fluid inclusions. In most porphyries, main-stage potassic alteration mineralization is connected with co-existing immiscible brine and vapour. Brine is dense (>1.3 g/cm^3) and saline (35 to >70 wt.% NaCl equivalent), with variable contents of K, Na, Ca, Fe, Mo and Cu chlorides (Klemm *et al.*875, 874; Rusk *et al.* 2008). The same elements are found in the multitude of daughter minerals in fluid inclusions (halite, sylvite, anhydrite, chalcopyrite, haematite, Fe-chloride, etc.) of minerals precipitated at this stage. Early fluids have metal ratios that correlate with those calculated for the whole deposit, which is a further argument for a magmatic derivation of the metals (Ulrich *et al.* 1999). Another confirmation is the observation that melt and fluid inclusions co-exist in early hydrothermal quartz (Harris *et al.* 2003). Cogenetic supercritical liquid brine and low-density vapour phase inclusions (D <0.1 g/cm^3) document boiling and unmixing of the fluids. Vapour collects acidic volatile species (SO_2, H_2S, CO_2, HCl, HF) and most Cu and Au, plus much of As, Ag, Sb, Te and B. Vapour flow dominates transport and precipitation. Cooling and contraction of magmatic vapour to a liquid appears to dominate copper mineralization (Klemm *et al.* 2007). Reduction can be induced by magnetite crystallization that triggers sulphate to sulphide conversion. Consequent sulphide precipitation promotes ore formation (Liang *et al.* 2009).

• extensive hydrothermal alteration and metals are vertically and cylindrically zoned in relation to the axis of the intrusion.

Host rocks of copper (Au-Mo) porphyry deposits are shallow (<4000 m), subvolcanic and mostly cylindric intrusions. The parent rocks are frequently calc-alkaline diorites (or the volcanic equivalent andesite-dacite), monzonites (latite) or granites (rhyolite) of I-type that occur in volcano-plutonic arcs above subduction zones, either on active continental margins or in island arcs. Subduction of topographic and thermal anomalies appears to favour copper porphyry genesis (Cooke 2005). Less frequent are Cu-Au porphyry and related epithermal gold deposits in continental collision zones and post-subduction settings (Richards 2009). The parent intrusives are geochemically not peculiar, apart from high oxidation (Mungall 2002). Their hydrous nature, elevated oxidation and SO_2 contents (including anhydrite as a magmatic phase) are the main differentiating characteristics in comparison to the numerous barren intrusions of convergent plate margins. Porphyries on active continental margins are marked by elevated Sn and Mo concentrations apart from copper, those of island arcs more often contain gold as the second metal. Porphyry deposits older than Early Tertiary are rare (Cooke *et al.* 2005) but the oldest date from the Archaean.

Several porphyry copper ore deposits display a strong epithermal signature (cf. "Volcanogenic Ore Deposits") and overprinting of earlier hydrothermal alteration, for example clays replacing potassic zone minerals, as at the gold deposit Ladolam on Lihir Island, Papua New Guinea. Such extreme telescoping is best explained by rapid erosion or a sudden collapse of the original volcano (Sillitoe 1994). Moderate telescoping is common, however, because of contraction of the hydrothermal system with progressive cooling.

Sources of the metals concentrated in porphyry copper ore deposits are probably deeper mafic magmas (Hattori & Keith 2001) but ultimately a fertile mantle. Melts and supercritical fluids that originate in the subducting oceanic crust are oxidizing and dissolve chalcophile metals. Absence of reduced sulphur in the source region is a precondition because otherwise sulphide melts would form, which must lag behind rising silicate liquid (Mungall 2002). In epizonal staging chambers, however, an intermediate step of sulphide melt formation may intervene (Halter *et al.* 2002, 2005). Transfer of the metals from mafic melt into the subvolcanic felsic magma may be enacted by highly metal-charged ore brines (Figure 1.34). The shallow felsic magma is the apparent source of ore fluids that cause brecciation, alteration and mineralization in the solid roof, concomitant with the dynamics of a rising stratovolcano. The activity of individual porphyry systems may last for ~100,000 to several million years; districts are active for 10–20 My (Sillitoe 2010).

