Sedimentary parameters for computer modeling

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Abstract The sedimentary parameters that are most important in modeling sedimentary sequences and geometry are accumulation rate, lag time, and accommodation space. Each parameter incorporates several other variables. Accumulation rate is the net result of sediment input and in situ production (for carbonates) less export through bypass or erosion. The appropriate accumulation rate to be chosen from the vast amount of data available will depend on depositional environment, basinal asymmetry, climate, tectonic setting, and the time increment being modeled. Lag time expresses the necessary condition for a transgressive sequence: that the initial sediment accumulation rate is less than the rate of submergence or accommodation. Mechanisms are not well understood; the potential for sediment production in shallow-water carbonate environments, for example, generally exceeds known rates of submergence. Biologic factors may reduce sediment production rates in shallow water, but a physical threshold, such as the wave base, above which accumulation is suppressed, seems more probable. Accommodation space is the increment of room available for sediment accumulation as determined by eustasy, subsidence, and erosion. Subsidence, in turn, incorporates tectonism, isostasy, and physical and chemical compaction. Lag time, compaction rates induced by pressure solution, and the interaction of siliciclastics and carbonates are probably the least constrained variables.

A treatise on improved parameter definitions logically begins with a review of simulation programs to extract the input parameters and critical assumptions, which can then be neatly arrayed in a table. The profusion of available programs, the result of exponential growth from roots in the 1960's [cf. Harbaugh (1966) and Harbaugh and Bonham-Carter (1970)], is such that the table alone would probably exceed the intended length of this article [cf. Aigner et al. (1988), Bice (1988), Bosence and Waltham (1990), Bridge and Leeder (1979), Demicco and Spencer (1989), Harris (1989), Helland-Hanson et al. (1988), Jervey (1988), Koerschner and Read (1989), Lawrence et al. (1990), Lerche et al. (1987), Read et al. (1986), Scaturo et al. (1989), Spencer and Demicco (1989), and Watney et al. (1989)]. The input end of simulation and the other ends as well are reviewed by Kendall et al. (this volume). In this article I focus on sources of precise values for the parameters that most influence the geometry and facies of a simulated sedimentary sequence: accumulation rate, lag time, and accommodation space. Each parameter incorporates other variables. Accumulation rate depends on sediment input and on in situ production in carbonate environments versus sediment removal by erosion and bypassing. Lag time may be the result of biologic or physical thresholds, and accommodation space is the net of eustasy, subsidence, and erosion.

Accumulation rates

All-purpose sedimentation models incorporate accumulation rates for both terrigenous and carbonate sediments. These rates differ in general; they also respond in quite different fashions to changes in many other parameters. A wealth of data is available on rates of sediment accumulation (table 1). Well-documented values span such a range of rates and environments that the problem is in shopping: What are the appropriate values for the conditions to be simulated? The values in table 1 are grouped by depositional setting to facilitate selection. Different groupings, for example, by tectonic setting, may prove more appropriate for some models.

The time span of observation is an important determinant of sedimentation rates [Kukal, 1971; Schindel, 1980; see also Barrell (1917)]. In modeling this means that the time increment of the simulation may be important in choosing the appropriate sedimentation rate. Short-term observations invariably emphasize maximum rates produced by short-term events, such as floods or growth of organisms. It appears as though some rates approach infinity as the time of observation approaches zero (fig. 1). Clearly this is not the case, but the extrapolation emphasizes that short-term observations are not the most relevant to long-term considerations. The importance of duration of observation varies greatly among environments. Environments that suffer few perturbations (e.g., pelagic realms) have essentially uninterrupted sedimentation at rates that are virtually constant. In contrast, environments in which episodic events such as storms, floods, or turbidity currents dominate sedimentation typically have extreme short-term rates but intermittent deposition that modulates long-term averages. Schindel (1980) elegantly illustrated this phenomenon by plotting period of observation versus rate of sedimentation for a variety of environments. Figure 1 presents a series of such plots from the expanded data base of table 1.

