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The changing vision of marine minerals

Peter A. Rona

Institute of Marine and Coastal Sciences and Department of Earth and Planetary Sciences, Rutgers University, New Brunswick, New Jersey 08901-8521, United States of America

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Abstract

Non-fuel marine minerals are reviewed from the perspective of resources and their value as active analogs that can advance understanding of types of ancient ore deposits that formed in marine settings. The theory of plate tectonics is the largest influence in expanding our vision of marine minerals and in developing our understanding of geologic controls of mineralization in space and time. Prior to the advent of plate tectonics, we viewed the ocean basins as passive sinks that served as containers for particulate and dissolved material eroded from land. This view adequately explained marine placer deposits (heavy minerals and gems), aggregates (sand and gravel), and precipitates (phosphorites and manganese nodules). Although numerous sites of placer mineral deposits are known on continental shelves worldwide, current activity pertains to diamond mining off southwestern Africa, tin mining off southeastern Asia, and intermittent gold mining off northwestern North America, which are all surpassed economically by worldwide recovery of marine sand and gravel, in turn dwarfed by offshore oil and gas. With the advent of plate tectonics, plate boundaries in ocean basins are recognized as active sources of mineralization in the form of hydrothermal massive sulfide deposits and proximal lower-temperature deposits hosted in oceanic crust (mafic at ocean ridges and felsic at volcanic island arcs), and of magmatic Ni–Cu sulfide, chromite and PGE deposits inferred to be present in the oceanic upper mantle–lower crust based on their occurrence in ophiolites. Some 300 sites of hydrothermal active and relict mineralization, most of them minor, are known at this early stage of seafloor exploration on ocean ridges, in fore-arc volcanoes, at back-arc spreading axes, and in arc rifts; deposits formed at spreading axes and transported off-axis by spreading are present in oceanic lithosphere but are virtually unknown. The TAG (Trans-Atlantic Geotraverse) hydrothermal field in the axial valley of the Mid-Atlantic Ridge (latitude 26° N) is considered to exemplify a major Volcanogenic Massive Sulfide (VMS) deposit forming at a spreading axis. The most prospective of these occurrences lie within the 200 nautical mile (370 km)-wide Exclusive Economic Zone (EEZ) of the nations of the volcanic island arcs of the western Pacific where metal content of massive sulfides (Ag, Au, Ba, Cu, Pb, Sb, Zn) exceeds that at ocean ridges. Plate tectonics early provided a framework for mineralization on the scale of global plate boundaries and is providing guidance to gradually converge on sites of mineralization through regional scales of plate reorganization, with the potential to elucidate the occurrence of individual deposits (e.g., Eocene Carlin-type gold deposits). Investigation of the spectrum of marine minerals as active analogs of types of ancient mineral deposits is contributing to this convergence. Consideration of questions posed by Brian Skinner (1997) of what we do and do not know about ancient hydrothermal mineral deposits demonstrates the ongoing advances in understanding driven by investigation of marine minerals.

Keywords: Marine minerals; Placers; Nodules; Crusts; Massive sulfides; VMS deposits; Magmatic sulfides; Hydrothermal processes; Plate tectonics; Marine environment; Marine mining

1. Introduction

Our vision of marine minerals is expanding rapidly as our understanding of the Earth advances. The theory of

E-mail address: rona@imcs.rutgers.edu.

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plate tectonics is recognized as the largest influence in expanding our vision. As noted by Brian Skinner (2004), “Plate tectonics has yet to influence economic geology the way it has influenced, even transformed, other areas of geology”. Perhaps nowhere has plate tectonics more potential to transform economic geology than through investigation of marine minerals as active analogs of various types of ancient deposits formed by inter- and intra-plate processes on and beneath the seafloor and preserved on land (Fig. 1; Table 1). Processes at plate boundaries have been effectively applied as a global framework to broadly classify certain types of ancient mineral deposits presently known from their occurrences on land (Table 2; Mitchell and Garson, 1981; Sawkins, 1984, 1990) and to show the evolution of metallogenic provinces through time (Kerrich, 1992; Kerrich et al., 2005). Whatever the outlook for their resource potential, investigation of marine mineral deposits provides an unprecedented opportunity to directly investigate active ore-forming processes that elucidate the genesis of those types of ancient deposits that were formed in marine settings related to plate boundaries. In this paper we briefly review the spectrum of non-fuel marine minerals from the perspective as products of plate interactions, of their roles as resources, and as active analogs of ancient ore deposits that formed in marine settings. We narrow the gap in application of plate tectonics to mineralization from a global to a regional scale by consideration of relations between geologic controls of mineralization and a global plate reorganization documented in the Eocene epoch.

2. Marine minerals and plate tectonics

Earth models prevalent prior to the advent of plate tectonics in the 1960’s minimized horizontal crustal movements and emphasized vertical movements. Continents were considered fixed in position and ocean basins were permanent features. Geotectonic cycles of mountain building were considered to follow a predetermined sequence of events involving a finite number of variations on the theme of sediment accumulation in subsiding belts (geosynclines) and subsequent crustal accretion involving vertical uplift with relatively small horizontal components of motion (Oreskes and Le Grande, 2001). Ocean basins were regarded as big bathtubs that passively contain the oceans and accumulate a large thickness of sediment under this relatively static and deterministic Earth model.

The pre-plate tectonic view adequately explained the types of marine minerals known at that time. These include placer deposits of dense metallic minerals and gemstones, and aggregates (sand and gravel) derived by mechanical erosion of terrestrial rocks and transported, sorted, and concentrated in beach and offshore sediments by flowing water (Fig. 1; Table 1). The marine minerals also include hydrogenetic (authigenetic) deposits of phosphorite on continental shelves and manganese nodules on abyssal plains (Cronan, 1980), as well as salts in seawater derived primarily from chemical erosion of terrestrial rocks.

The theory of plate tectonics changed our view of the ocean basins from passive sinks for material eroded from land to active sources of mineralization. Focus switched from terrestrial erosional processes to tectonic and magmatic processes at plate boundaries (divergent and convergent), largely submerged beneath the oceans. Seafloor mineralization is a byproduct of a global system of exchange of heat and chemicals among the mantle, crust and oceans at these plate boundaries. The ocean basins are leaky as containers, because the volcanic rocks of ocean crust are porous and permeable (permeabilities 10^{-9} to 10^{-10} m^2; Fisher, 2003; Becker and Davis, 2004; Becker et al., 2004) and subduction zones transport water into the mantle (Meade and Jeanloz, 1991). Thermoviscoelastic models suggest that thermal stress is sufficiently high to fracture 100 million year old oceanic lithosphere to a depth beneath the seafloor of at least 30 km providing pathways for hydration and serpentiniztion of the mantle (Korenaga, 2007). An estimated $\sim 10^{12}$ kg of pore water and $\sim 10^{12}$ kg of chemically bound water (altered basalt, gabbro, and serpentinized upper mantle) is subducted globally each year; approximately 5 to 20% of this water (0.1 to 0.4×10^{12} kg water/year) is subsequently incorporated into arc magmas (Peacock, 2004). The porosity of ocean crust contains an estimated nearly 2% of the total volume of the oceans (2×10^7 km^3 of 1.3×10^9 km^3; Johnson and Pruis, 2003). Where the dense, cold seawater flows in proximity to magma upwelling at plate boundaries, the seawater is heated, expands and buoyantly rises in sub-seafloor hydrothermal convection systems with discharge temperatures up to ~400 °C that dissolve and transport metals from the crust and that may entrain metal-rich magmatic fluids (Urabe, 1987; Hedenquist and Lowenstein, 1994; Yang and Scott, 1996; Williams-Jones and Heinrich, 2005). Serpentinization reactions produced when seawater hydrates peridotites of the upper mantle provide a secondary source of heat that can raise water temperature to 100 °C or higher depending on rates of heat production and extraction (Fyfe and Lonsdale, 1981; Francis, 1981; Macdonald and Fyfe, 1985; Lowell and

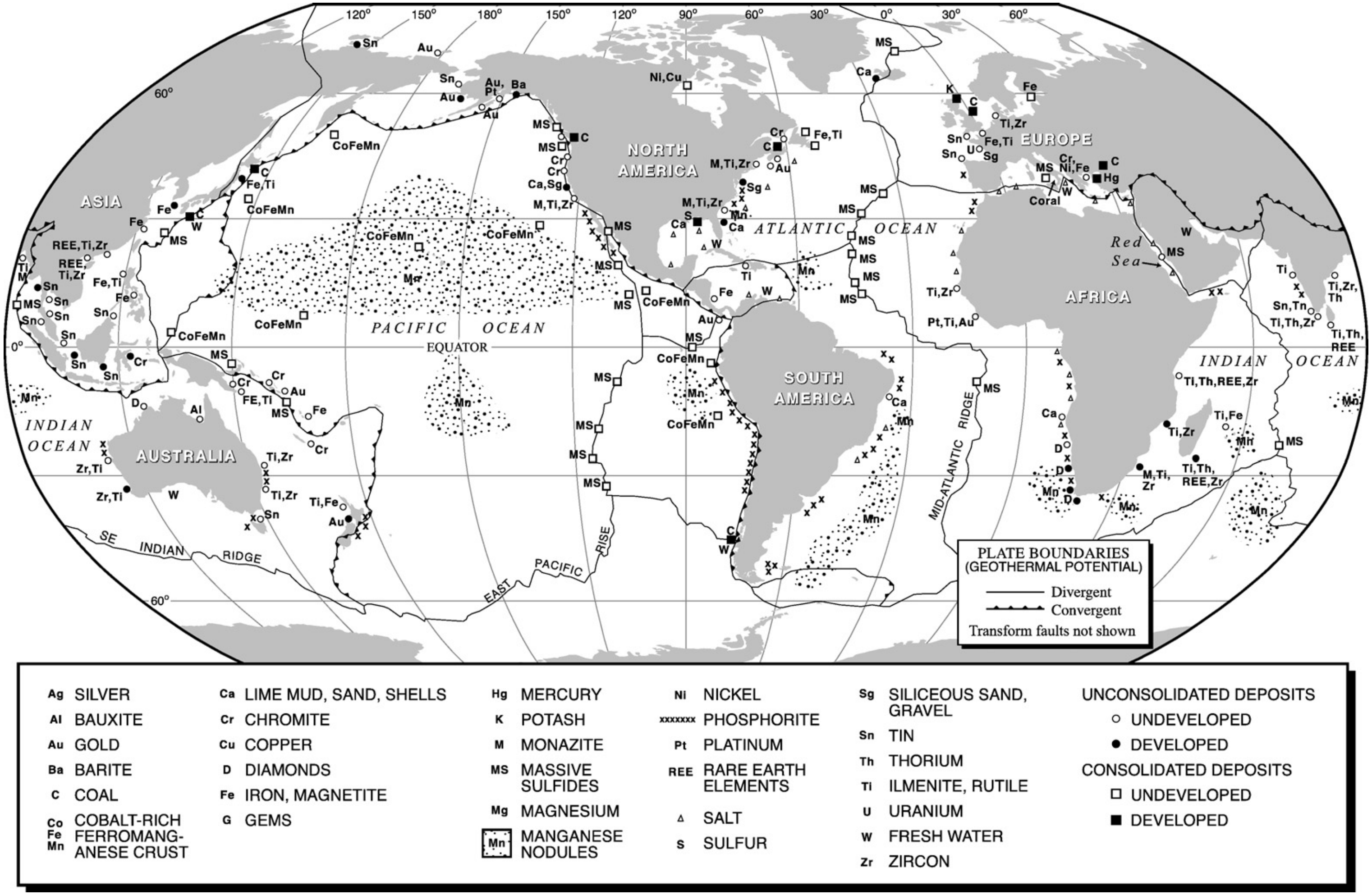

Fig. 1. Global distribution of known marine mineral resources (modified from Rona, 2003).

Rona, 2002; Schroeder et al., 2002; Blackman et al., 2006). The metal-rich hydrothermal fluids concentrate high-temperature (sulfides) and low-temperature (oxides, oxyhydroxides, silicates) mineral deposits at sites along plate boundaries. Magmatic deposits may form in the lower crust and upper mantle. The estimated current production rate of oceanic lithosphere is 2.94 km^2/year (Chase, 1972). A global mass rate of hydrothermal fluid flow of $1.3–2.7\times10^{17}$ g/year is calculated from the amount of cooling required to account for a discrepancy between the calculated amount of heat generated by emplacement of new lithosphere (Wolery and Sleep, 1976) and lower values of conductive heat flow actually measured at ocean ridges out to a distance from the spreading axis equal to an age of ∼65 million years. Values of hydrothermal fluid flow approach the flux of river water to the oceans of 3.8×10^{19} g/year (Mottl, 2003). An estimated 33% of the net hydrothermal heat flux occurs within 1 million years of the spreading axis (3.2×10^{12} W; Stein et al., 1995), with the balance occurring off-axis. The fluid flow rate cited implies that the entire mass of the oceans (13.7×10^{20} kg; Garrels and Mackenzie, 1971) cycles through the lithosphere at ocean ridges in 5 to 11 million years (Wolery and Sleep, 1976). Anomalous tectonic and magmatic conditions focus high-temperature fluid flow and concentrate mineral deposits at localized sites along plate boundaries within this overall scheme (Rona, 1988; Scott, 1997).

Marine minerals are considered in terms of types derived from sources on land, types from sources at submerged plate boundaries, and a combination of terrestrial and marine sources. In addition to natural boundaries, jurisdictional boundaries have been designated by the United Nations Convention on the Law of the Sea (UNCLOS), which recognizes a 200 nautical mile (370 km)-wide Exclusive Economic Zone (EEZ) under the jurisdiction of adjacent coastal states (McBryde, 1982; Broadus, 1987; Cronan, 1992), and an international zone designated the Area beyond that boundary declared as the "common heritage of mankind" with mineral resources under the jurisdiction of the United Nations (United Nations, 1997). UNCLOS was opened for signature in 1982 and entered into force in 1994, incorporating a provision for marine mining in the international Area that was renegotiated to meet industrial concerns, including protection of proprietary technology (United Nations, 1997, Part XI). The International Seabed Authority (ISA) is an independent international agency created by UNCLOS in 1994 upon the entry into force of the 1982 Convention through which States Parties to the Convention organize and control exploration for, and exploitation of the mineral resources of the deep seabed beyond the limits of national jurisdiction (International Seabed Authority, 1999, 2001a, 2003, 2006; Nandan et al., 2002; Antrim, 2005; website http://www.isa.org.jm/).

3. Marine minerals from terrestrial sources

3.1. Placer deposits

Numerous sites are known globally on continental shelves where placer deposits, primarily of metallic heavy minerals, have been mechanically concentrated by flowing water (Fig. 1) as a consequence of their higher density (>3.2 g cm^{-3}) relative to the bulk of detrital minerals consisting mostly of quartz and feldspar (2.5–2.7 g cm^{-3}). The resistance of a mineral (hardness, cleavage, density, solubility) to mechanical action during transport determines the distance it can be transported from its source without material change of state (Kudrass, 2000; Yim, 2000). The median distance of transport from a bedrock source to an offshore placer deposit is 8 km (Emery and Noakes, 1968). An outstanding feature of the distribution of placer deposits is the multitude of coastal sites known and the few of these sites of past or present mining (Table 3).

Three generic types of placer deposits are recognized (Kudrass, 2000): (i) disseminated beach placers usually containing light heavy minerals (density<6 g cm^{-3}; e.g., rutile, ilmenite, magnetite, monazite, zircon, sillimanite, garnet), which are concentrated by waves and longshore currents; (ii) drowned fluviatile placers comprising coarse sand and gravel overlying the bottom of river channels containing heavy metals (e.g., cassiterite, gold); and (iii) eluvial or lag deposits also containing heavy metals. Placer deposits may lie above, at, and below present sea level related to the history of regional and eustatic sea level change. In the geologic record, fluviatile placers are the presently most important from an economic point of view (Minter and Craw, 1999). For example, the Archean gold deposits in the Witwatersrand basin of South Africa (Frimmel et al., 2005) are interpreted as a fan deposit at the mouth of a river debouching into an intracratonic lake (Pretorius, 1991), although a hydrothermal origin has recently been championed (Law and Phillips, 2005).

3.2. Phosphorite

Phosphorites, consisting of varieties of the heavy mineral apatite, are distinguished from other sedimentary rocks by their higher phosphorous pentoxide (P_2O_5) contents (5 to 40%; Riggs, 1979; Bentor, 1980; Kudrass,

Table 1
Marine mineral resources (modified from Rona, 1983b; Cruickshank, 1998)

Region	Origin	Mode of occurrence			
		Unconsolidated (mineral)	Consolidated	Fluid	Heat
Continental Margin (shelf, slope and rise)	Terrigenous (derived by erosion and weathering of rocks on land)	Non-metals Beach deposit: siliceous sand and gravel (quartz) Placer deposit: diamond Metals Placer deposit: heavy mineral and native metal concentrates Barium (barite, witherite) Chromium (chromite) Gold Iron (hematite, magnetite, siderite) Rare-earth elements (monazite, basanite) Tin (cassiterite) Titanium (ilmenite, rutile) Thorium (monazite) Tungsten (scheelite, wolframite) Zirconium (zircon) Beach or placer deposit: iron sands (glauconite) Placer or solid layered deposit: phosphorite (apatite, fluorapatite, etc)			
			Cobalt–iron–manganese–platinum crusts Phosphorite Potash Salt (halite; sodium chloride) Sulfur (pure and as sulfate)		
	Diagenetic (produced by alteration of existing material)	Placer deposit: phosphorite (apatite, fluorapatite, etc.)	Solid layered deposit: (phosphorite)		
	Volcanogenic (derived from volcanoes)		Lode and vein deposits (all elements) Massive sulfide deposits (copper, iron, zinc, silver gold)		
				Freshwater Seawater solutes Salt (halite; sodium chloride) Magnesium	Geothermal energy

				Magnesium compounds Bromine Others Bicarbonate, boric acid, calcium, fluorite, potassium, strontium, sulfate, thorium	
	Biogenic (derived from organisms)	Beach deposit: lime (calcite, aragonite) mud and sand, shells Precious coral Pearl (primarily cultured)			
			Coal Limestone Gas hydrates (methane) Sulfur (pure and as sulfate)		
				Petroleum (oil and gas)	
	Hydrogenetic (authigenic; precipitated from seawater)	Lime (calcite, aragonite), mud and sand, shells			
	Meteoric (derived from the atmosphere)			Fresh water (desalination of seawater)	
Ocean Basin	Biogenic Hydrogenetic (authigenic)	Manganese nodules (manganese, iron, nickel, colbalt, copper)	Methane hydrate	Petroleum (oil and gas)	
			Cobalt–iron–manganese–platinum crusts		
	Volcanogenic	Metalliferous sediments (manganese, iron, copper, lead, zinc, gold, silver)			
			Manganese encrustations Massive sulfides (copper, iron, zinc, silver, gold) Copper–nickel sulfides, platinum-group elements, chromite deposits Sulfur (pure and as sulfate and sulfide)	Hydrothermal fluids (heat and metals)	Geothermal energy
	Magmatic (derived from magma)				
			Copper–nickel sulfides, platinum-group elements, chromite deposits		
				Magmatic fluids (heat and metals)	
					Geothermal energy

