CHAPTER 11

SEDIMENT-HOSTED MASSIVE ZINC-LEAD SULFIDE (SMS) DEPOSITS

11.1. Introduction

This class of Zn-Pb sulfide (±barite±Ag±Cu) deposits constitutes a major global resource of zinc (>50%) and lead (>60%), and contributes 31% and 25%, respectively, of world’s primary production of zinc and lead (Tikkanen 1986). This deposit class has been variously referred to as: Sullivan-type massive sulfide deposits (Sawkins 1976a); subclasses of stratiform sulfides of marine and marine-volcanic association (Stanton 1972); exhalative sedimentary group (Hutchinson 1980); sediment-hosted submarine exhalative deposits (SEDEX) (Large 1980, 1981, Carne and Cathro 1982); sediment-hosted Pb-Zn deposits (Badham 1981); sedimentary-type stratiform ore deposits in flysch basins (Morganti 1981); sediment-hosted stratiform lead-zinc deposits (Lydon 1983); syngenetic and diagenetic lead-zinc deposits in shales and carbonates (Edwards & Atkinson 1986); and shale-hosted deposits of Pb, Zn, and Ba (Maynard 1991b). By analogy with the volcanic-associated massive sulfide (VMS) deposits discussed earlier (Ch. 10), the descriptive term sediment-hosted massive sulfide (SMS) deposits is preferred, because it emphasizes the lithologic association of the deposits and excludes any genetic constraint. Most of the deposits included here are dominantly stratiform (i.e., the deposits are composed of sulfide layers parallel to the bedding of the host sedimentary rocks), but some are not, particularly those that have been highly deformed (McClay 1983). In addition, many deposits either contain, or are associated with, mineralization that is not stratiform. As has been pointed out by Large (1983), the term massive sulfide, which loosely describes mineralization containing more than 50% sulfides, separates this class from other classes of sediment-hosted sulfide deposits, such as the sediment-hosted (stratiform) copper deposits (see Ch. 12) and the Mississippi Valley-type Pb-Zn deposits (see Ch. 13); there are also significant differences in the lithologic association, nature of mineralization, and metal ratios among these three classes of sediment-hosted deposits. Also excluded from the present discussion are sediment-hosted barite deposits without significant base metal enrichment, such as those of the barite districts in Arkansas and Nevada (USA), although both deposit types are regarded as exhalative in origin. In contrast to SMS deposits, which are hosted by basinal clastics in dominantly intracratonic rift settings, barite deposits display geochemical signatures that indicate the influence of oceanic crust and appear to
have formed in compressional continental margin settings (Maynard 1991b), perhaps from cooler and shallower hydrothermal systems.

11.2. Distinguishing Features

The distinguishing features of an idealized, unmetamorphosed or mildly metamorphosed SMS deposit (Fig. 11.1) may be summarized as follows (Large 1981, 1983, Russell et al. 1981, Goodfellow et al. 1993):

(a) mineralization hosted mostly either by marine, clastic sediments of continental derivation, typically pyritic and carbonaceous shales, or by platform carbonate rocks, with thin (1 to 10 cm thick) tuff horizons in the sedimentary sequence in some cases;

(b) a zone of stratiform mineralization consisting of (i) stacked lens-like, concordant, tabular bodies of low-Cu massive sulfide ore, commonly composed of thin but remarkably continuous laminations of galena-sphalerite-pyrite that are conformable to the bedding of host rocks and interspersed with barren (meta)sedimentary units (the "bedded ore facies" of Goodfellow et al. 1993), and (ii) an outer margin of bedded hydrothermal products (barite, carbonates, Fe-oxides, chert, minor sulfides) of little economic value ("distal sedimentary facies" of Goodfellow et al. 1993);

(c) a footwall zone of minor stockwork and vein-type sulfide mineralization associated with hydrothermal alteration (predominantly silicification), similar to the alteration pipe underlying VMS deposits (but termed "feeder pipe" by Goodfellow et al. 1993);

(d) a distinct (Cu) ⇒ Pb ⇒ Zn ⇒ (Ba) lateral zoning of hydrothermally precipitated minerals and a less systematic (Cu) ⇒ Zn ⇒ Pb ⇒ (Ba) vertical zoning;

(e) spatial association with a synsedimentary fault system that was active at the time of mineralization and may have been reactivated during later deformation; and

(f) a general lack of demonstrable magmatic affiliation of the host rocks or of mineralization (although a few districts show a close temporal, and even spatial, relationship with mafic sills, dikes, and related volcanic rocks).

Apparently, SMS deposits have many features in common with VMS deposits (Ch. 10), especially of the Besshi-type, but there are significant distinctions as well (see Table 13.4). Compared with VMS deposits, SMS deposits show no clear evidence of a genetic relationship with volcanism, are characterized by stratiform (rather than stratabound) mineralization, have higher Pb:Zn and much higher (Pb+Zn):Cu ratios (Fig. 11.2), lack a consistent Cu ⇒ (Zn+Pb) vertical zoning, and typically contain much
higher combined (Pb+Zn) grades and tonnages, and higher contents of Ag and Co. The genesis of SMS deposits is more controversial (synsedimentary to syndiagenetic to epigenetic) and it is quite possible that they are not bound by a single genetic model. Even for SMS deposits believed to be exhalative in origin, there are notable differences in ore-solution chemistry and depositional environment compared with VMS deposits. Distinguishing between SMS and distal VMS deposits, however, may be quite difficult in some cases, because distal VMS deposits tend to be Zn-Pb-rich, and very few sedimentary basins are completely devoid of volcanic rocks.
11.3. Distribution

As a result of many recent discoveries and re-classification of previously known deposits, it is now realized that SMS deposits have a much wider geographical distribution (Fig. 11.3) than previously thought. Some of the largest Pb-Zn deposits in the world, containing between 10 and 50 million tonnes of (Pb+Zn) or more, belong to this class and a few also contain significant amounts of Cu and Ag. The well known examples of SMS deposits (Table 11.1) include those of Australia (Broken Hill, Mount Isa, McArthur River, Century, Lady Loretta, George Fisher, Hilton, Dugald River), Canada (Sullivan, Howard’s Pass, Anvil, Tom, Jason), Germany (Meggen, Rammelsberg), Ireland (Tynagh, Silvermines, Navan), South Africa (Gamsberg, Aggeneys, Big Syn), USA (Red Dog), and India (Zawar, Rajpura-Dariba). The Balmat-Edwards Zn-Pb district in northwestern New York (USA), earlier considered to be of high-temperature magmatic and/or metamorphic origin (Lea & Dill 1968) and now generally accepted as premetamorphic, may represent a highly metamorphosed variant of the SMS class (Lydon 1983, Whelan et al. 1984).

Host rocks of SMS deposits fall into two main age intervals, Early to Middle Proterozoic (=1700-1400 Ma) and Lower to Middle Paleozoic (=500-320 Ma). Most of

![Figure 11.2](image-url)  
*Figure 11.2. Cu:Pb:Zn proportions (wt%) of selected sediment-hosted massive Pb-Zn sulfide (SMS) deposits (after Lydon 1983). Note the lack of Cu compared with VMS deposits (Fig. 10.2).*
the giant deposits (e.g., Broken Hill, Mount Isa, McArthur River, Sullivan, Aggeneys, and Gamsberg, are hosted by Early to Middle Proterozoic rocks. Important examples of Paleozoic deposits are those of Ireland (e.g., Silvermines, Tynagh, Navan), Germany (Rammelsberg, Meggen), Kazakhstan (Tekeli, Mirgalimsai), Alaska (Red Dog), the Selwyn Basin in Canada (Howard's Pass, Tom, Jason, Anvil district), and South America (Aguilar). The known younger deposits are small and very few in number; they are mostly restricted to the Jurassic strata of Pakistan, Cuba, and Azerbaijan.

The Broken Hill deposit is hosted by a package of clastic and chemical sediments of Early Proterozoic age, and its original features remain a contentious issue because of the overprint of amphibolite to granulite facies and intense tectonic deformation. It has been classified as: (a) a VMS deposit (e.g., Stanton 1976, Plimer 1979, Laing et al. 1984), mainly on the ground that the Potosi gneiss, which occurs both stratigraphically and spatially close to the orebody, has chemical composition consistent with a felsic volcanic precursor; (b) a SMS deposit (Lambert 1983, Sawkins 1984, Wright et al. 1987); and (c) a distinct type of deposit with features that are common to both SMS and VMS deposits (Beeson 1990, Parr & Plimer 1993). It is excluded from further consideration in this chapter.
### TABLE 11.1. Selected sediment-hosted massive sulfide deposits/districts

<table>
<thead>
<tr>
<th>Host rocks</th>
<th>Average grade</th>
<th>Reserves (million tonnes)</th>
<th>Age</th>
<th>Metamorphic facies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pb (%)</td>
<td>Zn (%)</td>
<td>Cu (%)</td>
<td>Ag (g/t)</td>
</tr>
<tr>
<td><strong>Argentina</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>El Aguilar</td>
<td>50 6.5</td>
<td>8.5 9.8</td>
<td>0.2 175</td>
<td>?</td>
</tr>
<tr>
<td>Broken Hill</td>
<td>180 11.3</td>
<td>9.8 11.5</td>
<td>150 175</td>
<td>?</td>
</tr>
<tr>
<td><strong>Australia</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Broken Hill</td>
<td>* 180 11.3</td>
<td>9.8 11.5</td>
<td>150 175</td>
<td>?</td>
</tr>
<tr>
<td>McArthur River (HYC)</td>
<td>104 6.4</td>
<td>14.1 -</td>
<td>64</td>
<td>Dolomitic siltstones and shales #</td>
</tr>
<tr>
<td>Century</td>
<td>118 8.3</td>
<td>10.2 18.4</td>
<td>36</td>
<td>Silicic siltstones and shales #</td>
</tr>
<tr>
<td>Lady Loretta</td>
<td>107 5.4</td>
<td>11.1 125</td>
<td>93</td>
<td>Dolomitic siltstones and shales #</td>
</tr>
<tr>
<td>George Fisher</td>
<td>49 6.5</td>
<td>9.3</td>
<td>151</td>
<td>Dolomitic siltstones and shales #</td>
</tr>
<tr>
<td>Sullivan</td>
<td>155 6.6</td>
<td>5.7</td>
<td>68</td>
<td>Silicic siltstones, quartzites, turbidites</td>
</tr>
<tr>
<td><strong>Canada</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dagald River</td>
<td>38 2.1</td>
<td>13.0 42</td>
<td>151</td>
<td>Dolomitic siltstones and shales #</td>
</tr>
<tr>
<td><strong>Germany</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rammelsberg</td>
<td>30 9.0</td>
<td>10.0</td>
<td>103</td>
<td>Black shale (slate) #</td>
</tr>
<tr>
<td>Country</td>
<td>Location</td>
<td>Reserves (million tonnes)</td>
<td>Average grade</td>
<td>Host rocks</td>
</tr>
<tr>
<td>---------</td>
<td>----------</td>
<td>---------------------------</td>
<td>---------------</td>
<td>------------</td>
</tr>
<tr>
<td>India</td>
<td>Zawar</td>
<td>75</td>
<td>Pb + Zn = 5.5 - 6.7</td>
<td>Dolostone, phyllite, quartzite</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Dominant lithology</td>
</tr>
<tr>
<td></td>
<td>Rajpura-Dariba</td>
<td>152</td>
<td>Pb + Zn + Cu = 2.6 - 7.4</td>
<td>Dolomitic marble, chert, schists</td>
</tr>
<tr>
<td>Ireland</td>
<td>Tynagh*</td>
<td>9</td>
<td>6.2</td>
<td>Shallow marine carbonates</td>
</tr>
<tr>
<td></td>
<td>Silvermines*</td>
<td>18</td>
<td>2.5</td>
<td>Shallow marine carbonates</td>
</tr>
<tr>
<td></td>
<td>Navan*</td>
<td>70</td>
<td>2.6</td>
<td>Mixed shallow-water carbonate facies -- micritic, sandy, oolitic and argillaceous</td>
</tr>
<tr>
<td>South Africa</td>
<td>Gamsberg*</td>
<td>143</td>
<td>0.6</td>
<td>Pelitic schist interbanded with quartzite and calc-silicate rock (Gams Iron Formation)</td>
</tr>
<tr>
<td></td>
<td>Aggeneys</td>
<td>&gt;200</td>
<td>6.2</td>
<td>Pelitic schist interbanded with quartzite and calc-silicate rock (Aggeneys Ore Formation)</td>
</tr>
<tr>
<td>USA</td>
<td>Red Dog</td>
<td>77</td>
<td>5.0</td>
<td>Black shale</td>
</tr>
</tbody>
</table>

*Deposit with significant amount of barite.
# Volcanic rock present in the host rock sequence.
11.4. Types of Deposits

Based on the lithology of host rocks, SMS deposits are divisible into two types: (a) clastic-hosted, and (b) carbonate-hosted (Lydon 1983). Except for two possible Proterozoic examples — Balmat-Edwards (USA) and Black Angel (Greenland) — the carbonate-hosted deposits, including the important deposits of Ireland, are Phanerozoic in age. Clastic-hosted deposits occur in both Proterozoic and Phanerozoic sequences, but all the giant deposits are Proterozoic. Carbonate rocks are an important component of the host sequences of Proterozoic SMS deposits in Australia (e.g., McArthur River, Mount Isa, Lady Loretta), but these are carbonaceous and argillaceous dolostones compared with the poorly argillaceous limestones of the Irish deposits.

