluid circulation cells are established such that the fluid above the dome flows downward and that beside it flows outward (Fig. 2e).

2.6. Rock permeability

As indicated by Eq. (8), fluid flow rate is proportional to rock permeability. For a given driving force gradient, the larger the permeability, the higher the flow rate. Rock permeability is an intrinsic property of the rock, which is related to primary openings formed with the rock and secondary openings created after the rock was formed (Fetter, 2001). Rock permeability is not a constant and can change in space and time, depending on the geologic environments and processes. Permeability may increase or decrease through either mechanical or chemical processes, which may also be related to the driving forces (especially deformation), fluid flow and chemical reactions. Generally, a decrease in permeability may be caused by rock deformation (e.g., sediment compaction) or mineral precipitation or cementation, whereas an increase in permeability may be caused by deformation (e.g., fracturing) and mineral dissolution (Bethke, 1985; Ge and Garven, 1992; Xu et al., 2004; Cox, 2005; Zhao et al., 2008b, 2011). It has also been found that rock porosity and permeability may be influenced by temperature (Zhao et al., 1999a).

3. Megascopic features related to hydrodynamic regimes

The various driving forces discussed above may have disappeared for a long time in the geologic history, but they may have left some geologic records that can be used to infer their previous presence and to constrain the ancient hydrodynamic regimes. In many cases, such geologic records are of megascopic scales and can be studied in the field. Some of the more useful hydrodynamic indicators include hydraulic fractures, hydrothermal veins, clastic injection structures, and various breccias. These features and their relationships with hydrodynamic regimes are discussed below.

3.1. Hydraulic fractures and veins

Hydraulic fracturing or hydrofracturing generally refers to artificial fracturing of oil or gas reservoirs by pumping in high-pressure fluids in petroleum engineering (Hubbert and Willis, 1957; Bates and Jackson, 1980). However, this term is also widely used in geology, and generally refers to the production of tensile fractures when fluid pressure (P) is equal to or larger than the minimum principle stress (σ_3) and the difference ($P-\sigma_3$) is equal to or larger than the absolute value of the tensile strength of the rock ($|T|$) (Hubbert and Willis, 1957; Fyfe et al., 1978; Ingebritsen et al., 2006), as expressed in the following equation:

$$P - \sigma_3 \geq |T| \tag{11}$$

or

$$P \geq \sigma_3 + |T| \tag{12}$$

where tensile strength T is a negative value. If the rock does not have a tensile strength (e.g., across a pre-existing fracture), i.e., $T = 0$, then we have $P \geq \sigma_3$, which would be the minimum pressure condition for hydraulic fracturing.

In compressional tectonic environments, σ_1 and σ_2 are horizontal, and σ_3 is vertical, so the tensile fractures are horizontal. Because σ_3 is equal to lithostatic pressure, according to Eq. (12), fluid pressure must be lithostatic (if tensile strength is zero, e.g., if the horizontal fractures were pre-existing) or supralithostatic in order to open the fractures and form horizontal veins. Therefore, the occurrence of horizontal veins is an indication that the fluid was strongly overpressured during the formation of the veins, as seen in various orogenic type gold deposits (Fig. 3a) (e.g., Robert and Kelly, 1987; Sibson et al., 1988; Boullier and Robert, 1992; Cox, 1999; Chi and Guha, in press).

In some cases, subvertical veins are developed in compressional environments and are parallel to the foliation (Fig. 3b). This apparent incompatibility between vein formation and the stress field, i.e., opening along a plane perpendicular to σ_1, may be related to reduced tensile strength due to foliation (Kerrich, 1989). The combination of fluid pressure and principal stresses is such that ($P-\sigma_1$) is $\geq |T_{//}|$ (where $T_{//}$ is the tensile strength along the foliation), so tensile fractures are developed along the foliation, while at the same time ($P-\sigma_3$) is $< |T_{\perp}|$ (where $T_{\perp}$ is the tensile strength perpendicular to the foliation) and no tensile fractures can be developed perpendicular to the foliation (Kerrich, 1989). In this scenario, fluid pressure must be higher than σ_1, and so is supralithostatic. Alternatively, the occurrence of veins along the