An important consideration incorporated in some two-dimensional programs [e.g., Lawrence et al. (1990)] is whether

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Table 1. Modern sedimentation rates from various depositional settings[a]

Area	Rate (B)[b]	Period (yr)[c]	Reference
Fluvial environments			
*Lower Ohio R., natural levee, 1964 flood	460,000	1	Bridge & Leeder, 1979 (Alexander & Prior, 1971)
*Ohio R. floodplain, 1937 flood average	70,000	1	Bridge & Leeder, 1979 (Mansfield, 1938)
*Ohio R. floodplain, 1937 flood range	3,000–560,000	1	Bridge & Leeder, 1979 (Mansfield, 1938)
Lower Ohio R., natural levee	16,000	40	Bridge & Leeder, 1979 (Alexander & Prior, 1971)
Lower Ohio R., natural levee	10,000	750	Bridge & Leeder, 1979
Lower Ohio R., accretionary ridge	6,000	1,000	Bridge & Leeder, 1979
Ohio R. floodplain	4,500	150	Schindel, 1980 (Moore, 1971)
*Lower Ohio R., accretionary ridge, 1964 flood	3,200	1	Bridge & Leeder, 1979 (Alexander & Prior, 1971)
Lower Ohio R., swale	1,900	1,000	Bridge & Leeder, 1979
Lower Ohio R., accretionary ridge	270	1,000	Bridge & Leeder, 1979
Yuba R., California	100,000	1	Kukal, 1971
Sacramento R., California	75,000	1	Kukal, 1971
Cimarron R., Maryland, floodplain	51,000	12	Schindel, 1980 (Schumm & Lichty, 1963)
Western Run, Maryland, floodplain	16,300	50	Schindel, 1980 (Costa, 1973)
Nile R., floodplain	9,000	1	Kukal, 1971
Nile R., floodplain, range	9,100–12,200	1,000	Bridge & Leeder, 1979 (Leopold et al., 1964)
Delaware R. floodplain	140–1,150	6,000	Schindel, 1980 (Ritter et al., 1973)
Indus R.	200	4,500	Kukal, 1971
Wisconsin valley floodplain	1,000	6,070	Bridge & Leeder, 1979 (Knox, 1972)
Wisconsin valley floodplain	350	6,040	Bridge & Leeder, 1979 (Knox, 1972)
Blockhouse Creek, Wisconsin, floodplain	150–380	6,000	Bridge & Leeder, 1979 (Knox, 1972)
Little Tallahatchie R., Mississippi, natural levee	47,000–65,000	8	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Little Tallahatchie R., natural levee	13,000–20,000	31	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Little Tallahatchie R., crevasse splay	28,000–34,000	8	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Little Tallahatchie R., crevasse splay	27,000	31	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Little Tallahatchie R., abandoned channels	9,000–28,000	8	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Little Tallahatchie R., abandoned channels	7,000–10,000	31	Bridge & Leeder, 1979 (Ritchie et al., 1975)
Beatton R., British Columbia, floodplain	1,000–61,000	500	Bridge & Leeder, 1979 (Nanson, 1977)
Bobr & Strzegomka R., USSR, floodplains	1,000–5,000	≈10,000	Bridge & Leeder, 1979 (Teisseyre, 1977)
South Carolina Piedmont rivers, floodplain	8,000	150	Bridge & Leeder, 1979 (Happ, 1945)
Buck Run, floodplain	650	1,450	Bridge & Leeder, 1979 (Leopold et al., 1964)
Tigris & Euphrates, floodplain	200	5,000	Bridge & Leeder, 1979 (Leopold et al., 1964)
Cheyenne R., Wyoming, floodplain	41,000–61,000	60	Bridge & Leeder, 1979 (Leopold et al., 1964)
Dry Creek, Nebraska, floodplain	8,600	500	Bridge & Leeder, 1979 (Brice, 1966)
Upper Dry Creek, Nebraska, floodplain	4,600–5,500	33	Bridge & Leeder, 1979 (Brice, 1966)
Well Canyon, Nebraska, floodplain	15,500–20,000	40	Bridge & Leeder, 1979 (Brice, 1966)
Medicine Creek, Nebraska, floodplain	83,000	22	Bridge & Leeder, 1979 (Brice, 1966)
Medicine Creek, drainage basin average	25,000	22–500	Bridge & Leeder, 1979 (Brice, 1966)
Chemung R., New York, floodplain	4,600		Bridge & Leeder, 1979 (Nelson, 1966)
*Bijou Creek, Colorado, overbank, 1965 flood	61,000–3,600,000	1	Schindel, 1980 (McKee et al., 1967)
*Missouri R., levees, 1881 flood	$1.22–1.83 \times 10^6$	1	Bridge & Leeder, 1979 (Leopold et al., 1958)
*Kansas R., floodplain, 1951 flood	29,000	1	Bridge & Leeder, 1979 (Leopold et al., 1958)
*Farmington R., Connecticut, floodplain, 1955 flood	15,000	1	Bridge & Leeder, 1979 (Wolman & Eiler, 1958)
*Connecticut R., floodplain, 1936 flood	35,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., floodplain, 1938 flood	22,000	1	Bridge & Leeder, 1979 (Jahns, 1947)

