Morocco, and Oligocene ironstones in Kazakhstan. In hope of encouraging this effort we have now recorded the Phanerozoic ooidal ironstones around the globe, from eastern Australia, Asia, Europe, Africa, North America, to southern South America (Table 1), in terms of brief stratigraphic occurrence and sedimentary petrology. Most of these 58 countries include ironstones which had, or have, or probably will have economic iron ore (Table 2).

First we define the ooidal ironstones and speculate about their origin of accumulation. Then we collect the Phanerozoic ooidal ironstones around the world. In most instances the terminologies of Young (1989a) is followed, and the stratigraphic names and correlation should be checked with most recent publications.

Definition

Stratigraphic record

Banded ironstones were common during Precambrian time, but they were nearly absent in the Phanerozoic Eon. In contrast, only a few ooidal ironstones (for example, Transvaal in South Africa, The Roper Bar and Constance Range in northern Australia, the Sinian System in Northern China, the Gunflint in the Mesabi Range in North America, and the Sokoman in the Labrador trough; Taylor, 1969, p. 175) occurred during the earlier Proterozoic Eon, and none apparently developed in the later Proterozoic or early Cambrian ages – about 1000 to 570 Ma. Then more than 400 ooidal ironstones accumulated throughout most of the Phanerozoic Eon, from Middle Cambrian to Recent (Figure 1; Table 3). An increase and decrease in the number of ironstones through time resemble the warm and cool modes (Frakes et al., 1992, Fig. 11.1) and the higher and lower level of atmospheric CO_2 (Berner, 1990, Fig. 5) in the Phanerozoic Eon. These, in turn, are related to the varied amount of heat given off by the planet.

Phanerozoic ironstones were present in both low and high latitudes. They were especially common in the Ordovician and Devonian (45 degrees north to 65 degrees south of the paleoequator) and again in Jurassic and Cretaceous (10 degrees south to 70 degrees north to the paleoequator) times, when the greenhouse stage (Fischer, 1981) was marked by a high relative rate of marine sedimentation on the continents (Ronov and others, 1980, Fig. 4) and generally mild climate. In contrast, only a few occurred in the Cambrian, Carboniferous, Permian, Triassic, or in the late Cenozoic. Yet one of the richest ooidal ironstones developed in the Middle Pliocene sequence of Crimea in southern Ukraine.

Although most of the ironstones accumulated in warm climate some were deposited in a cooler mode, as in the Late Ordovician time of northern (marginal) Gondwana. Two ironstones also accumulated in cool temperate, humid climate in Late Permian time in eastern Gondwana. Here, however, the setting was a progressive warming from Early Permian glacial phase to latest Permian warm temperate to tropical conditions (Fawcett et al., 1994, p. 150–151).

Table 3. Stratigraphic record of Phanerozoic ooidal ironstones (numbers in parentheses are the numbers of deposits in the directory)