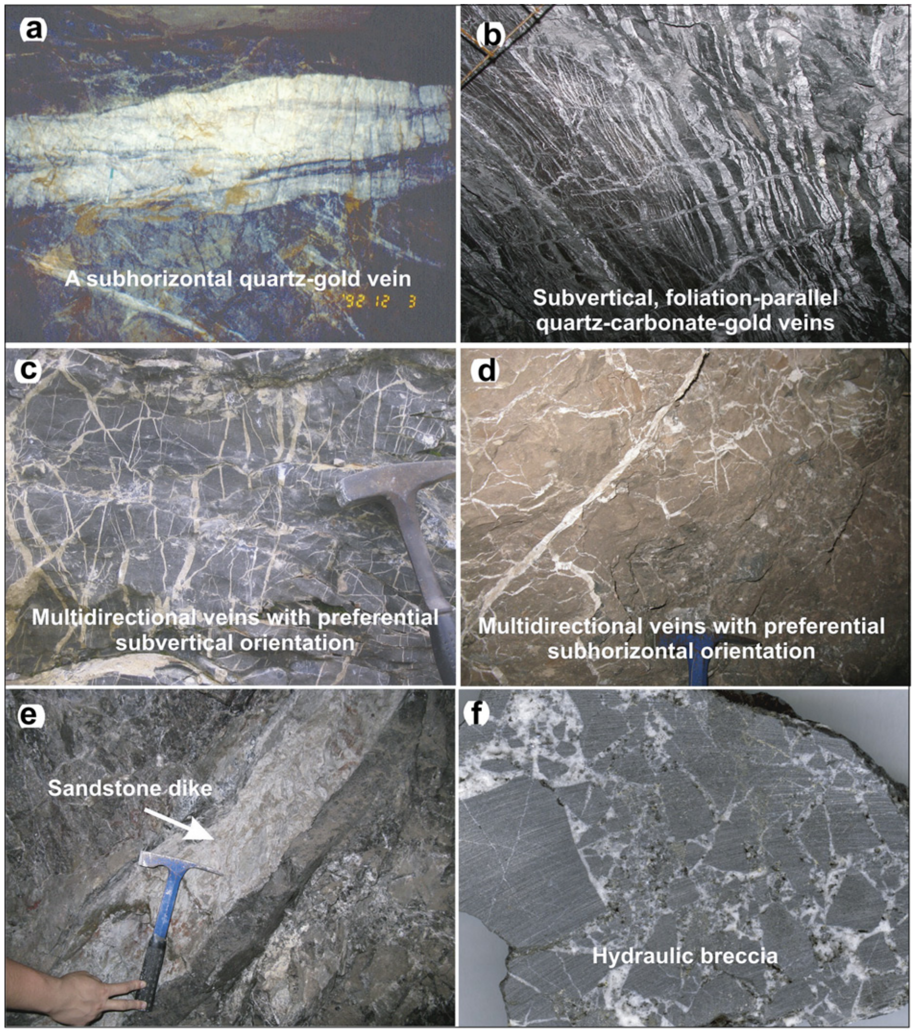

Figure 3 Megascopic features indicating high fluid pressures (generally lithostatic or supralithostatic). a) a subhorizontal quartz-gold vein from the Donalda gold deposit, Quebec, Canada); b) foliation-parallel carbonate-quartz-gold veins from the Campbell-Red Lake gold deposit, Ontario, Canada; c) multidirectional carbonate veins (with a preferential vertical orientation) from the Jinding Zn-Pb deposit, Yunnan, China; d) multi-directional carbonate veins (with a preferential horizontal orientation) from the Jinding Zn-Pb deposit; e) a sandstone dike from the Jinding Zn-Pb deposit; f) hydraulic breccias with calcite-pyrite-sphalerite cements from the Jinding Zn-Pb deposit.

foliation may be explained by episodic switch of σ_1 from horizontal to vertical orientation, so that the veins were actually formed in extensional environments. In such a scenario, the fluid pressure is not necessarily lithostatic or supralithostatic, as explained below.

In extensional tectonic environments, σ_1 is vertical, and σ_2 and σ_3 are horizontal, so the tensile fractures are vertical. In this case, σ_3 is smaller than the lithostatic load, so it is not necessary for fluid pressure to be $\geq$ lithostatic value in order to satisfy Eq. (12). Nevertheless, based on experiments of induced hydraulic fracturing it is suggested that fluid pressure reaching 1.25 times hydrostatic is generally required for the production of hydraulic fractures (Rojstaczer and Bredehoeft, 1988; Ingebritsen et al., 2006). Therefore, development of vertical hydraulic fractures also indicates that the fluid is overpressured, although the overpressure is not as strong as in the case of horizontal fractures.

If the rock is not differentially stressed, i.e., $\sigma_1 = \sigma_2 = \sigma_3$, tensile fractures may be developed in various orientations. In this case, it is likely that the principal stresses are all equal to the lithostatic load, as it is difficult to envisage situations where the principal stresses are less than lithostatic load in all orientations. It follows that the development of multidirectional hydraulic fractures, as recorded by variable oriented veins and veinlets, indicates that the fluid pressure is lithostatic or supralithostatic. However, strict conditions of $\sigma_1 = \sigma_2 = \sigma_3$ may not be met in most cases, therefore some preferential orientation may still be present even though the vein system may be overall multidirectional (Fig. 3c and d). Furthermore, it must be cautioned that multidirectional veins may have also resulted from opening of multi-generations of pre-existing fractures, and therefore do not necessarily indicate equal-stress conditions.

In all the cases discussed above, it is assumed that fluid pressure played an active role in the opening of fractures, thereby hydraulic fractures or hydrofractures. However, fluid participation is not a necessary condition for fracture opening. Open space may be produced through other mechanisms such as uneven rotation of fractured rocks (Ramsay, 1967), movement on non-planar shear surfaces (Guha et al., 1983), pulling apart along faults (Sibson, 1994;

Peacock and Sanderson, 1995; Kim et al., 2004; Cox, 2005), and inhomogeneous deformation (Hodgson, 1989). In these cases, fluid pressure is not necessarily high and may be sucked into the open space, i.e., seismic pumping (Sibson et al., 1975; Sibson, 1987, 1994).

A question that arises is how we can distinguish hydraulic fractures (or veins that fill hydraulic fractures) from those that are dilated without the active participation of fluids in the field. There are actually no straightforward answers to this question, and detailed structural analysis in the context of regional tectonic setting and structural evolution is generally required. For example, if a set of veins are located in a fault bend and the sense of movement along the fault suggests potential formation of dilational jogs, it may be inferred that the fractures filled by the veins are not hydrofractures and fluids flew into the open space passively. In some cases, especially in extensional, shallow, environments, veins (especially steep ones) may have resulted either from filling of pre-existing opened fractures or from hydraulic fracturing and vein filling. However, if the veins are wide, e.g., tens of centimeters or more, there is a high possibility that fluid played an active role in widening the fractures. Veins with crack-seal textures and structures (Ramsay, 1980; Cox and Etheridge, 1983b) formed from incremental opening of fractures may indicate hydraulic fracturing, although gradual structural dilation can also explain the phenomena. Other fractures or veins that are more likely of hydraulic nature, rather than structural dilation or pre-existing open space, include multidirectional fractures or veins and extensive, subhorizontal veins, both indicating lithostatic or supralithostatic fluid pressures.