A comparison of SMS deposits by dominant host lithology and age (Table 11.2) permits some generalizations (Lydon 1983): (a) for Phanerozoic deposits, there is virtually no difference in metal grades or ratios between the clastic and carbonate groups, suggesting little influence of the depositional environment on the chemistry of the deposit and derivation of ore fluids from similar geologic reservoirs in both cases; (b) the clastic-hosted Proterozoic deposits are the most enriched in Pb; and (c) the clastic-hosted deposits contain much larger tonnages of metals compared with the carbonate-hosted deposits, a difference attributable to greater abundance of argillaceous sediments (source of metals?) in the geological environments of the former.

### Table 11.2. Weighted average grades and tonnages of sediment-hosted Zn-Pb deposits according to age and host rock lithology (after Lydon 1983)

<table>
<thead>
<tr>
<th></th>
<th>Number of deposits</th>
<th>Tonnage (million tonnes)</th>
<th>Grade</th>
<th>Zn:(Zn+Pb)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pb %</td>
<td>Zn %</td>
</tr>
<tr>
<td>Phanerozoic</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clastic-hosted</td>
<td>14</td>
<td>67.0</td>
<td>2.86</td>
<td>7.24</td>
</tr>
<tr>
<td>Carbonate-hosted</td>
<td>9</td>
<td>14.5</td>
<td>2.46</td>
<td>8.12</td>
</tr>
<tr>
<td>Proterozoic</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clastic-hosted</td>
<td>13</td>
<td>90.0</td>
<td>5.09</td>
<td>7.14</td>
</tr>
<tr>
<td>Carbonate-hosted</td>
<td>2</td>
<td>12.2</td>
<td>1.78</td>
<td>11.51</td>
</tr>
</tbody>
</table>

11.5. Examples

11.5.1. McARTHUR RIVER (HYC), NORTHERN TERRITORY, AUSTRALIA

The McArthur River (also known as the HYC, a contraction of the expression "Here's Your Chance") deposit is one of many important SMS deposits located in Proterozoic sedimentary basins in northern Australia (Fig. 11.4). Particularly important are the
Figure 11.4. Geologic sketch map of the Precambrian rocks of the McArthur-Mt. Isa region, Northern Territory and Queensland, Australia, showing geologic domains and SMS deposits (after Williams 1980).

Batten Trough, the central structural element of the McArthur Basin, and the Leichardt River Fault Trough, both of which have been interpreted as intracratonic rifts that contain basinal facies sediments and are flanked by shallower-water sediments, predominantly shelf carbonates (Oehler & Logan 1977). The McArthur deposit, with resources of 237 million tonnes grading 9.2% Zn, 4.1% Pb, 0.2% Cu, and 41 g/t Ag (Gustafson and Williams 1981), is also one of the largest and least metamorphosed SMS deposits in the world. In the absence of metamorphic recrystallization, the sulfides are extremely fine-grained, diameters of sulfide grains seldom exceeding 10 microns compared with hundreds of microns in the more intensely metamorphosed Mount Isa deposit, and thus are not well suited for beneficiation. This has hindered commercial exploitation of the deposit, but the preservation of many original features of mineralization has rendered it an attractive target for investigating the genesis of SMS deposits (Croxford et al. 1975, Murray 1975, Walker et al. 1977, Oehler & Logan 1977, Lambert 1976, 1983, Williams 1978a, 1978b, 1979a, Eldridge et al. 1993, Perkins and Bell 1998, Large et al. 1998).

Zinc-lead-silver mineralization at McArthur occurs more than 3 km above the base of a thick (>5 km), Proterozoic sedimentary sequence that accumulated in the slowly
subsiding Batten Trough of the McArthur Basin (Fig. 11.4). The basinal sequence unconformably overlies an Early Proterozoic basement containing a thick pile of mixed mafic and felsic volcanic rocks, and is subdivided into three units (Williams 1979a): the Tawallah Group (Middle Proterozoic), consisting mostly of fluvialite and marginal marine sandstones in the upper part and volcanic rocks (dominantly mafic) in the lower part; the McArthur Group (Middle Proterozoic) dominated by evaporite-bearing dolomitic strata; and the Roper Group (Upper Proterozoic), composed of quartz-rich sandstone, micaceous siltstone, and carbonaceous shale.

Zn-Pb mineralization, restricted to the McArthur Group, is of two types: (a) a concordant (stratiform) type (HYC deposit) hosted by the carbonaceous and dolomitic HYC Pyritic Shale Member of the Barney Creek Formation (Fig. 11.5) dated by the U/Pb zircon method at 1640 ± 3 Ma; and (b) a relatively minor discordant type (Cooley and Ridge deposits), similar to carbonate-hosted Mississippi Valley-type deposits, hosted by brecciated dolostones. Williams and Logan (1981) interpreted the Barney Creek Formation to represent a mixture of shallow-water/ephemeral marginal marine or lacustrine sabkha deposition and turbiditic deposition at "intermediate" depths. On the other hand, Large et al. (1998) argued that water depth varied during deposition of the Pyritic Shale Member and that the HYC ore lenses were emplaced in the basal deep-water facies. Distant igneous activity during accumulation of the mineralized sequence is indicated by the presence of tuffaceous siltstones and many thin (8 to 30 cm thick), fine-grained, K-rich (up to 13% K₂O) tuffite beds. The tuffite has pseudomorphs of glass shards, now altered largely to K-feldspar (adularia) and, in some cases, replaced by sulfide minerals, probably during diagenesis. The deposit occurs close to the Emu fault (zone) that marks the eastern edge of the Batten Trough. The fault, interpreted as a growth fault that was active during mineralization, is believed to have provided the main conduit for mineralizing fluids (Williams 1978a).

Figure 11.5. a. Sketch map showing the location of HYC deposit in relation to major faults; b. Cross section of HYC deposit, 182900N. (After Hinman 1995.)
The shallow-dipping HYC deposit is conformable with its host sediments and is tabular in shape with an average thickness of 55 m. The deposit contains seven highly mineralized groups of beds or potential orebodies separated from each other by relatively metal-poor, dolomite-rich, and commonly brecciated inter-ore beds (Murray 1975). The main sulfide minerals are pyrite, low-Fe sphalerite, and galena, and essentially monomineralic sulfide bands without any systematic ordering are common. Textural evidence points to at least two stages of pyrite formation: an earlier stage pyrite (Py₁) that occurs as euhedral crystals and spherical clusters (framboids); and a later stage pyrite (Py₂) that occurs as overgrowths on the earlier stage pyrite. Pyrite is also abundant at several levels in the hanging-wall strata and moderately abundant laterally away from the Pb-Zn mineralization. The deposit is not strongly zoned with respect to base metals, but there appears to be two cycles of decreasing Zn:Pb ratio upward from the bottom to the top of the ore horizon (Murray 1975). At the northern end of the deposit, a partly brecciated zone of mineralization that contains only minor sphalerite, but is relatively rich in chalcopyrite, galena, and freibergite, probably formed close to a major point of ore-fluid influx (Lambert 1983).

The origin of the HYC deposit is quite controversial. The stratiform nature of sulfide mineralization and the common occurrence of soft-sediment deformation structures (such as slumps, slides, and scour) in the ore horizon, are compatible with a sedimentary-exhalative model (Croxford & Jephcott 1972, Lambert 1976, 1983, Large 1983) as conceptualized in Figure 11.6a. Some authors, however, have advocated a syndiagenetic replacement origin for the Zn-Pb sulfides (Williams, 1978a, 1978b, 1979a, 1979b, Williams & Logan 1986, Eldridge et al. 1993, Himman 1995). This interpretation (Fig. 11.6b) is based mainly on (a) textural relationships between the base metal sulfides and pyrite — the occurrence of galena and sphalerite as overgrowths on frambooidal pyrite and within dolomite nodules, and (b) a discontinuity in the sulfur isotopic composition of the pyrites. According to Eldridge et al. (1993), the growth of Py₁ (δ³⁴S = -13 to +15‰) and Py₂ (δ³⁴S = -5 to +45‰) likely involved H₂S generation by biogenic reduction of sulfate in a system closed to sulfate. The relatively restricted sulfur-isotopic variability (δ³⁴S = -5 to +14‰) of the base metal sulfides (sphalerite, galena, and chalcopyrite) was interpreted by these authors as evidence that the precipitation of the base metal sulfides did not involve biogenic H₂S, either directly through the use of residual microbially generated H₂S(aq) or by replacement of Py₁ or Py₂. They attributed the base metal sulfide emplacement to H₂S-bearing basinal brines that migrated along bedding planes. Recently, Large et al. (1998) have proposed a synsedimentary origin, a refinement of the exhalative model, to account for the sulfide sedimentary structures and the presence of Mn halos up to 20 km from HYC at the same stratigraphic position. Their interpretation of the sulfur isotope data may be summarized as follows: Py₁ was synsedimentary, formed with sulfur derived from dissolved H₂S in the anoxic bottom water; the galena-sphalerite layers were also synsedimentary, formed with sulfur derived from the anoxic bottom waters and from thermochemical reduction of SO₄²⁻ introduced in the metalliferous brine; and Py₂ was
diagenetic, formed with sulfur derived by biogenic reduction of $\text{SO}_4^{2-}$ introduced with the oxidized sedimentary and brine turbidites, which became trapped in the sediment-base metal sulfide layers. A further controversy has been introduced into the debate by the suggestion of Perkins and Bell (1998) that the HYC deposit formed by syntectonic replacement, an origin they favor for the Mount Isa and Hilton deposits.

11.5.2. MOUNT ISA, QUEENSLAND, AUSTRALIA

The sedimentary sequence in the Mount Isa region, the Mount Isa Group, is metamorphosed (lower greenschist facies) and deformed, but is broadly similar in character and age to that at McArthur. The Mount Isa Group, a $\equiv$5-km-thick package of mainly dolomitic and siliceous shales and siltstones, is part of the younger (1680-1600 Ma) of the two cover sequences in the Leichhardt River Fault Trough (Page & Sweet 1998). It unconformably overlies the Eastern Creek Volcanics, a pile of continental tholeiitic basalts with intercalated clastic sediments metamorphosed to amphibolite facies. This volcanic pile is believed to be the main source of copper for the Mount Isa copper sulfide mineralization (Smith & Walker 1971, Scott & Taylor 1982, Wilson et al. 1985, Wyborn 1987). In the mine area, the Mount Isa Group is truncated by a complex tectonic contact, the Paroo fault, which is underlain by altered Eastern Creek Volcanics (Bell et al. 1988). A minimum of three phases of deformation have been recognized regionally as well as in the mine area (Perkins 1984, Bell et al. 1988).
There are no talus or intraformational breccias in the mine, but thickness and facies changes provide evidence for syndepositional faulting and graben subsidence in the Mount Isa region (Smith 1969).