CAMBRIAN Canada (47), Morocco (187), United States (316–324)
ORDOVICIAN Algeria (1–6), Australia (29), Bolivia (40), Canada (48–52), China (63–65), Czech Republic (75–86), Estonia (93), France (94–98), Germany (110–111), Libya (164–166), Malaysia? (180), Morocco (188–204), Norway (219), Poland (225), Portugal (233–234), Russia – European (236) and Asian part (241), Spain (257–260), Sweden (269–272), Tunisia (278–280), Turkey (283), United Kingdom (289–292), United States (325–335)
SILURIAN Algeria (7–8), Argentina (25–27), Bolivia (41), Brazil (42), Canada (53–54), Guinea (144), Italy (157), Libya (167–168), Mauritania (185), Morocco (205), Spain (261), Turkey (284–285), United Kingdom (293–294), United States (336–341)
DEVONIAN Algeria (9–22), Argentina (28), Belgium (33–38), Brazil (43), Canada (55), China (66–67), Czech Republic (87–88), France (99–100), Guinea (145), Kazakhstan (158), Libya (169–172), Macedonia (177), Mali (183), Mauritania (186), Morocco (206–209), Nepal (211), Poland (226–227), Russia – European part (237) and Asian part (242), Saudi Arabia (249), Spain (262–263), Turkey (286), United States (342–348)
CARBONIFEROUS Czech Republic (89), Ireland (152), Libya (173–174), Morocco (210), United Kingdom (295–296), United States (349)
PERMIAN Australia (30), India (147)
TRIASSIC Canada (56), India (148–149), Nepal (212), Saudi Arabia (250), Slovakia (256), United States (350)
JURASSIC Algeria (23), Australia (31), Belgium (39), Bulgaria (44–46), Canada (57–59), China (68–69), France (101–108), Germany (112–138), Hungary (146), India (150), Iraq (151), Luxembourg (176), Malgash Republic (182), Nepal (213), Norway (220), Pakistan (222), Poland (228–231), Russia – European (238–239) and Asian part (243), Saudi Arabia (251), Slovakia (256), Spain (264–265), Sweden (273), Switzerland (274), Tunisia (281), United Kingdom (297–309), Yugoslavia (363–364)
CRETACEOUS Canada (60–62), Colombia (70–71), Egypt (91–92), France (109), Germany (139–142), Israel (153–156), Kazakhstan (159), Lebanon (163), Libya (175), Macedonia (178), Nigeria (217), Oman (221), Poland (232), Russia – European (240) and Asian part (244–247), Saudi Arabia (252–254), Sudan (266–268), Syria (276–277), United Kingdom (310–315), United States (351–354), Venezuela (357–358), Yugoslavia (365–366)
CENOZOIC Algeria (24), Australia (32), Colombia (72–74), Denmark (90), Germany (143), Kazakhstan (160–162), Malaysia (181), Mali (184), Niger (214–216), Nigeria (218), Pakistan (223), Philippines (224), Romania (235), Russia – Asian part (247–248), Saudi Arabia (255), Switzerland (275), Tunisia (282), Ukraine (287–288), United States (355–356), Venezuela (359–362)
RECENT Chad, Indonesia, Malawi, Venezuela