Table 2
Plate boundaries and associated mineral and energy occurrences

Type of plate boundary	Potential mineral occurrences	Potential energy occurrences	References
Divergent: spreading axes of ocean ridges and back-arc basins	Metalliferous sediments (Cu, Fe, Mn, Pb, Zn, Ba, Co, Ag, Au; e.g., Atlantis II Deep of Red Sea)	Geothermal	Naldrett (1989, 2004); Rona and Scott (1993); Hannington et al. (2004)
	Stratiform Mn- and Fe-oxides, hydroxides, and iron silicates (e.g., sites on Mid-Atlantic Ridge and Galapagos Spreading Center; fore-arc volcanoes and back-arc basins	Hydrocarbons at early stage of rifting and opening of an ocean basin (e.g., Red Sea; Atlantic continental margins; back-arc basins of western Pacific region)	
	Massive sulfides, including stockworks (Cu, Fe, Zn, Ag, Au; e.g., sites on East Pacific Rise, Galapagos Spreading Center, Mid-Atlantic Ridge, Central Indian Ridge, fore-arc volcanoes and back-arc basins of western Pacific)		
	Magmatic sulfides in disseminated or segregated form (Cu, Ni, PGE; upper mantle in ocean basins and volcanic island arcs)		
	Chromite: chrysotile: upper mantle in ocean basins and volcanic island arcs		
Convergent offshore: fore-arc volcanoes and back-arc basins in island arc systems	Massive sulfides and stratiform Mn- and Fe-oxides, hydroxides, and iron silicates (e.g. Tonga–Kermadec fore-arc volcanoes and Manus back-arc basin)	Geothermal hydrocarbons	Mitchell and Garson (1981); Sawkins (1990); de Ronde et al. (2003a)
Convergent onshore: volcanic island arcs and continental margins	Ophiolites containing same types of mineralization as at divergent plate boundaries (e.g., sites at former or present collisional plate boundaries	Geothermal	Mitchell and Garson (1981); Sawkins (1990); de Ronde et al. (2003a)
	Porphyry deposits (Cu, Fe, Mg, Sn,, Mo, Zn, Ag, Au; e.g., sites in the Andes)		
	Mineralization in granitic rocks (Au, Sn, and U)		
	Massive sulfides (Cu, Fe, Pb, Zn, Ag, Au, Ba; e.g., Kuroko-type deposits)		
Transform offshore: offsets of spreading axes	Generally only minor mineralization of similar types to those within spreading segments of ocean ridges		Bonatti (1981); Rona and Scott (1993); Hannington et al. (2004)

2000). Phosphorite is used mainly in the phosphate fertilizer industry. Phosphorite deposits are part of a biogeochemical cycle that involves dissolution and transport of phosphorous by rivers into the ocean, uptake by marine plankton, transfer into deep water masses by sinking and dissolution, return to surface water by upwelling, and deposition as hydrogenetic (authigenetic) precipitates and by diagenetic replacement of carbonates (Burnett, 1990). Phosporites occur in four seafloor settings (Hein et al., 2005): (1) continental shelves and slopes off the west coast of landmasses where easterly trade winds blow offshore (latitudes 30°N to 30°S) and induce upwelling; examples are phosphorite deposits of recent age at five localities: offshore Peru and Chile (Veeh et al., 1973; Burnett, 1977, 1990); offshore Namibia (Baturin et al., 1972; Veeh et al., 1974; Baturin, 1982); offshore eastern Australia (O'Brion et al., 1981); offshore Baja California in Mexico (Jahnke et al., 1983); and offshore the Atlantic margin of Morocco (Summerhayes and McArthur, 1990); (2) phosphorites formed by cementation and replacement of carbonates on submarine plateaus and banks like the Blake Plateau off southeastern United States (Manheim et al., 1980) and the Chatham Rise off New Zealand (Kudrass, 1984); (3) islands and atolls where the source of phosphorous is primarily guano and phosphorite has been mined; an example is the island of Nauru in the southwestern

Table 3
Developed marine mineral deposits (Lenoble et al., 1995)

Name (Latitude, longitude)	Commodity	Type of deposit	Water depth (m)	Location
Non-metals				
Coastal zone	Sand and gravel	Beach	0	Multiple sites worldwide
Coastal zone	Water/ice	Fluid and solid	0	Multiple sites worldwide
Groen River 30.5°S, 17.6°E	Diamond	Unconsolidated placer	25	South Africa
Chameis Bay 28.0°S, 15.7°E	Diamond	Placer	0–25	Namibia
Broadacres 31.6°S, 18.2°E	Diamond	Placer	0.5	South Africa
Casuarina Prospect (Inactive) 14.4°S, 127.8°E	Diamond	Placer	30	Australia
Hayward San Leandro 37.7°N, 122.1°W	Lime (shell fragments) 0	Beach San Francisco Bay, USA		
Laucala Bay 18.2°S, 178.5°E	Lime (coral sand)	Beach	0	Fiji
Faxa Bay 65.5°N, 22.5°W	Lime (shells)	Beach	35	Iceland
Vembanad 9.6°N, 76.3°E	Lime (shells)	Beach	0	India
Bahia Coast 13.0°S, 38.5°W	Lime (algae and shells)	Beach	0	Brazil
Cape Breton Islands 46.2°N, 60.9°W	Coal	Consolidated layers		Canada
Sunderland 54.9°N, 1.4°W	Coal	Consolidated layers		England
Torre de Geco 40.8°N, 14.5°E	Coral	Unconsolidated	5–300	Naples Bay, Italy
Metals				
Thai Muang 8.5°N, 98.2°E	Tin	Placer	10	Thailand
Tongkah Harbour 7.9°N, 98.5°E	Tin	Placer	20	Thailand
Takua Pa 9.0°N, 98.3′ E	Tin	Placer	0–18	Thailand
Copat Kelabat Bay 1.6°S, 105.7°E	Tin	Placer	0–1	Indonesia
Laut Tempilang 2.2°S, 105.7°E	Tin	Placer	10	Indonesia
Belitung (Billiton) 3.0°S, 108.2°E	Tin	Placer	10–20	Indonesia
Heinze Basin 14.7°N, 97.8°E	Tin, tungsten	Placer	16–30	Myanmar
Nome (Presently inactive) 64.5°N, 165.4°W	Gold	Placer	18–20	Alaska, USA
Bluff Soloman (Presently inactive) 64.6°N, 164.4°W	Gold	Placer	0–10	Alaska, USA
Gillespies Beach 43.4°S, 169.8°W	Gold	Placer	0–15	New Zealand
Richard’s Bay 28.8°S, 32.0°E	Titanium, zirconium	Placer	0–30	South Africa

(continued on next page)

Table 3 (*continued*)

Name	Commodity	Type of deposit	Water depth (m)	Location
(Latitude, longitude)				
Metals				
Fort Dauphin 25.0°S, 47.0°E	Titanium, thorium, rare earths, zirconium	Placer	0	Madagascar
Kanniyaknmari Manavalakurichi 8.2°N, 78.5°E	Titanium, zirconium, thorium	Placer	0	India
Chatrapur 19.4°N, 85.0°E	Titanium, zirconium, thorium	Placer	0	India
Castle Island 56.8°N, 133.0 W	Barium	Consolidated layered material	0–5	Alaska, USA
Sulawesi 2.0°S, 121.5°E	Chromite	Placer	0	Indonesia

Pacific; and (4) intra-plate seamounts which remain to be investigated. Most phosphorites on the present seafloor are of Miocene age. Present mining of phosphorite is from deposits on land that formed during past higher stands of sea level. The extensive untapped phosphorite deposits on westward-facing continental shelves within the trade wind belt constitute a potential future resource for agriculture in nations within that belt like India (Fig. 1).

3.3. Distribution of marine placers and phosphorite

Regional maps show the known distribution of deposits derived from terrestrial sources and concentrated on continental shelves by mechanical and chemical processes (Figs. 2–7; modified from Rona and Lenoble, 2004; Table 3). The realm of placer deposits can extend to continental slopes, rises, and abyssal plains where they are presently inaccessible. A primary source for construction of these maps is the MARMIN database (Lenoble et al., 1995) with additional references, as noted. Placer deposits that are undeveloped are distinguished on the maps from those deposits that are or have been mined (developed; Table 3).

North America and Central America (Fig. 2; Table 3): Gold is intermittently mined offshore Alaska contingent on market price (presently inactive). Placer deposits of gold derived from nearby primary deposits and re-deposited as glacial moraines in more widespread thin, lag gravels, were first mined in 1900 on the present beach of Nome (Garnett, 2000a). Gold was also mined from shallow submerged, buried paleobeaches and abrasion platforms cut into the moraines above bedrock. Glacial till with locally high gold grades extend nearly 5 km offshore in water depths up to 20 m. Gold is less mobile than other heavy minerals of equal particle size owing to its high density (15 to 19.3 g cm^{-3}), so that economic concentrations generally occur within kilometers of a source (Garnett, 2000a). A bucket-ladder dredge that had been operated as a cassiterite producer in calmer waters offshore Indonesia was adapted to perform much of the mining. Later, a track-mounted mining system deployed on the seabed from an anchored barge, the beach, or sea-ice enabled selective recovery of high grades. Although terminated in 1990, the Alaskan operations are the only example of commercial recovery of marine gold placer deposits (Garnett, 2000a), with the exception of a site intermittently mined offshore New Zealand (Fig. 7). Investigations of placer gold offshore New South Wales, Nova Scotia, West Africa, East Malaysia (Sarawak), and Siberia have yet to identify commercial deposits.

Placer deposits other than gold that occur offshore North America are largely undeveloped. Consolidated layers containing barium have been mined on Castle Island off Alaska. Extensive phosphorite deposits in the form of nodules and crusts lie in shallow water of the Pacific continental shelf of California and Baja California (Baturin, 1982; Jahnke et al., 1983). A province of relict phosphorites is present to water depths of 1 km on the Blake Plateau off southeastern North America (Manheim et al., 1980). Miocene deposits of southeastern North America (Riggs, 1979), deposited at former higher stands of sea level and presently exposed on land are mined. Lime (calcium carbonate) is dredged from shallow water areas where precipitating on the Bahama Banks, a shallow subsiding submarine carbonate plateau constructed of layers of limestone attaining a thickness of kilometers off southeastern Florida. An underground coal mine extends offshore in the Cape Breton Islands of Canada. Sand and gravel are recovered at many sites in the coastal zone of

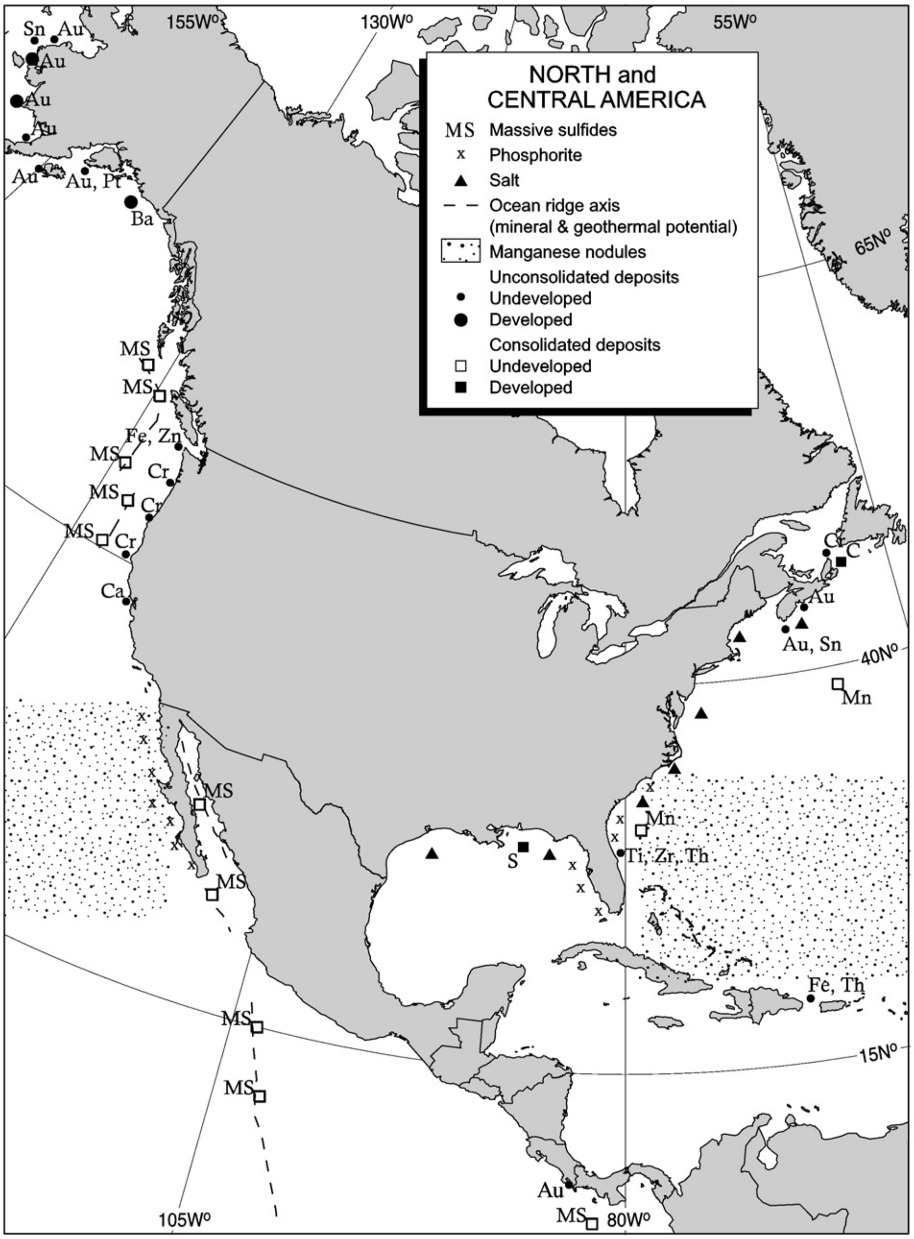

Fig. 2. Offshore mineral map of North America.

Canada and the United States primarily for beach restoration and shore protection. Beds of Jurassic (Tithonian) salt up to kilometers in thickness underlie sediments of the continental margin off eastern North America and the Gulf of Mexico.

The continental margins of Central America are practically unexplored for placer deposits with few exceptions (e.g., recycling of Late Tertiary placer gold off Costa Rica; Berrangé, 1989; Kriz, 1990). Porphyry mineralization at sites in volcanic belts along the west coast of Central America generated by subduction of the Cocos plate (Sawkins, 1990) suggests potential for metallic mineral placers (Fig. 2).

South America (Fig. 3): The continent of South America is noted for Andean Cu–Mo–Au porphyry and massive sulfide deposits related to volcanism generated by eastward subduction of the Nazca plate (e.g., Camus and Dilles, 2001). The middle Eocene to early Oligocene

Fig. 3. Offshore mineral map of South America.

belt of the central Andes contains the largest concentration of Cu resources known in the world (Sillitoe and Perello, 2005). The apparent absence of metallic placer deposits along the western continental margin seaward of the mineralized zones, except for a few placer gold occurrences offshore Ecuador, Chile, and Tierra del Fuego (R.H.T. Garnett, pers. comm..), is an artifact of the early stage of exploration. Although desert conditions presently prevail along a large section of western South America, rivers may have transported metallic minerals from the Andean deposits to the coast under former pluvial climates. Accordingly, the narrow western margin of South America may have significant potential for metallic placer deposits.

Phosphorite is precipitating on the Peruvian continental shelf from upwelling deep water (Burnett, 1977, 1990). Lime precipitated by marine plants is recovered from the Brazilian shelf. Kilometers-thick layers of

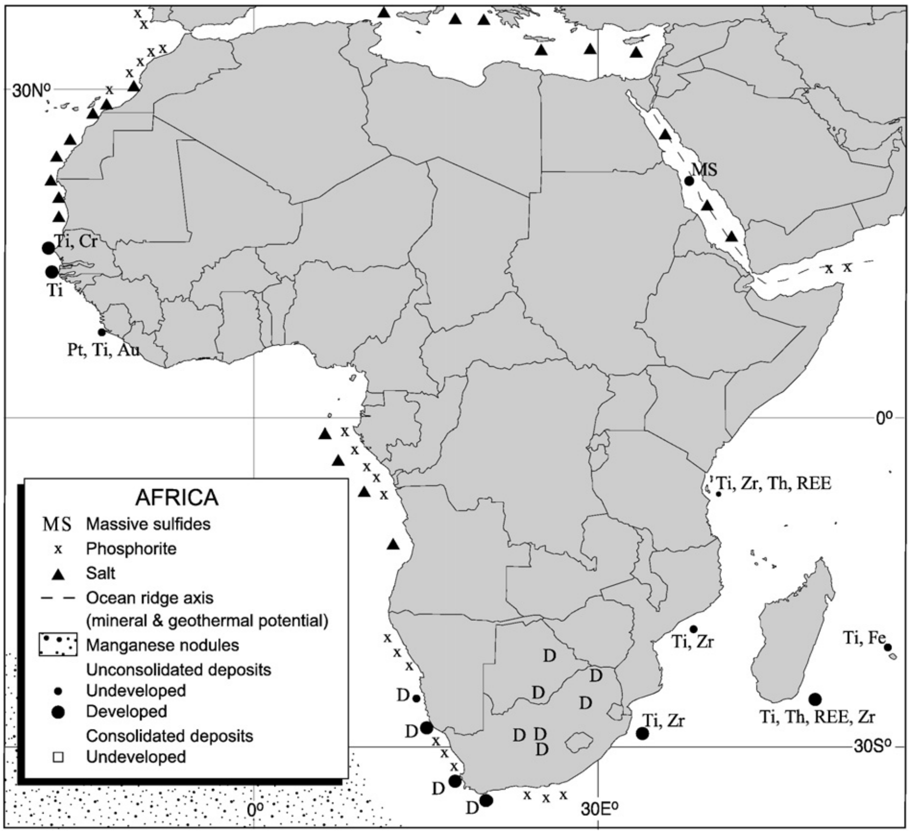

Fig. 4. Offshore mineral map of Africa.

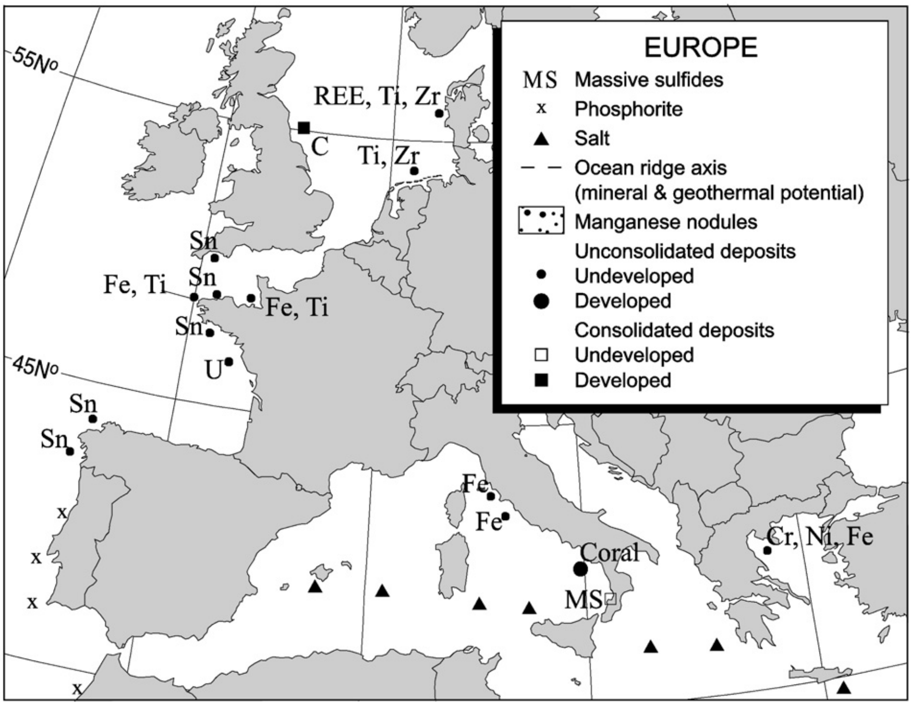

Fig. 5. Offshore mineral map of Europe.