Disseminated framboidal pyrite is found throughout the sedimentary sequence in the carbonaceous horizons, but the massive sulfide mineralization (Mount Isa, Hilton, and George Fisher deposits) occurs more than 3 km above the base of the sequence, within the 1,500-m-thick Urquhart Shale (Fig. 11.7), which is composed mainly of tuffaceous, carbonaceous, dolomitic shale, and siltstone. Fine-grained tuffite bands, readily identifiable by their high potassium and chert contents, are common in the sedimentary sequence, and some are distinctive enough to be used as marker horizons. Recent SHRIMP geochronology of tuff bands at the Mount Isa and Hilton mines gave ages of 1652 ± 7 Ma and 1654 ± 5 Ma, respectively (Page & Sweet 1998). The presence of flat pebble conglomerates, pseudomorphs after evaporite minerals, stromatolites, oncoids, and teepee structures indicates that the upper Mount Isa Group, including the

![Figure 11.7. Generalized cross section of the Mount Isa deposit, located in the Batten Trough (Northern Territory, Australia), showing the en echelon configuration of stratiform Pb-Zn orebodies within the Urquhart Shale and the distribution of copper orebodies within the "silica dolomite". Structural details have been omitted for simplicity. (Simplified from Bell et al. 1988.)](image)
Urquhart Shale, was deposited in shallow-water under hypersaline conditions (Neudert 1986).

The Mount Isa deposit, a major producer of base metals since the 1930's, comprises two physically distinct types of sulfide mineralization: stratiform Pb-Zn(-Ag) mineralization, and non-stratiform Cu mineralization. The stratiform mineralization occurs as conformable bodies that closely follow the sedimentary stratification and consist of individual laminae (0.5-10 cm thick) of fine-grained pyrite, galena, and sphalerite. Chalcopyrite, a very minor phase in the Pb-Zn ores, occurs as inclusions in Pb-Zn sulfide minerals. Where beds of economic sulfides are closely grouped together, they constitute an orebody. There are at least 14 distinct Pb-Zn orebodies in the deposit that stretch over a continuous strike length of more than 4 km. The inter-ore beds, composed mainly of dolomitic siltstone and shale, generally contain very little sulfide.

The stratiform Pb-Zn mineralization at Mount Isa is considered by most workers to be synsedimentary or syndiagenetic in origin (Stanton 1963, Mathias & Clark 1975, McClay 1983, Finlow-Bates & Stumpfl 1979, Russell et al. 1981, Neudert 1984, Valenta 1994). Perkins (1997, 1998), on the other hand, has argued for a syntectonic origin. His main arguments are: (a) in gross terms, fine-grained pyrite distribution in the Mount Isa mine is discordant with the stratigraphy (including time markers provided by tuff bands), indicating a large-scale, post-sedimentation replacement process; (b) the fine-grained pyrite appears to have overprinted the main tectonic cleavage (nearly bedding-parallel, S2); and (c) where there is an overprinting relationship between the fine-grained pyrite and the base metal sulfides, the pyrite is paragenetically earlier and either overgrown, or replaced, by the base metal sulfides, implying that the Pb-Zn mineralization was a late tectonic phenomenon. However, Valenta (1994) has reported that at the nearby Hilton mine, the fine-grained pyrite is overprinted by the tectonic cleavage and was deformed along with the Pb-Zn sulfides.

A special feature of the Mount Isa deposit (as well as of the Hilton deposit; Valenta 1994), compared with other giant Proterozoic Zn-Pb deposits, is the presence of major copper mineralization as irregular veinlets and patches. The copper orebodies, composed mainly of chalcopyrite, pyrrhotite, and cobaltite, are strata-bound within the Urquhart Shale and are hosted by the so-called "silica dolomite", an altered facies of the Urquhart Shale. The silica dolomite, as used in mine mapping, includes several related rock types characterized by brecciation and/or recrystallization, and either coarsely crystalline dolomite or high silica content (Mathias & Clark 1975, Perkins 1984). It truncates bedding, but lobes of alteration follow selected beds over distances of hundreds of meters and locally overprint Pb-Zn mineralization. The silica dolomite and the copper mineralization are most prominently developed in the southern part of the mine over the shallow-dipping contact with the Eastern Creek Volcanics, whereas toward the north the silica dolomite is split into a number of lobes (Fig. 11.7). Although the locus of the copper mineralization is south and downdip of the bulk of the Zn-Pb mineralization, individual Zn-Pb and Cu orebodies interfinger and both types of ores may be recovered from the same shaft.

The proposed models for the origin of the Cu ores and their relationship to the Zn-
Pb ores fall into five categories:

(a) Both Zn-Pb and Cu ores are submarine exhalative in origin, but formed in different sedimentary environments; later tectonic processes produced brecciation of the silica dolomite and selective redistribution of the copper mineralization (Stanton 1963, Mathias & Clark 1975).

(b) Both Zn-Pb and Cu ores are products of the same hydrothermal system; the Zn-Pb ores represent exhalative mineralization on the sea-floor, whereas the silica dolomite and Cu ore represent alteration and mineralization along conduits for hydrothermal solutions (Finlow-Bates & Stumpfl 1979), a model similar to that discussed earlier for VMS deposits, or replacement of the Zn-Pb mineralization (Goodfellow et al. 1993).

(c) The Zn-Pb and Cu ores are cognetic, formed during or soon after sediment deposition. Neither formed on the sea-floor and their physical separation simply reflects differences in the relative solubilities of these metals. Subsequent tectonic and hydrothermal activity resulted in textural modification and redistribution of the primary Cu mineralization, but the Zn-Pb orebodies, farther removed from the basement contact, were not affected (McGoldrick & Keays 1990).


(e) Both the Zn-Pb mineralization and Cu mineralization were produced by epigenetic replacement during tectonic deformation (Perkins, 1997, 1998, Perkins and Bell 1998).

Textural and structural studies (Perkins 1984, Swager 1985) provide convincing evidence that the silica dolomite is a syntectonic hydrothermal alteration product of the Urquhart Shale hosting the Pb-Zn orebodies and that the copper mineralization resulted mainly through replacement, with or without attendant silicification, of previous coarsely crystalline dolomite. That the copper mineralization was a later event than the Zn-Pb mineralization is also indicated by the localized sulfur isotopic equilibrium among the major sulfides of the Zn-Pb orebodies, presumably due to metamorphism, and a lack of such equilibrium between pyrrhotite and chalcopyrite in the copper orebodies (Smith et al. 1978). Fluid inclusion and stable isotope studies (Heinrich et al. 1989) suggest that the dolomitic alteration was produced by chemical interaction between the Urquhart Shale and moderate amounts of two fluids — a CO₂-rich, local metamorphic fluid of the Urquhart Shale, and a CaCl₂-rich fluid introduced from the underlying greenstones. This was followed by silicification and copper mineralization by a NaCl-rich fluid under conditions of high fluid:rock ratios and near-lithostatic fluid pressures. Thus, the Mount Isa copper mineralization appears to be epigenetic and unrelated to the Zn-Pb ores. In fact, a study of many copper prospects in the Mount Isa
Block led Van Dijk (1991) to conclude that syndeformational dolomitization and copper mineralization, as recognized at Mount Isa, was a regional event.

11.5.3. ZAWAR, RAJASTHAN, INDIA

The productive massive sulfide belt of Rajasthan (Deb & Sarkar 1990) is located in the Aravalli-Delhi orogenic belt that extends some 700 km in a southwesterly direction from Delhi (Fig. 11.8). Supracrustal rocks of the region comprise an early to middle Proterozoic (1.8 to 1.1 Ga) sequence of metamorphosed sedimentary and volcanic rocks that is divisible into six lithostratigraphic domains: Bhilwara, Aravalli, Jharol, North Delhi, South Delhi, and Vindhyan. The orogen is divided into two distinct terranes by the Rakhabdev lineament, a possible suture. The eastern terrane is underlain by an Archean basement composed of Banded Gneissic Complex (BGC), containing rocks as old as 3.5 Ga, and the large body of Berach Granite (2.6 Ga). The tectonic evolution of the Proterozoic sequence began with rifting of the Archean basement around 2.2 Ga and culminated with the development of an island-arc complex, represented by the South

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Figure 11.8. Simplified geologic map of a portion of the Aravalli-Delhi orogenic belt, Rajasthan, India, showing the location of major sediment-hosted Pb-Zn (± Cu) deposits: Ambaji-Deri (AD), Rajpura-Dariba (RD), Rampura-Agucha (RA), and Zawar (ZA). BCG = Banded Gneissic Complex, RL = Rakhabdev lineament. (Simplified from Deb & Sarkar 1990.)
Delhi belt, around 1.1 Ga. The Aravalli and Jharol belts (=1700 Ma) represent two segments of a single basin. The Aravalli belt comprises a sequence of carbonate-bearing shelf sediments that were deposited on a passive continental margin at the rifted western edge of the Archean basement complex and subsequently folded and metamorphosed to greenschist facies. The carbonate-free deep-sea facies of this basin represents the Jharol belt (Roy et al. 1984). Mafic rocks occurring in the basal part of the Aravalli sequence have been interpreted as ocean-floor basalts (Deb & Sarkar 1990). Granitic intrusives are rare in the Aravalli and Jharol belts, which host all the massive Zn-Pb sulfide deposits of the region. The deposits have undergone appreciable deformation and metamorphism, but are believed to be exhalative in origin, formed in extensional tectonic settings, such as intracratonic and back-arc basins.

The Zawar district, located in the Aravalli belt, is the most important Zn-Pb mining district in India. The district includes several deposits (Mochia, Balaria, Zawar Mala, and Baroi), with pre-mining reserves estimated at about 75 million tonnes of ore with a combined (Pb+Zn) content of 5-6%. The main rock types of the district are quartzite, metagreywacke, phyllite, carbonaceous phyllite, and dolostone. The ore mineralization is restricted to the dolostone and the mineralized lenses terminate at the contact with phyllites. The principal ore minerals are sphalerite, galena, and pyrite, with minor amounts of pyrrhotite, arsenopyrite, argentite and native silver. The ores contain only trace amounts of copper as chalcopyrite and native copper. The pyrite occurs as both frambooidal and non-framboidal forms, exhibiting various generations of diagenetic recrystallization, and is closely associated with sphalerite and galena suggesting a syngenetic origin of the sulfides (Chauhan 1984).

The Zn-Pb-Fe sulfide mineralization is of two types (Poddar 1965): (a) replacement bodies in dolostone along shear planes and later tensional fractures in zones of intense shearing, where primary stratification has been obliterated by deformation and recrystallization; and (b) stratiform mineralization, with well-preserved sedimentary structures, in dolostone free from considerable shearing and recrystallization. In the Mochia and Balaria mines, the mineralization is dominantly structurally controlled and was ascribed earlier to ascending hydrothermal solutions (Mookherjee 1964). The present consensus is in favor of synsedimentary-syndiagenetic sulfide mineralization with remobilization into structural planes during later deformation (Poddar 1965, Mukherjee & Sen 1980, Chauhan 1984).

11.5.4. SULLIVAN, BRITISH COLUMBIA, CANADA

The Sullivan deposit (Fig. 11.9), located near Kimberley, southeastern British Columbia, is the largest of several conformable Pb-Zn deposits (e.g., Kootenay King, North Star, and Vulcan) hosted by the Aldridge Formation of the Middle Proterozoic Purcell Supergroup. The Purcell Supergroup and its laterally equivalent Belt Supergroup in USA constitute a thick (>10,000 m) prism of dominantly clastic sediments that was probably deposited in an intracratonic basin (Hamilton et al. 1982). The Aldridge Formation, the lowest unit of the Purcell Supergroup exposed in the
Sullivan area, has been divided into three mappable units: the Lower, Middle, and Upper Aldridge. The Lower Aldridge is composed of a rhythmic succession of thin- to medium-bedded, very fine-grained quartz wacke that are commonly graded and interlayered with argillite and laminated sequences of dark mudstone. The Middle Aldridge is typically more arenaceous and coarse-grained, and has been interpreted as a turbidite sequence (Bishop et al. 1970). Thin-bedded to laminated carbonaceous mudstone is the dominant lithology of the Upper Aldridge. Pyrrhotite occurring as laminations and blebs is particularly abundant in the Lower Aldridge. The pyrrhotite is believed to be sedimentary, rather than metamorphic, in origin as its distribution throughout the Aldridge Formation is independent of metamorphic grade.