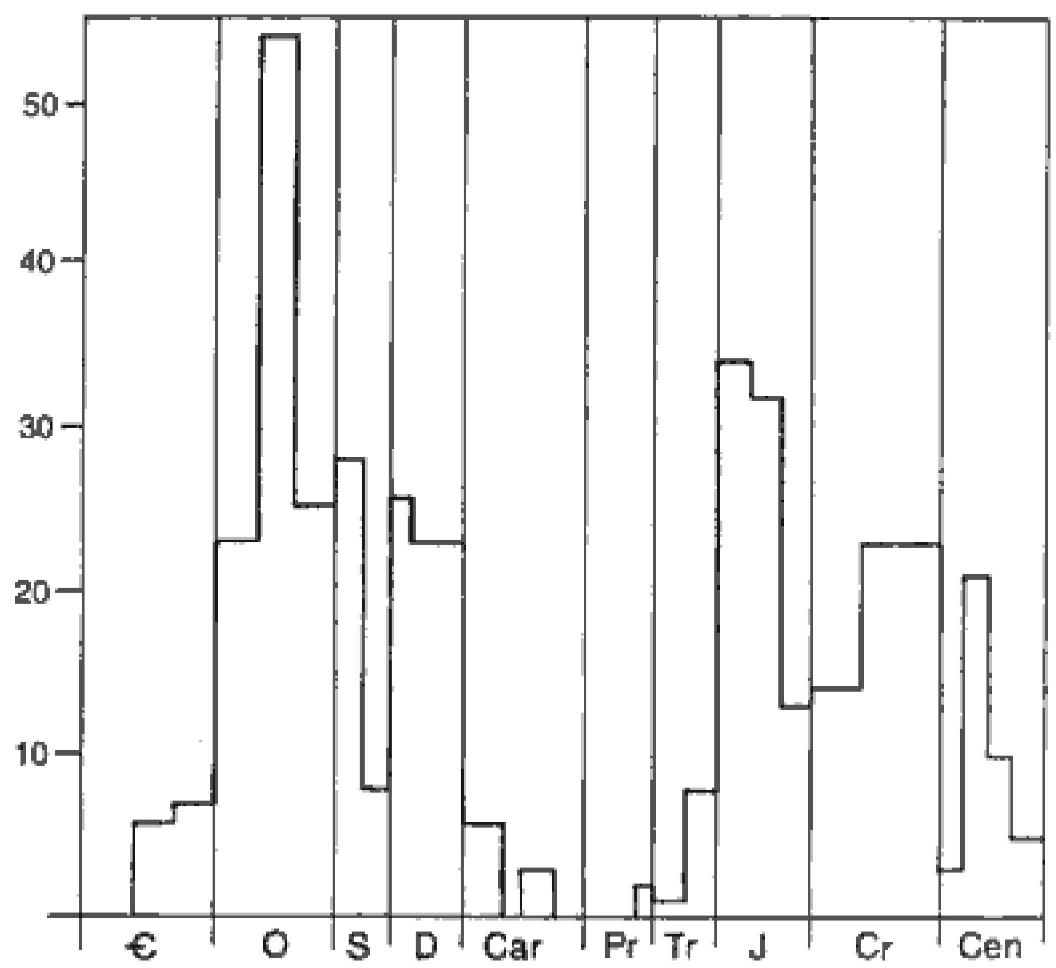

Figure 1. Stratigraphic distribution of Phanerozoic ooidal ironstones, in localities for early, middle, and late subdivisions of each period.

Tectonic setting

Phanerozoic ooidal ironstones accumulated mainly on large lithospheric plates in three different kinds of cratonic basins in an anorogenic or diminished orogenic episode. In addition, some were formed on small drifting and later docked microplates, like the Ordovician ironstones of Newfoundland. Absent from the directory are ironstones that never were deposited in Antarctica, southern India, between southern China and Siberia, and in southern Africa.

1) Many developed in anorogenic basins normally dominated by prolonged stability, as the Ordovician ironstones of central United States, and sometimes in a complex setting of extensional tectonics like the middle and late Mesozoic ironstones of Europe (Sellwood and Jenkyns, 1975).

2) Some developed along the cratonic margin at the time of divergence or initial convergence of lithospheric plates, as the Jurassic ironstones of the northern Indian plate. Deposits of this sort are difficult to reconstruct because many of them were later deformed by mountain building.

3) Other ironstones were accumulated on the inner side of a mobile belt at times of diminished deformation and curtailed detrital influx, as the Silurian ironstones of central-east United States. But none of these ironstones were developed along the Pacific orogenic rim (except one in northernmost Philippines), or within the interior of the major mountains like the Alpine, the Himalayan, and the Andean ranges.

Major patterns

Almost all of the ooidal ironstones are marine sedimentary rocks. But a few are nonmarine lacustrine or alluvial deposits. Most of the marine ironstones accumulated in interdeltaic or broadly embayed shallow seas. They may interfinger or replace sandy and shelly marine glauconitic sediments, or glauconitic peloids may occur as cores of berthierine ooids, as in some early Jurassic ironstones of England (Taylor and Curtis, 1995, p. 362).

They developed during relatively long periods of open circulation, reduced sedimentation input, abundant burrowers, and normal fauna and micro-organisms. In detail, the ironstones developed with reduction or interruption of sedimentation, commonly in the latest regression or in initial transgression following briefer times of shoaling or progradation. Some also occur in omission discontinuities (Burkhalter, 1995).

The ironstones are commonly associated with phosphates, coal measures, hard ground, or even evaporites. Laterites and bauxites may also be associated with ironstones, as in Middle and Late Creataceous of western Siberia (Nagorskiy, 1981), in Late Cretaceous of southeastern Egypt and Sudan (Schwarz and Germann, 1993), and in Middle and Late Oligocene of west-central Kazakhstan (Zitzmann, 1977c, p. 363–367). Moreover, some authorities believe that ooidal ironstones were derived from laterites and bauxites (Siehl and Thein, 1989; Bardossy, 1994, p. 285).