1.1.9 Hydrothermal-metasomatic ore deposits

The term "metasomatic ore deposit" implies that the mass of ore was formed by hydrothermal chemical conversion and replacement of a pre-

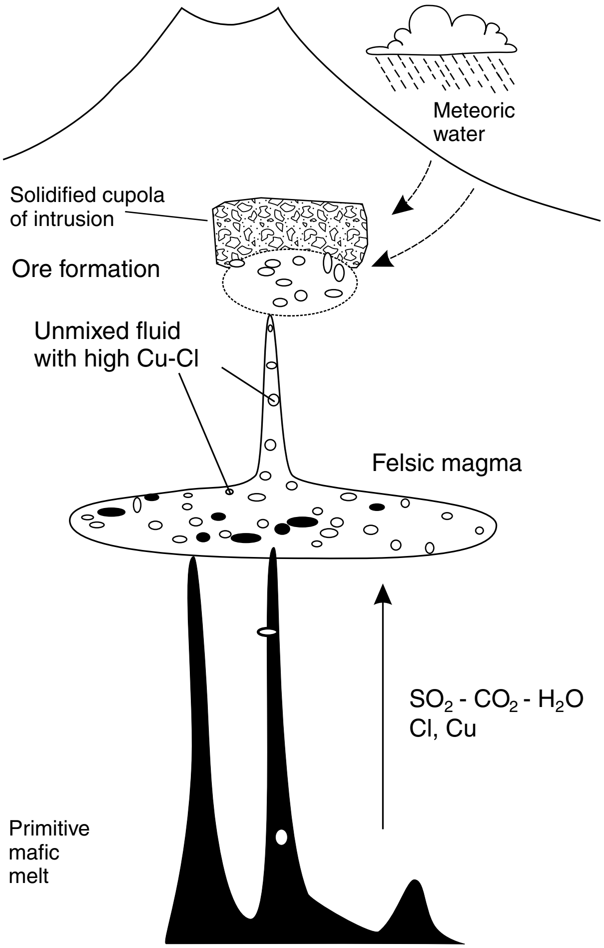

Figure 1.34 Genetic concept of deep magmatic processes preparing the formation of a porphyry copper deposit (modified from Hattori & Keith 2001). With kind permission from Springer Science+Business Media.

existing rock. Local metasomatic processes, for example in ore veins, are ubiquitous co-products of hydrothermal activities, but do not justify attribution to this genetic group. Metasomatic ore formation near igneous intrusions is extremely frequent (cf. Skarn- and Contact-Metasomatic Ore Deposits). In order to avoid repetition, we focus here the discussion on metasomatic ore deposits that have no recognizable genetic relation to magmas. Non-magmatic hydrothermal metasomatic ore formation is usually due to the passage of diagenetic and metamorphic fluids, and especially of evaporative and salt-solution brines (cf. Sections 1.4 and 1.6).

Typically, the metasomatized rocks are marine limestones. This preference can be demonstrated with numerous examples (e.g. many lead-zinc orebodies, gold as at Carlin, USA, magnesite and siderite: Pohl 1988). Main controls of the replacement process include the reactive surface and permeability of the precursor rock, pH and Eh of the mineralizing solutions, and the relative solubility of the participating minerals. A simplified equation 1.7 describes the metasomatic formation of siderite rock (an iron ore) from limestone.

Metasomatic formation of siderite from limestone:

$$CaCO_3 + FeCl_2(aq) \Rightarrow FeCO_3 + CaCl_2(aq) \quad (1.7)$$

In this case, cation exchange is the dominant mechanism, replacing each molecule of calcite

Figure 1.35 The metasomatic siderite deposit Erzberg ("ore mountain" in German) in Austria. The active quarry faces away from the camera. The view shows disposal of barren carbonate host rock.

with one of siderite. Because of the smaller molecular volume of siderite, the resulting ore rock should have a reduced volume. This can be observed in the form of drusy ore at a number of siderite deposits (Erzberg, Austria, Figure 1.35; Ouenza, Algeria), but in many other cases metasomatism exchanges equal volumes. Indicators for this process include the conservation of intricate textures and structures of the replaced rocks (e.g. layering, banding, fossils, stylolites, etc.). Generally, however, metasomatism induces a dramatic change in both chemistry and fabric of the precursor. Some metasomatic ores appear to replace newly deposited soft sediment (e.g. in some magnesite deposits: Pohl & Siegl 1986), but well consolidated precursor rocks are more common.

