C

COLORS OF SEDIMENTARY ROCKS

The colors of sedimentary rocks have complex origins and in cases are secondary. However, colors are commonly primary and reflect important aspects of depositional environments including redox conditions and rates of deposition of organic matter. These conditions in turn affect fauna and thus a strong correlation exists between color and biofacies patterns, including macrofaunal distributions and burrowing type and depth (Leszcynski, 1993), particularly in deposits of oxygen-stratified basins (Myrow and Landing, 1992). Color may be useful for the interpretation of variations in such factors as relative sea level, oceanic circulation, sedimentation rate and primary productivity.

Colors are generally controlled by accessory minerals and compounds of iron and organic carbon (see reviews by Pettijohn, 1975; Potter *et al*., 1980; and Myrow, 1990). Compounds of other transition metals (e.g., Ti, Mn, Co, Cu, and Zn) also impart a pigment in some cases. For the most part the colors of sediment and sedimentary rock fall within two spectra: green-gray to red and olive-gray to black (Figure C70). The first color spectrum is controlled by the oxidation state of iron, specifically the Fe^{+3}/Fe^{+2} ratio in various minerals, and not the total iron content (Tomlinson, 1916; McBride, 1974). In particular, the color is a function of the mole fraction representing the proportion of iron in the +2 state ($Fe^{+2}/Fe^{+2} + Fe^{+3}$) per gram of rock. In relatively unoxidized strata, green colors result from the presence of iron-bearing phyllosilicates such as chlorite, illite, and in cases, glauconite. Kaolinite and smectite provide for white to light neutral colors. Red coloration is due to the presence of hematite, whereas less common yellows and browns generally result from limonite and goethite, respectively. Red hematitic staining is generally an early diagenetic phenomenon resulting from: (1) dehydration reactions in which limonite stains on detrital particles are altered to hematite; (2) dissolution of iron silicates and precipitation of the released iron; and (3) direct oxidation of magnetite and ilmenite grains (Hubert and Reed, 1978). Very low levels of hematite can impart a deep red coloration to a rock. The conversion from red to green colors occurs by reduction of the iron, which is either carried away in solution or reprecipitated as iron-rich clay minerals such as chlorite (Thompson, 1970). The strength of red or green color is a function of grain size, with fine-grained rocks having higher iron content and thus more intense color. This relationship is due to the presence of amorphous or poorly crystalline iron oxides attached to clays (McPherson, 1980). Changes in the oxidation state of iron, and hence the color, result from interactions with altering fluids at any time in the post-depositional history of a unit, and are thus controlled in part by permeability. Oxidation or reduction spots are due to incomplete diagenetic alteration of a layer.

The green-gray to black color spectrum in rocks is a function of total organic carbon (TOC), with darker colors corresponding to higher carbon content. This relationship is empirically confirmed in studies of modern sediment (e.g., Sheu and Presley, 1986). Main controls on organic carbon content include accumulation rate of organic matter, sediment accumulation rate, rate of decay of organics, and oxygen levels (Potter *et al*., 1980). Diagenetic loss of carbon, which leads to lighter colors, is an irreversible process. The carbon reservoir acts to shield the iron-bearing minerals from oxidation until most of the carbon is oxidized, at which point the green to red color spectrum is found. Oxidative loss of carbon, and red coloration, occurs when bottom water and/or pore-fluids are highly oxygenated. Green colors result from deposition of sediment with low organic content and weakly reducing to oxidizing conditions. Gray to black colors are associated with high TOC and dysaerobic to anaerobic bottom waters.

Gray colors also occur as a result of the presence of disseminated pyrite. During burial diagenesis metastable minerals such as mackinawite (FeS) and greigite (FeS_4) are transformed into sub-micron sized framboids of pyrite (Berner, 1971). Thermal maturity may also be an important factor in some rock. For instance, Lyons (1988) demonstrated that the black color of some limestone is due to carbonization of very small quantities of organic matter (<0.06 per cent TOC). Darker colors may also be imparted by Mn oxides or Fe-Mn oxides such as pyrolusite. Shale with siderite is gray to