Most of the ironstones have no direct relation to volcanism. A few, however, have a remarkable synchronism with the volcanics of nearby areas, as for example in the Early Ordovician ironstones of Nova Scotia (Murphy and others 1980), the Middle Ordovician ironstones of Sweden (Sturesson, 1992) and of northern Spain (Garcia-Ramos and others, 1984), the Late Devonian ones of Belgium (Dreesen, 1989), and Late Cretaceous ironstones of southeastern Egypt (Van Houten and others, 1984, Fig. 4, 5).

Those marine ironstones that are thin, lean, and sporadic, and may have dwarfed faunas, commonly accumulated in prograding mudflats at the site of origin. Others are thicker and are concentrated in nearshore bars and sandwaves as condensed deposits, or carried offshore and collected in sheets across distal muddy facies (Bhattacharyya, 1989). In these offshore muddy deposits intraclasts and bioclasts are common.

In a characteristic shoaling-upward setting the ironstones lay at the top of the sandbed, whereas the mudstone lay below. One or several shoaling-upward sequences by small-scale regressive oscillations produced ironstones (Fig. 2). These ironstones are commonly condensed deposits, as in storm reworking (Young and others, 1991), while thicker related deeper sediments, including a black shale facies, have accumulated elsewhere. In the Phanerozoic Eon black shale and ooidal ironstones developed most readily during the Ordovician-Devonian and the Jurassic-Paleogene times characterized by relatively rapid sea-floor spreading and global high stand of sea level. The succession of some of the marine ooidal ironstones above the regression or at the beginning of transgression points to a link between sequence stratigraphy and the history of the basin. The eustatic control of the sea-level changes has been outlined in some cases (Van Houten and Arthur, 1989), as well as to Milankovitch patterns (Van Houten, 1986; Van Buchem and others, 1992, p. 999).