3.2. Clastic injection structures and breccias

Dikes composed of clastic rocks such as sandstones and conglomerates are found in some mineral deposits (Bryant, 1968; Chi et al., 2007, in press). The clear boundaries between the dikes and the host rocks (Fig. 3e) indicate that the host rocks had been solidified and fractured when the dikes were formed. Petrographic studies indicate that clastic grains are cemented by hydrothermal minerals (Chi et al., 2007, in press), suggesting that sand grains and rock fragments were injected into fractures. Because the clastic grains or fragments are heavier than hydrothermal fluids, such structures cannot be formed in normal (near hydrostatic) fluid pressure and slow fluid flow conditions unless a downward flow is indicated. Instead, they indicate high fluid pressure and high fluid flow velocity (Jolly and Lonergan, 2002). The upward transport of sand grains by fluids into fractures requires very steep hydraulic gradients, which result in a fluid flow velocity so high that upward dragging force exceeds the gravity (Richardson, 1971).

Many mineral deposits are related to or hosted in breccias (Sillitoe, 1985; Jebrak, 1997). There are different genetic types of breccias, including tectonic, hydraulic, intrusive, volume expansion, volume reduction, impact, collapse, and replacement (Bryant, 1968; Sillitoe, 1985; Jebrak, 1997). Among them the breccias of hydraulic, intrusive, and volume expansion origins are closely related to the hydrodynamic regimes, whereas the others are of relatively passive nature, e.g., as ground preparation for mineralization fluid flow due to enhanced permeabilities.

Hydraulic breccias result from multidirectional hydraulic fracturing, the intensity of which is so high that the rock is divided into fragments separated by fractures or hydrothermal minerals, with the fragments being angular and in situ (Fig. 3f). Therefore, the discussion in 3.1 about the relationships between hydraulic fracturing, fluid pressure and stress regimes also applies here, and the development of hydraulic breccia indicates high fluid pressure (most likely lithostatic or supralithostatic) and low differential stress.

Intrusive breccias are defined by Bryant (1968) as “a heterogeneous mixture of angular to rounded fragments in a matrix clastic material that has been mobilized and intruded into its present position along pre-existing structures”, which lack evidence of control by magmatic intrusions. They may have resulted from explosive intrusion of liquefied sands and breccias due to high hydraulic gradients, or from slow intrusion of semi-coherent sand bodies due to their lower density or high pressure than surrounding sediments (Chi et al., in press). The first mechanism is similar to that of the clastic injection structures as discussed above, and the second one mainly applies to loose or semi-consolidated sediments. In either case, high fluid overpressures are implied. Unlike the hydraulic breccias, the intrusive breccias are characterized by transported fragments which may show variable degrees of rounding.

Breccias are most commonly developed in mineral deposits related to magmatic activities (Sillitoe, 1985). Except for breccias resulting from movement of magmas, e.g., intrusion breccias (note, not intrusive breccias as described above), most breccias in magmatic hydrothermal deposits are related to volume expansion during the magmatic hydrothermal processes (Jebrak, 1997). The mechanisms of volume expansion include exsolution of magmatic fluids from the magmas either through first boiling or second boiling (Fig. 1d), and heating of groundwater by the heat of magmatic intrusions (Fig. 2d). In either case, extremely high fluid pressures (mostly supralithostatic) are developed, and the sharp hydraulic gradients result in explosive movement of fluids accompanied by fragmentation of rocks (through hydraulic fracturing) and transport of rock fragments, forming breccias. The breccias may consist of rock fragments from various sources with variable sizes and roundness, and can occur as pipes or various irregular shapes, as is common in porphyry and epithermal mineralization systems (Norton and Cathles, 1973; Sillitoe, 1985; Hedenquist and Henley, 1985; Jebrak, 1997).

In summary, various megascopic geologic features in mineral deposits may be linked to the hydrodynamic regimes during mineralization, which are related to fluid pressure and stress field. Generally low fluid pressure and slow fluid flow would leave relatively little impact in terms of hydrodynamics recognizable in the field, whereas high fluid pressure and fast fluid flow may be recorded by prominent features such as horizontal veins, multidirectional fractures (veins), sand injection dykes, and various breccias, depending on the tectonomagmatic environments.

4. Microscopic studies as constraints on fluid flow models

As discussed above, fluid pressure and stress field play important roles in fluid flow. Although such information may be broadly inferred from field observations, more quantitative constraints relies on laboratory studies, including geothermometers, geobarometers, and microstructures. This section discusses the applications of these methods, with an emphasis on fluid inclusion studies.

4.1. Temperature constraints

Fluid temperature is an important aspect of a hydrodynamic system, because fluid flow inevitably induces temperature change, which depends on the fluid flow rate and direction. The study of fluid temperature and its spatial change may provide important constraints on fluid flow models.

There are many geothermometers that can be used to estimate the temperature conditions of mineralization fluids, among which fluid inclusion microthermometry is the most widely used. The homogenization temperatures of primary fluid inclusions represent the minimum formation temperatures of minerals; only in special conditions, i.e., fluid immiscibility, are homogenization temperatures equal to the formation temperatures (Roedder and Bodnar, 1980; Roedder, 1984). However, at shallow crustal levels, where fluid pressure is small, the homogenization temperatures are fairly close to the formation temperatures (Roedder and Bodnar, 1980; Roedder, 1984).

For some mineral deposits, especially those formed in shallow parts of sedimentary basins, the homogenization temperatures of fluid inclusions provide important constraints on the fluid flow velocities and driving forces. For example, fluid inclusion studies of the Mississippi Valley-type (MVT) Zn-Pb deposits commonly show homogenization temperatures from 80 to 150 °C (Roedder, 1984; Anderson and Macqueen, 1988); such temperatures at a relatively shallow formation depths (<2 km), based on stratigraphic reconstruction, implies that the fluid flow must have been fairly fast, otherwise the fluids would have been cooled to the background temperature of around 60 °C. It has been shown that fluid flow driven by disequilibrium sediment compaction is too slow to explain the heat anomaly associated with the mineralization (Cathles and Smith, 1983; Bethke, 1985), and either topographic relief (Garven et al., 1993) or episodic release of overpressured fluids (Cathles and Smith, 1983; Cathles and Adams, 2005) may have been responsible for the fluid flow, although higher-than-normal regional heat flow rate (100 mW/m^2) may still be required (Deming and Nunn, 1991).