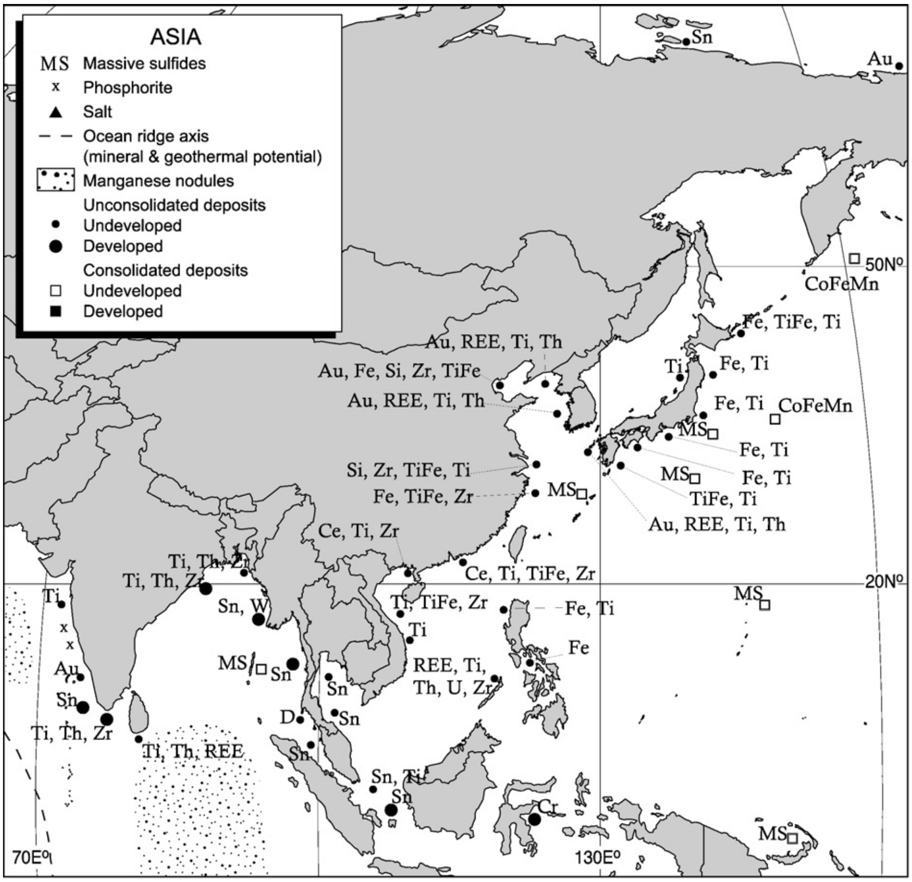

Fig. 6. Offshore mineral map of Asia.

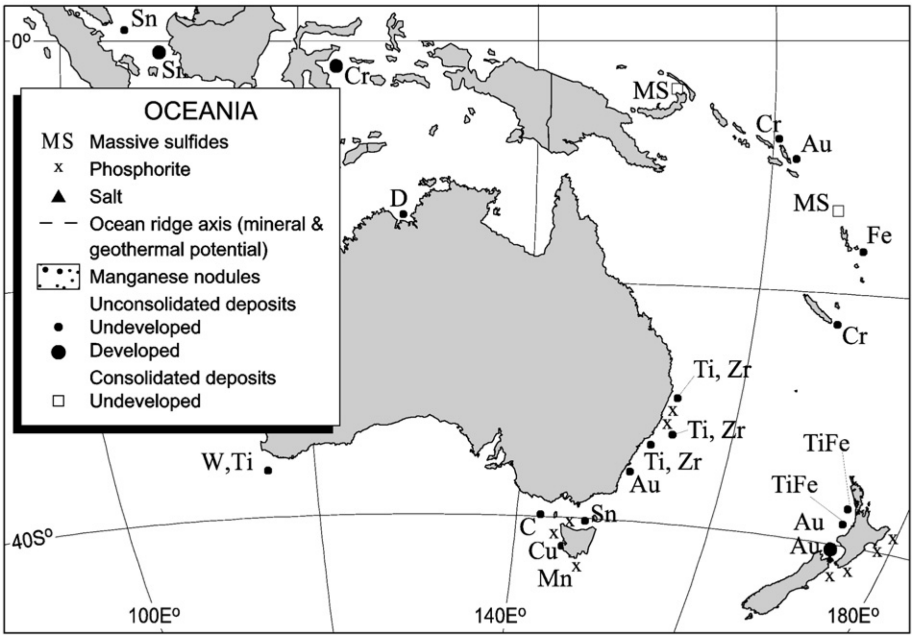

Fig. 7. Offshore mineral map of Oceania.

Aptian salt are present in basins underlying the Brazilian margin where salt structures are associated with offshore petroleum (Rona, 1982).

Africa (Fig. 4; Table 3): A large placer diamond province encompasses beaches and the adjacent continental shelf from 100 m above to at least 200 m below sea level and extends between 450 km south and 300 km north of the present Orange River that bounds Namibia and South Africa (Garnett, 2000b, 2001). More than 75 million ct of diamonds have been produced from this section of the Namibian coast ("Sperrgebeit" = "forbidden territory" from German heritage; also known as Diamond Area 1 extending 5.5 km offshore and 20–35 km inland) over nearly 100 years, 95% of which are gem quality, making this the richest diamond placer known (Schneider and Miller, 1992; Oosterveld, 20003). The diamonds (density 3.5 g cm^{-3}) were eroded from numerous Cretaceous kimberlites in the interior of southern Africa and transported hundreds of kilometers westward to the South Atlantic coast by a system of rivers represented by paleo-channels and the present Orange River since at least Middle Eocene times reflecting a Late Cretaceous regional sub-continental uplift that initiated deep fluvial incision and has continued intermittently through much of the Cenozoic (Spaggiari et al., 2006). The diamonds were redistributed from paleo-river mouths onto ancient beaches, and seaward of the beaches by waves and longshore currents during sea level changes inferred to have ranged from ca. +180 m to −120 m relative to present mean sea level (Spaggiari et al., 2006). Diamonds decline in size and value with distance from source (Garnett and Bassett, 2005). Within the overall scheme, diamond grades tend to increase from onshore to offshore related to properties of the diamonds and the amount of reworking by water motions, which concentrated the diamonds in drowned paleobeaches and trapped them in shallow depressions in bedrock underlying thin overburden. In Namibia recovery by a contractor (De Beers Marine; 50:50 joint venture with the Namibian Government Namdeb Diamond Corporation Ltd.) of high-quality marine diamonds using seafloor mining machines and drills on 4 offshore production vessels increased from 30,000 to >570,000 ct (1 ct=0.2 g)/year over the last decade (EMI Placer Stockfile website http://www.mine.mn/Placer_Stockfile_De_Beers.htm). The offshore recovered grade is 0.2 ct/m^2 at US$298/ct with estimated current production value>US$1700 million. Average diamond values per carat for a given diamond occurrence may vary by three orders of magnitude (US$1 to $1000/ct; Gurney et al., 2005). The estimated total value of global diamond production (marine and non-marine) in 2004 was US$11,800 million (Business Day, 2005). Exploration for marine diamonds has also been conducted offshore West Africa, Australia, and Indonesia, but without commercial success.

Placer deposits containing titanium, thorium, rare earth elements and zirconium have been mined at a location on the southeast coast of Madagascar (Fig. 5; Table 2). The Corridor Sands (1765 million tonnes containing 73 million tonnes of ilmenite at an estimated average ilmenite grade of 4.14%; Mining Review Africa, 2003) and the Moma disseminated beach deposits (estimated 60 million tonnes of ilmenite; Planet Ark, 2003) onshore near the coast of Mozambique are both under development and are considered, respectively, the world's largest and second largest undeveloped resources of titanium dioxide (TiO_2). With reference to phosphorite, both relict (principally Miocene) and modern deposits occur on the northwestern continental shelf of Africa offshore Morroco (Summerhayes and McArthur, 1990), Namibia (Thomson et al., 1984), and South Africa (Birch, 1980; McArthur et al., 1988).

Salt layers up to several kilometers thick lie buried beneath western and eastern South and North Atlantic continental margins and intrude overlying sediments as diapirs (Figs. 2, 3 and 4). The salt was deposited at early stages of opening of the North Atlantic in the Jurassic period and of the South Atlantic in the Aptian stage of the Cretaceous of the South Atlantic. At those times the Atlantic was a sea with circulation restricted by the positions of the surrounding continents, causing evaporation to exceed inflow and precipitation of salt and organic matter (Rona, 1969, 1982). The salt is associated with petroleum production and potential at sites on continental margins of West Africa (e.g., Meyers et al., 1996), eastern South America, eastern North America and the Gulf of Mexico.

Europe (Fig. 5): Numerous marine metallic mineral placer deposits are identified, but none have been developed. Layers of Miocene salt up to kilometers in thickness are buried beneath sediments under large areas of the Mediterranean Sea, where the salt was deposited under former conditions of restricted ocean circulation (Hsü, 1983). Undeveloped phosphorite deposits lie adjacent to areas of deep ocean upwelling on the Atlantic continental shelf of northern Spain (Lamboy and Lucas, 1979). An underground coal mine extends seaward on the east coast of England. Coral recovery is an industry in the Bay of Naples and salt recovery from evaporation of seawater is practiced at many places around the Mediterranean and western France.

Asia (Fig. 6; Table 3): A diverse suite of marine metallic placer deposits exists on the continental margins of Asia. Of these various deposits, placers of the tin mineral cassiterite offshore Southeast Asia are the principal deposits that have undergone sustained mining. The Southeast Asian tin deposits generally occur in belts genetically related to terrestrial granitoids or their volcanic equivalents and were concentrated by fluvial processes involving selective or hydraulic sorting (Yim, 2000). Cassiterite with high density (~7), moderate hardness (6–7), a brittle nature, and poor cleavage, has low transportation resistance and occurs near the source rock. The deposits lie offshore Myanmar, Thailand, Malaysia, and Indonesia at sites adjacent to outcrops of Carboniferous and Mesozoic granites on the east and west sides of the Malaysian peninsula, respectively (Hosking, 1971). The estimated value of annual tin production from marine tin placers offshore Thailand and Indonesia in 1968 was US $24 million (Cruickshank et al., 1968). Titanium-rich placer magnetite has been mined from the northwestern coast of New Zealand (North Island) and Indonesia (Java), the Philippines (Luzon), and Japan (Hokkaido; Kudrass, 2000). Numerous undeveloped placer deposits of light heavy minerals (ilmenite, rutile, magnetite, zircon, garnet, and monazite) are present on beaches and offshore the Indian sub-continent (Roonwal, 1986; Rajamanickam, 2000) and P. R. China (Institute of Marine Geology, 1988; Tan et al., 1996). Gold and tin placers on the Arctic shelf of Siberia are in an early stage of investigation (Patyk-Kara, 1999).

Oceania (Fig. 7; Table 3): The titanium minerals rutile and ilmenite have been mined from beach sand in southeast and southwest Australia (Roy, 1999). The other Australian coasts are relatively unexplored for such deposits (CSIRO, 2006). Phosphorites in the form of nodules occur on the outer shelf and upper slope of eastern Australia. A placer gold deposit has been intermittently mined off New Zealand. The iron–titanium-rich placer magnetite has been mined from the northwestern coast of New Zealand (North Island) and Indonesia (Java; Kudrass, 2000). Placer chromite has been produced from a site on the east coast of the Indonesian island of Suluwesi. An extensive field of phosphorite nodules with an average P_2O_5 content of 22% lie between water depths of 350 m to 450 m on the crest of the Chatham Rise east of New Zealand (Kudrass, 1984; von Rad and Kudrass, 1984; Cullen, 1986; Exon et al., 1992).

3.4. Marine sand and gravel

Marine sand and gravel are transported to the coast by rivers and concentrated in beaches and longshore bars by waves and longshore currents. The sand and gravel are presently the most extensively mined, utilized and valued of all marine mineral resources because aggregates (sand and gravel) are universally used for construction (concrete), and sand is dredged for beach restoration, shore protection and deepening of channels in harbors (Cruickshank, 1988). Aggregate sites are too numerous to show on the maps (Figs. 1–7). Estimated global annual production value of marine sand and gravel for 2000 was about US$3000 million based on production amounts (millions of tonnes) at US$15/tonne (UK 24, Netherlands 36, Denmark 18, Belgium 3, France 3, Poland 0.5, Norway 0.1, Germany 7, Japan 70; D.J. Harrison, British Geological Survey, pers. comm.). A conservative estimate of United States annual production of marine aggregate of 31 million tonnes is based on the average volume used for beach nourishment between 1990 and 2007 from the Program for the Study of Developed Shorelines database (21,069,482 yd^3; http://psds.wcu.edu/1038.asp; A. Coburn, Western Carolina University, pers. comm.), with the addition of 10% for marine aggregate used for construction material, and conversion from cubic yards to tonnes (density for dry mixture of sand and gravel 1727 kg/m^3; Ontario Stone, Sand and Gravel Association, 2006). Canada is not a big producer of aggregate from the marine environment (D. Pangapko, Natural Resources Canada, pers. comm.).

3.5. Marine solutes

A number of materials are extracted from seawater at some 300 coastal operations in 60 countries, including rock salt (sodium chloride), magnesium metal, magnesium compounds, and bromine (Table 1). Of these, freshwater extracted from seawater by desalination processes is the most critical mineral, in light of the global need for an adequate and safe supply of water for consumption, agriculture and industry. Desalination by reverse osmosis and other processes is energy intensive. The oceans are the largest reservoir for water on Earth. Production of freshwater from seawater is expected to exceed that from all other marine minerals in importance and value as need continues to grow and alternative energy sources for the desalination process are developed (Revenga et al., 2001; Newton et al., 2006; UNESCO, 2006; UNESCO Water Portal at http://www.unesco.org/water).

4. Marine minerals from sources in ocean basins

4.1. Metalliferous sediments

The metalliferous sediments of the Atlantis II Deep discovered in 1965 at the spreading axis of the central

Table 4
Resource potential of metalliferous sediments of the Atlantis II Deep, Red Sea

Metal	Grade (wt.%, dry salt-free basis)	Weight (tonnes; dry salt-free basis)
Metalliferous sediments		89,500,000
Zn	2.06	1,838,000
Cu	0.45	404,000
Ag	38.4 g/tonne	3432
Au	0.5 g/tonne	45

Data from Mustafa et al. (1984), Nawab (2001).
1 tonne = 1 metric ton = 1000 kg; Co, and Cd also recoverable.

Red Sea are the first hydrothermal deposit found at a submerged divergent plate boundary (Figs. 1 and 4; Swallow and Crease, 1965; Miller et al., 1966; Degens and Ross, 1969), and remain the most efficient ore-forming system and the largest seafloor hydrothermal deposit found to date. The Atlantis II Deep at a water depth of 2 km is a parallelogram-shaped basin with sides 12 km long aligned with faults parallel to the spreading axis and 5 km wide aligned with cross-axial transform faults. It is the largest of a series of basins that lie at the spreading axis between the African and Arabian plates. Hydrothermal discharge in the Atlantis II Deep is the source of spillover to two adjacent basins (Discovery and Chain deeps). Rifting between Africa and Saudi Arabia started at about 10 million years ago (Miocene) and is generating oceanic lithosphere at a slow-spreading full-rate (2 cm/year) having opened the central Red Sea to its present 200 km width. The margins of the Red Sea are underlain by evaporites up to several kilometers thick that were deposited at an earlier stage of opening when circulation was restricted by the surrounding landmasses and evaporation exceeded inflow of seawater through narrow straits. The hydrothermal ore-forming system present is inferred to comprise downwelling of seawater through kilometers of permeable transitional and ocean crust, nearly an order of magnitude increase in salinity above that of normal seawater (35 ppt) by flow in proximity to the evaporites (salt is impermeable), upwelling of the thermally expanded hyper-saline solutions, enhanced acquisition and transport of metals from the ocean crust and possibly magmatic effluents as chloride complexes; and discharge into the basin (Scholten et al., 2000). The density increase from dissolved salts more than compensates for the thermal expansion resulting in stable density stratification of the venting solutions trapped within the Atlantis II basin. Metallic sulfides, oxides, and hydroxides precipitate from the ponded solutions and settle to form layers of unconsolidated metalliferous

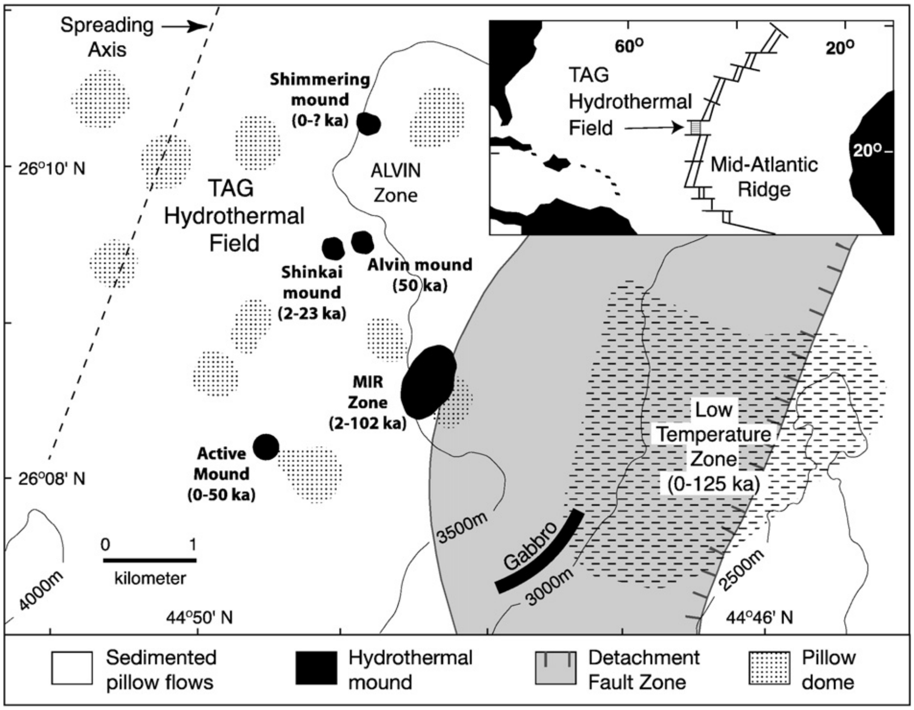

Fig. 8. Map of the TAG hydrothermal field in the axial valley of the Mid-Atlantic Ridge near 26°N, 45°W (radiometric ages of massive sulfides and oxides in parentheses; Zonenshain et al., 1989; Rona et al., 1993, 1998; deMartin et al., 2007).

sediments up to 30 m thick hosted in basaltic ocean crust (Bäcker, 1980; Blissenbach and Nawab, 1982; Mustafa et al., 1984).

The Saudi–Sudanese Joint Red Sea Commission oversees the development of the Atlantis II Deep deposit located within the overlapping 200 nm-wide Exclusive Economic Zones of the bordering coastal states (Mustafa, 1979; Mustafa et al., 1984; Nawab, 1984, 2001). The Commission sponsored a survey by the German geophysical company Preussag (Amann, 1985) followed by a pre-pilot mining test in 1979, which demonstrated the feasibility of using hydraulic dredging from a modified offshore petroleum drilling vessel (Sedco 445) to recover a sample of the metalliferous sediments (15,000 tonnes) and shipboard flotation techniques to separate the metalliferous component (4 tonnes of concentrate containing up to 30 wt.% Zn, 4 wt.% Cu and 600 g/tonne Ag). According to Nawab (2001), "The gold content is low; it has not been systematically determined in core samples, but has been recovered in bulk sulfide-flotation concentrates". An estimate of gold content is 0.5 g/tonne (Mustafa et al., 1984). Estimated grade and tonnage based on statistical analysis (two-dimensional kriging) of the composition of 605 cores are presented in Table 4 (Mustafa et al., 1984; Nawab, 2001). The resources are considered adequate to support an average annual production of 60,000 tonnes of Zn, 10,000 tonnes of Cu, 100 tonnes of Ag, and 1 tonne of Au for a period of approximately 20 years, with initiation of mining considered contingent on primarily market conditions (Nawab, 2001). The small size of the particles (clay) may prove difficult to refine.