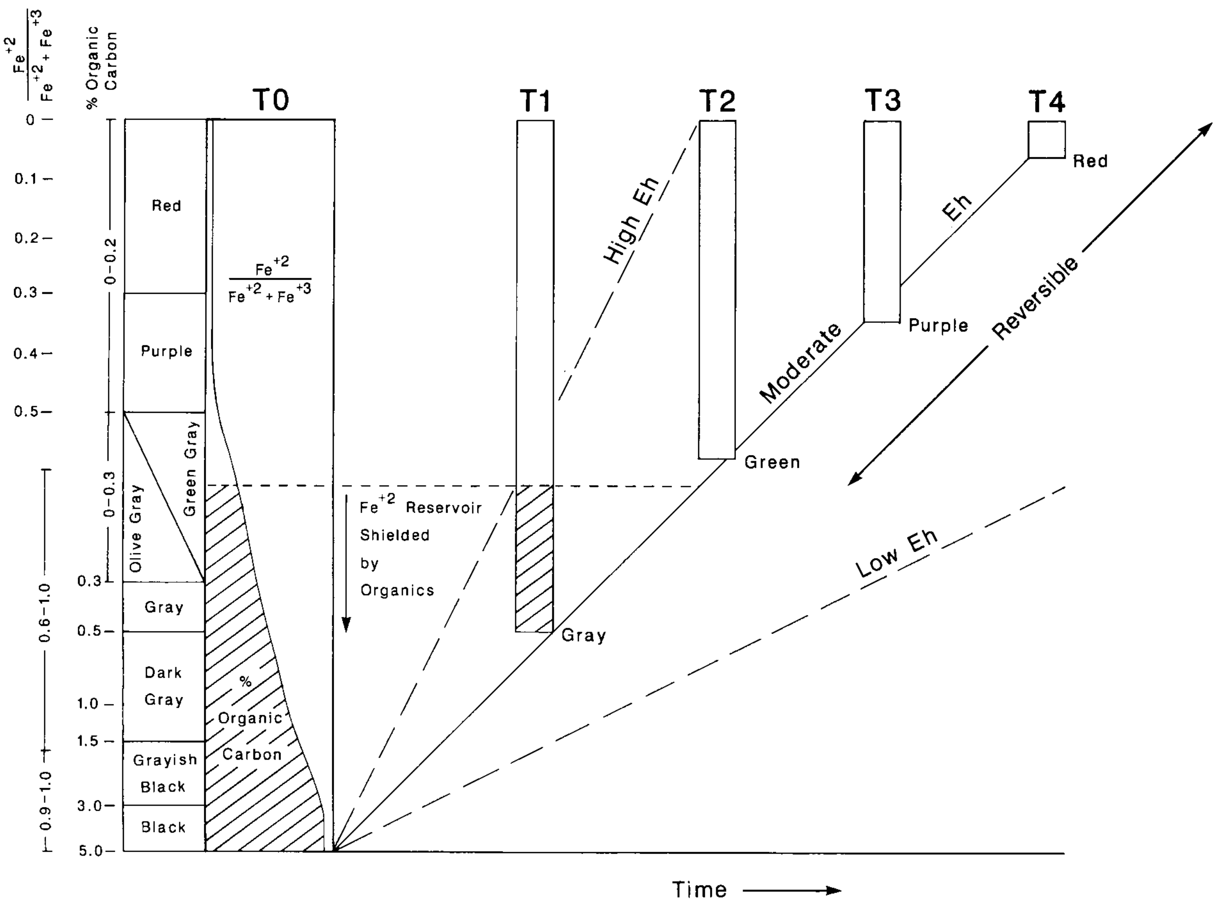

Figure C70 Graph relating color of rock, organic content and oxidation state of iron to time. "Time" refers to the length of time that pore fluids interact with sediment prior to lithification. Dark colors, olive-gray to black, are controlled by organic content. Green-gray to red colors, controlled by the oxidation state of iron, are reversible. The color history of sediment is dependent upon the Eh conditions. Three representative paths (high, moderate and low Eh) are shown for sediment with initially high organic content. Figure from Myrow (1990).

bluish on fresh surfaces but altered to brown tones by weathering (Pettijohn, 1975).

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Cross-references

Black Shales
Chlorite in Sediments
Red Beds
Sulfide Minerals in Sediments