Leach and Rowan (1986) demonstrated that fluid inclusion homogenization temperatures from MVT deposits in the Ozark district in the mid-continent of the United States systematically decrease northward, away from the Ouachita Orogenic Belt, which appears to support the topographic relief-driven fluid flow model. It has been shown that there is a temporal and spatial correlation between orogenic belts and MVT in North America (Sangster et al., 1994). However, in other MVT deposits that are not clearly related to orogenic belts, such as in the Maritimes Basin in eastern Canada, it appears that the fluid inclusion homogenization temperatures are not systematically varied, suggesting that the fluid flow may have been controlled by individual sub-basins and driven by episodic release of overpressured fluids (Chi et al., 1998; Chi and Savard, 1998).

Spatial variation of fluid inclusion homogenization temperatures can also be used as an indicator of fluid flow direction in other types of mineral deposits. For magmatic hydrothermal deposits where the source magmatic intrusions can be identified, temperature decrease away from the source is expected. For hydrothermal deposits without obvious fluid source, however, the spatial variation of fluid inclusion homogenization temperature can be revealing for fluid flow direction. For example, in the Jinding Zn-Pb deposit, Yunnan, which is hosted in a domal structure composed of sandstones adjacent to a regional-scale steep fault in the east, the homogenization temperatures of fluid inclusions in various ore and gangue minerals systematically decrease away from the fault, suggesting that the mineralizing fluids ascended from the subvertical fault and then migrated westward (Xue et al., 2007; Fig. 4a).

4.2. Fluid pressure estimation

Fluid pressure is directly related to the driving force of fluid flow, therefore knowing fluid pressure is extremely useful for hydrodynamic studies of mineralization. However, there are few geobarometers that can be used to calculate fluid pressures, of which fluid inclusion analysis is the most widely used (Roedder and Bodnar, 1980; Roedder, 1984).

The principle of fluid inclusion geobarometer is to construct an isochore, which is a line showing the correlation between temperature and pressure under constant-volume conditions, based on data of homogenization temperature and fluid composition derived from microthermometric studies; once an isochore is constructed, and the fluid temperature is estimated independently (e.g., with stable isotope geothermometer, or mineral assemblages), the fluid pressure can be calculated. The accuracy of pressure calculation depends on those of the temperature estimation and isochore construction.

The isochores of aqueous inclusions are generally steep in the temperature (horizontal axis) – pressure (vertical axis) space. They are not suitable for pressure calculation because a minor error in temperature estimation would lead to a large error in pressure. Therefore, fluid inclusions with relatively gentle isochores are desired for fluid pressure estimation. Such inclusions include CO_2-dominated, CH_4-dominated, and various hydrocarbon inclusions (Brown, 1989; Thiery et al., 2000; Bakker, 2003).

Ideally, a systematic study of fluid pressure around a mineral deposit would readily indicate the fluid flow direction and flow rate. However, although such studies have been reported in oil fields (e.g., Swarbrick et al., 2000), they are difficult to realize because the fluctuation of fluid pressure at a given point in a mineral deposit can be much larger than the pressure gradient between two points. On the other hand, the pressure fluctuation derived from fluid inclusion studies is an important indicator of the hydrodynamic regime by itself. If the maximum and minimum fluid pressures estimated from fluid inclusions for a given generation of mineral at a given point of a mineral deposit is different by 2.5 times or more, it is likely that the fluid pressure regime fluctuate between hydrostatic and lithostatic (or supralithostatic). There are numerous reports of such fluid pressure fluctuation in orogenic gold deposits (e.g., Robert and Kelly, 1987; Chi et al., 2003), which provide strong support to the fault-valve model (Sibson et al., 1988). Similar phenomena have also been reported in other types of deposits, e.g., the Jinding Zn-Pb deposit (Chi et al., 2005) and the Jinman Cu deposit (Chi and Xue, in press), both in the Lanping-Simao basin, Yunnan Province, where fluid overpressures were also indicated by other studies (Chi et al., 2005, in press).

In all these examples, fluid pressures are mainly estimated from isochores of carbonic (without visible aqueous phase) and carbonic-aqueous inclusions (Fig. 4b). Generally the isochores of the carbonic inclusions are gentler than the carbonic-aqueous ones (Fig. 4b), and fluid pressure fluctuation estimated from them are considered to be more reliable, whereas the isochores of the carbonic-aqueous inclusions may be significantly affected by the inaccuracy of the estimation of carbonic to aqueous phase ratios. Hydrocarbon inclusions (including oil and gas) also have relatively gentle isochores, so they are preferably used for estimating fluid pressure if they are present in the mineral deposits studied.

4.3. Fluid inclusion plane (FIP) analysis

Microfractures are well developed in minerals in various mineral deposits, especially those occurring in tectonically active