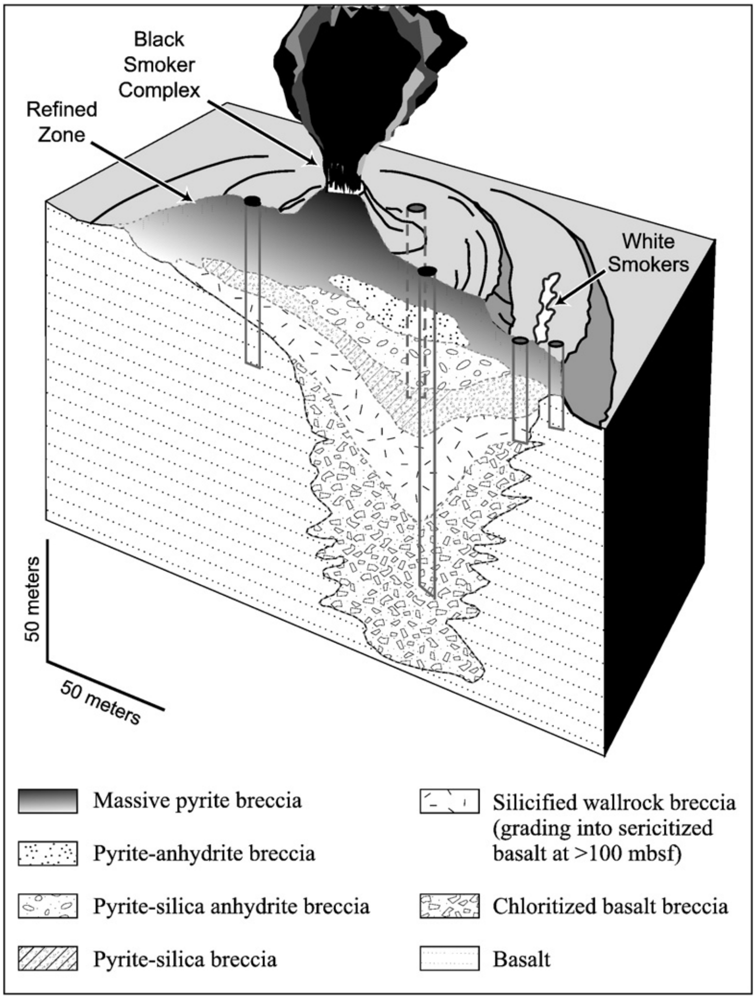

Fig. 9. Schematic cross-section through the TAG active massive sulfide mound (Fig. 8) showing characteristic surface and sub-surface features of a seafloor VMS deposit based on results of Ocean Drilling Program Leg 158 (Rona, 1992; Hannington et al., 1995; Humphris et al., 1995a,b).

As seafloor spreading continues, the Red Sea may widen into an ocean like the Atlantic. Relict metalliferous hydrothermal deposits may be present in ocean crust aligned along flow lines of seafloor spreading extending away from the spreading axis depending on the position, persistence, and episodicity of axial hydrothermal ore-forming systems (Rona, 1973, 1985). Metalliferous sediment deposits like those forming at the Atlantis II Deep are expected to be present buried kilometers beneath sediments at sites along rifted continental margins of ocean basins like the Atlantic that underwent a Red Sea stage early in their opening. Metalliferous sediments are also a byproduct of hydrothermal systems at ocean ridges and volcanic island arcs (Gurvich, 2006).

4.2. High-temperature massive sulfides

Volcanogenic massive sulfide (VMS) deposits were first found in mafic ocean crust of an ocean basin at the intermediate-spreading rate (full-rate 6 cm/year) axis of the East Pacific Rise at 21°N, 103°W in 1978 (CYAMEX, 1979; Francheteau et al., 1979; Hekinian et al., 1980). In the following year high-temperature metal-rich solutions (350 °C) were found discharging from sulfide chimneys as "black smokers" at the same location (RISE, 1980). The scientific consensus at that time was that the thermal regime of intermediate- to fast-spreading ocean ridges in the Pacific was required to drive high-temperature hydrothermal activity that concentrates VMS deposits and that size of a deposit may vary directly with spreading rate (use of the term "massive" refers to mineralization of at least 60% sulfide and carries no textural connotation; Sangster and Scott, 1976). Slow-spreading ocean ridges (full-rate < 4 cm/year) were ruled out for high-temperature hydrothermal activity in spite of the earlier discovery of the Atlantis II Deep deposit at the slow-spreading axis of the Red Sea, which was then considered an anomaly of an early stage of opening of an ocean basin. Accordingly, initial VMS exploration at that time focused on the axial zone of intermediate-to-ultrafast spreading portions of the East Pacific Rise between 21°N and 32°S (full-rate 6 to 18 cm/year), which were considered most prospective (Bäcker, 1980; Bäcker et al., 1985; Marchig et al., 1987, 1988). Although an estimated 66% of hydrothermal heat and water flow occurs on the flanks of ocean ridges out to a distance equal to a crustal age of 65 Ma (Stein et al., 1995), it is presumably low-temperature flow (e.g., Benjamin and Haymon, 2006). Magmatic systems that drive high-temperature hydrothermal activity that concentrates sulfides on ocean ridges primarily underlie the axial zone where the lithosphere forms.

The discovery, in 1985, of the first black smokers, VMS deposits and vent biota in the Atlantic Ocean at the TAG (Trans-Atlantic Geotraverse) hydrothermal field on the Mid-Atlantic Ridge near 26°N, 45°W (Rona et al., 1986; Figs. 8 and 9), opened to hydrothermal exploration the slow-spreading ocean ridges in the Atlantic, Indian, and Arctic oceans, comprising more than half the ~55,000 km global length of ocean ridges. Since that time massive sulfides have been found in a variety of tectonic settings at the full range of spreading rates (Fig. 1; Rona and Scott, 1993; Herzig and Hannington, 1995; Hannington et al., 2004) including sediments of mafic ocean ridges (Escanaba Trough of the Gorda Ridge, Morton et al., 1995; Fouquet et al., 1998; Middle Valley of Juan de Fuca Ridge, Davis et al., 1992; Fouquet et al., 1998); felsic rocks at sites at spreading axes in back-arc basins (Halbach et al., 1989;

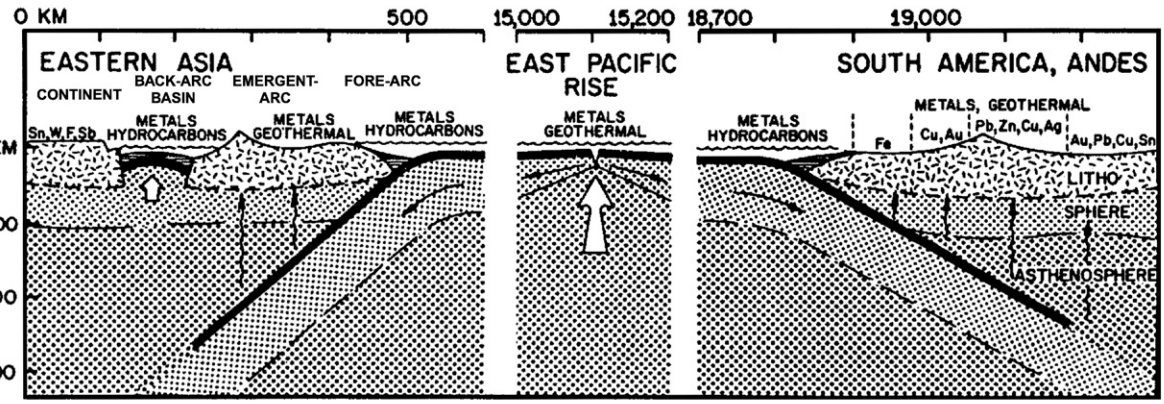

Fig. 10. Diagrammatic east–west cross-section of the Pacific Ocean between South America and southeast Asia showing settings for hydrothermal mineralization associated with convergent (subduction zones) and divergent (ocean ridges) plate boundaries and metallogenic zones of the Andes (modified from Rona, 1983b; Sillitoe and Perello, 2005).

Table 5
High-temperature hydrothermal systems with massive sulfide mineralization drilled by the Ocean Drilling Project (ODP)

Leg	Year	Location	Site	Results
I. Mafic-hosted volcanogenic massive sulfide (VMS) system				
ODP 158	1994	Active high-temperature sulfide mound in TAG hydrothermal field, Mid-Atlantic Ridge near 26°08′N, 44°50′W	957	Drilled 17 holes to a maximum of 125 mbsf (meters below sea floor) in lens and stockwork of sulfide and sulfate hosted in basalt (Humphris et al., 1995a,b)
II. Felsic-hosted volcanogenic massive sulfide (VMS) system				
ODP 193	2000–2001	PACMANUS active high-temperature hydrothermal system including sulfides in Manus back-arc basin in the western Pacific near 3°43′S, 151°40′E	1188–1191	Drilled and logged 13 holes to a maximum of 387 mbsf with minor sulfides hosted in altered dacitic to rhyodactic rocks (Binns et al., 2002)
III. Sediment-hosted massive sulfide system				
ODP 169	1996	Central Hill, an uplifted block in the sediment-filled Escanaba Trough of the southern Gorda Ridge near 41°00′N, 127°30′W	1038	Drilled 9 holes to a maximum of 404 mbsf terminating in basalt, indicating that massive sulfide forms only a thin veneer (5–15 m) over the sediment sequence along the faulted margin of Central Hill (Fouquet et al., 1996)
ODP 139	1991	An inactive massive sulfide body (Bent Hill) and hydrothermal recharge and reservoir zones in the sediment-filled axial valley of the northern Juan de Fuca Ridge near 48°26′N, 128°41′W	855–858	Drilled 8 holes to a maximum of 122 mbsf through massive sulfide and sediment into a basalt sill (Davis et al., 1992)
ODP 169	1996	An inactive massive sulfide deposit (Bent Hill) and an active hydrothermal field (Dead Dog) hosted in the sediment-filled axial valley of the northern Juan de Fuca Ridge near 48°26′N, 128°41′W	1035 (Bent Hill; 9 holes); 1036 (Dead Dog; 5 holes)	Drilled 9 holes to a maximum of 497 mbfs (re-entry and logging of Leg 139 hole 856) through sediment underlying the massive sulfide deposit and into basalt flows at Bent Hill; and 5 holes to a maximum of 296 mbsf through sediments at Dead Dog (Fouquet et al., 1998)

Binns and Scott, 1993; Ishibashi and Urabe, 1995); and in fore-arc volcanic calderas of volcanic island arcs associated with convergent plate boundaries of the western Pacific (Fig. 10; Iizasa et al., 1999; Herzig, 1999; Glasby et al., 2000; de Ronde et al., 2005; Stoffers et al., 2006). The Ocean Drilling Program (http://www-odp.tamu.edu/) has obtained information on the third dimension of certain of these deposits by drilling three types of high-temperature hydrothermal systems (Table 5) comprising massive sulfides hosted in: 1) mafic ocean crust (TAG on the Mid-Atlantic Ridge); 2) sediments over mafic ocean crust (Bent Hill in Middle Valley of the northern Juan de Fuca Ridge); and 3) felsic ocean crust (PACMANUS hydrothermal field in the Manus back-arc basin). The present tally of ~300 sites of active and relict hydrothermal mineralization known in these various settings is an artifact of this early stage of exploration when only about 5% of prospective seafloor settings have been investigated in sufficient detail to find deposits (Rona and Scott, 1993; Hannington et al., 2004). Only some 100 of these sites host massive sulfides and only two of these massive sullfide deposits are known to be over 1×10^6 tonnes (TAG on the Mid-Atlantic Ridge and Magic Mountain on the Explorer Ridge shown in Fig. 11). In terms of jurisdictional boundaries, about 80% of the global length of ocean ridges lies in the international "Area" under the UNCLOS with mineral resources under the management of the ISA (Fig. 11). The balance of the global length of ocean ridges lies within the 200 nautical mile-wide EEZ (Fig. 11; Table 6) under the jurisdiction of coastal states. All the volcanic island arc systems lie within the EEZ of coastal states.

Size, composition, and distribution of VMS deposits is influenced by similarities and differences in morphologic, tectonic, and magmatic characteristics that exist between intermediate- to fast-spreading ocean ridges and slow-spreading ocean ridges (Small, 1998; Carbotte and Scheirer, 2004). Mineralization at ocean ridges, in turn, exhibits differences from that at volcanic island arcs. However, exceptionally large, high-grade deposits are anomalies that can occur in any of these settings at any spreading rate depending on the geologic controls

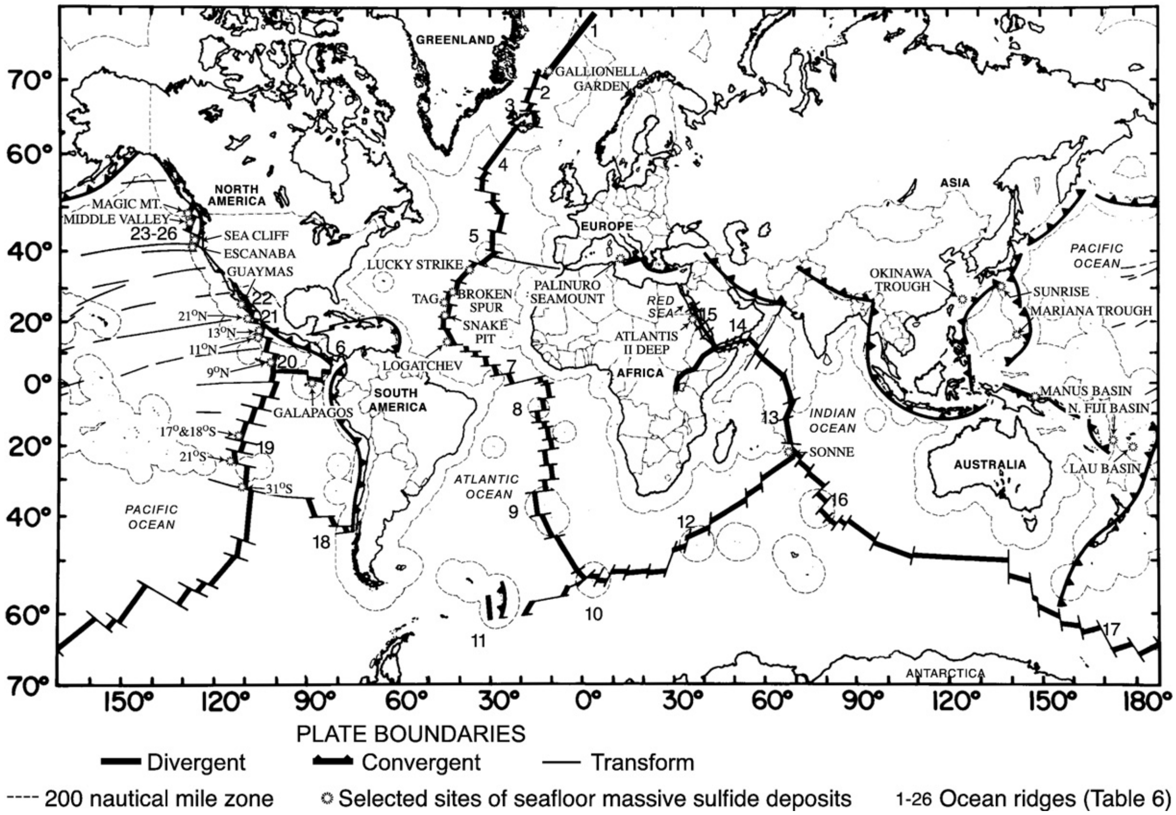

Fig. 11. World map showing plate boundaries, the 200 nautical mile-wide (370 km) Exclusive Economic Zone (EEZ; dashed line surrounding land), and the international Area seaward of the EEZ. Sites are shown at plate boundaries where some of the larger massive sulfide deposits have been found (Rona, 1983b; Rona and Scott, 1993; Andreev et al., 2000; Hannington et al., 2004). Only about 5% of submerged plate boundaries have been explored in sufficient detail to find axial deposits and virtually none of the oceanic lithosphere away from present plate boundaries to find relict deposits transported off-axis by seafloor spreading.

that concentrate the mineralization (e.g., Schardt et al., 2006). Salient characteristics of these settings are summarized, as follows.

4.2.1. Intermediate- to fast-spreading ocean ridges (full-rate of spreading 6 to 18 cm/year)

Ocean ridges at all spreading rates exhibit an axial zone of magmatic intrusion and volcanic extrusion flanked by marginal zones of extension (Fig. 12). The typical cross-sectional form of fast-spreading ocean ridge crests like the southern East Pacific Rise and intermediate-spreading ridge crests like the Juan de Fuca Ridge off northwest North America is that of a rise some tens of kilometers wide with hundreds of meters relief attributed to thermal expansion and volcanic construction (Fig. 12; Macdonald, 1986; Perfit and Chadwick, 1998; Small, 1998). A neovolcanic zone of extrusion bisects the rise ranging from a narrow axial summit trough (AST tens to hundreds of meters wide with walls tens of meters high) on fast-spreading ridges, to a wider axial valley (1 to 5 km wide) with faulted walls 50 to 1000 m high on ridges spreading at intermediate rates (Kappel and Ryan, 1986; Macdonald, 1998; Perfit and Chadwick, 1998). The axial summit trough and axial valley are floored predominantly by basaltic lobate and sheet flows representing relatively high effusion rates (Bonatti and Harrison, 1988; Head et al., 1996). A systematic variation exists on intermediate- and fast-spreading ocean ridges such that lobate/sheet flows dominate at segment centers and pillow flows and lava domes are more common on segment ends suggesting higher eruption effusion rates and magma pressure and lower magma viscosity at segment centers relative to segment ends (White et al., 2002, 2006; Soule et al., 2005; Macdonald, 2005). Multi-channel seismic profiling at sites along the spreading axis of intermediate- to fast-spreading ocean ridges has imaged magma bodies 1 to 2 km beneath the seafloor interpreted to consist of a thin lens (tens of meters thick) of partial melt overlying a thicker zone of crystal mush (Sinton and Detrick, 1992;

Table 6
Submerged divergent plate boundaries (ocean ridges) within 200 nautical miles (370 km) of land within Exclusive Economic Zones of coastal states (location identified by number in Fig. 11; modified from Rona, 1983b)

Location	Ocean	Oceanic ridge	Land	Country
1	Greenland Sea−Norwegian Sea	Mohns Ridge	Svalbard Islands	Norway
			Jan Mayan Island	Norway
2	Greenland Sea−Norwegian Sea	Iceland-Jan Mayan Ridge	Jan Mayan Island	Norway
			Greenland	Denmark
			Iceland	Iceland
3	Greenland Sea−Norwegian Sea	Kolbeinsey Ridge	Iceland	Iceland
4	North Atlantic	Reykjanes Ridge	Iceland	Iceland
5	North Atlantic	Mid-Atlantic Ridge	Azores Islands	Portugal
6	Caribbean Sea	Cayman Spreading Center	Cayman Islands	Jamaica
			Swan Island	Honduras
7	Equatorial Atlantic	Mid-Atlantic Ridge	St. Peter and St. Paul Rocks	Brazil
8	South Atlantic	Mid-Atlantic Ridge	Ascension Island	United Kingdom
9	South Atlantic	Mid-Atlantic Ridge	Tristan de Cunha	United Kingdom
			Gough Island	United Kingdom
10	South Atlantic	Mid-Atlantic Ridge	Bouvet Island	Norway
11	South Atlantic−Scotia Sea	Scotia Spreading Center	South Sandwich Islands	United Kingdom
12	Indian Ocean	Southwest Indian Ridge	Prince Edward Islands; Marion Island	Republic of South Africa
13	Indian Ocean	Carlsberg Ridge	Chagos Archipelago	United Kingdom
14	Indian Ocean	Gulf of Aden Spreading Center	Democratic Republic of Yemen	Democratic Republic of Yemen
15	Red Sea	Red Sea Spreading Center	Yemen	Yemen
			Saudi Arabia	Saudi Arabia
			Ethiopia	Ethiopia
			Sudan	Sudan
			Egypt	Egypt
16	Indian Ocean	Southeast Indian Ridge	Amsterdam Island	France
			St. Paul Island	France
17	Antarctic Ocean	Pacific-Antarctic Rise	Antarctica	Australian Claim
			Antarctica	New Zealand Claim
18	Pacific Ocean	Chile Rise	Chile	Chile
19	Pacific Ocean	East Pacific Rise	Easter Island	Chile
20	Pacific Ocean	Galapagos Spreading Center	Galapagos Islands	Ecuador
21	Pacific Ocean	East Pacific Rise	Mexico	Mexico
22	Pacific Ocean	Gulf of California Spreading Center	Baja California	Mexico
23	Pacific Ocean	Gorda Ridge	Oregon, California	USA
24	Pacific Ocean	Juan de Fuca Ridge	British Columbia	Canada
25	Pacific Ocean	Endeavour Segment	British Columbia	Canada
26	Pacific Ocean	Explorer Ridge	British Columbia	Canada

Detrick et al., 1993). A relatively high continuity of magma supply and frequency of eruption rate (tens of years) along axis is related to continuity of ridge morphology, and with primarily axial focus of high-temperature hydrothermal discharge and associated mineralization (Lutz and Haymon, 1994; Perfit and Chadwick, 1998).