Hydrothermal-metasomatic ore deposits are often stratabound and occur in the same stratigraphic level across large regions. For example, siderite deposits in North Africa and in northern Spain occur in Early Cretaceous micritic limestone, and lead-zinc ores of the Viburnum MVT district of mid-western USA are confined to stromatolite reefs of the Late Cambrian Bonneterre Formation. Zinc-rich MVT ores in the Tri-State district occur in Mississippian carbonates. In the past, observations like this were believed to prove synsedimentary mineralization. Today, a hydrothermal and epigenetic emplacement is undisputed, but the causes of the preference for certain strata are not always clear. Possibilities include chemical, hydraulic and mechanical parameters.

The emplacement of metasomatic ore is favoured by low-permeability rock horizons (e.g. shales) that form a physical barrier to upward flow (similar to petroleum traps). Focused hydrothermal solutions react more intensively with the carbonate host. Model calculations show that metasomatism is promoted by: i) large temperature differences in flow direction; ii) large focused quantities of fluids; and iii) anomalous phase equilibria (Ferry & Dipple 1991).

The form of many metasomatic deposits is characterized by irregular reaction fronts ("reaction fingering") between ore and host rocks. Higher permeability tongues may be caused by the commonly smaller volume of the product ("induced permeability"). Orebodies are

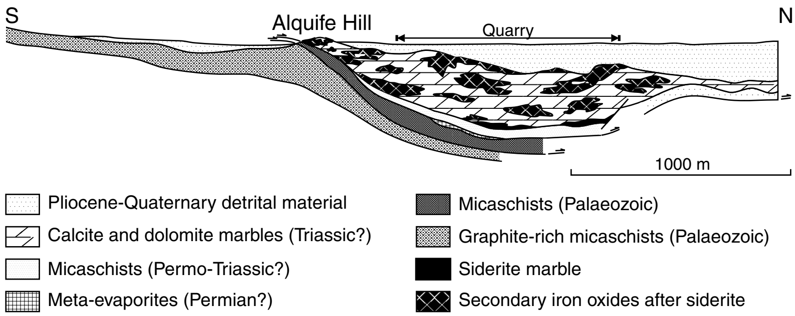

Figure 1.36 Geological cross-section of the Alquife metasomatic iron ore district, Betic Cordillera, southern Spain. Modified from Torres-Ruiz, J. 2006, Society of Economic Geologists, Inc., *Economic Geology* Vol. 101, Figure 3, p. 670. For location refer to Figure/Plate 1.89.

stratabound and cloudy masses with irregular outlines, but some borders mimic structures of the precursor rock (joints, faults, bedding planes). Other ores may take the shape of extensive stratiform bodies (mantos). Both are demonstrated at Alquife in southern Spain (Figure 1.36). In this region, metasomatic siderite was formed during the Triassic by acidic and reducing solutions that leached iron from graphite-rich micaschists in the Palaeozoic basement and ascended along synsedimentary extensional faults into Permo-Triassic cover carbonates. Host rocks and ore were later deformed and metamorphosed by a polyphase Alpine orogeny. In the Tertiary, most of the siderite bodies were affected by supergene alteration and now consist of goethite and haematite.

The flow direction of hydrothermal solutions generates two boundary types of very different nature:

1 In the flow direction, various reactions of mineral neogenesis and isotope exchange migrate with different speed in the form of a chromatographic model (Korzhinskii 1970); assuming kinetically fast reactions, several spatially separate reaction fronts should be the result.

2 Boundaries that laterally limit flow channels (e.g. massive impermeable limestone constricting flow in jointed limestone) can display direct contacts between totally unaltered host rock and high-grade metasomatites (Yardley & Lloyd 1995).

Metasomatism can be accompanied or closely followed in time by mineralization with the character of simple open space filling (veins, fissures, druses, karst cavities). In the North African siderite deposits, this role is played by quartz-barite-fluorite veins, in Bleiberg, Austria by cave-filling ore. Cave ores may include bedded sediments and are a rare, fascinating aspect of hydrothermal systems (cf. "Karst").