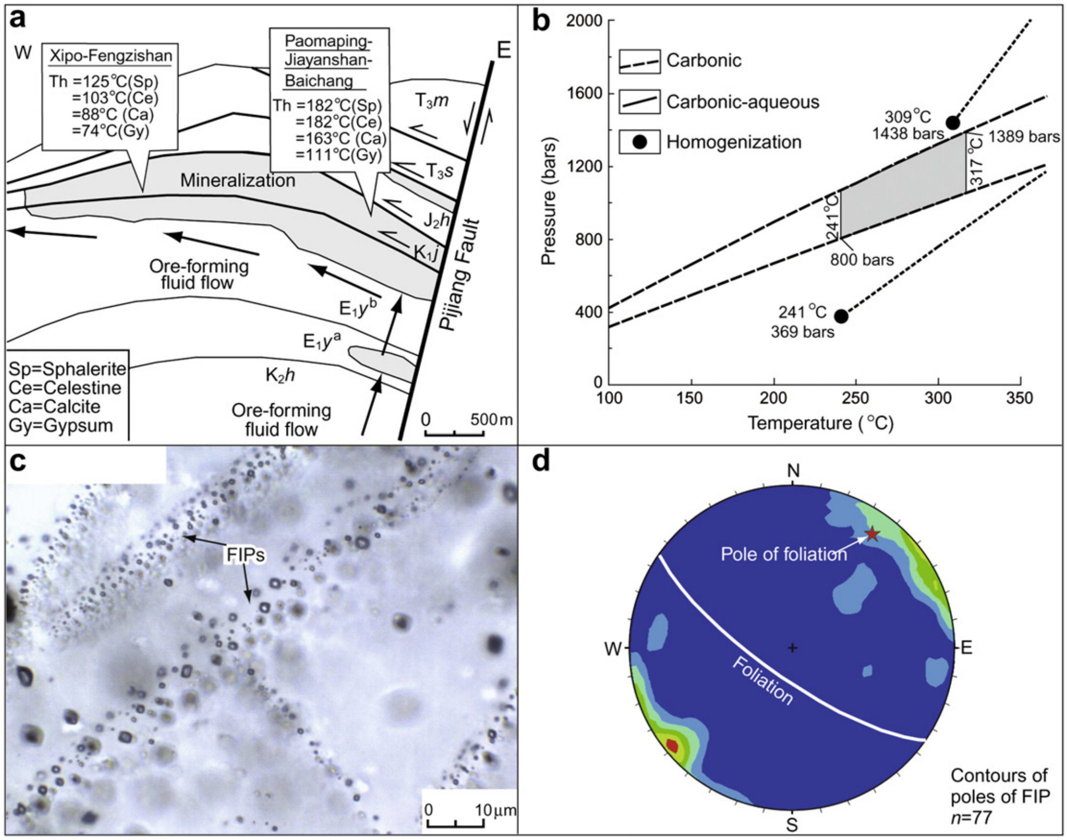

Figure 4 a) A schematic cross section of the Jinding Zn-Pb deposit, Yunnan, China, showing systematic decrease of fluid inclusion homogenization temperatures from east to west, suggesting westward fluid flow direction of the mineralizing fluids (based on data from Xue et al., 2007); b) isochores of carbonic and carbonic-aqueous inclusions from the Jinman Cu deposit, Yunnan, China, showing relatively gentle slopes of the isochores of the carbonic inclusions and possible range of fluid pressure during mineralization (after Chi and Xue, 2011); c) fluid inclusion planes (FIPs) in vein quartz from the Campbell-Red Lake gold deposit (from Liu, 2010); d) a stereonet showing the orientation of FIPs in ore stage quartz in comparison with that of the foliation in the Red Lake mine (from Liu, 2010).

environments, and can provide information about the stress field during and after the formation of the deposits. Many of these microfractures are filled with fluid inclusions and are called "fluid inclusion planes" or FIPs. Although these inclusions are of secondary origin, it does not necessary mean that they are late and unrelated to mineralization. In fact, gold mineralization in orogenic gold deposits commonly occurs in microfractures along with secondary inclusions (e.g., Robert and Kelly, 1987; Liu, 2010; Chi and Guha, in press), suggesting that the secondary inclusions recorded the mineralizing fluids. Therefore, studies of FIPs, including their orientations and microthermometric characteristics, may provide information both on pressure-temperature conditions of the fluids and on the stress field during mineralization (Lespinasse and Pecher, 1986; Boullier and Robert, 1992; Lespinasse, 1999; Liu, 2010; Chi and Guha, in press), making it a powerful tool in hydrodynamic studies of mineralization.

For FIP studies, samples must be oriented in the field, and oriented doubly polished thin sections are required. Although a universal stage may be used to measure the orientations of the FIPs, it is possible to use ordinary petrographic microscopes to measure the dip direction and dip angle of individual FIPs, as used in many previous studies (e.g., Boullier and Robert, 1992), and described in detail by Chi and Guha (in press) and Liu (2010).

As an example of application of the FIP method, Liu (2010) studied the microstructures and fluid inclusions in the Red Lake mine trend, Red Lake greenstone belt, Canada, which hosts the world-class Campbell-Red Lake gold deposit. The gold mineralization is characterized by silicification of carbonate-quartz veins which are mostly parallel to the foliation (Fig. 3b). Fluid inclusions are dominated by carbonic inclusions which indicate significant pressure fluctuation (Chi et al., 2003, 2006a; Liu, 2010). Fluid inclusion planes are well developed in ankerite and quartz in the veins (Fig. 4c), and they are preferentially oriented parallel to the foliation of the host rocks (Fig. 4d). Because the FIPs are mostly of extensional origin, or are mode-I microfractures (Tuttle, 1949; Brace and Bombolakis, 1963; Tapponnier and Brace, 1976; Krantz, 1979; Lespinasse and Pecher, 1986), their orientations are at odds with the stress field inferred from the foliation. Although the mechanical anisotropy due to foliation may be invoked to explain the megascopic feature (i.e., foliation-parallel veins), as proposed by Kerrich (1989), it is difficult to explain similar phenomena at microscopic scale, i.e., FIPs, because no microscopic mechanical anisotropy was observed. Therefore, it may be inferred that the development of the foliation-parallel veins and microfractures resulted from episodic change of local stress field, accompanied by episodic ascent of overpressured fluids.

5. Numerical modeling

5.1. General principles

It would be ideal if we could physically simulate the fluid flow processes related to mineralization in the laboratories, in which we could control the driving forces and fluid flow direction, rate and duration, and chemical reactions to investigate the favorable conditions for mineralization. Unfortunately, fluid flow related to mineralization is generally of such large scale (in terms of hundreds of meters to hundreds of kilometers) and long duration (in terms of thousands to millions of years) that it is impossible to replicate it in laboratories. Therefore, although physical simulation may be carried out to simulate local fluid flow processes, it is generally not an effective method for hydrodynamic studies of mineralization, and so mathematic simulation is necessary.