Detailed surveys with temperature and suspended particulate matter sensors towed along sections of intermediate- to fast-spreading axes have found a direct linear relation between spreading rate, magma supply, and the spatial frequency of hydrothermal venting (Baker and German, 2004; Baker et al., 2004). The frequency of venting sites based on plume incidence along the spreading axis ranges from about 1.5 to 4.5 sites per hundred kilometers for fast- to ultra-fast-spreading rates on the East Pacific Rise at sites surveyed from about 18°N to about 32°S, and intermediate-spreading rates on the Juan de Fuca Ridge (Fig. 21 of Baker and German, 2004). Examination of the seafloor at many of these sites reveals that the sources generally are isolated groups of active sulfide chimneys venting high-temperature fluids (350° to 405 °C; combined thermal output ~ 10 MW; combined fluid flux 132–2885 $cm^3\ s^{-1}$; Table 2 in Hey et al., 2006) and occupying small areas (hundreds of m^2; Bäcker, 1980; Bäcker et al., 1985;

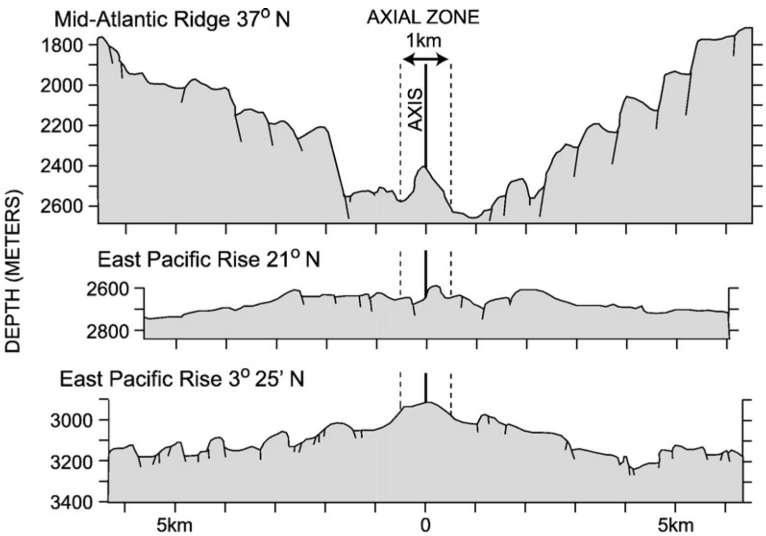

Fig. 12. Topographic cross-sections of ocean ridge crests showing axial zone of volcanic intrusion–extrusion and variation of morphology with spreading rate (vertical exaggeration 4:1; Macdonald, 1986).

Marchig et al., 1987, 1988; Holler, 1993; Rona and Scott, 1993). As stated by Marchig (2000), "Although hydrothermal activity is high in the southern part of the East Pacific Rise, massive sulfide ores do not occur in large amounts or as big edifices, first, because most of the precipitates from the hydrothermal solutions get dispersed to the surrounding sediment, and second, because the edifices formed become covered with fresh lava flows".

Exceptions include a large relict VMS deposit on the flank of an off-axis seamount near 13°N on the East Pacific Rise (Hekinian and Fouquet, 1985; Fouquet et al., 1996), and the active Magic Mountain VMS mound near 49°N on the Southern Explorer Ridge off British Columbia (250 m diameter, 18 m thick; estimated 5 million tonnes; Scott et al., 1990). The northern Gorda Ridge offshore the U.S. state of Oregon, is an intermediate-spreading-rate ridge (full-rate 6 cm/year), which exhibits morphology characteristic of slow-spreading ocean ridges. This variation is attributed to the role of a hotter mantle beneath typical intermediate- to fast-spreading ridges, like the East Pacific Rise and the Juan de Fuca Ridge, producing greater amounts of decompression melting, a thicker crust and a thinner, weaker lithosphere than that along the Gorda Ridge (Hooft and Detrick, 1995). The high-temperature Sea Cliff hydrothermal field, perched on a fault block on a wall of northern Gorda Ridge 2.6 km east of the spreading axis (Rona et al., 1990; Von Damm et al., 2006), is capped by a hydrothermal crust that is inferred to seal an underlying VMS deposit. The Sea Cliff deposit is considered analogous to the basalt-hosted ~6 million tonne Turner–Albright VMS deposit in the ophiolite belt of Oregon that was obducted from a back-arc basin during the Jurassic period (Zierenberg and Schiffman, 1990; Zierenberg et al., 1995).

4.2.2. Slow-spreading ocean ridges (full-rate of spreading < 4 cm/year)

The typical cross-sectional form of slow-spreading ocean ridge crests is an axial ("rift") valley with a floor up to 10 km wide, flanked by walls up to several kilometers high constructed of outward dipping normal fault blocks, and widths up to several tens of kilometers between topographic highs at the top of each wall (Fig. 12). Axial valley cross-sections are generally asymmetric, manifesting differences in spreading rates, offsets of the neovolcanic zone of intrusion and extrusion from the geometric center of the valley, and the occurrence of detachment (normal) faulting that involves extension and exposure of lower crust in a wall of the axial valley (Tucholke and Lin, 1994; Schroeder and John, 2004).

Acoustic interference from rough ridge morphology and bounding faults has impeded interpretation of seismic records for magma chambers beneath slow-spreading ocean ridges. A recent multi-channel seismic survey over active hydrothermal vents in the axial valley of the central section of a segment encompassing the Lucky Strike hydrothermal field near 37° North on the Mid-Atlantic Ridge imaged a bright reflector interpreted as a magma chamber extending about 5 km along axis and at a depth of 3 km beneath an axial volcano (Singh et al., 2005). The magma chamber is inferred to provide the magma source for the volcano and the heat source for the high-temperature hydrothermal circulation.

In contrast, an active-source seismic experiment and passive record of microseismicity at the TAG hydrothermal field situated in the axial valley of the Mid-Atlantic Ridge near 26°N, indicate that the active high-temperature sulfide mound situated 2.4 km east of the spreading axis (Figs. 8 and 9) is offset horizontally by several kilometers from a magmatic heat source (melt zone) inferred to lie at least 7 km beneath a neovolcanic zone on the western half of the axial valley. This geometry implies fluid flow pathways from the heat source to the active mound along intervening faults beneath the axial valley (Kleinrock and Humphris, 1996; Canales et al., 2007; deMartin et al., 2007). Diversity in depth of heat sources and geometry of fluid pathways is a characteristic of slow-spreading ocean ridges.

The frequency of eruptions on slow-spreading ridges is considered to be hundreds to thousands of years (Perfit and Chadwick, 1998). The frequency of venting sites based on plume incidence detected along different sections of axes of slow-spreading ocean ridges ranges from about 1 per 400 km to 3 per 100 km (~1 per spreading segment) based on a few surveys (Klinkhammer et al., 1984; Baker and German, 2004; Fig. 21 in Baker et al., 2004).

Active high-temperature venting sites and hydrothermal deposits diverge from the axial alignment generally observed at intermediate- to fast-spreading ridges, and have been found at distances of at least kilometers off-axis. For example, the Snake Pit hydrothermal field occurs on a neovolcanic ridge at the Mid-Atlantic Ridge axis near 23°N (Fouquet et al., 1993), whereas the TAG field encompasses actively venting high- and low-temperature zones at distances between 2 and 8 km to the east of the axis (Rona et al., 1993); the neovolcanic zone occurs on the western half of the axial valley (Kleinrock and Humphris, 1996). This divergence of the distribution of actively forming deposits on slow-spreading ridges from axial linearity indicates the increased role of faulting in providing extensive pathways for charging and discharging hydrothermal systems. The faults act as conduits to distribute high-temperature hydrothermal solutions from magmatic heat sources beneath the axial valley; and lower-temperature solutions alternatively produced by conductive cooling, exothermic reactions associated with serpentinization of peridotites, and by deep penetration to tap geothermal gradients.

Upper mantle peridotites are especially accessible to hydration in regions of crustal thinning attributed to low magma budgets, like the Mid-Atlantic Ridge crest near 15°N (Rona et al., 1987, 1992; Cannat et al., 1992), and/or tectonic unroofing by detachment faulting (Francis, 1981; Cannat et al., 1992, 1997). Seafloor hydrothermal systems inferred to be driven primarily by serpentinization reactions or deep geothermal gradients, like the Lost City field situated 15 km west of the Mid-Atlantic Ridge axis on the north wall of the Atlantis transform fault at 30°N, precipitate calcium carbonate and magnesium hydroxide from alkaline solutions up to 75 °C devoid of sulfides and of Fe–Si or Mn oxyhydroxides (Kelley et al., 2001; Schroeder et al., 2002). Serpentinization reactions can act in concert with magmatically-driven systems, as inferred for the Rainbow (Fouquet et al., 1997) and Logatchev (Krasnov et al., 1995; Mozgova et al., 2005) fields near 36°N and 15°N on the Mid-Atlantic Ridge, respectively. Analysis of surface samples from these two sites that involve hydrothermal circulation in both mafic and serpentinized ultramafic rocks indicates that the sulfides exhibit higher contents of Au, Cd, Co and Ni than sulfides from seafloor hydrothermal sites solely hosted in basalt (Mozgova et al., 2005).

As a consequence of magmatic and tectonic differences, sites of high-temperature hydrothermal venting and mineralization are farther apart and more irregularly distributed on slow- than on intermediate- to fast-spreading ocean ridges. However, hydrothermal fields and VMS deposits tend to be larger on the former (Rona, 1988; Fig. 28 in Hannington et al., 1995) although, as noted, large deposits can form at all spreading rates (Rona, 1988). Reasons for this tendency include up to a factor of ten longer residence time of a parcel of oceanic crust near heat sources beneath the axial valley on slow- than on intermediate- to fast-spreading ocean ridges, to build large deposits by superimposition of multiple ore-forming cycles energized by magmatic cycles on a time scale of hundreds to thousands of years with fluid pathways maintained by ongoing seismicity (Rona, 1987). In sum, slow-spreading ocean ridges exhibit a diversity of crustal structure, heat sources, geometry and distribution of hydrothermal systems in comparison to intermediate- to fast-spreading ocean ridges which typically are more homogeneous with reference to these features (Karson, 1998; Perfit and Chadwick, 1998).

4.2.3. Volcanic island arcs

The settings found to be most favorable for concentration of massive sulfide deposits at volcanic island arcs, like those arcs associated with convergent plate boundaries of the western Pacific, are volcanic craters and calderas of seafloor volcanoes in the fore-arc region, and sites at spreading axes in back-arc basins (Fig. 10; Ishibashi and Urabe, 1995), including sites where the spreading axis is rifting continental crust (e.g., Woodlark Basin; Binns et al., 1993). Although the hydrothermal processes are similar, the composition of volcanic rocks varies from basalt of mid-ocean ridges (MORB) to calc-alkaline felsic lavas (andesite, dacite, rhyolite) in island arcs. Rhyolite-hosted massive sulfide deposits may occur in a variety of extensional settings (Sillitoe, 1982). The metal contents of deposits associated with the volcanic island arcs of the western Pacific are systematically higher than those at ocean ridges (Table 7). This increase in metals is attributed to a combination of contribution of metals from subducting oceanic crust (Hedenquist and Lowenstein, 1994), and to a larger contribution of metals from magmatic fluids, in addition to those leached from the ocean crust by circulating seawater (Urabe, 1987; Yang and Scott, 1996, 2002, 2005). Examples of hydrothermal deposits in volcanic island arc settings in the submarine fore- and back-arc settings are, respectively.

Fore-arc setting: The Sunrise VMS deposit (mound ~400 m in diameter and 30 m high; estimated 9×10^6 tonnes; water depth 1300 m) is accumulating from solutions focused by upwelling at the faulted east wall of the caldera (3.5 km in diameter) of the Myojin Knoll silicic active fore-arc volcano in the Izu–Bonin Arc (or Izu–Ogasawara Arc) south of Japan (Fig. 11; Usui and Glasby, 1998; Iizasa et al., 1999; Glasby et al., 2000; Yamazaki, 2006a). Highly fractionated hydrothermal manganese (todorokite) crusts (thickness ~1 cm) lie along the northern and western margins of the Myojin Knoll caldera and are found on active and inactive seamounts associated with the volcanic front in this arc (Usui and Glasby, 1998). Similar hydrothermal manganese crusts are broadly distributed on active submarine volcanoes and at back-arc spreading axes (Usui and Someya, 1997; Glasby et al., 2000. A deposit within a ~200 m diameter area of the fissured summit of Conical seamount (water depth 1050 m) in the fore-arc region offshore New Ireland (Figs. 11 and 13), contains the first indications found on the seafloor of gold-rich epithermal-style vein mineralization hosted in ankaramitic altered basalt with gold content (maximum 230 ppm Au; average 26 ppm; $n = 40$ random grab samples), comparable to that of the Ladolam gold deposits being mined on neighboring Lihir island (Herzig, 1999; Herzig and Hannington, 2000; Petersen et al., 2002). On Brothers submarine volcano in the Kermadec fore-arc immediately northeast of New Zealand two relatively large active areas of chimneys and sub-cropping sulfide (each area ~600 m by at least 50 m) lie at the faulted margin of the dacitic caldera (basal diameter ~3 km; floor at water depth of ~1850 m; de Ronde et al., 2003a, 2003b, 2005). Neptune Minerals, a company which holds Prospecting Licenses within New Zealand's EEZ, reports (http://www.neptuneminerals.com) that, "Although active SMS (seafloor massive sulfide)-forming processes are observed in widespread sulphide chimneys there is no accumulation at or near the surface of commercial quantity of sulphide mineralization in the areas drilled", based on 23 core holes up to 14 m deep drilled in 2005. Their report identifies the calderas of Brothers and adjacent Healy submarine volcanoes as "drilling targets".

Back-arc setting: The Jade VMS deposit hosted in altered rhyolite of a caldera-like structure at 1650 m water depth in the Okinawa Trough intra-continental back-arc basin (Halbach et al., 1989, 2003) is considered to be an analog of the type Miocene Kuroko stratiform massive sulfide deposits hosted in felsic calc-alkaline volcanic rocks (rhyolite and dacite) of Honshu island, Japan (Scott, 1980; Urabe, 1987). The PACMANUS (Papua New Guinea–Australia–Canada–Manus) hydrothermal field in the eastern Manus intra-oceanic back-arc basin of the Bismarck Sea of Papua New Guinea contains a surface area at least 800 m × 3 km of discontinuous massive sulfide deposits hosted in dacitic and andesitic volcanic rocks between water depths of 1650 m and 1800 m (Fig. 13; Binns and Scott, 1993). Ocean Drilling Program Leg 193 studied alteration in the host rocks of the PACMANUS deposits (Shipboard Scientific Party, 2002; Bach et al., 2003; Vanko et al., 2004). In 1997 Nautilus Minerals Corporation, an Australian mining company, leased two sites in this field (area 5000 m^2; Fig. 13) within territorial waters of Papua New Guinea from that government to evaluate for mining (Fig. 13; Broad, 1997; Malnic, 2001;

Table 7
Bulk chemical composition of seafloor massive sulfides from ocean ridges and back-arc spreading axes (Herzig and Hannington, 1995)

	Ocean ridges	Back-arc spreading axes	
	Volcanic-hosted	Intra-oceanic	Intra-continental
n	890	317	28
Fe (wt.%)	23.6	13.3	7.0
Zn	11.7	15.1	18.4
Cu	4.3	5.1	2.0
Pb	0.2	1.2	11.5
As	0.03	0.1	1.5
Sb	0.01	0.01	0.3
Ba	1.7	13.0	7.2
Ag (ppm)	143	195	2766
Au	1.2	2.9	3.8

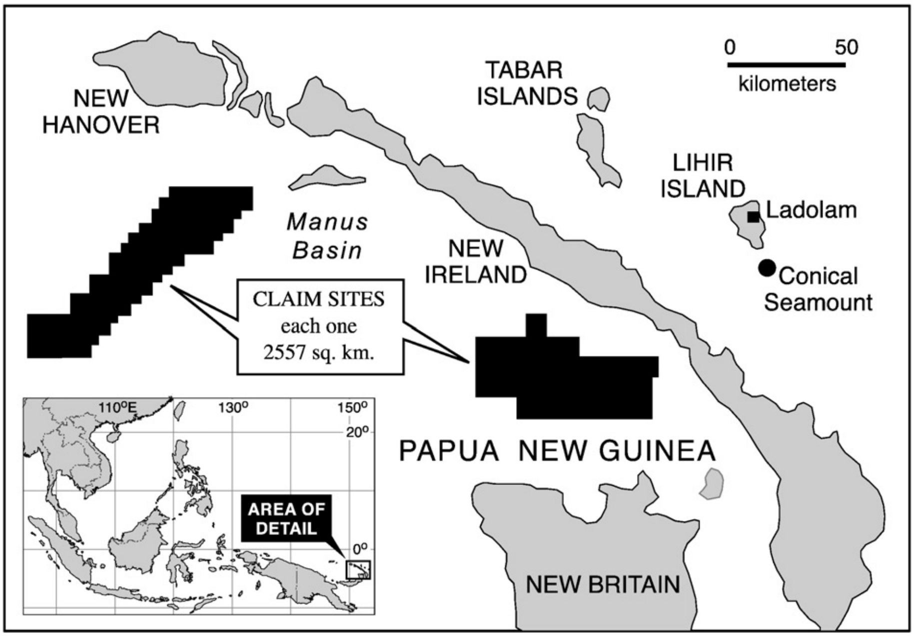

Fig. 13. Map showing the location of two tracts of the PACMANUS hydrothermal field containing active and relict massive sulfide deposits (water depth 1500 to 2500 m) in the Manus back-arc basin in the Bismarck Sea of Papua New Guinea (PNG). The PNG government leases the tracts and additional areas to Nautilus Minerals Inc., an Australia-based mining company, for exploration and development (modified from Broad, 1997; Herzig, 1999; Malnic, 2001). The crater of a volcanic seamount (Conical Seamount; water depth 1050 m) in the fore-arc region northeast of New Ireland contains hydrothermal minerals with variable gold content (Petersen et al., 2002, 2005) comparable with the operational Ladolam gold deposits on neighboring Lihir island (Herzig and Hannington, 2000).

International Marine Minerals Society, 2002, 2005). They report (http://www.nautilusminerals.com, news releases dated 16 November 2005 and 21 February 2006) that 39 samples dredged in 2005 from inactive deposits in the “Suzette” area (280,000 m^2; now part of their Solwara-1 Project) contain a weighted average of 15.52 g/tonne Au, 12.24 wt.% Cu, 4.20 wt.% Zn, 1.77 wt.% Pb, and 256 g/tonne Ag.

Volcanic island arcs are clearly favored for initial potential mining of seafloor VMS deposits by virtue of the generally higher metal contents of deposits in the fore- and back-arc settings than those deposits formed at ocean ridges and the situation of these settings within the EEZ of sovereign coastal states (Herzig, 1999).