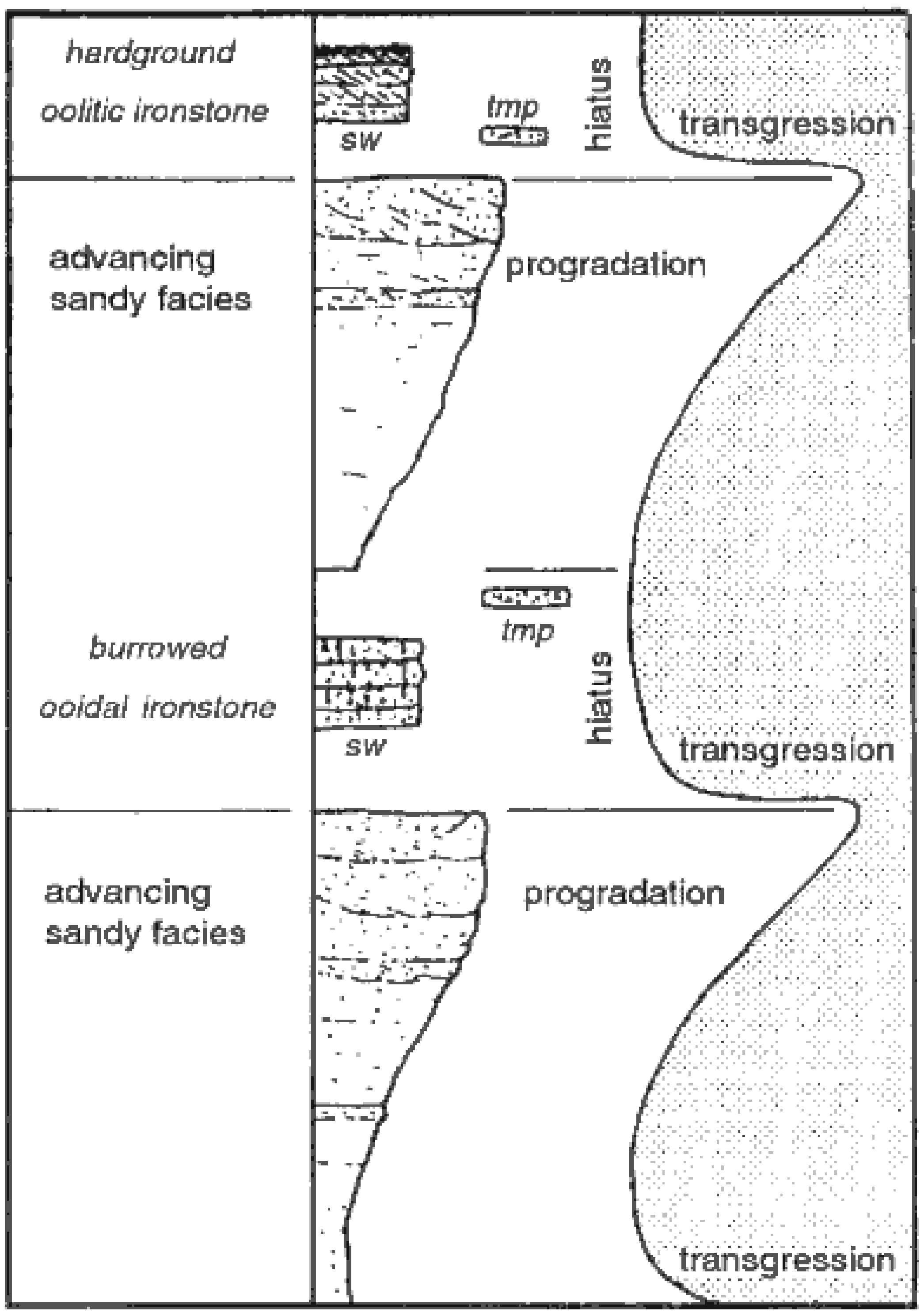

Figure 2. Facies model of shoaling-upward regressive siliciclastic sequence commonly a few to several tens of meters thick, with capping ooidal ironstones developed during reduction of sedimentation or hiatus. Reconstruction portrays initial transgression and development of ironstones before detrital influx. sw – sandwave. tmp – tempestite. (Van Houten and Arthur, 1989).

Sedimentary petrology

Ooids are multicoated granules less than 2 mm in diameter; pisoids are larger than 2 mm and they are generally rare. Ooids are commonly simple, but sometimes several to many ooids were wrapped together in a sheath to form multiple or composite ooids.

The sheaths of the ooids are either random or statistically tangentially arranged, and some sessile microbialites may build into the sheaths (Burkhalter, 1995, p. 62). Ooidal shape may be spherical, ellipsoidal (possibly an original shape), or disturbed to a spastolithic form. The disturbed ooids are common in siliciclastic sequences, and spastolithic ones are generally restricted to ooids that have a green clay berthierine or chamosite constituency. Some ooids were broken in transport, and locally these two parts were again sealed together.

The core or nucleus may be a grain, mostly quartz, a tiny clay particle, a minute shell fragment, a broken ooid, or a very small peloid. The matrix is largely quartz grains and lesser feldspar, clay minerals, including goethite or hematite. The cement is largely berthierine or chamosite, calcite, siderite.

The main clay minerals present in the ooids consist of some of the following: kaolinite, ferrous berthierine (7Å serpentine) or ferric odinite-endmember berthierine, ferrous chamosite (14A chlorite) or ferric chamosite (thuringite is discarded), and rarely nontronite (14A smectite), green smectite, illite, or green clay if it is unspecified. The berthierine or chamosite clays may also occur other than as ooids, for example in Coal Measures claystones, but these are not considered here.