As discussed in Section 2.1, fluid flow in porous media is generally governed by Darcy's law, which describes fluid flow rate as a function of hydraulic head or potential in one dimension. Fluid flow in 2D and 3D space may be represented by partial differential equations derived from the mass conservation or fluid continuity principle as well as Darcy's law, which describe the change of hydraulic head or potential in the x, y, and z direction (Freeze and Cherry, 1979; Domenico and Schwartz, 1998; Fetter, 2001; Ingebritsen et al., 2006; Zhao et al., 2008a, 2009) and in time. These equations may be solved analytically to express hydraulic head or potential as a function of x, y, and z only for some regular geometric domains (e.g., Toth, 1963). However, in most real geological conditions, especially due to the irregular geometry of the study areas, such analytical solutions are impossible.

In addition to complexities raised by system geometry, geological fluid flow is commonly accompanied by heat transfer and rock deformation, therefore the fluid flow (or mass conservation) equation needs to be coupled with the heat flow (or heat energy conservation) and rock deformation (medium continuity) equations. Also, because chemical reactions can change porosity and permeability due to mineral precipitation or dissolution, mass balance calculations need to be coupled with the other equations too. Many problems deal with the change of fluid pressure and temperature with time (transient), rather than just the spatial distribution of these parameters (steady-state). Furthermore, fluid properties (e.g., density and viscosity) are generally related to temperatures and pressure, making the fluid flow and heat flow equations nonlinear. All these factors add to the difficulty of analytical solutions, and so in general the partial differential equations can only be solved with numerical methods (Zhao et al., 2009).

The basic principle of numerical solution is to divide the study area (or volume) into grids or elements, and convert the partial differential equations into algebraic equations for each of the discrete points or nodes that define the grids or elements, and then use computer to solve the large number of algebraic equations. There are generally two methods of converting the partial differential equations into algebraic equations. One is the finite difference method, in which the derivatives in the partial differential equations are replaced with the differences between neighboring nodes. The other one is the finite element method, in which the partial differential equations are first integrated over the individual elements and then expressed in algebraic forms for the points defining the elements. For steady-state problems, boundary conditions (e.g., known hydraulic potentials or temperatures or their gradients along the boundaries) are required to solve the equations. For transient problems, initial conditions (e.g., known fluid pressures or temperatures at the beginning of the processes) are required in addition to the boundary conditions. The result of the numerical solution is the distribution of such parameters as hydraulic head or potential, fluid pressure, temperature, density and flow rate in discrete points in space and time. Because these parameters define the model of fluid flow systems and processes, the procedure of numerical solution is generally called numerical modeling.

Numerical modeling of fluid flow can be done with commercial or published software, or home-made computer programs. Generally the former has the advantage of stability, graphic output and technical support, whereas the latter has the flexibility of modifying the governing equations as required by specific problems. Some of the more used commercial or public software includes MODFLOW (McDonald and Harbaugh, 1984; Harbaugh, 2005), Basin2 (Bethke et al., 1993, 2007), TOUGH2 (Pruess, 1992), TOUGHREACT (Xu et al., 2004), FEFLOW (Diersch, 2002), FLAC2D and FLAC3D (Itasca, 1995, 2000, 2004).

MODFLOW is mainly used for modern groundwater flow, and has been used for modeling fluid flow in relatively shallow mineralization systems such as tabular sandstone-type uranium deposits (Sanford, 1994). Basin2 is mainly used for modeling fluid flow during the evolution history of sedimentary basins, including those related to fluid overpressures and topographic relief (e.g., Bethke, 1986; Chi et al., 2006b; Xue et al., 2010). A home-made program similar to Basin2 but with additional capacity taking into consideration the effect of hydrocarbon generation on fluid overpressure development has been developed and used in the study of various sedimentary basin-related mineralization systems (Chi and Savard, 1998; Chi, 2001; Chi et al., 2010; Xue et al., 2011). TOUGH2 and TOUGHREACT have been used in various hydrothermal systems, especially those involving multiple phases and water−rock interactions (e.g., Todesco et al., 2003; Mannington et al., 2004), FEFLOW for modeling various fluid convection systems (e.g., Cui et al., 2010; Ju et al., 2011), and FLAC2D and FLAC3D mainly for modeling coupled deformation and fluid flow (Zhang et al., 2006, 2011; Ju and Yang, in press).

Although most of the software mentioned above are mainly used for simulating fluid and heat flow (physical processes), some of them can also simulate chemical reactions. For example, Basin2 can be used to calculate quartz cementation (Bethke et al., 2007), and TOUGHREACT can be used to calculate the distribution of hydrothermal minerals (Xu et al., 2004). A program called RST2D, developed by Raffensperger (1993), has been used to calculate the distribution of metallic and alteration minerals in mineralization systems (Raffensperger and Garven, 1995a, b; Garven and Raffensperger, 1997; Appold and Garven, 2000). Zhao et al. (1999b) have developed a general interface between two commercial computer software, FIDAP (Fluid Dynamics International, 1997) and FLAC (Itasca, 1995), to solve a fully coupled problem between material deformation, fluid flow, heat transfer, mass transport and chemical reactions in a mineralization system (Zhao et al., 1999b, 2000).

5.2. Examples of numerical modeling of fluid flow

As stated earlier, it is not the purpose of this paper to review the various fluid flow models related to mineralization. However, a couple of examples of numerical modeling are presented here, in order to demonstrate how numerical modeling may help answer some questions related to specific mineral deposits or districts, what the procedures are, and what the results look like. We choose to show our own work as examples, using the software of Basin2 (Bethke et al., 1993, 2007), with which we are more familiar.