5. Modern and ancient VMS deposits and proximal low-temperature deposits

5.1. Exploration criteria, setting and genesis

Aspects of active seafloor hydrothermal systems that have particular potential for elucidating outstanding questions concerning exploration for and genesis of ancient VMS deposits pertain to spatial and temporal interactions of mineralization with tectonic, magmatic, and biological processes (Skinner, 1997; Scott, 1997; Hannington et al., 2005; Franklin et al., 2005). The TAG hydrothermal field situated in the axial valley of the slow-spreading Mid-Atlantic Ridge in the central North Atlantic near 26°N, 45°W is selected to address some of these questions, because it contains an assemblage of large VMS deposits that have been the subject of investigation by the international community since discovery in 1985 as the first hydrothermal field found on a slow-spreading ocean ridge (Fig. 8; Rona et al., 1986), refuting the consensus of the scientific community at that time that high-temperature hydrothermal systems could only occur on intermediate- to fast-spreading ocean ridges. At least four active and relict hydrothermal zones containing large massive sulfide mounds (diameter 100 m to ~1 km) are present in the field, spanning a radiometrically dated age range of 140,000 years (Lalou et al., 1995, 1998). The state of the mounds ranges from young-hot to old-cold and encompasses the evolution of a hydrothermal ore-forming system from origin to fate as the hydrothermal zones are formed in the axial valley, uplifted in one wall of the axial valley and rafted away by the spreading seafloor. These high-temperature deposits, as well as proximal low-temperature deposits, lie within a 5 km by

5 km area of the eastern floor and lower wall of the axial valley between water depths of 2400 and 3700 m. Relevant findings at the TAG field are presented in context of a systematic exploration procedure of progressively closing range from far to near by sensing chemical and physical properties of an active hydrothermal source and by deciphering the geologic controls of mineralization (Rona, 1978a, 1983a, and in Table 2 of Rona, 1999–2000; Fornari et al., 1997), as follows.

5.1.1. Exploration procedure

Approaching the TAG field the first hydrothermal signal detected is that of weak acid soluble iron and manganese oxide particles suspended in the water column at the depth of the Mid-Atlantic Ridge crest (2000 m) at a distance of 750 km west of the crest. This may be an integrated signal from hydrothermal fields in the axial valley of the Mid-Atlantic Ridge transported westward by deep currents that flow through fracture zones that breach the walls of the valley (Rona, 1978a; Fig. 10 in Rona, 1980a). The signal of suspended hydrothermal particulates detected to the west of the Mid-Atlantic Ridge is absent to the east of the ridge, indicating that the deep currents that are transporting the particulates are unidirectional to the west (Rona, 1980a), consistent with simulated flow trajectories (Fig. 14.5 in Speer et al., 2003). Continuing eastward toward the ridge, the concentration of Fe and Mn in particulate and dissolved phases and of the conservative primordial gas ^3He (expressed as δ^3He) released from the mantle through seafloor hydrothermal systems would be expected to gradually increase, leading to a neutrally buoyant hydrothermal plume in the axial valley (Jenkins et al., 1980). The neutrally buoyant plume between about 200 and 450 m above the source vents is vertically connected to the vents by a narrow (ca. 15 m diameter) buoyant plume discharging from a group of black smokers vigorously venting high-temperature (365 °C) solutions (Rona et al., 1986; Klinkhammer et al., 1986; Rona and Speer, 1989; Speer and Rona, 1989; Edmond et al., 1995; Rudnicki, 1995; Chiba et al., 2001) from multiple chimneys. The chimneys top a massive sulfide mound 200 m in diameter and 35 m high with base at a water depth of 3670 m on the floor of the axial valley. Measurements of thermal output of the plume discharging from the TAG active sulfide mound range from 86 MW (Goto et al., 2003), to 120 MW (Rona et al., 1993), to 500–940 MW (Rudnicki and Elderfield, 1992), to 1.7 GW (Wichers et al., 2005) with estimated fluid flux to the neutrally buoyant plume of 1460–2740 $m^3\ s^{-1}$ (Rudnicki and Elderfield, 1992), reflecting real variations of output with time and different measurement methods. The neutrally buoyant plume discharging from the active sulfide mound is laterally advected by tidal forcing and extends tens of kilometers along the axial valley at an altitude of several hundred meters above the seafloor (Rudnicki, 1995). The corresponding signal on the seafloor is a trail of metalliferous sediments produced by fallout from the water column increasing from a trace component diluted by normal deep-sea sediment at distances of tens to hundreds of kilometers to dark brown, centimeter-scale iron-rich layers in a light tan carbonate lutite matrix within kilometers of the source vents (Shearme et al., 1983; Metz et al., 1988), to Fe-rich (20–40% Fe) red-brown mud near the base of and on the mound derived from mass-wasting of oxidized sulfides and settling of suspended particulate matter from the overlying hydrothermal plume with distinctive isotopic and REE/Fe distributions (German et al., 1993).

5.1.2. Geologic setting

The TAG active high-temperature sulfide mound lies 2.4 km east of the spreading axis, near the center of a 40 km-long NE–SW-trending spreading segment adjacent to a marginal fault at the base of the east wall of the axial valley. The massive sulfide mound laps onto an adjacent old pillow mound of similar size (Fig. 8). The massive sulfide and pillow mounds are situated at the intersection of axis-parallel with axis-oblique fault systems within a faulted and fissured zone that occupies the eastern half of the 7 km-wide axial valley floor (Karson and Rona, 1990; Kleinrock and Humphris, 1996; Bohnensteihl and Kleinrock, 1999). The fault intersections are inferred to act as past conduits for magma and present conduits for hydrothermal solutions, and are kept open to fluid flow by ongoing microearthquake seismicity (Kong et al., 1992; Smith et al.; 2005; deMartin et al., 2007). Patchy glassy sheet flows with dimensions of tens of meters and a volcanic seamount lie along the spreading axis (Eberhart et al., 1988). A neovolcanic zone that exhibits constructional volcanic morphology of linear volcanic ridges and hummocks occupies the western half of the axial valley (Kleinrock and Humphris, 1996).

5.1.3. Structure

The walls of the axial valley are asymmetric with reference to morphology and crustal structure. The east wall that hosts the TAG field is higher (2400 to 3600 m), steeper, and smoother than the west wall (2800 to 3600 m), where hydrothermal activity is absent (Karson and Rona, 1990; Kleinrock and Humphris, 1996). The crustal structure of the west side of the axial valley is typical of other volcanically constructed Mid-Atlantic Ridge segments, but the eastern side is underlain by a large, high-velocity seismic anomaly indicating the presence of lower crustal and/or serpentinized upper mantle rocks at anomalously shallow depths (deMartin et al., 2007). The central portion of the east wall, where some of the hydrothermal zones are located, exhibits an anomalous salient that projects about 3.5 km westward over the floor of the axial valley. Zonenshain et al. (1989) observed sheeted dikes and recovered gabbro from mid-depth (~3000 m) on the east wall indicating vertical uplift of ocean crust and upper mantle. They ascribed this uplift to expansion by serpentinization of the underlying peridotites of the upper mantle by circulating hydrothermal fluids accompanied by thinning of the basaltic layer of ocean crust. A ~50% deficit of boron measured in hydrothermal fluids sampled from vents on the TAG active high-temperature sulfide mound relative to seawater evidences uptake by hydration during ongoing serpentinization (Palmer, 1996). Particles of native nickel in sediments of the active sulfide mound may be derived from the upper mantle (Dekov, 2006). Using conservative estimates of the volume of serpentinite Palmer (1996) and Germanovich et al. (2006) estimate about 100 m of uplift from serpentinization, which can account for the westward salient in the east wall but not for the full uplift.

5.1.4. Magnetic signatures of detachment faulting and hydrothermal alteration

The east wall of the axial valley of the TAG segment is associated with an elongate axis-parallel low in crustal magnetization within the normal polarity Brunhes anomaly (McGregor et al., 1977; Wooldridge et al., 1992; Tivey et al., 1996). The active high-temperature sulfide mound and the relict sulfide mounds of the Alvin zone to the north (Fig. 8) are located at a transition between the low and a magnetization high related to the neovolcanic zone on the western side of the axial valley (Tivey et al., 2003). The magnetization low at the east wall is modeled as a product of a 4 km-wide zone of crustal thinning exposing the gabbros and sheeted dikes observed by Zonenshain et al. (1989) in the footwall of a detachment fault zone (Tivey et al., 2003). The modeling indicates that normal displacement on the detachment fault has produced about 4 km of horizontal extension within the past 350,000 years corresponding to the zone of crustal thinning (Fig. 8). The active and relict zones of the TAG hydrothermal system are located on the hanging wall of this fault. Seismic refraction and microearthquake data indicate that the upper portion of the detachment fault dips west toward the spreading axis at an angle of ~20°, passing under the active sulfide mound. At ~3 km beneath the seafloor the detachment fault rolls over into a steep (70°) west-dipping fault plane that can be traced 7 km downward associated with microearthquake hypocenters into a region interpreted as a melt zone (deMartin et al., 2007). The upper low-angle section of the detachment fault may be explained by flexure and isostatic uplift from an initial high-angle congruent with the deeper portion of the fault plane (Garces and Gee, 2007). Silica geobarometry of high-temperature solutions discharging from the active sulfide mound indicates that the fluids last equilibrated with quartz at 2 to 3 km below the seafloor (Campbell et al., 1988), possibly at a mid-crustal magma body fed by magmatic intrusions from the inferred melt zone at and below 7 km beneath the seafloor. Seismic refraction data, a seismic velocity model, and the distribution of earthquake hypocenters effectively preclude large crustal magma chambers (Canales et al., 2007; deMartin et al., 2007). However, smaller crustal magma bodies unresolved by the seismic studies may be periodically replenished from larger deep sources. Time-series measurements of high-temperature fluid chemistry at the TAG active sulfide mound (years sampled: 1986, 1993, 1995, 1998, 2003; Edmond et al., 1995; Edmonds et al., 1996; Chiba et al., 2001; Parker et al., 2005) indicate compositional stability which, in turn, may relate to depth of circulation. Direct observations suggest that the volume flux and discharge zones of the mound are changing on a time scale of years. Normal displacement on the detachment fault accommodates lithospheric extension of the east side of the axial valley. Movement on the detachment fault zone may episodically increase the permeability of the hanging wall preferentially reactivating the hydrothermal zones that lie on the hanging wall with mostly asynchronous hydrothermal events and at least one synchronous field-wide hydrothermal event (~50,000 years ago).

Short-wavelength (<100 m) lows in magnetic intensity were measured near the seafloor associated with the active high-temperature sulfide mound and with two of the relict sulfide mounds adjacent to the broader more intense magnetic low at the east wall. These short-wavelength magnetic lows are attributed to hydrothermal alteration of the magnetic mineral component in pipe-like up-flow alteration zones under the individual mounds through the host basalts, as distinguished from Curie point effects (Rona, 1978b; Tivey et al., 1993, 1996). A similar magnetic low was measured at the Agrokipia B sulfide ore body in the Cretaceous Troodos ophiolite of Cyprus and is also related to hydrothermal alteration of magnetic minerals (e.g., magnetite) to less magnetic minerals (e.g., titanomaghemite, sphene) in cores recovered from drill holes in the deposit (Johnson and Pariso, 1987). Magnetic lows have been recorded at certain other ancient VMS deposits, such as Kuroko-type massive sulfides (Scott, 1980).

Thus, the asymmetric uplift, detachment fault zone, and salient in the east wall of the TAG segment of the axial valley are interrelated products of tectonic extension and serpentinization (deMartin et al., 2007; Germanovich et al., 2006). Pathways for fluid circulation intermittently opened by tectonic activity are inferred to have tapped multi-level magmatic heat sources to drive hydrothermal episodes that deposited the assemblage of massive sulfide mounds of the TAG hydrothermal field within the overall structural control of the detachment fault zone and intersections of the zone with cross faults, as inferred for many ancient VMS deposits (Figs. 12 and 19 in Cox et al., 2001; Cox, 2005; Fig. 13 in Franklin et al., 2005). The mantle uplift, crustal thinning, and long-wavelength magnetic low on a scale of kilometers to tens of kilometers associated with the entire TAG field and the short-wavelength magnetic lows on a scale of tens to hundreds of meters attributed to alteration pipes that underlie individual sulfide bodies are seismically and magnetically detectable exploration targets in modern and ancient VMS deposits hosted in ocean crust.

5.2. 3-D form and overall composition of TAG active high-temperature sulfide mound

The three-dimensional form and overall composition of the active high-temperature sulfide mound on the floor of the rift valley in the TAG field, was determined primarily by 17 holes up to 125 m beneath the seafloor drilled by the Ocean Drilling Program Leg 158 in 1994 (Fig. 9; Humphris et al., 1995a). This mound (200 m in diameter, 35 m high) contains 3.9 million tonnes of massive sulfides comprising a lens-shaped body with ~2% Cu underlain by a stockwork or stringer zone with ~1% Cu (Humphris et al., 1995b; Herzig et al., 1998; Hannington et al., 1998; Petersen et al., 2000). The form and overall composition of the mound is similar to that of many ancient VMS bodies preserved in ophiolites (Fig. 9). Analogous VMS deposits include those of the Troodos ophiolite in Cyprus, the Semail ophiolite in Oman, and the Newfoundland Bay of Islands ophiolite. A striking difference from ancient VMS deposits is the presence in the TAG mound of a significant volume (~30%) of anhydrite as matrix of a massive sulfide breccia in the sulfide lens and as veins in the stockwork zone. The anhydrite forms by reaction involving calcium in hydrothermal solutions and sulfate in seawater entrained within the mound and conductively heated to temperatures in the stability zone of anhydrite (>150 °C; Tivey et al., 1995). Dissolution of the anhydrite due to its retrograde solubility (solution temperatures <150 °C) can explain its absence in many ancient VMS deposits, and the presence of massive sulfide breccias consolidated by collapse of anhydrite matrix-supported clasts (Hannington et al., 1998) and enhanced by local seismicity. The apparent stratigraphy with depth within the TAG mound from massive pyrite breccia, to anhydrite-supported sulfide breccias, though underlying pyrite–silica anhydrite breccias, to quartz-cemented breccias, reflects a general process of progressive infilling and replacement of the breccias by anhydrite, followed by quartz (Fig. 9; Hannington et al., 1998).

5.3. Zone refinement in TAG active high-temperature sulfide mound

The bulk of the TAG active sulfide mound is composed of massive pyrite and anhydrite-cemented sulfide breccias (Fig. 9). Metal enrichment (Cu, Zn, Ag, Au, Sb) in the mound is restricted to the upper 5 m of the deposit (Hannington et al., 1998). Internal steep vertical zonation suggests that a long history of hydrothermal reworking (at least 10,000 years; Lalou et al., 1998) has effectively stripped the constituents that are soluble at lower temperatures from the massive sulfides (Zn, Au, Ag, As, Pb, Sb) and concentrated them at the top of the deposit. Zone refining by upwelling of hot (>350 °C) hydrothermal fluids resulted in a pyritic massive sulfide body at depth and in the enrichment of selected trace elements and Zn near the surface of the mound away from the central high-temperature up-flow zone (Petersen et al., 2000; Koski et al., 2003). This zonation is similar to the strong co-enrichment of Cu and Zn observed at the top of many Cyprus-type deposits (Constantinou and Govett, 1972, 1973), and was actually applied to target mining of an oxide cap enriched in gold on the mid-Cretaceous Tambo Grande 1 VMS deposit in northern Peru (Tegart et al., 2000; J. M. Franklin et al., 2005 and pers. comm.).

5.4. Clustered mode of massive sulfide mounds in the TAG field

A similar distribution and size of massive sulfide bodies as that in the TAG field (Fig. 8) is observed in many ancient VMS deposits. For example, in the Noranda Archean felsic province of Canada one to three massive sulfide mounds per kilometer occur on two successive horizons, where hydrothermal alteration extends into the lava flows on top of the lower deposits and overprints younger deposits stacked vertically above the up-flow zone (Fig. 14; Fig. 14 in Knuckey et al., 1982; Fig. 20 in Galley et al., 1995; Fig. 26 in Hannington et al., 1995). In the Cretaceous Troodos ophiolite of Cyprus massive sulfide mounds tend to cluster in 5 by 5 km areas like the TAG assemblage (e.g., Skouiotissa–Phoenix–Mavrovouni–Apliki, Agrokipia–Kokkinoyia–Memi–Kokkinopezoula, Kambia–Kapedhes–Peristarka–Pytharochoma, Mathiati–Sha, Limni–Kinousa–Evoymeni–Uncle Charles; Bear, 1960; Constantinou and Govett, 1972; Rona, 1973; Geological Survey Department, 1982). In the Hokuroko district of Japan, the massive sulfide and major vein deposits tend to form groups with internal spacings of 1 to 2 km, while the groups (e.g., Hanaoka–Matsumine–Shakanai) have spacings ranging between 8 km (e.g., Furotope–Kosaka) and 17 km (e.g., Hanaoka–Furotobe (Figs. 1 and 3 in Sato et al., 1974; Solomon, 1976). The Rosebery and Mt. Lyell massive sulfide bodies in western Tasmania exhibit a linear distribution with spacing between bodies of kilometers (Solomon, 1976). The cluster of five orebodies that comprise the Neves–Corvo VMS deposit of the Iberian pyrite belt (siliciclastic-felsic setting in epicontinental back-arc basin; Franklin et al., 2005), exhibit spacing of hundreds of meters to kilometers (Relvas et al., 2006).

5.5. Chronology of sulfide mounds in the TAG field

The resolution of radiometric dating precludes distinguishing hydrothermal events in individual sulfide bodies of ancient clusters. A combination of Pb- and U-series isotopes has been applied to radiometrically date individual sulfide bodies in the TAG field (Figs. 8 and 9; Lalou et al., 1990). The active sulfide mound was initiated about 50,000 years ago. Since that time, high-temperature hydrothermal activity has been episodic with active periods of relatively short duration (tens to hundreds of years) separated by hiatuses of 3000 to 5000 years (Lalou et al., 1990, 1993, 1995, 1998; You and Bickle, 1998). The current high-temperature episode began in about the year 1930 (Lalou et al., 1990). The evidence for episodic activity from radiometric dating is consistent with estimates from thermal and chemical balances based on direct measurements of the size, chemical composition, heat flux, and fluid composition of the TAG active sulfide mound and the apparently analogous size, composition, and reaction zone of the Skouriotissa VMS deposit on Cyprus (Humphris and Cann, 2000). They estimated that about 2×10^{19} J of energy supplied at high temperature and extraction of metals from a reaction zone 1–2 km^3 in volume, with the addition of seawater sulfate, can form the observed massive sulfide deposit consistent with earlier estimates (Lowell and Rona, 1985). The sulfur isotope composition of the sulfides (+4.4 to 8.2‰ $\delta^{34}S$; average 6.5‰) evidences the introduction of heavy seawater sulfate to the hydrothermal fluid (Petersen et al., 2000). The 50,000 year hydrothermal event that initiated the active high-temperature mound is recorded in the other massive sulfide bodies in the TAG field indicating that this event was field-wide, while other events apparently affected individual mounds separately.