Common types of assemblages in ooids include berthierine and chamosite (or green clay) plus kaolinite, green clay plus goethite, green clay plus goethite plus hematite, kaolinite plus goethite, kaolinite plus goethite plus hematite, goethite, goethite plus hematite, and hematite (Mücke, 1994). Berthierine and chamosite are essentially the same in chemical composition so that their presence relays on some other method of identification. Generally berthierine is very early diagenetic with organic activity, and it is unstable chemically and structurally relative to chamosite (Jahren and Aagaard, 1989, p. 169). As known now, clay mineral berthierine is present in many Mesozoic and Cenozoic ooids, and chamosite is the clay mineral in the Paleozoic ooids. But the persistence of both berthierine and chamosite in some Silurian ironstones (Schoen, 1964) and of chamosite in early Cenozoic (Paleocene) ironstone (Maynard, 1986) is evidence against a pervasive process of this sort. Goethite is also more important in Mesozoic and Cenozoic ironstones whereas hematite is more important in Paleozoic ones.

In some alluvial deposits, as in the early Cenozoic ooids of Western Australia, the ferric oxide in them may have come directly from laterites. In some other hematitic ooids, as in the Cambrian ones in western United States, they may have been originally glauconite.

Diagenetic and post-diagenetic assemblages

Normally the minerals in the ooidal ironstones are not deducible from the primary minerals. The present assemblages are diagenetic and post-diagenetic, and in careful work the interest must be both in the opaque and non-opaque iron minerals as well as clay minerals. The physical sedimentary environment is the main control style of early diagenesis in the ironstones (Taylor and Curtis, 1995). In earliest marine-formed ooids the minerals are probably iron oxide and/or berthierine, kaolinite, and locally apatite, with other minerals added later in diagenesis with continuing slow accumulation (Young, 1993, p. 464). In early diagenesis the ooidal minerals are replaced mainly by apatite, siderite, and amorphous silica. Some of the multicoated granules, as in several Carboniferous sequences, may actually be sideritic spherulites rather than ooids.

In later diagenetic or post-diagenetic effects the ooidal minerals may be replaced by kaolinite, goethite or hematite, siderite, Fe-calcite, ankerite, apatite, chamosite, illite, magnetite, maghemite, pyrite, or stilpnomelane. In many cases it is not clear whether goethite was primary or whether it replaced berthierine, but generally hematite is a late diagenetic alteration of goethite. Magnetite is produced by metamorphism of iron oxide or by generation of hydrocarbons (Young, 1993, p. 468). These results progressively destroy the concentric fabric and increase the proportion of iron oxide.

This mineralogy has to work from the present diagenetic or post-diagenetic results backward to an original deposit. Today identifying these ooids is advanced by the use of X-ray diffraction, transmission and scanning microscopy, Mössbauer infra-red spectra, backscattered electron imagery, and electron microprobe analyses. Clues to the problem are helped by analyses such as those by Hughes (1989), Kearsley (1989), and Mücke (1994). In the Late Devonian Wadi Shatti ooids of central Libya (Turk et al., 1980, Fig. 16–27) the diagenetic and post-diagenetic results are illustrated, including siderite, magnetite, and maghemite.

Recent ooidal ironstones

Most of the Recent ferruginous green or brown granules are peloids, some of which are faecal pellets, such as the ones from the shelf off Trinidad, Venezuela, and Guiana of South America, the one from the Loch Etive in western Scotland, the ones from the shelf off Senegal, Guinea, Nigeria, and Gabon of Africa, and the one from the shelf off Sarawak (Odin, ed., 1988), as well as peloids from interdistributary area of the Mahakam Delta in eastern Kalimantan, Indonesia (Allen et al., 1979). So far only a few Recent sediments have yielded iron-rich ooids. These are described briefly.

Ooids accumulated in some sediments in the southernmost Lake Malawi in Malawi (Muller and Forstner, 1973). Geothermal springs erupt along the lake. The ooids were encountered in a grab sample. The surrounding sediments are sand and gravel. The ooids contain amorphous hydrous feric oxide and opal, commonly with a superficial shell of nontronite. The ooids chemical composition (%): Fe_2O_3 50.5, FeO 0.21, SiO_2 20.4, Al_2O_3 3.2, CaO 0.84, MgO 0.31, MnO 0.31, P_2O_5 0.82, TiO_2 0.15.