The Purcell sediments are associated with three igneous suites (Ethier et al. 1976, Hamilton et al. 1983). These are: (a) the Moyie (or Purcell) Sills (1433 ± 10 Ma, Zartman et al. 1982), a group of tabular gabbroic bodies that is most abundant in the Lower Aldridge Formation (the sill-like body of hornblende diorite in the Sullivan area belongs to this group); (b) the Nicol Creek Formation (or Purcell Lavas), dated at 1075 Ma (Hunt 1962), which occurs about 7,000 m above the Sullivan ore horizon (no flows have been found in the Aldridge Formation); and (c) the Hellroaring Creek Stock and related pegmatitic bodies (1300 Ma; Ryan & Blenkinsop 1971) that intrude metamorphosed and deformed Aldridge rocks and Moyie Sills in an area about 15 km southwest of the Sullivan mine. Based on the Pb-isotope data (Le Couteur 1979), the age of the Sullivan deposit is estimated to be ~1500 Ma (Campbell & Ethier 1983).

The Sullivan deposit, about 2,000 m (N-S) x 1,600 m (E-W) in lateral extent, is located near the top of the Lower Aldridge Formation, on the folded and faulted eastern limb of a broad north-trending anticline, and has the shape of an inverted and tilted saucer (Hamilton et al. 1983). It consists of a western proximal zone and an eastern distal zone, which differ significantly with respect to footwall rocks, style of mineralization, and associated hydrothermal alteration (Fig. 11.9). Pb:Zn ratios decrease from the western to the eastern zone.

The western zone, commonly greater than 60 m in thickness, is composed of a lower subzone of massive pyrrhotite and an upper subzone of massive to laminated pyrrhotite - sphalerite - galena (Fig. 11.9). In the latter, layering is defined by contorted wispy concentrations of galena and sphalerite in a pyrrhotite matrix and, toward the top part, by layers of fine-grained clastic sedimentary rock (Hamilton et al. 1983). According to Shaw and Hodgson (1980), the lower massive pyrrhotite subzone was originally composed of Pb-Zn-Fe sulfides, and the Pb and Zn leached from this zone by later, ore-stage fluids were redeposited at the sediment-seawater interface forming the overlying Pb-Zn-Fe ores. The eastern zone (maximum thickness 36 m) is conspicuously stratiform. It is characterized by sharply defined contacts between sulfide layers that differ strikingly in mineralogy and are conformable with the bedding of the intercalated and enclosing sedimentary rocks. Sedimentary features, such as soft-sediment slumps, graded beds, and sedimentary boudins, are observed in the eastern laminated ores. The ores in this zone are composed principally of pyrrhotite, sphalerite, and galena, but there is a gradual increase of pyrite relative to pyrrhotite
Figure 11.9. Schematic east-west cross section of the Sullivan deposit, British Columbia, Canada, showing the typical distribution of lithologic units and hydrothermal alteration (after Ethier et al. 1976). po = pyrrhotite.
quantities of tin, mainly as cassiterite, which occurs in the footwall fracture systems.

Also present in the footwall of the western ore zone is an extensive zone of tourmalinite, an extremely hard rock containing 30% or more tourmaline as felted masses of accicular to stubby crystals in a matrix of quartz, chlorite, K-feldspar, plagioclase, garnet, and minor muscovite (Hamilton et al. 1983). The tourmalinite zone is roughly funnel shaped with an elliptical top of approximately 1,400 m x 900 m, steep sides, and a vertical extent of at least 400 m. Within this zone, textures of breccia, footwall conglomerate, and normal bedding are well preserved, and crosscutting sulfide veins and discordant zones (1 to 30 cm wide) of albite - chlorite - pyrite - carbonate are common. In some places, the ore zone rests directly on top of the tourmalinite zone; elsewhere up to 5 m of tourmaline-poor sedimentary rocks separate the two. In the ore zone, tourmalinite occurs as tectonically rafted blocks, but more commonly disseminated tourmaline and rare patches of tourmalinite are erratically distributed in the hanging-wall overlying the western zone (Hamilton et al. 1983). Campbell and Ethier (1983) argued that the tourmaline formed from boron-rich fluids that invaded the Lower Aldridge Formation prior to the ore-forming event and attributed the source of boron to either leaching from the sedimentary pile or degassing of a granitic magma at great depth (probably the magma that ultimately produced the Hellroaring Creek stock and satellites). Hamilton et al. (1982, 1983), on the other hand, considered the footwall tourmalinization as a product of hydrothermal alteration by the ore-forming fluids.

The hanging-wall of the Sullivan deposit, comprising Middle Aldridge sedimentary units with interbedded Fe or Fe-Pb-Zn sulfide layers, is intensely altered in the western zone to a coarse-grained rock composed of albite, chlorite, carbonate and pyrite (Fig. 11.9). This alteration zone extends for as much as 100 m above the ore zone and has a gradational contact with unaltered Middle Aldridge rocks. The alteration also extends into the top part of the ore zone, where the normal pyrrhotite-rich Zn-Pb ore is altered to a pyrite - chlorite - carbonate assemblage containing only minor amounts of pyrrhotite, sphalerite, and galena. The hanging-wall alteration clearly represents a continuation of the hydrothermal activity beyond the sulfide deposition event.

The Sullivan deposit formed from hydrothermal exhalations on the sea-floor (Cominco staff 1972, Ransom 1977, Campbell et al. 1978, LeCouteur 1979, Hamilton et al. 1982, Campbell & Ethier 1983, Nesbitt et al. 1984). This model is consistent with the sulfur isotope data for the ore sulfides suggesting a marine sulfate origin for the sulfur (Campbell et al. 1978, 1980), lead isotope data suggesting a crustal source for the lead and, by analogy, of the other metals (LeCouteur 1979), and the footwall alteration associated with the deposit (Hamilton et al. 1982, 1983).

11.5.5. RAMMELSBERG, HARZ MOUNTAINS, GERMANY

The Rammelsberg deposit is the largest sediment-hosted, “stratiform” Zn-Pb-Ag-barite deposit in Central Europe. It had been worked to shallow levels for silver over many centuries before its real extent and potential was recognized in late 19th century. A
detailed account of this highly deformed deposit is provided by Hannak (1981), which is the main source of the summary presented below.

The Lower Devonian (Emsian) sequence in the Harz Mountain, deposited in a littoral-neritic environment, contains minor syngenetic and epigenetic sulfide mineralization. During Middle Devonian (Eifelian), differential subsidence of the Variscan Trough led to two facies of sedimentation: a shallow-water facies of sandy and calcareous shales in the West Harz Rise; and a deeper-water, basinal shale facies (Wissenbach Shales) in the Goslar Trough. The Rammelsberg deposit, hosted by the Lower Wissenbach Shales, is believed to have formed in a local depression on the seafloor close to the hinge zone between the Goslar Trough and the West Harz Rise. The shales that are contemporaneous with the ore horizon contain completely pyritized, dwarfed forms of brachiopods, goniatites, and tentaculites. Such dwarfed forms are typical of a biologically hostile environment, which was the consequence of metal-bearing sulfide solutions contaminating the seawater. More than 20 horizons of tuff, felsic to intermediate in composition, have been identified in the Lower and Middle Devonian strata below the ore horizon, but none above. A second period of regional volcanism, represented by diabasic greevstones, occurred during the late Middle Devonian to late Devonian time.

The Rammelsberg deposit consists of two sulfide orebodies which are designated, according to the relative dates of discovery, as the Old Orebody an the New Orebody. The Old Orebody has a strike length of 600 m, extends to a depth of over 3,000 m, and is approximately 15 m thick. The New Orebody extends to a depth of 500 m and has a true stratigraphic thickness of 10 m, but the thickness increases to about 50 m in the lower part of the orebody due to isoclinal folding. It is not certain whether the two orebodies are tectonically separated parts of an originally single orebody or they formed as separate entities. A further complication is the occurrence of a barite orebody ("gray ore") that overlies the Old Orebody and is physically separated from the sulfide orebodies. At present, the host rocks as well as the orebodies are part of a complicated isoclinal fold zone (the Orebody Syncline) overturned to the northwest, and the contacts between the orebodies and the host rocks are mostly marked by faults. The tectonic deformation, which occurred during the Variscan orogeny (late Carboniferous), has produced very complicated exposure patterns and fabrics within the orebodies as well as at their contacts with the host rocks.

Both orebodies are composed of two types of sulfide ores (Fig. 11.10): (a) massive ore, comprising layers with different proportions of pyrite, chalcopyrite, sphalerite, galena, and barite; and (b) banded ore, consisting of interlaminated stratiform sulfides (pyrite, sphalerite, galena, chalcopyrite) and black shale. The massive ore is characterized by colloform texture, including recrystallized frambooidal pyrite, and an overall Fe (bottom) \( \Rightarrow \) Cu \( \Rightarrow \) Zn \( \Rightarrow \) Pb \( \Rightarrow \) Ba (top) zoning. Sedimentary structures such as graded bedding, cross-bedding, current-erosion channels, load casts, and slump folds are preserved in the banded ores. Gray ore in the New Orebody consists of massive barite with sphalerite, galena, and pyrite; the Gray Orebody above the Old Orebody is characterized by barite-shale alternation with scattered pyrite.
The Rammelsberg orebodies are underlain by a discordant zone of "Kneist", a hard, partly brecciated and irregularly mineralized siliceous rock in which sulfides (sphalerite, galena, chalcopyrite, and pyrite) occur as veins and fracture-fillings. The Kneist has been interpreted by some as the discordant mineralization along the feeder zone to the overlying stratiform ore (Ramdohr 1953, Gunzert 1969), an interpretation consistent with the presence of a pronounced cone-shaped alteration zone in the footwall rocks with Fe-chlorites in the core and Mg-chlorites at its margin. Others have argued that the tectonic fabric and the lack of hydrothermal alteration in the Kneist indicate mineralization or remobilization subsequent to the formation of stratiform sulfides during a later tectonic event (Kraume 1960, Hannak 1981). A reconstruction of the

Figure 11.10. Cross section of the Rammelsberg New Orebody below the 9th level showing ore types and internal folds (after Kraume 1960).

Figure 11.11. Reconstruction of the pre-deformation disposition of the Rammelsberg deposit, assuming that the deposit originally comprised a single orebody (after Ramdohr 1953). Vertical zoning in the massive ore (from bottom to top): Cu-rich ore ⇒ Zn-Pb ore ⇒ Pb-Zn-barite ore.
Rammelsberg deposit prior to deformation, assuming a single original orebody, is presented in Fig. 11.11.

11.5.6. NAVAN, IRELAND

A number of sediment-hosted Zn-Pb±Ba±Ag deposits (e.g., Navan, Lisheen, Tynagh, Silvermines, Galmoy) occur in the Lower Carboniferous, dominantly carbonate rocks, of the Irish Midlands. The deposits are generally restricted to three stratigraphic levels of a marine shelf succession: the Navan Group and equivalents; the Waulsortian facies; and the Supra-Waulsortian facies (Hitzman & Large 1986). The Navan Group in central Ireland is a mixed sequence of peritidal carbonate, clastic, and locally evaporitic sediments. The Waulsortian facies is a distinctive lithologic package composed mainly of poorly fossiliferous, micritic limestone derived from a mudbank complex. Rocks of the Supra-Waulsortian facies are shallow-water shelf limestones and deeper-water carbonate and argillites. Upper Carboniferous rocks in the Irish Midlands are generally terrigenous and record the transition from marine to non-marine sedimentation.

Massive Zn-Pb mineralization in the Navan Group and Waulsortian facies is broadly strata-bound (truly stratiform mineralization is present only in a few Irish deposits such as Silvermines and Navan), and is commonly associated with crosscutting, stockwork, and vein-type mineralization that extend from the base of the carbonate sequence (e.g., Ballinalack deposit). Most of the Irish deposits are located adjacent to faults that were active during mineralization and served as fluid conduits. Mineralization in the Supra-Waulsortian rocks generally occurs as crosscutting zones, commonly in breccia bodies.