1.1.10 Hydrothermal vein deposits

For a long time in the past, ore veins were the most important deposit type. Practice and theory of mining and geosciences grew with the challenges of vein mining, as shown by fundamental books from Agricola (1556) to Lindgren (1933). More recently, the economic relevance of vein mining decreased compared to large-tonnage low-grade operations such as those based on copper porphyries. However, several high-grade base metal vein deposits successfully compete with the mechanized giants.

Veins are tabular bodies of hydrothermal precipitates that typically occupy fissures. Less often, veins originate by metasomatic replacement of rock (replacement veins), propagating from a joint or shear plane. Vein walls range from clean parting planes to "frozen" contacts. Many veins develop upwards into a fan of thinner veins and veinlets (Figure 1.15), which resemble a branching tree. At district scale, veins tend to occur in groups that form vein systems.

Thickness, vertical extent and horizontal length of veins vary widely. Less than 0.5 m thickness may allow profitable mining of high-grade gold and silver ore veins, whereas tin and tungsten require a width of 1 m, barite and fluorite a minimum of 2 m. The world's longest veins may be those of the Mother Lode system of California, with 120 km strike length. Most veins, however,

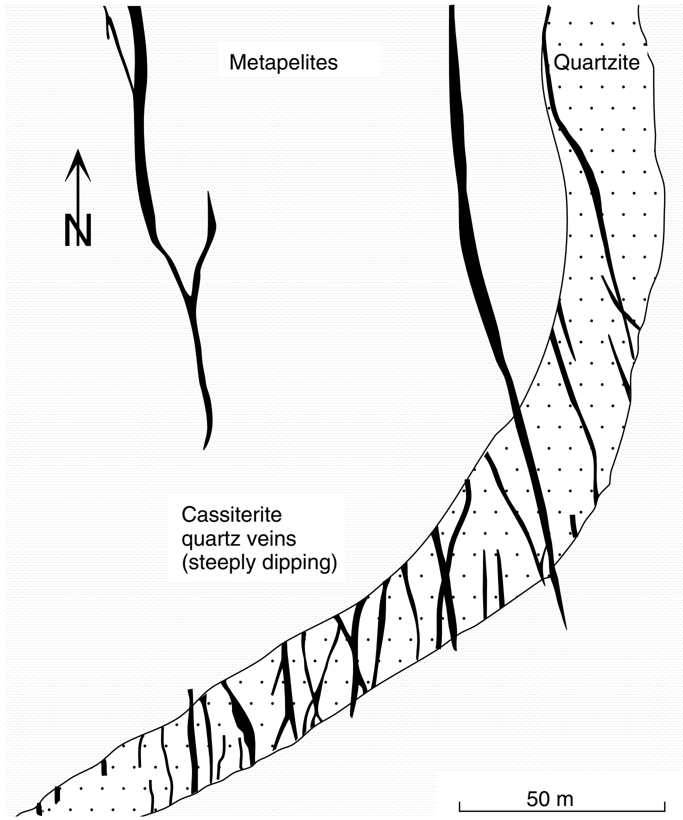

Figure 1.37 Geological map of one level in Rutongo tin mine, Rwanda, showing closely spaced parallel quartz-cassiterite veins (for location, compare Figure 1.16).

have lengths between a few tens to several thousand metres. The disposition of veins in space ranges from horizontal to vertical, but because hydrothermal solutions have a general tendency to flow upwards (more precisely towards lower hydraulic potential) steeply dipping veins are in the majority.

Mechanical properties of host rocks are the most important controls of vein formation, in contrast to metasomatic ore deposits that depend first on chemical properties. Fractures form more readily in competent rocks than in ductile material. Therefore, the cassiterite–quartz (muscovite, arsenopyrite, tourmaline) veins at Rutongo, Rwanda occur preferentially in vitreous quartzites and few cut across low-grade metamorphic schists (Figure 1.37). Very brittle rocks such as dolomite, rhyolite and quartzite are prone to form a network of short fractures instead of spatially separated longer ones (Figure 1.38). In that case, hydrothermal activity may result in stockwork ore. "Stockwork orebodies" consist of numerous short veins of three-dimensional orientation, which are so closely spaced (e.g. 10–30 veins/m in the tin deposit of Tongkeng-Changpo, South China) that the whole rock mass can be mined.