Ooids accumulated in shallow, brackish open water areas north of the Chari delta in southern Lake Chad in western Chad (Lemoalle and Dupont, 1973). The deposits are surrounded by mud. The brown ooids contain goethite and nontronite (Pedro et al., 1978). The average chemical composition (%): Fe_2O_3 34.5–49.5, SiO_2 33.0–45.2, Al_2O_3 1.75–4.0, CaO 1.18–1.69, MgO 1.05–1.69, MnO 0.11–0.40, TiO_2 0.09–0.23.

Ferric-rich ooids and peloids are deposited in a very shallow sea along the coast of Cape Mala Pascua in northern Venezuela (Kimberley, 1994). The region lies in an eastward fault zone in the coastal range. The pale green to brown ooids are enclosed in greenish mud. The ooids contain ferric silicate odinite-endmember berthierine. Their chemical composition (%): FeO 6.89–7.55, Fe_2O_3 16.22–23.49, SiO_2 25.70–29.72, Al_2O_3 4.76–5.04, CaO 1.97–8.72, MgO 10.89–12.97, MnO 0.02, P_2O_5 0.17–0.26, TiO_2 0.08–0.10.

Further research on these Recent iron-rich ooids and their origin is of primary significance.

Some problems of origins of ooidal ironstones

Time and space

Ooidal ironstones and their equivalents occur in all Phanerozoic periods. Yet some intervals exist that are noted for their conspicuous ferruginous accumulations (Strakhov, 1947; Van Houten and Bhattacharyya, 1982). Very significantly increased deposition of ooidal ironstones falls into (1) Ordovician and the subsequent part of Silurian, (2) Devonian, and (3) Jurassic-Cretaceous. On the other hand, long intervals also exist that are lacking major accumulations of ooidal ironstones. This applies especially to Carboniferous, Permian, and Triassic. Why? The rivers were steadily bringing iron from land into the aqueous basins and oceans, and the atmosphere and hydrosphere did not differ substantially from the preceding or succeeding periods. Neither organic life that witnessed such a notable expansion on land in Carboniferous time could have hardly been a sufficient reason for such a radical restriction of ferruginous deposition. The acid waters of extensive swamps should have provided conditions for enhanced leaching of iron from regoliths and subsequent transportation of iron into the depositional realms, but the content of CO_2 in the Carboniferous atmosphere (Berner, 1990) was low.

Source of iron

The source of iron has usually been sought in the weathering processes upon land, especially in tropical climate and the subsequent introduction of iron by rivers into seas and oceans, mostly in the adsorbed form (adhering to clay and other particles) or as colloidal suspensions. In the present writers' opinion only a small portion of iron can be ascribed to hydrotherms issuing upon the ocean bottoms or to submarine weathering and leaching of basic lavas and their pyroclastics. Opposite view is held by Kimberley (1989) who assumes that iron formations are attributable to submarine exhalations of fluids and the origin of ooidal ironstones specifically is attributed by him to hypersaline fluids which have risen to marine bottom along deep faults. According to Stakhovitch (1986), the iron of European Mesozoic and Cenozoic ooidal ironstones had been brought by hydrothermal ore-bearing solutions rising during the active phase of continental rifting.

Some authors seek the source of iron in the sediments underlying the ooidal ironstones (for example Aldinger, 1957; Lipayeva and Pavlov, 1986). The interstitial solutions enclosed in the underlying sediments (for example in petroliferous basins; Pavlov, 1989) rose to the sediment/water interface where their iron became bound in the ironstone. According to Borchert (1964), iron had been leached out from sediments in a "CO_2-zone" situated between a near-surface oxygenated zone and the deeper lying H_2S-zone.

The conspicuous and frequent association of ooidal ironstones with black shales may be an indication of some