The Navan deposit (Andrew & Ashton 1985, Ashton et al. 1986) in east-central Ireland, with initial ore reserves of 70 million tonnes averaging 12.7% (Zn+Pb), is the largest Zn-Pb deposit in Ireland. It was discovered in 1970 as a result of regional, shallow soil geochemical sampling and has been in production since 1977. About 97% of the ore at Navan is hosted by the Pale Beds (Navan Group), which comprise a 200 m-thick suite of pelletal, oolitic and bioclastic calcarenites, locally containing significant quantities of quartz sand and darker argillaceous layers. The Pale Beds-hosted ore is subdivided vertically into 5 lenses. The rest of the ore occurs within the Boulder Conglomerate, which has been interpreted as submarine debris-flow material deposited on a pre-Arundian erosion surface. No mineralization has been found in the Upper Dark Limestones overlying the Boulder Conglomerate. The Navan orebody occurs on the northwestern flank of a complex anticlinal structure that has been severely dislocated by a major NE-trending fault system, termed the A-C Fault Complex (Fig. 11.12). Although this fault system has post-ore movement of presumed Hercynian age, it was active during pre-Carboniferous time and during the mineralization event in the early Carboniferous time (Ashton et al. 1986).

Sphalerite with lesser amounts of galena comprise the main ore minerals of the Navan orebody; Fe-sulfides (pyrite and/or marcasite) are locally dominant in the ores of Conglomerate Group and in some parts of the Pale Beds. Barite and carbonates are the main gangue minerals.
Ores hosted by the Pale Beds show both bedding-parallel (stratiform) and crosscutting styles of mineralization. In its simplest form, the bedding-parallel mineralization consists of locally well-laminated sphalerite interlayered with smaller quantities of galena and, occasionally, thin bands of barite. The layering, however, is frequently disrupted by contortions and pull-apart structures with dislocation and rotation of broken fragments. In the absence of evidence for the presence of precursor material, such as evaporite, amenable to selective replacement by sulfide, the combination of well-developed laminated sulfide and soft-sediment deformation has been interpreted as evidence for the deformation of synsedimentary sulfide layers (Ashton et al. 1986). Crosscutting mineralization occurs in discrete veins and breccia zones, which predominantly strike NE to ENE and cut across the sedimentary stratification. The larger veins locally show crustification banding of galena and sphalerite, suggesting fracture filling-type, rather than replacement-type, mineralization. Some veins, particularly those containing pyrite or marcasite, postdate bedding-parallel sulfides, but many merge into bedding-parallel sulfide layers. The increase in thickness of the bedding-parallel sulfide layers near the zone of intersection suggests that the veins, formed by filling of hydraulic fractures (Phillips 1972), acted as feeders for the bed-parallel sulfide layers (Ashton et al. 1986).
The conglomerate-hosted mineralization is patchy and dominated by pyrite and/or marcasite of variable textures, ranging from fine-grained framboidal laminae of presumed syngenetic/early diagenetic origin to complex intergrowths with carbonate minerals. The highest grade mineralization takes the form of massive sulfides with irregular patches and locally laminated sphalerite and subordinate amounts of galena. This mineralization is also considered to be synsedimentary to early diagenetic (Ashton et al. 1986), and its feeders probably are represented by the pyrite-rich veins that pass through the stratiform sulfides hosted by the Pale Beds.

11.6. Ore Composition

11.6.1. MINERALOGY AND TEXTURES

SMS deposits typically contain one (e.g., Meggen) or more (e.g., Navan, Mount Isa) tabular bodies of stratiform sulfides up to a few tens of meters in thickness and hundreds to thousands of meters in lateral dimensions. In strongly deformed deposits (e.g., Rammelsberg), it is difficult to determine if the observed orebodies were originally separate bodies emplaced at the same or different stratigraphic levels, or they represent dismembered fragments of an originally continuous orebody. Some deposits (e.g., Sullivan) contain more than one orebody at the same stratigraphic level and may represent proximal and distal accumulations from the same mineralization event. Stockwork and vein-type sulfide mineralization is commonly found underlying or adjacent to the stratiform ore and the contact between the two styles of mineralization is usually quite distinct (Large 1983). As discussed earlier, the crosscutting mineralization in some deposits (e.g., Sullivan, Tynagh, Silvermines, Navan) is believed to represent epigenetic mineralization in the feeder system to the overlying stratiform ores; in others (e.g., Mount Isa, Rammelsberg), its relationship to the stratiform ores is controversial.

The dominant sulfide minerals of the bedded ore facies of stratiform mineralization are pyrite and/or pyrrhotite, sphalerite, galena, and minor chalcopyrite, arsenopyrite, and marcasite. The ratio of iron sulfides to base-metal sulfides varies from <1:1 to >5:1 (Goodfellow et al. 1993). The pyrrhotite is probably a primary phase rather than a product of metamorphism (Finlow-Bates & Large 1978, Hamilton et al. 1983). The most common non-sulfide gangue minerals are barite and chert, but carbonate minerals (siderite, ankerite, calcite, dolomite, wetherite and other barium carbonate minerals) are quantitatively important in some deposits. Chert is spatially restricted to the ore deposits and probably represents an exhalative-hydrothermal silica phase (Large 1983). Layering defined by monomineralic layers of hydrothermal minerals, millimeters to tens of centimeters in thickness, is a conspicuous feature of the bedded ore facies. The distal sedimentary facies, which commonly has a gradational and assay-defined contact with the bedded ore facies, is simply a chemical sedimentary rock dominated by the more mobile or more stable and/or more oxidized, non-ore components of the bedded...
ore facies. The vent complex (see Fig. 11.1), where present (e.g., Tom and Jason deposits, Canada), appears to be the product of partial replacement of previously deposited bedded ores by upward-flowing hydrothermal fluids. Commonly, the complex is a heterogeneous mixture of massive sulfide, replacement patches, and irregular veins and/or disseminations of sulfides, carbonates, and silicates (mostly quartz). The mineral assemblage is dominated by pyrite, pyrrhotite, galena, sphalerite, ferroan carbonate, dolomite, quartz, and tourmaline, and lesser amounts of muscovite, chlorite,chalcopyrite, arsenopyrite, and sulfosalt minerals (Goodfellow et al. 1993).

Many SMS deposits, especially those of Phanerozoic age, are associated with significant amounts of barite (see Table 11.1). The barite may occur stratigraphically above the sulfides (e.g., Rammelsberg), peripheral to the sulfide horizon (e.g., Meggen), as thin interbeds within stratiform sulfides (e.g., Tom, Navan), or as massive bodies within the sulfide zone (e.g., Silvermines). Some SMS deposits have no associated barite (e.g., HYC, Mount Isa, Sullivan). There is no apparent correlation between the Zn:(Zn+Pb) ratios and barite content in SMS deposits (Lydon 1983). The uneven distribution of barite, even among deposits of approximately the same age and tectonic setting (e.g., Mount Isa and Lady Loretta), is an unresolved issue.

Metal zoning is common in stratiform ores, but generally less systematic than in VMS deposits. The general sequence of zoning suggested by Large (1983), \((\text{Cu}) \Rightarrow \text{Pb} \Rightarrow \text{Zn} \Rightarrow \text{(Ba)}\) in laterally zoned deposits and \((\text{Cu}) \Rightarrow \text{Zn} \Rightarrow \text{Pb} \Rightarrow \text{(Ba)}\) in vertically zoned deposits, is an oversimplification. Lateral zoning, with Zn:Pb ratios increasing away from faults and/or breccia zones that have been interpreted as “feeder zones”, is characteristic of both clastic-hosted and carbonate-hosted SMS deposits (Hamilton et al. 1983, Mathias & Clark 1975, Hitzman & Large 1986). The vertical zoning is much more variable. For example, upward in the stratigraphic sequence, Zn:Pb ratios tend to decrease at Rammelsberg, increase at Tom, Sullivan, and Navan, but show no systematic trend in HYC and Howard deposits (Lydon 1983, Ashton et al. 1986). With the exception of parts of the Rammelsberg deposit, chalcopyrite is usually not a significant component of the stratiform ores, but quite commonly an important phase in the stockwork part of the deposits. SMS deposits show no consistent trend in Cu:Zn or Cu:(Zn+Pb) ratios comparable to that in VMS deposits.

In addition to layering, the zone of stratiform mineralization commonly exhibits an array of sedimentary structures, such as graded bedding, cross-bedding, scours, slump folds, and intraformational breccia. Detailed structural analysis of Mount Isa and Sullivan deposits (McClay 1983) has shown that the major and minor structures in these deposits are overwhelmingly tectonic in origin, but synsedimentary folds produced by slumping can still be recognized. The latter are characterized by random structures with no particular geometric relationship to large-scale folds and a lack of axial planar sulfide fabrics or mobilization of sulfides to fold hinges. Evidence in favor of syngenetic emplacement of the stratiform sulfides is provided by the presence of intervals, at scales of a few centimeters or more, of irregularly folded, conformable sulfide laminae that are succeeded by undisturbed sulfide laminae (Lambert 1983).
11.6.2. BULK COMPOSITION

Almost all SMS deposits are bimetallic in Zn and Pb, but statistical correlation between the two metals is weak. Average Zn grades range between 2 and 10 wt%, with a maximum of about 19 wt%; Pb grades cluster between 1 and 4 wt%, with a maximum of just over 11 wt% (Sangster 1990). An analysis of bulk compositions of SMS deposits by Lydon (1983) has revealed several interesting features. The average grades and tonnages of SMS deposits (see Table 11.2) are much higher than those of VMS deposits. For example, the weighted average of 144 VMS deposits is 7.78 million tonnes of ore grading 1.25% Cu, 0.73% Pb, and 3.76% Zn, giving 0.45 million tonnes of combined metal; the weighted average of 38 SMS deposits is 59.57 million tonnes of ore grading 0.07% Cu, 2.98% Pb, and 7.29% Zn, which works out to 6.75 million tonnes of combined metal. Both VMS and SMS deposits have a bimodal distribution of Zn:(Zn+Pb) ratios. For VMS deposits, the two populations, with modes at, 0.70-0.85 and >0.90, correspond to felsic volcanic and/or sedimentary rock association and mafic volcanic rock association, respectively. Metal ratios of SMS deposits also have a bimodal distribution, with mean Zn:(Zn+Pb) ratios of approximately 0.50 and 0.75 for the two populations (Sangster 1990). According to Lydon (1983), SMS deposits with the lower ratios tend to be hosted by clastic-dominated sedimentary sequences, whereas those with higher ratios occur in carbonate-dominated sequences.

There is no systematic difference in the distribution of Ag between the SMS and VMS deposits (Gustafson & Williams 1981). The Ag contents of SMS deposits are highly variable. For example, of the 38 SMS deposits listed by Lydon (1983), Ag grades are not reported for 10 deposits, probably because they are too low (<10 g/t) to be of economic significance, and 6 deposits have >100 g/t of Ag, the highest being ~150 g/t in the Hilton deposit (Australia). An overall positive correlation between Ag and Pb in SMS deposits indicates galena to be an important carrier of Ag, but a significant amount of Ag is also contained in other sulfide and sulfosalt minerals in some deposits, such as Mount Isa, Broken Hill, and Sullivan. The most important of these are tetrahedrite, pyrargyrite, and argentite (Riley 1974, Both & Stumpf 1987).