Many vein deposits are spatially and genetically associated with brecciated rock bodies that may host rich ore (Jebrak 1997). As a rule, fissures and breccias have a much higher permeability than most consolidated rocks, which are commonly aquitards. Therefore, these structures focus fluid flow at all scales, from single veins to large structures such as rift faults and crustal shear zones (Weinberg *et al.* 2004). Many fissures, for example the tin ore veins of Cornwall,

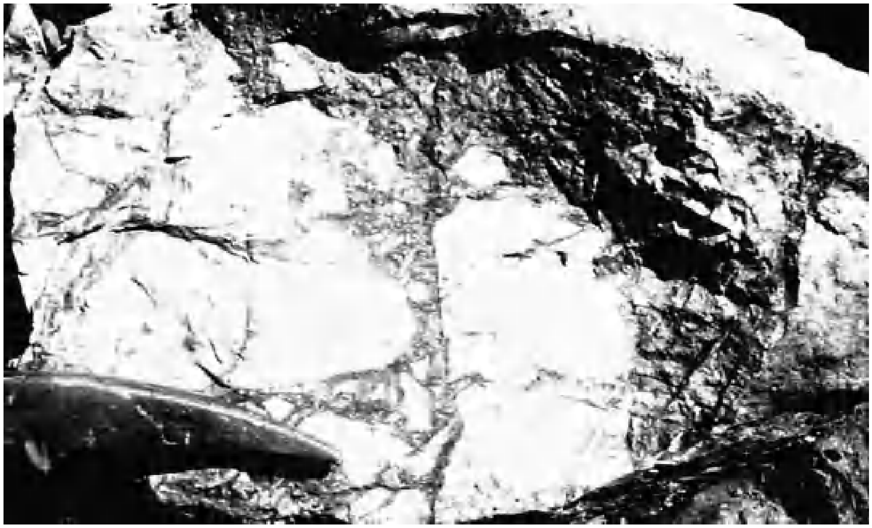

Figure 1.38 Breccia ore at Mammoth copper mine near Mt Isa, Queensland. Supergene secondary chalcocite (grey near hammer) replaces primary chalcopyrite and pyrite in veinlets within quartzite (white).

England, are surrounded by broad zones of intensive micro-fracturing that provide access to fluids from the main flow channel (Dominy *et al.* 1996). It is well-known in the oil industry that in contrast to common permeable faults, clay and shale gouge can transform faults into very effective barriers to fluid flow (Egholm *et al.* 2008). Unconsolidated rocks (clay, sand) are unlikely to fracture but deform ductilely. Typically, this reduces their permeability so that such faults restrict flow.

Veins are channels of former fluid flow. Principally, hydraulic mass flow is a function of hydraulic conductivity and velocity. Flow in a single fissure is controlled by its aperture and secondary properties such as morphology and roughness of the walls. Note that the hydraulic aperture of a vein was hardly ever equal to its present thickness, because most veins were filled while gradually opening. For cases of sheeted vein systems (Figure 1.37), the total cross-sectional permeability can be estimated by incorporating the distance between veins (Lee & Farmer 1993; eq. 1.8). Only in favourable cases, the flow velocity can be measured, for example when upflowing water deposited suspended sediment or hydrothermal precipitates (cf. eq. 1.4 in section "Mineral Succession").

Hydraulic permeability (k) of a single vein and of a sheeted vein system:

$$\begin{aligned} &\text{Single fissure} \quad && k_{vein} = (\rho \cdot g/\mu) \cdot (a^2/12) \\ &\text{Sheeted fissures} \quad && k_{sum} = (\rho \cdot g/\mu) \cdot (a^3/12d) \end{aligned} \tag{1.8}$$

ρ = density of the fluid (g/cm^3), g = gravitational acceleration (m/s^2), μ = dynamic viscosity (m^2/s), a = aperture (m), and d = distance between veins (m).