The first example comes from the Jinding Zn-Pb deposit in the Lanping basin, western Yunnan, China (Xue et al., 2007). Field and laboratory studies indicated that the mineralizing system was episodically strongly overpressured (Chi et al., 2005), but the mechanisms of fluid overpressure development were not well understood. Chi et al. (2006b) carried out a numerical modeling of fluid pressure evolution in the Lanping basin, taking into consideration the contribution of disequilibrium compaction and thrust faulting to fluid overpressure development. A conceptual cross section of the basin was constructed based on regional geological data, where the strata were divided into several hydrogeologic units with different porosity–permeability parameters and time durations. The structural nappes were simulated as additional, solidified strata contributing to compaction. The results (Fig. 5a) indicate that the basin was not overpressured in normal sedimentation processes, while strong overpressures were developed underneath the nappes during the thrusting process, but dissipated shortly afterward. Because mineralization took place after the thrusting, it is difficult to invoke thrusting as the main mechanism for fluid overpressuring during mineralization. Previous studies had indicated that mantle-derived fluids were involved in the mineralization processes of the Jinding deposit (Xue et al., 2003), so Chi et al. (2006b) concluded that extra-basinal fluids, possibly of mantle origin, may have been mainly responsible for the development of fluid overpressure during mineralization. Later work (Chi et al., 2007, in press) further supports the proposal that the Jinding mineralization system was episodically strongly overpressured, although the detailed mechanisms, e.g., degassing directly from the mantle or through magmatic intrusions hidden beneath the deposit, still remain to be investigated.

The second example deals with the localization of uranium deposits in the Ordos basin, northern China, where uranium mineralization appears to be concentrated in the Upper Jurassic Zhiluo Formation. Although it has been generally recognized that hydrocarbons may have played an important role in uranium deposition, which may have controlled the localization of uranium deposits, it was not well understood why the deposits were preferentially located in the Zhiluo Formation. Xue et al. (2010) carried out a numerical modeling of fluid flow in the Ordos basin considering both disequilibrium compaction and topographic relief as the driving forces. Two fluid flow systems, a downward flow system in the upper part, driven by topographic relief and an upward flow system in the lower part of the basin, driven by disequilibrium compaction, were developed (Fig. 5b). The downward flow system was oxidizing and uranium-carrying, the upward flow system was reducing, and the interface of the two systems represents the redox front where uranium precipitation took place. It was found that the location of the interface depends on the interplay between sediment compaction and topographic relief, and that this interface can be maintained at a fixed stratigraphic level (e.g., the Zhiluo Formation, Fig. 5b) over a prolonged period of time, which is favorable for mineralization; this is possible if the dissipation of fluid overpressure in the lower part of the basin is just compensated by a gradual reduction of topographic relief (Xue et al., 2010). This model remains valid even if other factors of fluid overpressure development, such as hydrocarbon generation (Xue et al., 2011) are considered, although the rate of topographic relief change will need to be adjusted so fix the interface of the two fluid flow systems at the same stratigraphic level.

6. Discussion: significance and limitation of hydrodynamic studies of mineralization

Many hydrodynamic studies of mineralization appear abstract and their usefulness may be questioned. However, it is increasingly recognized that hydrodynamic model is an integrated part of any hydrothermal mineralization model, and hydrodynamic studies can be directly or indirectly employed in mineral exploration. On the other hand, hydrodynamic studies have some limitations, some of which are inherently related to the complexities of the fluid flow systems.

The formation of hydrothermal deposits generally consists of three steps, i.e., extraction of ore-forming materials from the source, their transport, and deposition (Guilbert and Park, 1986). Fluid flow is involved in each step, especially the second one. A comprehensive mineralization model should include the fluid flow

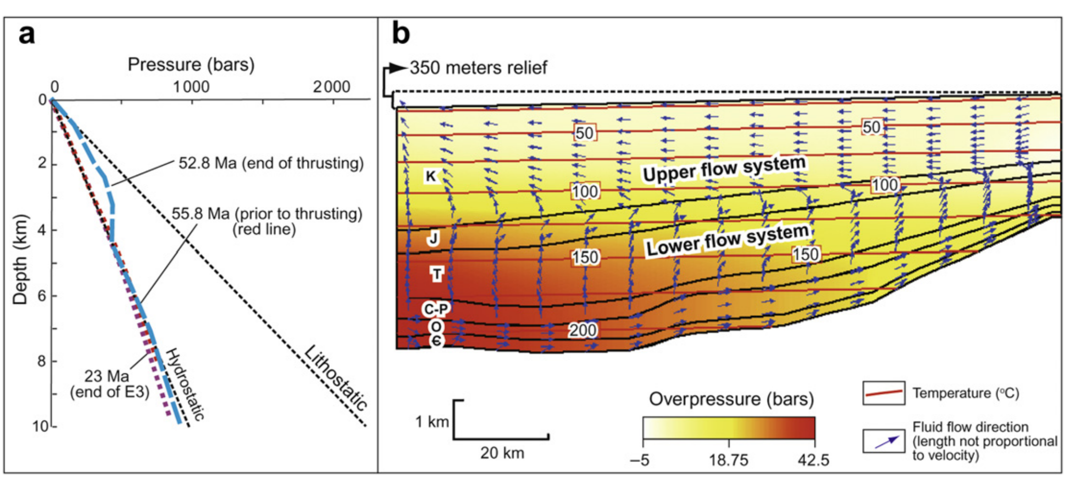

Figure 5 a) A pressure–depth profiles based on numerical modeling results of the Lanping basin, Yunnan, China, showing the development of fluid overpressure due to thrusting (from Chi et al., 2006b); b) numerical modeling results of the Ordos basin in northern China showing two fluid flow systems related to topographic relief and sediment compaction (from Xue et al., 2010).