11.7. Hydrothermal Alteration

Compared with VMS deposits, mineralization-related hydrothermal alteration associated with SMS deposits, generally, is quite subtle. This is so perhaps because of the less reactive nature and lower permeability of the fine-grained siliciclastic sediments that underlie most SMS deposits (Goodfellow et al. 1993). Examples of subtle hydrothermal alteration include the development of muscovite at the expense of chloride and biotite in the immediately adjacent wallrocks of the Anvil district (Shanks et al. 1987), and an increase in dolomite:calcite ratio and in Mn and Fe contents of dolomite toward the deposits such as HYC, Mount Isa, Silvermines, and Navan (Lambert 1983,
As discussed earlier, the "silica dolomite" of the Mount Isa deposit and the Kneist of the Rammelsberg deposit may represent zones of discordant, Cu-enriched mineralization and hydrothermal alteration comparable to those found in VMS deposits. In fact, the Rammelsberg deposit is associated with a cone-like, chlorite-rich footwall alteration zone (Renner & Brockamp 1985). The most pronounced hydrothermal alteration occurs in the footwall of the Sullivan deposit (Hamilton et al. 1983). Although tourmaline is a minor mineral in the alteration assemblages associated with many massive sulfide deposits (Taylor & Slack 1984, Slack & Coad 1989), the extent of tourmalinization at Sullivan is atypical of VMS and SMS deposits.

Hydrothermal alteration of the hanging-wall sequence (e.g., the albite - chlorite alteration at Sullivan) and the occurrence of low-temperature hydrothermal sediments (such as barite, phosphate, silica, and pyrite) in some deposits indicate that hydrothermal systems responsible for the formation of SMS deposits continued to operate, albeit at a lower intensity, for some time after ore deposition.

### 11.8. Origin

#### 11.8.1. GENETIC MODEL

SMS deposits are believed by most workers (e.g., Hutchinson 1980, Large 1980, 1983, Lambert 1983, Sangster 1990, Goodfellow et al. 1993) to be dominantly syngenetic in origin, formed in an exhalative environment (Fig. 11.13). Such a model is compatible with the following characteristic features of SMS deposits:

(a) the stratiform nature of sphalerite - galena - pyrite - barite mineralization (i.e., compositional layering parallel to bedding in the enclosing sediments), especially at the peripheral parts of orebodies;
(b) the great lateral extent of the finely laminated sulfides at a single stratigraphic level;
(c) the sharp contacts between the sulfide layers composed entirely of hydrothermal minerals and interbeds containing only trace amounts of the same hydrothermal minerals;
(d) the associated sedimentary structures such as graded bedding, cross-bedding, scours, and sulfide clasts;
(e) intervals of disharmonic folding of sulfide layers that are overlain by undisturbed layering; and
(f) lateral and vertical metal zoning that is not consistent with a replacement origin, and discordant footwall mineralization and alteration in some deposits that may be reasonably interpreted as conduits of fluid discharge.

The majority of SMS deposits are not underlain by a discordant feeder/alteration zone
Some features of SMS deposits, especially those associated with carbonate rocks, appear to be syndiagenetic rather than synsedimentary or synmetamorphic. These include: contortion of crosscutting sulfide veins during sediment compaction at Navan (Ashton et al. 1986); association of stratiform sphalerite and galena with diagenetically recrystallized pyrite at Zawar (Chauhan 1984); replacement of stratiform pyrite, barite, and carbonate by Zn-Pb sulfide minerals at Silvermines (Andrew 1986); the occurrence of Zn-Pb sulfides as internal sediments in dissolution cavities at Tynagh (Boast et al. 1981); and the development of coarse-grained patches and large crystals of sulfides within fine-grained ores at Mount Isa (Lambert 1973). Perhaps the best case for predominantly syndiagenetic mineralization in a SMS deposit is provided by the HYC deposit in the McArthur basin of Australia.

The Irish deposits contain a variable combination of synsedimentary, syndiagenetic, and epigenetic mineralization, all related to the same mineralization event, and this may be a logical consequence of the ore-forming environment. Whether the mineralization was synsedimentary or syndiagenetic at a given locality was probably controlled by the relative rates of sedimentation and sulfide accumulation. For example, at Navan synsedimentary mineralization of an exhalative nature appears to have stayed abreast of rapid sedimentation, but mineralization also occurred at the same time in a range of diagenetic environments in the shallow-water carbonate sediments undergoing lithification (Andrew & Ashton 1985). The minor, vein-type Zn-Pb±Cu mineralization present in the older footwall sequence of many deposits (e.g.,
Silvermines, Navan) is clearly epigenetic.

Finlow-Bates and Large (1978) proposed that exhalative mineralization in shallow marine environments would tend to have a large proportion of epigenetic crosscutting mineralization with correspondingly less stratiform mineralization (e.g., Silvermines); the opposite would be the case in deep marine environments (e.g., Rammelsberg). This is because boiling of the hydrothermal fluids within the feeder zone would cause a rapid drop in temperature and, possibly, hydraulic fracturing, thus promoting crosscutting mineralization. However, fluid inclusion data do not support boiling as a major cause of mineralization at Silvermines (Samson & Russell 1987, Hitzman & Beaty 1996). The lack of evidence for boiling of the ore fluids, combined with the fluid inclusion data on temperature and salinity, were used by Sangster (1990) to constrain the minimum water depth of sulfide precipitation in selected SMS deposits to about 50 to 1,500 meters. The common occurrence of epigenetic mineralization in many Irish deposits, compared with typical SMS deposits, appears to be related to the presence of permeable and reactive carbonate host rocks (Lower Carboniferous) that consumed the mineralizing fluids before they could reach the sea-floor (Hitzman & Beaty 1996).

Most SMS deposits are associated with synsedimentary faults (Large 1983, Goodfellow et al. 1993). Evidence for contemporaneous fault movements at or close to the site of stratiform mineralization includes facies and thickness variation of host sediments across the fault, slump breccias and conglomerates, soft-sediment deformation indicated by local zones of folding and faulting of greater intensity relative to regional deformation, and the presence of sulfide ore fragments, sometimes with contorted sulfide laminae, within the breccias (Large 1983). The synsedimentary faults are believed to have controlled the generation of morphological traps (third-order basins) for accumulation of sulfides and provided near-surface, high-permeability channelways for ore fluids from deeper regions, probably through faults or fracture zones associated with local rifting or rapid differential subsidence of underlying sedimentary strata (Carne & Cathro 1982, Lydon 1983). They may also have initiated or accelerated convective circulation of overlying seawater to deeper levels of the crust (Russell et al. 1981).

11.8.2. ORE-FORMING FLUIDS

Very little fluid inclusion data, the main source of direct information on ore fluids, have been reported from the clastic-hosted SMS deposits because of the fine-grained nature of the primary sulfide minerals, the lack of associated coarse-grained gangue minerals, and the recrystallization of hydrothermal minerals by diagenesis and burial metamorphism. Estimates of temperature for these deposits, based on sulfur and/or oxygen isotopic ratios of coexisting mineral pairs, range from a low of <100°C for the Meggen deposit (Nielsen 1985) and ≈150°C for the Sullivan deposit (Nesbitt et al. 1984) to a high of ≈300°C for the Mount Isa and HYC deposits (Smith & Croxford 1973, Smith et al. 1978, Rye & Williams 1981). The lack of significant amounts of copper in most SMS deposits suggests that the temperature of the ore-forming fluid was generally below 300°C. [The solubility of chalcopyrite is <1 ppm Cu below 300°C at seawater
chloride concentrations and an oxygen partial pressure within the stability fields of sulfides (Lydon 1986). This is consistent with the average homogenization temperatures of ≈260°C and salinities of ≈9 wt% NaCl equivalent reported for Tom (Ansdell et al. 1989) and Jason (Gardner & Hutcheon 1985) deposits in Canada.

Fluid inclusion studies on the carbonate-hosted Irish deposits, generally limited to coarse-grained sphalerite and hydrothermal gangue minerals (dolomite, quartz, barite), indicate wide ranges of fluid temperature (<100° to ≈250°C) and salinity (<10 to ≈30 wt% NaCl equivalent). From an examination of all available fluid inclusion and stable isotopic thermometric data, Hitzman and Beaty (1996) concluded that the temperatures and salinities of fluids responsible for the mineralization in the Irish deposits were ≈100-190°C and ≈10-23 wt% NaCl equivalent, respectively. An overall negative correlation between homogenization temperature and salinity for quartz-hosted fluid inclusions from the Silvermines deposit has been interpreted as evidence of fluid mixing during the mineralizing process, both within and below the ore zone (Samson & Russell 1987). The fluids involved were Ca-depleted, one a high-temperature fluid of lower salinity (8-13 wt% NaCl equivalent) and the other a low-temperature fluid of high salinity (18-22 wt% NaCl equivalent). The presence of two fluids differing in temperature and salinity has also been inferred for mineralization at Lisheen (Thompson et al. 1992) and Tynagh (Banks & Russell 1992).

11.8.3. HYDROTHERMAL SYSTEMS

Many SMS deposits are associated with thin tuffite horizons (see Table 11.1) indicative of minor, penecontemporaneous volcanism. The eruptive centers of the tuffites have generally not been identified, but the fine-grained nature of the tuffites and the lack of lapilli in them suggest that the volcanic source must have been distant from the site of mineralization. In some areas, such as Ireland and the Rhenish Basin (Meggen deposit), there is evidence of an increase in volcanic activity during the mineralization event, but the two are not spatially related. A genetic connection between SMS deposits and magmas appears unlikely in view of a lack of coeval plutons either of sufficient size to have sustained hydrothermal convection for sufficient duration or of appropriate composition to have exsolved large quantities of Zn-Pb-rich fluids. Nevertheless, the time-frame of igneous activity in many areas containing SMS deposits does indicate higher than normal geothermal gradients during the mineralization event.

Considering the large tonnages of metals contained in individual SMS deposits and the low solubility of metals in saline hydrothermal fluids, the formation of a typical SMS deposit must have involved sustained flow of a large volume of ore-forming fluids. Moreover, the large ranges of fluid inclusion homogenization temperature and salinity for SMS deposits, often in the scale of an individual deposit (e.g. Silvermines), emphasize the complexity of the associated hydrothermal systems. Proposed models of fluid circulation can be grouped into three end-members: (a) one-pass flow along pressure gradients of basinal fluids expelled by sediment compaction (Walker et al. 1977, Williams 1978b, Badham 1981, Carne-Cathro 1982, Lydon 1983, Sawkins
1984); (b) thermally driven convection of seawater within an extensional basin to progressively deeper parts of the crust (Russell et al. 1981, Russell 1983); and (c) gravity-driven flow in a foreland basin (Hitzman & Beaty 1996), the model favored by many for Mississippi Valley-type deposits (see Ch. 13). The compaction driven fluid flow (stratal aquifer model) appears to be the more popular one at present. Modeling by Person and Garven (1994) suggests that compaction-driven groundwater flow dominates in the onlap facies during the thermal cooling (flexural) stage of the evolution of a continental-rift basin (when laterally extensive onlap facies are developed), as well as in the lacustrine facies deposited in the initial (stretching) phase of rifting.

The temperature and composition of the fluids derived by basin dewatering depends on the geothermal gradient and the minerals with which the fluid is in equilibrium. Dissolution of evaporites or the evaporation of marine or continental waters is the most direct way for achieving high salinities in basinal waters. The likely importance of this process is underscored by the observation that most SMS deposits lie within about 30° of the equator in paleogeographic reconstructions of land masses (Turner 1992). A possible alternative for producing high-salinity fluids is the reaction of connate fluids with the enclosing sediments. Computations by Lydon (1983) showed that under reasonable geothermal gradients connate fluids in a reservoir of arenaceous and argillaceous rocks would be buffered to a pH of about 4.5, attain temperatures of 100-200°C and salinity up to ~25 wt% NaCl equivalent, and be capable of carrying high concentrations of Zn, Pb, Ba, Fe, and Ag as chloride complexes, but not much Cu at temperatures below 200°C. The metal ratios in the fluids and in the resulting SMS deposit would depend mainly on the fO2 of the reservoir (reaction) zone.