Methods of fractal analysis (Mandelbrot 1982) were conceived for studying structures built of many elements. Examples of vein-related topics include the statistical distribution of vein thickness and the spacing of veins, and extrapolation of results into adjacent blocks or to different scales (Roberts *et al.* 1998, Marrett *et al.* 1999). In some cases, the fractal distribution of tectonic features and of mineralization, that is their similarity from microscopic to regional scale, can be used for the definition of new exploration targets (Weinberg *et al.* 2004).

Rock mechanic and tectonic interpretation of vein systems play a central role in vein mining, from creating exploration targets to predicting the shifted position of a vein behind a fault. Mining districts display individual controls that must be identified by careful mapping and structural analysis at all scales. Neogenesis and opening of fissures, shear planes and faults reflect spatial orientation and the ratios of principle tectonic stresses during vein formation (Figure 1.39). The opening of fissures is only possible if normal stress is negative (simple tension) or if fluid pressure (u) is high enough to counteract normal stress (effective stress $\sigma_{eff} = \sigma - u$). This condition is often realized by injection of high-pressured hydrothermal fluids. Many veins display pronounced banding (Figure 1.28) that indicates a correlation between pressure increase of the fluids, movement of the fissure walls and precipitation of hydrothermal fill. This is called "seismic pumping" or "fault valve cycling" (Sibson 1990).

Very large faults and shear zones are rarely the site of hydrothermal mineralization. Orebodies occur rather in clusters of fractured rock near jogs or bends in the large structures. This can be explained by the time-integrated evolution of rock mass permeability after faulting (Sheldon & Mikleswaithe 2007): Main faults that experience a high displacement seismic event enter a healing regime that rapidly reduces permeability. The rock mass at some distance from the fault, however, undergoes a period of weakening that may result in seismic aftershocks. Because this causes elevated permeability for weeks or months after an earthquake, the time-integrated fluid flow will be larger than in the main fault, favouring the formation of ore.

Tectronic control of ore veins

The tectonic control of ore veins may be neogenesis of new fractures or opening of older structures. Both allow an examination of the tectonic processes that were active during vein formation. Not unexpectedly, many veins are associated with large-scale tensional tectonics including rifting (e.g. silver-lead-barite ore veins near the Tertiary Upper Rhine Rift, Figure 1.28; silver veins at Kongsberg near the Permian Oslo Graben; fluorite in the Tertiary East African Rift near Naivasha, Kenya) and late-orogenic relaxation of orogens (e.g. silver veins near Freiberg, Saxony, lead-zinc veins in the Harz Mountains, Germany). However, veins may also originate during convergent tectonics, synchronous with shearing, folding,

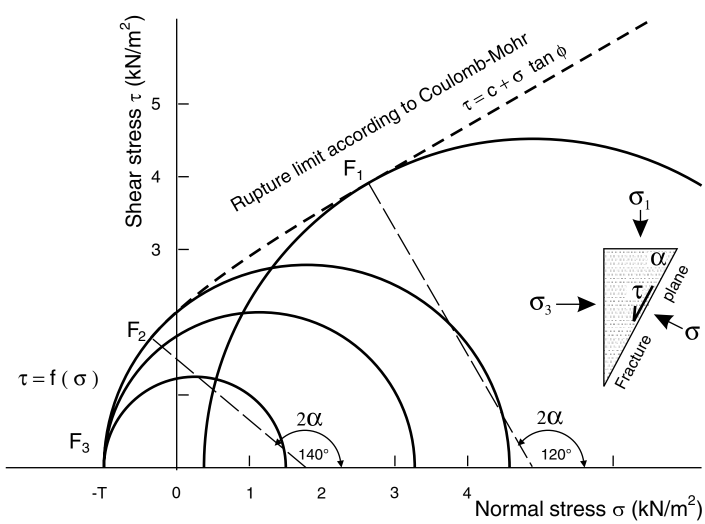

Figure 1.39 The formation of shear fractures (F1), shear fissures (F2) and tensile fissures (F3) with the corresponding angle of fracturing (α), and tectonic stresses $\sigma_1 > \sigma_2 > \sigma_3$ in the Mohr diagram.The inset triangle on the right depicts a section of the upper half of an originally cylindrical sample specimen with the stress geometry of fracture case F1. Fracturing is often assisted by high fluid pressure (u) which reduces strength according to $\sigma_{effective} = \sigma - u$.