processes in addition to geologic and geochemical ones. For example, in the porphyry Cu mineralization model, it is generally agreed that: 1) much of the metals and sulfur were derived from the porphyry intrusions and their parent magmatic chambers; 2) these ore-forming elements are partitioned into the ore-forming fluids through exsolution of magmatic fluids or fluid–rock interaction; 3) the ore-forming elements were precipitated near the margin of the intrusion or near the contact zone due to temperature drop, fluid boiling, and fluid-rock interactions; and 4) zonation of metals, mineralization styles and alteration types is formed around the porphyry intrusion (see Robb, 2005). In addition to these geochemical processes, the physical processes of fluid flow also constitute part of the mineralization model, including: 1) "first boiling" (exsolution of magmatic fluids caused by decrease of magma pressure due to emplacement at shallow crustal level), and "second boiling" (fluid exsolution due to crystallization of anhydrous minerals in the intrusions), 2) development of fluid overpressure caused by volume expansion due to magmatic fluid exsolution, 3) hydraulic fracturing and outward expelling of magmatic fluid driven by sharp hydraulic gradients, 4) groundwater circulation around the intrusion driven by heat anomaly, and 5) collapse of the hydrothermal system and incursion of shallow groundwater (Whitney, 1975; Norton and Cathles, 1979; Burnham, 1997; Cathles, 1997; Robb, 2005). These fluid flow models greatly enriched the general porphyry Cu mineralization model. Similarly, the seawater circulation models driven by heat anomaly or salinity variation have become an integrated part of the VMS and SEDEX mineralization models (Cathles, 1981; Robb, 2005), and basinal-scale fluid flow is considered to be part of the classical MVT mineralization model (Bethke and Marshak, 1990; Garven, 1995; Garven and Raffensperger, 1997).

Perhaps the question that is of most practical significance is how hydrodynamic studies of mineralization can help mineral exploration. In general, a refined mineralization model including hydrodynamic processes is helpful for mineral exploration because it determines what kind of geologic conditions or their combination are best for mineralization. In practice, the roles of hydrodynamic studies in mineral exploration may be divided into two types: indirect and direct. "Indirect roles" means that the hydrodynamic studies can provide information about favorable geologic conditions for mineralization without pointing out specifically where the mineral deposits may be located, whereas "direct roles" means that the results of hydrodynamic studies are used to predict the potential locations of mineralization.

As an example of the indirect roles of hydrodynamic studies of mineralization, we may consider the Jinding and other base metal deposits in the Lanping basin, Yunnan. The mineralization models of the basin may be classified into two types in terms of hydrodynamics, one is basinal-scale fluid flow driven by topographic relief as for classical MVT deposits, and the other is episodic release of overpressured fluid along major fault which may be linked to deep-seated fluid and metal sources (Chi et al., 2007). The exploration strategy would be different for the two mineralization models. For the first model, favorable mineralization conditions would be a good combination of sedimentary facies and structures (e.g., Qin and Zhu, 1991) which facilitated basinal fluid flow in a passive way (i.e., providing high permeabilities), whereas the second model suggests that the pathways of the mineralization fluids were proactively generated and maybe linked to the overpressured fluid sources through major faults, so the favorable mineralization locations would be those areas that are potentially connected with deep fluid sources and that have indicator structures of fluid overpressure release such as sandstone dykes (Chi et al., 2007).

Most hydrodynamic studies of mineralization, especially numerical modeling of fluid flow, are of generic nature and cannot be directly used for mineral exploration. However, in recent years, there have been increasing attempts in using numerical modeling to guide selection of exploration targets (e.g., Ord et al., 2002; Schaubs et al., 2006; Potma et al., 2008; Liu et al., 2010; Zhang et al., 2010, 2011). It is known that many mineral deposits are controlled by structures, especially faults or fracture zones, but it is also known that mineral deposits only occur in some of the structures, while most structures are barren. It is therefore of practical interest to study what structures, or what segments within a structure, are potentially good for mineralization within a given region. Generally, it is thought that the structures or segments of structures where fluid flow are focused for a prolonged period of time are favorable for mineralization, and 2D or 3D numerical modeling can predict such areas. As explained by Zhang et al. (2011), a general practice is to compare the numerical modeling results with the occurrences of known mineral deposits, and try to find areas that have similar hydrodynamic attributes as the mineralized areas as targets.

It must be pointed out however, that numerical modeling of hydrodynamic conditions of mineralization as a tool for exploration has certain limitations. This may be related to four major factors. Firstly, the physical models used for numerical modeling, including the geometry of the systems and the distribution of rock properties such as porosity and permeability, are much simpler in the models than those in the real world. Secondly, the driving forces of fluid flow may have multiple origins in the real geologic situations, whereas in the numerical models, typically only one or two are assumed to be operating. Thirdly, the deposition of ore-forming elements is controlled by many factors, especially geochemical processes (e.g., fluid mixing, fluid phase separation, and fluid-rock interactions) in addition to fluid flow rate and duration. Finally, most of the mineralizing hydrothermal systems are not active anymore today, and so numerical modeling results cannot be really tested, unlike modern groundwater water flow systems.

An ideal numerical model to simulate mineralization processes would be one which has a 3D configuration of the geometry of the system, rock and fluid properties, and initial and boundary conditions close to the real world, and which can solve fully coupled equations of material deformation, fluid flow, heat transfer, mass transport and chemical reactions. Unfortunately, owing to the complicated and complex nature of mineralization systems, there is no computer software available in the current world that has all these capabilities; developing such a computational platform is one of the future research directions in the field of hydrodynamics of mineralization. Nevertheless, current hydrodynamic studies of mineralization are still valuable in constraining exploration models with considerations that other geologic and geochemical methods cannot provide. The results of hydrodynamic studies may generate new targets that may not have been proposed with other methods, or may help reduce the risk for targets proposed by other methods. After all, mineral exploration is a high-risk business, and risk management is a key element in any exploration program.

7. Conclusions

Fluid flow constitutes an integral part of the formation of any hydrothermal ore deposits. Hydrodynamic studies of mineralization deal with analyzing the driving forces, fluid pressure regimes, fluid flow rate and direction, and their relationships with localization of mineralization. The methods of hydrodynamic studies of mineralization include recognition of megascopic features indicating high fluid overpressures in the field, microscopic studies of fluid temperature, pressures and stress analysis, and numerical modeling. These studies can help understanding the formation processes of hydrothermal deposits, and the results can be used directly or indirectly in mineral exploration.

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