(table notes on p. 76)

Area	Rate (B)[b]	Period (yr)[c]	Reference
*Connecticut R., banks, 1936 flood	259,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., banks, 1938 flood	173,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., tributary banks, 1936 flood	200,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., tributary banks, 1938 flood	107,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., artificial levees, 1936 flood	91,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Connecticut R., artificial levees, 1938 flood	43,000	1	Bridge & Leeder, 1979 (Jahns, 1947)
*Ob' R., USSR, point bar, 1969 flood	≤1,500,000	1	Bridge & Leeder, 1979 (Velikanov & Yarnykh, 1970)
*Ob' R., crevasse splay & levee, 1969 flood	≤600,000	1	Bridge & Leeder, 1979
*Ob' R., USSR, flood basin, 1969 flood	200–30,000	1	Bridge & Leeder, 1979
San Joaquin River, California, (Holocene)	15,000	≈10,000	Bull, 1972
Mississippi R., floodplain	1,400	30,000	Bridge & Leeder, 1979 (Fisk, 1944)
Upper Mississippi R., artificial backwater	25,000–35,000	20	Bridge & Leeder, 1979 (McHenry et al., 1976)
*Mississippi R., point bar, 1973 flood	860,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., point bar, range	130,000–3,000,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., natural levee, 1973 flood	530,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., natural levee, 1973 flood	100,000–840,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., levee back, 1973 flood	125,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., levee back, 1973 flood	60,000–270,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., abandoned channels, 1973 flood	60,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., abandoned channels, 1973 flood, range	40,000–90,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., backswamp, 1973 flood	11,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
*Mississippi R., backswamp, 1973 flood	5,000–25,000	1	Bridge & Leeder, 1979 (Kesel et al., 1974)
Mississippi River (Miocene)	32–53	20×10^6	Rainwater, 1966
Diablo Range, California, Holocene alluvial fan	26,000	≈10,000	Bull, 1972
Eolian environments			
Southern Peru	2,000,000	3	Bigarella, 1972
Sahara Desert, Algeria	6,500	4,000	Galloway & Hobday, 1983 (Wilson, 1973)
Southern Sahara Desert	800–1,700	12,000	Breed et al., 1979
Grand Erg Oriental, Algeria, average	19	1,350,000	Breed et al., 1979
Grand Erg Oriental, Algeria, max.	87	1,350,000	Breed et al., 1979
Kalahari Desert	300–3,300	≈10,000	Breed et al., 1979
Cerchen Desert, PRC, average	60,000	1,500	Breed et al., 1979
Navajo Sandstone (E. Jurassic), USA	53	17×10^6	Galloway & Hobday, 1983
Loess	200–1,000		Kukal, 1971
Loess, central Alaska	15–193		Kukal, 1971 (Péwé, 1968)
Lacustrine environments			
Lacustrine, average	3,000		Kukal, 1971
Vierwaldstättersee, Switzerland, calcareous clays	10,400–31,700		Kukal, 1971
Vierwaldstättersee, Switzerland	3,500–5,000		Schindel, 1980 (Schwarzacher, 1975)
Brienz, Switzerland, calcareous clays	31,700		Kukal, 1971
Léman, Switzerland	1,200	≈150	Schindel, 1980 (Krishnaswami et al., 1971)
Lunz, [Léman?], mouth of Rhone	17,900		Kukal, 1971
Lunz, [Léman?], average	2,500		Kukal, 1971
Lunz, Austria, average	1,800		Schindel, 1980 (Schwarzacher, 1975)