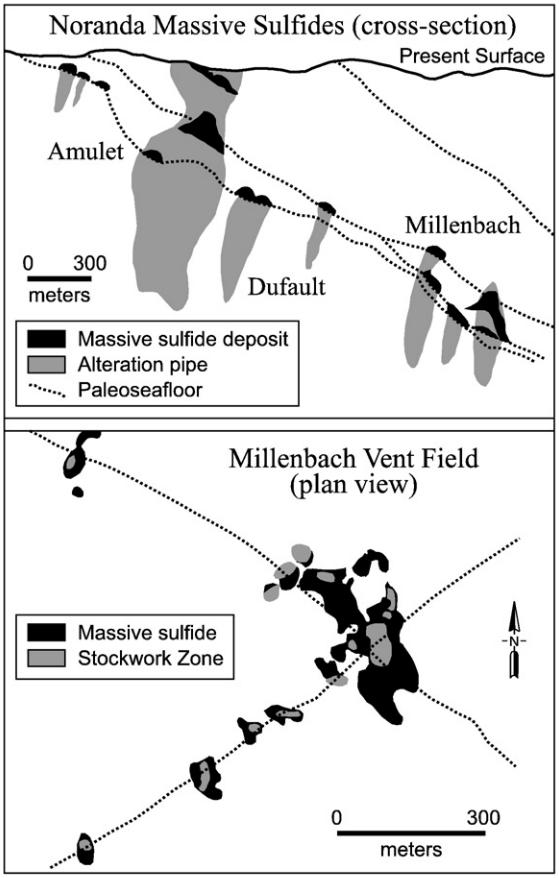

Fig. 14. Cross-section (top) of massive sulfide deposits in the Archean Noranda volcanic complex (~2.7 thousand million years old) of eastern Canada showing clustered distribution, stacking of massive sulfide bodies and underlying alteration pipes (stockwork or feeder zone). Map view (bottom) of hydrothermal massive sulfide mounds clustered along an inferred fault intersection in a 2 km × 2 km area of the Millenbach vent field of the Noranda volcanic complex (Knuckey et al., 1982; Hannington et al., 1995). The size, clustered distribution and the alteration pipes of the massive sulfide bodies are similar to those of the TAG hydrothermal field on the Mid-Atlantic Ridge (Figs. 8 and 9).

5.6. Proximal low-temperature deposits in the TAG field

A zone about 3 km in diameter encompassing patchy low-temperature Fe–Si and Mn oxyhydroxide deposits and diffusely venting low-temperature (<25 °C) solutions lies within the detachment fault zone on normal fault blocks of the east wall of the rift valley 5 to 8 km east of the spreading axis and 3 km east of the active sulfide mound in the TAG field (Fig. 8; Rona et al., 1984; Thompson et al., 1985; Humphris et al., 2003).

The isolated position of the zone at mid-depth (2300 to 3100 m) on the east wall and the low-temperature venting and mineralization suggests that the heat source is magmatic with conductive cooling of solutions during flow along kilometers-long fault pathways produced by the detachment faulting (Fig. 8). The deposits primarily comprise manganese oxides as laminated crystalline birnessite up to several centimeters thick exhibiting extreme fractionation (40% Mn with only trace amount of other metals in contrast to hydrogenetic deposits; Scott et al., 1974) and relatively rapid radiometrically measured deposition rates (~200 mm/million years) one to two orders of magnitude faster than deposition rates of hydrogenetic manganese crusts and nodules (Scott et al., 1974). Also present are massive green earthy Fe-rich silicate (nontronite); and red earthy amorphous Fe-oxides (Thompson et al., 1985). The manganese oxides range in age from 0 to 125,000 years (Lalou et al., 1990). A similar zone of low-temperature deposits and diffuse flow partially overlies relict massive sulfides along the eastern margin of the Mir zone where this zone laps onto an old pillow dome with high heat flow (Fig. 8; Rona et al., 1996). Manganese oxides with similar hydrothermal properties (extreme fractionation and relatively high accumulation rates) occur on active submarine volcanoes and at back-arc spreading centers in the western Pacific (Usui and Someya, 1997; Usui and Glasby, 1998). The TAG and western Pacific fractionated manganese deposits differ from the Cyprus umbers that are not fractionated and are interpreted as fallout from distal smokers, rather than direct precipitates from low-temperature vents (Boyle, 1990). Many of the largest manganese deposits in the geologic record occur apparently independently of volcanism (Guilbert and Park, 1985). Low-temperature mineral facies proximal to high-temperature sulfides, like those at the TAG field, are vulnerable to dissolution but, if preserved in the record, may be useful indicators of proximity to massive sulfides.

6. Seafloor hydrothermal minerals and microbes

The TAG active sulfide mound, as well as other actively accumulating hydrothermal deposits, host ecosystems of a growing global inventory of some 712 species of macrofauna (visible to naked eye; 71% of species known exclusively from vents) including 185 families (14 of these families found at vents only) and 12 phyla that exhibit geographical diversity along the ocean ridge system and between oceans (Tunnicliffe 1991, 1992; Tunnicliffe and Fowler, 1996; Tunnicliffe et al., 1998; Van Dover, 2000; Wolff, 2005; Gjerde, 2006; Deybruyeres et al., 2006). The species diversity at seafloor hydrothermal vents is low relative to the species diversity in the deep ocean exclusive of hydrothermal vents (estimated number of species in excess of 10 million exclusive of microbes, comparable to that in rain forests; Grassle and Maciolek, 1992; May, 1994; Van Dover, 2000). However, the lower number of species present at hydrothermal vents is represented by the high diversity of major animal groups (taxonomic phyla) and high biomass (Van Dover, 2000). Endemism at hydrothermal vents far exceeds that found elsewhere in the marine environment.

Chemosynthetic microbes are primary producers at the base of the food chain that supports the macrofauna and are hosted in the massive sulfides and volcanic substrate of sub-seafloor hydrothermal systems (Deming and Baross, 1993; Kelley et al., 2002; Juniper, 2004; Sogin et al., 2006). The chemosynthetic aerobic and anaerobic microbes utilize redox reactions with carbon, hydrogen, sulfur and metals in the ore-forming solutions (primarily oxidation of H_2S) as a source of chemical energy utilizing CO_2 dissolved in seawater and water to manufacture carbohydrates for nourishment (Jannasch, 1997). The microbes are inferred to be part of a sub-seafloor biosphere involving the hydrothermal ore-forming systems as the source of chemical energy (Deming and Baross, 1993). Microbial activity may continue with aging of a VMS deposit. Submarine volcanic glass exhibits alteration attributed to microbial activity in the upper 300 m of ocean crust that has been found in nearly all ocean basins and in certain ophiolites and greenstone belts back to 3.5 thousand million years ago (Staudigel et al., 2006).

Studies of the role of the microbes in hydrothermal mineralization are at an early stage (e.g., Zierenberg and Schiffman, 1990; Zierenberg et al., 1995; Edwards et al., 2005; Southam and Saunders, 2005). The identification of structures interpreted as vent worms and other macrobiota in certain ancient massive sulfide bodies indicates that microbial interactions have been associated with hydrothermal mineralization through geologic time (Banks, 1985; McGoldrick, 1999; Rasmussen, 2000; Fallick et al., 2001; Little and Vrijenhoek, 2003; Boyce et al., 2003).

The microbes are a living resource of value to science and to industry (Fenecal, 1993, 1996; Jannasch, 1995; Cary et al., 2004). Analysis of their genomes indicates that certain of the heat-loving microbes (thermophiles; domains Archaea and Bacteria) exhibit genetic characteristics that place them at the base of the tree of life and that their study may elucidate the origin of life (Russell et al., 2005). Enzymes found in certain of these microbes in terrestrial and submarine high-temperature hydrothermal systems are already employed to replicate

DNA for forensic and other genetic "fingerprinting" purposes (polymerase chain reaction). Emerging applications include high-temperature–pressure industrial processes such as enhancing flow in deep oil wells, and use in detergents and food preservatives. Bioactive compounds produced by certain of the microbes are being tested as pharmaceuticals with potential for treatment of cancer and other diseases. The microbes themselves are bioreactors with potential applications such as enhancing the refining of metallic mineral ores. The fact that certain of these microbes are being utilized commercially before the seafloor hydrothermal mineral deposits that host them is cited as, "the deepest of ironies" (Glowka, 1996).

Scientific, legal, and policy aspects of bioprospecting for genetic resources at hydrothermal vents and other settings in the international Area of the ocean beyond the limits of national jurisdiction are presently outside the scope of the UNCLOS (Arico and Salpin, 2005; UNU-IAS, 2005; UNEP/GRID-Arendal, 2006). Deliberations are underway to extend UNCLOS to encompass biological resources including conservation and sustainable use of genetic resources of the international Area, possibly through the United Nations Convention on Biological Diversity that entered into force in 1993 (Glowka et al., 1994). Research protocols are being instigated to sustain living and non-living resources at all seafloor hydrothermal sites (International Marine Minerals Society, 2001a,b; InterRidge, 2006; Devey et al., 2007). Marine Protected Areas are being initiated where hydrothermal sites are present within the EEZ of coastal nations (e.g., Mullineaux et al., 1998; Mann Borgese, 1998; SOPAC, 1999; Dando and Juniper, 2001; Santos et al., 2003; Conley, 2006).

7. Magmatic deposits

The metallic mineral potential of ultramafic rocks of the upper mantle in ocean basins is largely unknown because of limited exposures and sampling (Fig. 15). Mantle exposure occurs principally in regions of low magma budget (volume of magma per unit of plate separation) and/or heterogeneous magmatic accretion across axial valleys and detachment faulting on slow-spreading ocean ridges, as between 14°N and 16°N in the axial valley of the Mid-Atlantic Ridge (Rona et al., 1987, 1992; Cannat et al., 1992; Bougault et al., 1993; Cannat, 1993; Cannat et al., 1997). Tectonic windows to mantle are less common on intermediate- to fast-spreading ocean ridges where the magma budget appears to be consistently higher and magmatic construction dominates (Karson, 1998). The types of deposits hosted in gabbros of oceanic lower crust and upper mantle peridotites (usually serpentinized) are known from terrestrial exposures, as in the Oman and Troodos ophiolites (Nicolas, 1995). These types of deposits include chromite and nickel- and platinum-group element (PGE)-rich sulfide mineral phases of magmatic origin (Naldrett, 1989, 2004). According to Arndt et al. (2005), "magmatic Fe–Ni–Cu+/−PGE sulfide deposits form where mantle-derived mafic and ultramafic magmas become saturated in sulfide and segregate immiscible sulfide liquid, commonly following interaction with crustal rocks." Ni–Cu sulfide deposits generally occur at the base of mafic and ultramafic bodies (Barnes and Lightfoot, 2005). PGE, chromium and vanadium deposits on land generally occur in mafic intrusions (Cawthorn et al., 2005) According to Zhou and Robinson (1997) concentration of podiform chromites is favored beneath volcanic island arcs where refractory melts react with thick sections of old lithospheric mantle.

Magmatic deposits may exhibit clustering on the same scale as that observed in massive sulfides in ocean crust. For example, magmatic chromite and sulfide deposits of the Troodos ophiolite (e.g., Kokkinorosos, Kannoures, and Hippavlou chromite mines) lie within a 5 km by 5 km area in the Mount Olympus peridotites (Geological Survey Department, 1982). The apparently analogous clustering of hydrothermal and magmatic ore deposits suggests that a spatial and temporal relation may exist between certain magmatic ore-forming systems in the upper mantle and hydrothermal ore-forming systems in the overlying ocean crust that may be driven by latent heat of crystallization of the underlying magma bodies. In addition, magmatic fluids may contribute metals to the VMS deposits. For example, the native nickel particles noted in sediments of the TAG active sulfide mound (Dekov, 2006) may indicate the presence of magmatic nickel-sulfide deposits in the mantle and lower crust underlying the TAG hydrothermal field. In this way, some VMS and magmatic deposits may be cogenetic. According to Naldrett (1989), nickel-sulfide deposits in extensive Archean greenstone belts like the Abitibi belt of the Superior Province of the Canadian Shield, are "an enigma in terms of present day plate tectonics", with reference to oceanic or continental settings.

8. Marine mineral deposits from combined terrestrial and deep ocean sources

8.1. Manganese nodules

Manganese nodules have been the signature marine mineral since they were first recovered from the deep seafloor by the *HMS Challenger* Expedition of 1872–

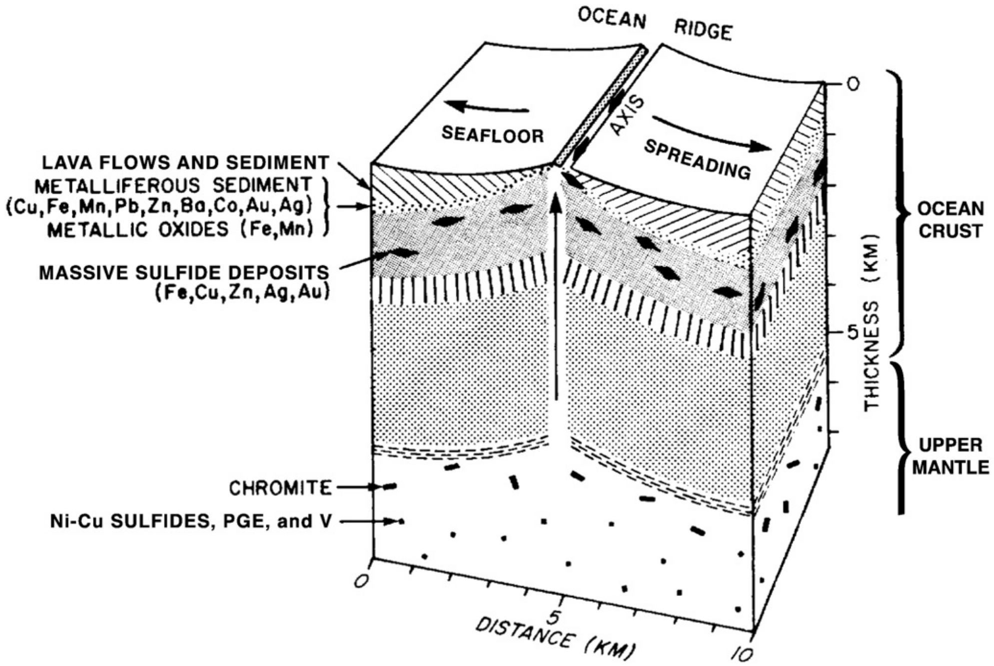

Fig. 15. Schematic block diagram showing potential occurrence of metallic mineral deposits formed at a spreading axis in ocean crust (shaded) and the underlying upper mantle (Bonatti, 1981; Rona, 1983b). The deposits are covered by lava flows, deformed by faulting, and transported away from the axis by ongoing seafloor spreading. Certain VMS deposits (ocean crust) may form together with underlying magmatic deposits (mantle and lower crust), which supply the heat to drive hydrothermal activity and contribute metals, as discussed in text.

76 (Murray and Renard, 1891). Mero (1965) brought attention to the potential economic value of manganese nodules when he estimated the *in situ* value of the suite of metals contained in the nodules without accounting for the substantial costs of recovery and refining. His estimates created a “gold rush” mentality at that time, which contributed to drive the development of the UN Convention on the Law of the Sea (UNCLOS) to assure sharing of the anticipated wealth between developed and developing nations. The golf- to tennis-ball sized nodules lie on and in sediment on vast expanses of the abyssal plains that cover some 70% of the ocean basin at a typical water depth of about 5 km (Fig. 1). Todorokite is the dominant mineral phase present in the nodules. The portion of the nodules that protrudes above the surface of the seafloor sediment is slowly precipitated hydrogenetically from metals dissolved in seawater, while the underside accumulates diagenetically from metals dissolved in the pore water of the sediment over millions of years (Cronan, 1980; Halbach et al., 1988). Formation rates vary depending mostly on the flux of Mn(II) to the deposit site with typical radial growth rates in the deep ocean between 2 and 10 mm/million years (Morgan, 2000). Calculation of geochemical mass balances indicates that fluxes of Mn from river input and from seafloor hydrothermal discharge are approximately equal (Edmond et al., 1979), and that the flux from each source is sufficient to account for the Mn in nodules (combined with lesser amounts of other metals). The resulting mixture of Cu, Ni, Mn, Co, and Fe in the nodules varies in different regions of the seafloor related to proximity to sources including rivers, hydrothermal vents, and plankton that take up the metals in the photic zone (upper ~300 m of the water column). The metals are released where the tests and fine-grained inorganic materials incorporated into the fecal matter of the plankton accumulate in seafloor sediment, enhanced by high surface productivity in equatorial zones (Table 8; Seibold, 1978; Cronan, 1980). The metals are reduced through consumption by benthic biota, and adsorbed on the manganese oxide surfaces of the nodules (Verlaan et al., 2004). The total amounts of Ni, Co, and Mn contained in manganese nodules globally are considered to exceed terrestrial reserves, while terrestrial reserves of Cu are considered to exceed those in the nodules (Thiel et al., 1997). The nodules also contain trace quantities of almost all the metallic elements in seawater (Au, Ag, Pt, Ti, Mo, Zn, etc.). Heterotrophic bacteria capable of mobilizing and immobilizing manganese have been identified on manganese nodules (Chandramohan et al., 1987). A number of studies have been made of the environmental

impact of mining the nodules (e.g., Burns, 1980; Sharma, 2005). The vast two-dimensional distribution of manganese nodules on abyssal sediment has yet to be recognized in the geologic record.

Between 1974 and 1982, several consortia of private companies and government organizations spent at least \$US 1 billion in failed ventures to mine manganese nodules. The failures are attributed to inflated evaluations of the potential resource, high costs of metallurgical extraction, political interference, and drops in metal prices (Glasby, 2000, 2002; Scott et al., 2006). In 2001, the ISA granted to seven national and industrial groups of "pioneer investors" (P.R. China, Japan, Korea, France, Interoceanmetal Joint Organization, Russian Federation, Germany) exclusive 15-year exploration contracts for tracts of the region of the most prospective manganese nodules (combined Cu, Ni, Co content $>2.5\%$ by weight, abundance >10 kg m^{-2}) between the Clarion and Clipperton fracture zones in the eastern equatorial Pacific Ocean (Fig. 16; Morgan, 2000; International Seabed Authority, 2001b). This grade is similar to that of terrestrial Ni–Cu sulfide ore such as Sudbury, Ontario, Canada (Exon et al., 1992; Scott et al., 2006). Preliminary assessment suggests that at least 34 thousand million tonnes of these deposits occur within the Clarion-Clipperton zone, containing 7.5 thousand million tonnes of manganese, 340 million tonnes of nickel, 265 million tonnes of copper and 78 million tonnes of cobalt (Table 6.3 in Morgan, 2000). A similar procedure by the ISA granted India 1,500,000 km^2 in the Central Indian Ocean (Fig. 16; Jauhari and Pattan, 2000).

Assuming a nodule recovery efficiency of 70% with the first generation of mining technology, the area covered would be ~ 1 km^2/day, or 6000 km^2 over the 20-year life of a mine site (Thiel, 1991; Thiel et al., 1997). As a potential resource, manganese nodules contain about 10% of known land reserves of copper. Manganese constitutes 25 to30% of the higher grade nodules and is an essential element in steel making and other industrial uses. The key constituent for potential commercial extraction from the nodules is nickel, which occurs in concentrations between 1.2 and 1.45 wt.% in the Clarion-Clipperton zone (CCZ) deposits (Scott et al., 2006). Krigue-block calculations estimate 340×10^6 tonnes of nickel in the CCZ nodules that could potentially provide a significant proportion of world land production estimated at 900,000 tonnes/year in 1998 (Table 6.3 and reference in Morgan, 2000). Environmental guidelines for prospecting and exploration of nodules have been developed by the ISA (International Seabed Authority, 1999), although mining awaits favorable market conditions vis-à-vis sources on land like the Voisey's Bay Ni–Cu–Co deposit in Labrador, Canada (Naldrett, 2000).

8.2. Cobalt-rich ferromanganese crusts

Like manganese nodules, cobalt-rich ferromanganese crusts of the deep ocean precipitate hydrogenetically over millions of years from metals dissolved in cold ambient seawater derived from terrestrial dissolved input and seafloor hydrothermal discharge (Hein and Morgan, 1999; Hein et al., 2000; Hein, 2004). The crusts grow at rates of <1 mm to about 10 mm/million years. The slow growth rates allow for adsorption of large quantities of trace metals by the oxyhydroxides at the crust surface (Hein et al., 2000). The crusts are nearly ubiquitous on hard-rock substrates that have been swept clean of sediments over millions of years by ocean currents on seamounts, ocean ridges, plateaus and abyssal hills. The most favorable setting for the occurrence of these crusts lies in the region of the volcanic island arcs of the central equatorial Pacific and on volcanic seamounts in the equatorial Indian Ocean where extensive volcanic substrates exist (Figs. 1 and 11; Banakar et al., 2000; Hein and Morgan, 1999; Hein et al., 2005).