Conditions favorable for the stratal aquifer model are (Lydon 1983, 1986): (a) a syn-rift clastic sequence containing adequate connate water; (b) an impermeable cap of argillaceous sediments that helps to maintain elevated temperatures in the underlying reaction zone by minimizing conductive and convective heat loss and to promote geopressuring of the hydrothermal reservoir; and (c) extensional faulting that provides a conduit for cross-stratal upward flow of the hydrothermal fluids to the sea-floor. As the basin dewatering is likely to be episodic (Cathless & Smith 1981), the stratal aquifer model offers a mechanism for the formation of stratigraphically stacked ore lenses separated by barren layers of host rock, each ore lens representing a separate influx of ore fluids (Sawkins 1984), although alternative mechanisms such as intermittent precipitation from a virtually stagnant brine pool or periodic influxes of sediments are possible (Lydon 1983). The model is also compatible with the close spatial association of SMS deposits with synsedimentary faults, because upward movement of high-salinity, hence high-density, brines by release of an over-pressured reservoir system or by seismic pumping (Sibson et al. 1975) would be facilitated by faulting. A criticism of the brine expulsion model is that the temperatures of ore fluids for SMS deposits range up to about 300°C, significantly higher than would be predicted by the model (Russell 1983). Lydon (1986) has pointed out that hydrothermal fluids for the Silvermines deposit in Ireland record an evolutionary path characterized by decreasing salinity with increasing temperature. This is the trend expected from progressive
tapping of a geopressed stratal aquifer, as saline pore waters of the reservoir are progressively diluted by water released from clay minerals with increasing temperature.

Both models envisage extensional structures as paths of fluid discharge, but they predict different spatial relationship among deposits in a given region. A cluster of deposits would reflect the distribution of favorable structural or stratigraphic traps according to the basin-expulsion model, but would require the operation of a number of convection cells, each with its own area of influence, for the convection model. A large mineralization trend, such as the 5 km-long trend in the Navan area, is difficult to reconcile with the convection model, but is compatible with the basin-expulsion model as representing suitable traps along an aquifer (Hitzman & Large 1986).

An attractive feature of the convection model (see Fig. 2.27e) is that it allows for variable temperature and composition of the hydrothermal fluids depending on the depth of penetration of the convection cell. For example, the later copper mineralization at Mount Isa may simply reflect significant (~1 ppm) copper solubility in fluids that had attained higher temperatures because of penetration into basement rocks at greater depths. Using a thermal balance model of hydrothermal circulation, assuming a maximum temperatures of 200-500°C for the fluids and a porous medium, Sterns et al. (1987) computed that such a system would be capable of forming very large Zn-Pb deposits, provided that the metals were available to the system. But the question is whether rocks in a sedimentary basin are permeable enough for downward penetration of connection cells to depths of 6 to 10 km as envisaged in this model. The downward propagation of microfracturing necessary for this process requires the presence of extensively jointed rock (Pine & Batchelor 1984), but the sediments underlying SMS deposits were probably not lithified enough to be jointed. On the other hand, Mills et al. (1987) have argued that a progression to less radiogenic lead with time, as shown in their study of the Navan deposit, is best explained by the convection cell model that envisages extraction of metals from progressively deeper levels of the crust. The convection cell model is also consistent with the lead isotopic data for the Irish deposits suggesting a dominantly basement source for the lead (Caufield et al. 1986, LeHuray et al. 1987). A convection-cell model has also been proposed for the sea-floor sulfide accumulations in Middle Valley and the Guaymas Basin, although in these cases the convection cells appear to be driven by the heat of high-level magma bodies.

The foreland basin model proposed by Hitzman and Beaty (1996) for the Irish deposits envisages that the hydraulic head for gravity-driven flow was provided by the Lower Carboniferous uplift of the Hercynian mountains south of Ireland. The Old Red Sandstone (Upper Devonian to Lower Carboniferous) served as the principal aquifer for fluid migration. The migrating formation waters were heated either by relatively deep flow paths, or by the extension-related high heat flow along the flow path. Gravity-driven models, however, do not explain the correlation between SMS deposits and continental rifting and the localized submarine discharge of hydrothermal fluids at cratonic margins (Goodfellow et al. 1993).

A notable aspect of the giant SMS deposits (e.g. HYC, Mount Isa, Sullivan) is the considerable stratigraphic interval over which conformable mineralization occurs. In
addition, the hanging-wall in many SMS deposits shows hydrothermal alteration and precipitation (see Fig. 11.1). Thus, irrespective of the model, the hydrothermal system for these deposits must have remained active intermittently for millions of years. This is in sharp contrast to VMS deposits which probably formed over time periods of only hundreds of years (Sawkins 1983).

11.8.4. ORE ACCUMULATION

The mechanism of ore accumulation in SMS deposits is less clear than for VMS deposits, because modern analogs of SMS deposits are rather limited. That the process for proximal SMS deposits might have been initiated by a VMS-style build-up of sulfide mounds where feeders debouched on to the basin floor derives some support from the recent discovery of fossil hydrothermal chimneys at Silvermines (Boyce et al. 1983) and Tynagh (Banks 1986). These are tabular, concentrically zoned growth forms of pyrite-marcasite up to 5 cm long and 2 cm in diameter, resembling miniature versions of 'black smoker' chimneys of the East Pacific Rise at 21°N. Alternatively, the bedded ores may represent either a shower of sulfides onto the sea-floor from buoyant hydrothermal plumes or an apron of clastic sulfides shed from sulfide mounds (Goodfellow et al. 1993). The tabular morphology of the stratiform mineralization as well as the great lateral continuity of individual beds and laminae, however, are more in accord with sulfide accumulation of in brine pools formed at topographic lows on sea-floor, similar to the accumulation of metalliferous sediments in the Atlantis II Deep, Red Sea (see Ch. 2). Changes in the geochemical parameters (P, pH, $f_O^2$, sulfide concentration, salinity) required for precipitation of the observed sulfide assemblages in SMS deposits have been modeled assuming appropriate boundary conditions (e.g., Russell et al. 1981, Russell 1983, Lydon 1983, Large et al. 1998), but these models do not uniquely define the mechanism(s) of sulfide precipitation.

11.9. Metallogenesis

11.9.1. HOST ROCKS

Large (1981, 1983) characterized the geologic settings of SMS (SEDEX) deposits in terms of first-order, second-order, and third-order basins, a hierarchy of sedimentary basins of decreasing size. The first-order basins may be either epicratonic embayments into continental margins, usually initiated during extension of a passive continental margin, or intracratonic basins. They are commonly fault-bounded, half-graben features of lateral dimensions in excess of 100 km, containing thick sequences of clastic and/or carbonate sediments that were deposited during a prolonged period of crustal stability. The sedimentary facies vary from shallow-marine evaporite-bearing carbonates and sandstones (e.g., Batten Trough, Leichardt River Fault Trough), to delta-front sandstones and siltstones (e.g. Belt-Purcell Basin), to deeper marine turbidities, pelagic
carbonates, and shales (e.g., Selwyn Basin). Synsedimentary differential subsidence within first-order basins gives rise to second-order basins of tens of kilometers in lateral dimensions flanked by rises (e.g., the Goslar Trough and the West Harz rise within the Variscan Trough). The presence of second-order basins can be identified by variation in sedimentary facies and thickness revealed through detailed geologic mapping. Third-order basins are relatively small depressions, several hundred meters to a few kilometers in lateral dimensions, and they represent actual depocenters of massive sulfide formation. Sediments that host SMS deposits in third-order basins may be viewed as a combination of "autochthonous" lithologies (fine-grained clastics and carbonates) and "allochthonous" lithologies (conglomerates, intraformational breccias, and coarse clastic sediments that are commonly interbedded with fine-grained sediments). The recognition of locally derived allochthonous lithologies helps in the identification of third-order basins, an important step in the exploration of SMS deposits. Sulfide deposits often occur close to the fault-controlled margins of the first- or second-order basins. As mentioned earlier, these faults were active during sedimentation and are believed to have provided the main conduit for the ore-forming fluids. Reactivation of these faulted margins may produce distinctive lineaments, such as the Emu fault (Fig. 11.5), the Mount Isa fault (Fig. 11.7), and the Balve-Overscheld fault zone at the western end of the Meggen basin (Werner 1989).

It is apparent from an examination of the host rocks that SMS deposits have formed both in the deep-water ('black shale') environment as well as in shallow basins dominated by carbonates and evaporites (Sangster 1990). Also, there is no simple relationship between host lithology and other features of SMS deposits. For example, the host lithology and tectonostratigraphic setting of Rammelsberg and Mount Isa deposits are quite different, but they are very similar in many other respects. The control of host lithology on SMS mineralization appears to be indirect, reflecting factors such as the distribution of organic matter and/or fortuitous proximity to plumbing systems (Gustafson & Williams 1981).

11.9.2. SOURCE OF SULFUR

From a comparison of $\delta^{34}$S values of sulfides and barite with coeval seawater sulfate (see Fig. 4.4), Sangster (1990) concluded that the ultimate source of sulfur for SMS deposits was seawater sulfate. The interpretation of the sulfur isotopic data for individual deposits/districts (Fig. 11.14), however, is more complicated.

The close agreement between $\delta^{34}$S values of coeval seawater and barite at Tynagh, Silvermines, Rammelsberg, Meggen, and in the deposits of Selwyn Basin suggests that the sulfate in these deposits was derived from marine sulfate in formation waters. The unusually high $\delta^{34}$S of barite at Lady Loretta is probably a reflection of the restricted nature of the depositional basin in which the $\delta^{34}$S of the residual seawater became higher after the precipitation of diagenetic pyrite.

In contrast to barite, $\delta^{34}$S values of sulfide phases show considerable variation among deposits and even within a single deposit. The generally large variation in $\delta^{34}$S
Figure 11.14. The range of $\delta^{34}S$ values in sulfides and barites of selected SMS deposits. Sources of data: compilation by Large (1981), Boast et al. (1981), Deb & Sarkar (1990), Eldridge et al. (1993), and Goodfellow et al. (1993).

of pyrite and an increase in $\delta^{34}S$ of pyrite stratigraphically upward suggest the precipitation of pyrite by biogenic reduction of seawater sulfate in a restricted basin (see Ch. 4). This interpretation is consistent with the observed $\delta^{34}S$ variation in the Selwyn Basin deposits (Anvil, early Cambrian; Howrad's Pass, early Silurian; Tom, Jason, Cirque, and Driftpile, late Devonian) that indicates the precipitation of pyrite with highly positive $\delta^{34}S$ values (about +30‰ or higher) under periodic anoxic
bottom-water conditions on a global scale in the Phanerozoic oceans (Goodfellow 1987). The highly variable isotopic composition of pyrite in the Rammelsberg deposit, which is not systematically correlated with the stratigraphy, is probably related to varied microenvironments at the site of biogenic pyrite precipitation (Large 1983) or, as argued by Goodfellow et al (1993), to a prolonged episode of bacterial sulfate reduction in the ambient water column.

The isotopic compositions of sphalerite and galena show no consistent pattern. For example, $\delta^{34}S$ values of galena and sphalerite systematically decrease upward in the Sullivan deposit (Campbell et al. 1978), increase upward in the Rammelsberg main ore bodies (Anger et al. 1966), and show no obvious stratigraphic relationship in the HYC and Lady Loretta deposits (Smith & Croxford 1973, Carr & Smith 1977). The frequency distribution of $\delta^{34}S$ values of galena and sphalerite are also highly variable, from modes centered around 0‰ (e.g., Sullivan, Broken Hill, Silvermines ‘Lower G’ orebody) to strongly positive values (e.g., Mount Isa, Rammelsberg, Meggen) and to strongly negative values (e.g., Silvermines ‘B’ and Tynagh ‘Upper G’ ore bodies).

In some SMS deposits, $\delta^{34}S$ values of sphalerite are higher than those for coexisting galena, as would be expected for a sphalerite-galena pair precipitated in equilibrium from the same solution (see Ch. 4). In contrast, there is commonly no evidence for isotopic equilibrium in the pyrite - galena and pyrite - sphalerite pairs, implying that the pyrite-sulfur had a different origin or that the pyrite precipitated under different physico-chemical conditions. Some interpretations of the sulfur isotope data for SMS deposits favor a dual sulfur-source model (Anger et al 1966, Smith & Crawford 1973, Large 1980,1983): (a) seawater sulfate at the site of accumulation for the barite and biologically reduced seawater sulfate for most of the pyrite; and (b) a “deep-seated” source for the sulfur in galena, sphalerite, pyrrhotite, and some of the pyrite, sulfur that was introduced into the site of mineralization in the same hydrothermal fluids that transported the metals. The coexistence of biogenic and hydrothermal sulfide may be explained by one or both of the following situations: (a) the discharge of hot hydrothermal fluids on the seafloor was episodic, allowing the bacteria to be active during the intermittent periods, and (b) the sphalerite and galena formed later than the bulk of the pyrite, as suggested by the textural relationships in many SMS deposits. The ultimate origin of the ‘deep-seated’ sulfur might be magmatic, but more likely, considering the paucity of contemporaneous volcanism, it was seawater sulfate that was inorganically reduced at high temperature (<200°C), perhaps during circulation of seawater through the underlying sedimentary prism. Other suggested sources of sulfur in sphalerite and galena include seawater sulfate at the site of mineralization (Campbell et al. 1978, Campbell & Either 1983, Shanks et al. 1987) and diagenetic pyrite and/or evaporite in the sediments within the reaction zone of a sea­floor hydrothermal system (Hitzman & Large 1986).

A hydrothermal sulfur source may have been important in some SMS deposits, such as the Rammelsberg deposit, which have sulfide-rich feeder zones, but most deposits appear to have obtained their reduced sulfur (H$_2$S) from the sulfate of the ambient seawater column (Turner 1992, Goodfellow et al. 1993). This conclusion is
supported by the low H₂S content of modern metal-rich basinal and geothermal waters, and the evidence for anoxic ambient environments during the formation of SMS deposits at least during the lower Paleozoic. The evidence includes non-bioturbated laminated host rocks, high S:C ratios in coeval sediments, preservation of graded pyrite framboid beds, and the highly positive δ²⁹S values of the pyrite. This anoxic condition was also the likely reason why SMS-type sulfide accumulations on the seafloor were protected from destruction by oxidation.

11.9.3. SOURCE OF ORE METALS

There is little direct evidence for the metal source for SMS deposits, because the hydrothermal reservoir (or reaction) zone of the hydrothermal system responsible for a SMS deposit has not been positively identified in any specific case (Goodfellow et al. 1993). Lydon (1983) presented arguments, based on the geochemistry of the leaching process and the composition of oil field brines, that the source of metals for SMS deposits was primarily the underlying sedimentary pile. The lead isotope data (Fig. 11.15), however, are inconsistent with such a conclusion in some cases and equivocal in others. For example, the best fit line for the Irish deposits corresponds to a source 'age' of about 2350 Ma; Caulfield et al. (1986) interpreted this distribution to indicate mainly a lower crustal origin for the lead in Irish deposits. The isotopic ratios for the Australian deposits suggest either a common, homogeneous source of lead or, more likely, a smoothing of the isotopic variation in different source materials by a regionally extensive and long-lived hydrothermal system (Gulson 1985). Lambert (1983) has reported that the present-day lead isotopic ratios in the sedimentary strata underlying the HYC deposit define a linear array which extrapolates to pass through the ore lead. This relationship is compatible with derivation of the ore metals from such strata, but could be explained as well by addition of ore lead to the strata.

With minor exceptions, the lead isotope ratios of SMS deposits fall between the mantle and upper crust growth curves (Fig. 11.15). According to Doe and Zartman (1979), the lead isotopic characteristics of SMS deposits are more likely the result of a complex process involving the mixing and homogenization of lead from several isotopically heterogeneous sources during the orogenic melting of crustal rocks, the reworking of sediments, the leaching of sediments by circulating hydrothermal fluids, and the transportation of lead to the site of mineralization in these fluids. A major limitation in the use of lead isotopic ratios for deciphering the source of ore metals in SMS deposits is the uncertainty regarding the initial lead isotopic compositions of the associated sediments.

Metal leaching by hydrothermal reactions, which require high water:rock ratio, is unlikely to result in a hydrothermal fluid with more than a few ppm of (Pb+Zn), (see Ch. 2) — not enough to form a significant SMS deposit. Therefore, sialic crystalline crust is not a good candidate for the reaction zones of SMS-forming hydrothermal systems. On the other hand, pore fluids of a sedimentary sequence, undergoing the first cycle of burial and metamorphism, may achieve high enough concentrations of
(Pb+Zn) to qualify as ore-forming fluids. Goodfellow et al. (1993) have outlined three favorable scenarios for the generation of viable ore-forming fluids for SMS deposits: (a) the burial and compaction of argillaceous sediments containing evaporites, in which thermal metamorphic reactions are driven by an increase in temperature due to burial; (b) a sedimentary sequence containing highly saline formation waters in which local prograde reactions are driven by the heat from a high-level magma body; and (c) the subsidence of an intracontinental rift to the point of allowing a marine transgression. The last one, they conclude, provides the best opportunity for forming SMS deposits.

A related question is whether the Proterozoic sediments contained enough lead to account for the vast quantities of lead in giant SMS deposits. The question is less critical to the generation of major SMS deposits in Phanerozoic time, because by then the concentration of lead in the upper crust was probably much higher. Sawkins (1989) argued that a critical factor in the formation of giant SMS deposits during the Proterozoic was a major outbreak of pre-mineralization, anorogenic felsic magmatism that brought significant amounts of lead (>30 ppm) to the surface of the Proterozoic cratons. Detritus derived from these volcanics transferred large amounts of the lead into sedimentary basins, rendering the basin-fill sediments a potential source of lead.

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**Figure 11.15.** Lead isotopic ratios of SMS deposits in relation to the mantle, orogene, and upper crust growth curves of Zartman and Doe (1981). Source of data: compilation by Sangster (1990).
11.9.4. PALEOTECTONIC SETTINGS

Inferred tectonic settings proposed for SMS deposits include: (a) intracontinental rifts (e.g., Mount Isa, Lambert 1983; McArthur River, Williams 1978a, b; Sullivan, Kanasewich 1968b, Stewart 1972); (b) continental margin side of back-arc basins (e.g. Howard’s pass, Badham 1981); (c) passive continental margins transected by incipient rifts or grabens (e.g., Meggen, Werner 1989); and (d) reactivated wrench fault systems in the basement of intracontinental basins (e.g., Navan and Silvermines, Andrew & Ashton 1985). Intracratonic grabens, characterized by elongate sediment-filled depressions bounded by normal faults that trend parallel to the axis of extension, appear to be the most frequent setting for the SMS deposits, especially the giant Proterozoic deposits (Gustafson & Williams 1981, Sawkins 1983). A particularly suitable environment for the formation of SMS deposits may have been aulcogens (failed arms of successful spreading systems) as these are sites of rapid clastic sedimentation, block faulting, and high heat flow (Sawkins 1976b).

Within the framework of a rift setting, the formation of SMS deposits appears to be related to tensional tectonism during the ‘post-rift’ thermal subsidence (Large 1992). This may be marked by: (a) the development of local basins and rises along sedimentary faults, as recognized by local facies changes, lateral variations in sedimentary thickness, and locally derived sedimentary conglomerates and breccias, (b) onset of rapid subsidence as evidenced by anomalously thick sedimentary sequences and sudden changes in sedimentary environment; and (c) development of a thermal anomaly along deeply penetrating structures, as indicated by volcanic rocks or tuffs.

11.9.5. AGE DISTRIBUTION

SMS deposits first appear at about 2000 Ma, attain their maximum development during early to middle Proterozoic, and become rare after the Mississippian. A conspicuous feature of this age distribution pattern (Fig. 11.16) is an antipathetic relationship between the temporal distribution of VMS and SMS deposits (Hutchinson 1980). There is a marked paucity of major VMS deposits (the “volcanogenic gap” of some authors) in the middle Proterozoic during which all the giant SMS deposits formed. Also, the younger Archean belts that contain abundant Noranda-type VMS deposits appear to be practically devoid of SMS deposits. A possible explanation may be that the dominant style of tectonism in the middle Proterozoic consisted of rifting and the formation of sedimentary basins, which contrasts with the compressional tectonism and associated prolific volcanism in the younger Archean greenstone belts and the Phanerozoic island arc environments that host most of the VMS deposits (Sawkins 1983). This, however, does not explain the lack of SMS deposits during post-Mississippian time despite the development of submarine rift environments in many parts of the Earth.

An alternative explanation may be that SMS deposits formed during intervals of anoxic bottom water condition in a stratified ocean that provided an adequate supply of
SMS deposits first appeared around 1900 Ma (Goodfellow et al. 1993) and the oldest known VMS deposits are those of the Pilbara craton hosted by 3.46-Ga volcanics (Barley 1992). (Modified from Goodfellow et al. 1993).

Figure 11.16. Age distribution of SMS and VMS deposits. SMS deposits first appeared around 1900 Ma (Goodfellow et al. 1993) and the oldest known VMS deposits are those of the Pilbara craton hosted by 3.46-Ga volcanics (Barley 1992). (Modified from Goodfellow et al. 1993).

reduced sulfur. As mentioned earlier, there is good evidence for four world-wide anoxic events (early Cambrian, middle Ordovician to early Silurian, early Devonian, and late Devonian), which correlate well with times when SMS and sediment-hosted barite deposits formed, whereas intervening periods correspond with the formation of barite deposits (Goodfellow 1987). The abundance of Proterozoic SMS deposits may be related to similar anoxic events, but there are no data to make such a case. The absence of Archean SMS deposits has been ascribed by Goodfellow (1992) to “the limiting effect of high reduced iron contents on the activity of reduced sulfur in anoxic oceans,” presumably because of Fe-sulfide precipitation. Under these conditions, metals in hydrothermal fluids vented into the water column were dispersed because of a lack of reduced sulfur to precipitate them. With the oxidation and deposition of iron as banded iron formations, the buildup of sulfate in the oceans during the early Proterozoic (Cameron & Hattori 1987b) and bacterial reduction of sulfate, anoxic oceans became H₂S-rich, and SMS-type sulfide deposition occurred when metalliferous fluids were discharged into euxinic basins (i.e., where H₂S was present in the water column).
11.10. Summary

SMS deposits include some of the largest Zn-Pb deposits of the world, and several contain significant amounts of barite and recoverable Ag. The host rocks are shallow to deep-water, marine clastics and carbonates, often with thin bands of K-rich tuffites as the manifestation of contemporaneous, albeit distant, volcanism. Contemporaneous igneous activity in the form of dikes and sills are present in a few districts, raising the possibility of a magmatic heat source for driving ore-fluid circulation, but there is no evidence of the involvement of magmatic fluids in the formation of SMS deposits.

The deposits typically consist of stacked lenses of massive, laterally zoned, Zn-Pb±Ba ore, characterized by stratiform mineralization of syngeneric, probably exhalative, origin. The massive ore zone may be underlain by a zone of discordant, epigenetic mineralization and hydrothermal alteration related to the same mineralization event. The ore-forming fluids were moderately hot (≈80° to ≈300°C) and saline (≈10 to 30 wt% NaCl equivalent). The H₂S for sulfide precipitation was derived mainly through bacterial reduction of seawater sulfate (although evaporite minerals of host rock sequences and dissolution of pyrite may have been important in some cases) and the ore accumulation occurred in localized brine pools within third-order basins. Synsedimentary faults played a key role in providing conduits for hydrothermal fluids to sites of mineralization. The deposits were emplaced in an extensional, intracontinental rift or continental margin wrench-fault setting.

Many aspects of SMS deposits are not well understood. These include the mechanism of stratiform mineralization, the source of metals, the irregular distribution of barite, and the relatively subdued footwall alteration in many large deposits that must have involved hydrothermal systems of considerable size and duration. The relatively restricted spatial and temporal distribution of SMS deposits is a puzzle, because compaction-driven and/or convective circulation of metalliferous brines should be a normal consequence of sedimentary basin development. Perhaps, the critical combination of large-scale metalliferous brine generation, efficient plumbing system, and favorable depositional environments did not occur too frequently (Sawkins 1984). The presence of euxinic ocean basins appears to have been a first-order control on the spatial and temporal distribution of SMS deposits, as well as for their preservation.

The SMS deposits, compared with VMS Deposits, are characterized by stratiform mineralization, much higher (Zn+Pb):Cu ratios, a lack of volcanic association, less pronounced footwall alteration, synsedimentary faulting, more saline hydrothermal fluids, a hydrothermal system of much longer duration, and bacterial reduction of seawater sulfate as the source of reduced sulfur (see Table 13.4).

11.11. Recommended Reading