The metals in the crusts comprise cobalt, nickel, platinum, and titanium in addition to iron and manganese depending on proximity to different sources (Table 9). These metals accumulate to thicknesses up to about 25 cm in crusts on the hard-rock substrates between ocean depths of 400 and 4000 m, encompassing the oxygen minimum zone depleted in oxygen by decay of sinking planktonic organisms. The mean dry bulk density of the crusts is 1.3 g cm^{-3}, mean porosity is 60%, and the high mean surface area (300 m^2 g^{-1}) enhances adsorption. The dominant crystalline phase of

Table 8
Average concentration of metals in manganese nodules from different oceans (Ghosh and Mukhopadhyay, 2000)

Element	Atlantic	Pacific	Indian	World oceans
Manganese (wt.%)	13.25	20.10	15.25	18.60
Iron	16.97	11.40	14.23	12.40
Nickel	0.32	0.76	0.43	0.66
Copper	0.13	0.54	0.25	0.45
Cobalt	0.27	0.27	0.21	0.27
Zinc	0.12	0.16	0.15	0.12
Lead	0.14	0.08	0.10	0.09
Iridium (ppm)	9.32	6.64	3.48	–
Uranium (ppm)	7.4	7.68	6.20	–
Palladium (ppm)	5.11	72	8.76	–
Thorium (ppm)	55.00	32.06	40.75	–
Gold (ppb)	14.82	3.27	3.59	–

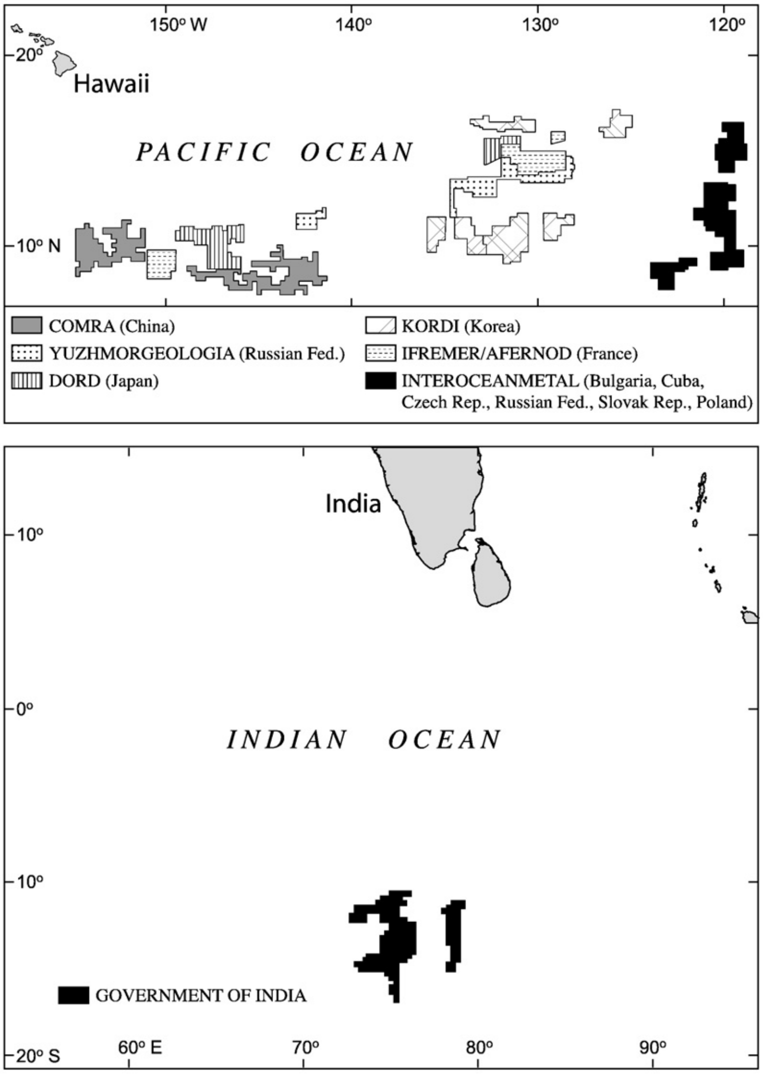

Fig. 16. Maps showing manganese nodule contract areas in the Clarion-Clipperton zone of the northeastern equatorial Pacific Ocean (top) and of the Indian Ocean (bottom; courtesy of the International Seabed Authority).

the crusts is ferruginous vernadite (Fe-rich δ-MnO_2) and X-ray amorphous iron oxyhydroxide, with incorporated detrital minerals such as quartz, feldspar, pyroxene, phillipsite, and authigenic carbonate fluorapatite. Most thick crusts that formed in the open Pacific consist of a phosphatized older growth generation underlying a younger non-phosphatized generation (Hein et al., 2000). Cobalt in the crusts (range 0.5 to 2 wt.%; average 0.7%) is enriched by a factor of two to five times its typical concentration in manganese nodules and is the primary value metal in the crusts, although minor amounts of V, Mo, Pt, Ti, Ce and other rare-earth elements, Zr, Ni, Te, Tl and Cu could be recovered as byproducts (Hein et al., 1986; Halbach et al., 1988). Cobalt, Ni and Pt contents are highest in crusts from the central Pacific and lowest in crusts from spreading centers in the southeast Pacific, the continental margins, and volcanic island arcs of the western Pacific (Hein et al., 2000). Cobalt contents in Atlantic and Indian Ocean crusts are roughly equivalent to those in crusts from the western margin of North America. An estimated 25% of the annual global need for cobalt could potentially be produced from a single mine site on a volcanic seamount (Mangini et al., 1987; Hein et al., 2000).

A technique is needed to systematically map the thickness of the crusts over large areas. Recovery of the crusts in terms of removal and refinement is more challenging than that of manganese nodules that lie loose

Table 9
Range of mean concentration of metals in cobalt-rich ferromanganese crusts from the Pacific, Atlantic, and Indian oceans (Hein et al., 2000)

Element	Range
Iron (wt.%)	15.1–22.9
Manganese	13.5–26.3
Nickel (ppm)	3255–5716
Copper	713–1075
Cobalt	3006–7888
Zinc	512–864
Barium	1494–4085
Molybdenum	334–569
Strontium	1066–1848
Cerium	696–1684

on sediment, or massive sulfides, that form compact mounds, because the thin crusts are difficult to separate from the large areas of hard-rock substrate to which they adhere both physically and in terms of refining (Yamazaki, 2006b). Like manganese nodules preliminary environmental impact and engineering studies of ferromanganese crusts have been undertaken, and potential mining awaits additional studies and favorable market conditions. Analogs of cobalt-rich ferromanganese crusts have yet to be recognized in the geologic record.

9. Discussion

9.1. Marine minerals and plate tectonics

Processes at plate boundaries have been effectively applied as a global-scale framework to classify certain types of ancient mineral deposits presently known from their occurrences on land (Table 2; Mitchell and Garson, 1981; Sawkins, 1990). As Skinner (1997) points out, numerous papers have recognized the increased frequency of mineralizing events related to tectonic changes in the lithosphere through geologic time (e.g., Damon and Mauger, 1966; Song, 1983; Meyer, 1988; Titley, 1993; Groves et al., 2005) and at peak times in the Wilson cycle of supercontinent aggregation (Barley and Groves, 1992; Kerrich, 1992). However, the challenge remains to bridge the gap in order to elucidate how the settings relate to the geologic controls of individual deposits on regional and local scales (Skinner, 2004). A step in this direction is to relate mineral occurrences to reorganization of plate motion that occurs on a global scale at intervals of tens to hundreds of millions of years, and is ongoing on a local scale.

A particularly promising time to search for relations between plate tectonics and mineralization is a global plate reorganization that occurred during the Eocene epoch (55 to 34 million years ago). The preservation of the record of plate motions in the magnetic polarity reversal sequence in oceanic crust of Eocene age make this the most complete documentation of a global plate reorganization in geologic history (Rona and Richardson, 1978; Rona, 1980b). The Eocene plate reorganization involved an initial seven-fold increase in length of collisional convergent plate boundaries oriented east–west to accommodate north–south convergence between the Eurasian, African and Indian plates primarily in the region of the former Tethys Sea (Table 10). The plate motion then reoriented to predominantly east–west accommodated by enhanced subduction in zones around the Pacific. Resistance to south–north convergence by the increase in length of collisional convergent plate boundaries apparently fed back to the mantle convection cells driving the plates, forcing the cells to reorient to the east–west direction of least resistance (Rona and Richardson, 1978; Anderson, 2002). In other words, mantle convection cells drive the plates, but stress fields produced by interaction of the lithospheric plates are inferred to orient the cells in a self-organizing system. Emerging relations between mineralization and plate tectonics during the Eocene plate reorganization include the following:

(1) The erosion associated with the Alpine–Himalayan mountain building driven by the collisional plate boundaries along the former Tethys Sea increased both the particulate and dissolved terrestrial input to the ocean, with implications for formation of fluviatile and marine placer deposits and for contribution of dissolved material to form precipitates (phosphorites, manganese nodules, cobalt–ferromanganese crusts).
(2) More VMS deposits formed (1.0/million years) during the Eocene epoch than at any other time except the Miocene epoch (1.4/million years) and the Ordovician period (1.4/million years; the time of an earlier global plate reorganization) based on analysis (Figs. 7 and 8 in Rona, 1988) of a compilation of information on known Archean,

Table 10
Eocene global plate reorganization (55 to 34 million years ago; Rona and Richardson, 1978)

Worldwide total length (km)	Before	After	Net change
Convergent plate boundaries (subduction and collision)	52,000	74,500	+22,500
Convergent plate boundaries (collision)	2500	19,000	+16,500
All divergent plate boundaries	50,000	50,000	0

Proterozoic, and Phanerozoic VMS deposits hosted in volcanic rocks presently on land (508 deposits; Mosier et al., 1983; age-frequency distribution of deposits uncorrected for exhumation; Kesler and Wilkinson, 2006). Paleomagnetic, geochemical, and tectonostratigraphic evidence indicate that plate tectonics has been active at least since 3.8 thousand million years ago (Kusky, 2004; Cawood et al., 2006; Furnes et al., 2007). Massive sulfide deposits hosted in rhyolitic rocks inferred to have formed in volcanic island arc- and rift-related tectonic settings (56% of 508 deposits) predominate over basalt- and sediment-hosted massive sulfides inferred to have formed in mafic rocks of ocean ridges (<26% of 508 deposits).

(3) The global incidence of VMS deposits increased in volcanic island arc systems of the western Pacific related to the continuation and lengthening of subduction zones during the Eocene plate reorganization (Mitchell and Bell, 1973; Sawkins, 1990; Hall, 2002). Ophiolites were emplaced at sites along collisional convergent plate boundaries in the Alpine–Himalayan belt (Ding et al., 2005) and other localities (Dilek and Robinson, 2003), increasing the prospects for VMS and magmatic deposits of the ocean crust and upper mantle in the ophiolites.

(4) Porphyry copper–molybdenum–gold deposits increased in incidence along the western margin of North America (Sillitoe, 1976) and belts of these deposits formed along the western margin of South America (Sillitoe and Perello, 2005; Fig. 2) correlate with the change to a large east–west component of Eocene plate motion and enhanced subduction around the Pacific. Porphyry deposits formed in the western North Pacific and South Pacific (Garwin et al., 2005), eastern and western Caribbean, and Mediterranean regions (Sawkins, 1990), also correlate with the increase in convergent plate motion and subduction in the Eocene.

(5) A global change in stress field from north–south compression to east–west extension associated with the Eocene reorientation of plate motion had regional to local effects on mineralization. For example, Seedorff and Barton (2004) report that Carlin-type Au–Ag deposits in northeastern Nevada "formed between 42 and 36 million years ago, following a long period of contraction and crustal thickening of the miogeocline. The ages of deposits coincide with the initiation of extension in this region…". According to Cline et al. (2005), "North-northwest-trending paleo-normal faults and northeast-trending paleo-transform faults, preferentially dilated during Eocene extension, controlled the regional position, orientation, and alignment of the deposits". Dikes oriented north–south from an underlying plutonic complex intruded the north–south Carlin trend between ~40 and 36 million years ago consistent with east–west extension (Ressel and Henry, 2006). The igneous activity is inferred to have provided magmatic heat to drive mineralization. This change from north–south compression to east–west extension as part of the Eocene global plate reorganization may have opened pathways for magmatism and associated mineralization at terrestrial and marine sites in this and other regions of the world. The Eocene Carlin-type sediment-hosted disseminated Au deposits of Nevada (hundreds to thousands of tonnes Au) are the largest Au deposits known after those of the Archean Witwatersrand region of South Africa (Cline et al., 2005). As a product of a global plate reorganization, other regions of Carlin-type deposits may remain to be found guided by now largely known exploration criteria. For example, like the Carlin deposits the Muteh gold orebodies in westen Iran are inferred to have been emplaced along northwest-oriented normal faults (in exhumed metamorphic rocks) during or after late stages of Eocene (55.7–38.5 million years ago) east–west brittle extension and intrusive activity, in this case related to the closure of the Tethys ocean during late stages of the Zagros orogeny (Moritz et al., 2006).

10. Perspective and conclusions

Marine minerals are being selectively developed as resources with sand and gravel recovery economically leading that of placer diamond, tin, and gold (Figs. 1–7; Scott, 2001). Seafloor VMS deposits, manganese nodules, and cobalt–ferromanganese crusts are being evaluated as potential future resources under national (EEZ) and international (UNCLOS and ISA) regimes (Figs. 11, 13 and 16). The deep seabed has the potential to become a major source of metals including Ni, Cu, Co, Mn, Pb, Pt, Li, Au and Ag that are essential to developed and developing economies (Broadus, 1987; Herzig, 1999; Antrim, 2005, 2006). The value of precious metals (Au, Ag, Pt), as well as the supply of those metals that is dwindling due to use relative to their abundance in the lithosphere (Pt, Cu, Zn; Gordon et al., 2006) will selectively drive marine mining. The technologies for extraction of placers including diamonds, and of aggregates are operational. For phosphorites, manganese nodules, cobalt–ferromanganese crusts and massive sulfides (near spreading axes) the technology is within reach (Scott, 2001; Scott et al., 2006). Presently inaccessible massive sulfide deposits in older oceanic crust away from spreading axes in ocean basins and back-arc basins and as well as Ni–Cu–PGE magmatic sulfides in lower crust and upper mantle at and away from spreading axes will become accessible as appropriate exploration and exploitation methods are developed in response to scientific interest and economic need. Recovery of fresh water from seawater may become the ultimate marine mineral resource.

Skinner (1997) has posed questions of what we do and do not know about ancient hydrothermal mineral deposits. It is instructive to expand his questions to include all marine minerals including seafloor VMS deposits, to consider how what we are learning from investigation of various types of marine minerals may elucidate their ancient counterparts, as follows:

What we do know: Sources (solutions for hydrothermal deposits; other sources for other types of deposits), channel-ways or transport paths, and precipitation or deposition:

(1) For marine placer deposits, we can clearly follow fluviatile transport, and mechanical sorting between source and sink for diamond deposits from kimberlites in southern Africa to the southwest African continental margin; from alluvial gold to the shelf in northwestern North America; from granites to the continental shelf for cassiterite deposits of southeast Asia. We can predict prospective regions of placer deposits from the juxtaposition of mineralized zones and likely transport paths and depositional areas, as for the western margin of Central and South America; Figs. 2–7).
(2) For seafloor VMS deposits, we can sample and measure physical and chemical properties of active hydrothermal circulation systems and their host rocks, decipher tectonic and magmatic controls that concentrate the deposits at sites at plate boundaries, and apply this understanding to guide exploration for ancient deposits that have been emplaced on land. For example, insights gained from investigation of active seafloor VMS deposits have contributed to a recent surge of discoveries of ancient VMS deposits in P. R. China (Rona and Hou, 1999; Hou et al., 1999). VMS deposits in oceanic lithosphere that formed at seafloor spreading axes and have been transported away along flow lines of spreading remain an exploration challenge. Exploration criteria for large relict off-axis VMS deposits can be developed from characteristics of large deposits at spreading axes, such as magnetic signatures of crustal thinning associated with detachment faults and of hydrothermal alteration as deciphered at the TAG hydrothermal field. The magnetic low of hydrothermal alteration, first identified in mafic crust formed at ocean ridges (Rona, 1978b), also works in felsic crust formed at convergent plate boundaries, although the signal strength is lower (M. A. Tivey, pers. comm.).
(3) Magmatic deposits (chromite and Ni–Cu–PGE sulfides) may be next to be discovered in areas where low magma budgets and detachment faulting have resulted in exposure of lower crust and upper mantle on slow-spreading ocean ridges (Fig. 15); these deposits hosted in gabbros, peridotites and serpentinites may contribute to elucidation of the nature of vast greenstone belts and their deposits (e.g., Abitibi-type deposits; Naldrett, 1989, 2004).

What we do not know: Geological age and duration of an ore-forming system, the bounds and magnitudes and depth of ore-forming systems, the importance of source-rock enrichment, and the constraints of space and time. Contemporary marine mineral deposits have an advantage over ancient deposits in providing at least preliminary answers to these questions, as follows:

(1) Quaternary sea level fluctuations were particularly important for the concentration of present placer deposits (2 to 0 million years), although erosion of source rocks like the Cretaceous kimberlites of southern Africa and the gold-bearing rocks of northwestern North American may have occurred on much longer time scales.
(2) Manganese nodules and cobalt–ferromanganese crusts have accumulated hydrogenetically over millions of years from a combination of terrestrial riverine and seafloor hydrothermal input of metals, but apparently lack analogs in the geological record.
(3) Investigation of seafloor VMS deposits by surface and sub-surface (drilling and coring) methods is yielding answers to these questions. An example presented is the TAG hydrothermal field that comprises a cluster of large massive sulfide mounds occupying a 5 by 5 km area of the axial valley of the Mid-Atlantic Ridge. As noted, the sulfide mounds progressively range in age/stage from young-hot to old/cold. The mounds were formed episodically at radiometrically-determined intervals of high-temperature activity (tens to hundreds of years),

alternating with intervals of quiescence (thousands of years) for at least 100,000 years, with a hydrothermal event that was field-wide (50,000 years) and other events apparently limited to individual mounds. The primary heat source for high-temperature hydrothermal activity continues to be magmatic with subsidiary contribution from lower-temperature exothermic sources. The superposition of multiple episodes of high-temperature hydrothermal activity at intervals of thousands of years has built the TAG active sulfide mound. Zone refinement by circulation within the mound has concentrated the interesting metals in the upper few meters. The alignment and episodic timing of mineralization of the assemblage of mounds in the TAG field is related to changing circulation pathways between magmatic heat sources and the deposits apparently controlled by the development of a major detachment fault in the wall of the axial valley accompanied by ongoing seismicity that maintains circulation pathways. The size, grade, zone refinement and 3-D anatomy of the TAG active high-temperature sulfide mound and the spatial distribution of the assemblage of sulfide mounds in the TAG field are analogous to ancient clustered counterparts (Figs. 8, 9 and 14) and provide a key to understanding how such clusters form. Accordingly, the most prospective seafloor VMS deposits for marine mining will be those old enough to have grown and to have been refined through superposition of multiple high-temperature hydrothermal episodes. Large, high-grade mineral deposits of any type are anomalies that require special spatial and temporal conditions to concentrate. Marine mining is anticipated to begin selectively for VMS and other types of marine minerals when anomalously large volume high-grade deposits can be mined at a profit. Exploration to find anomalously large deposits and drilling to determine grade and tonnage are technically challenging and costly at sea. Large, high-grade VMS deposits are not renewable resources because they take thousands of years to form.

(4) The theory of plate tectonics early provided a framework for mineralization on the scale of global plate boundaries, and its application is gradually converging to elucidate geologic controls of mineralization from regional to local scales of plate reorganization with the potential to eventually elucidate controls on the scale of individual deposits. Investigation of the spectrum of marine minerals as active analogs of types of ancient mineral deposits is contributing to this convergence.

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