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Lacustrine environments (cont.)			
Salton Sea, California, saline	5,000–14,000	≈50	Schindel, 1980 (Arnal, 1961)
Wallensee, Switzerland, calcareous clays	11,300		Kukal, 1971
Onega, USSR, clays	7,100		Kukal, 1971
Ladoga, USSR, clays	6,120		Kukal, 1971
Trout and Mendota, Wisconsin	6,000	<100	Schindel, 1980 (Bruland et al., 1975)
Trout Lake, Wisconsin	4,000	≈100	Schindel, 1980 (Koide et al., 1972)
Olof Jone Damm, Sweden, peat	5,300		Kukal, 1971
Shinji, Japan	3,000–5,000	<100	Schindel, 1980 (Matsumoto, 1975)
Shinji, Japan	1,200	9,500	Schindel, 1980 (Mizuno et al., 1972)
North German lakes, marl	1,000–3,000		Kukal, 1971
Swedish lakes, gyttja	1,000–2,000		Kukal, 1971
Pavin, France	1,300	≈100	Schindel, 1980 (Krishnaswami et al., 1971)
Tahoe, USA	1,000	≈100	Schindel, 1980 (Koide et al., 1972; Bruland et al., 1975)
Titicaca, Bolivia	1,000		Schindel, 1980 (Bruland et al., 1975)
Zürich, Switzerland	700		Kukal, 1971
Neuchâtel, Switzerland, varved calcareous clays	700		Kukal, 1971
Maxinkuckee, Canada, marl, eutrophic	300		Kukal, 1971
Great Lakes, N. America, varved mud	150		Kukal, 1971
Michigan, varved calcareous clays	3,000		Kukal, 1971
Michigan	100–1,200	<100	Schindel, 1980 (Robbins & Edgington, 1975)
Superior	100–600	<100	Schindel, 1980, (Bruland et al., 1975)
Constance (Bodensee), mouth of Rhine	22,400	15,000	Müller & Gees, 1970
Constance (Bodensee)	1,500–6,000	15,000	Müller & Gees, 1970
Diatomite (average)	300–1,000		Kukal, 1971
Varved glacial lakes			
Weistriztal, Czheckoslovakia	60,000–100,000		Reineck & Singh, 1975 (Schwarzbach, 1940)
Burks Falls, Ontario	2,000–17,000	620	Antevs, 1925
Espanola, Ontario	1,000–83,000	985	Antevs, 1925
Tishaming, Ontario	4,000–65,000	1,335	Antevs, 1925
Huntsville, Ontario	2,000–45,000	760	Antevs, 1925
Bracebridge, Ontario	3,000–92,000	511	Antevs, 1925
Bracebridge, Ontario, average	10,900	112	Antevs, 1925
Ancient lakes			
Lake Bonneville, Pleistocene, Utah	1,800	10^6	Feth, 1964; Picard & High, 1972
Unita Formation, Eocene, Wyoming	5.7	13.3×10^6	Feth, 1964; Picard & High, 1972
Green River Formation, Eocene, Wyoming and Colorado	150	4×10^6	Feth, 1964; Picard & High, 1972
Flagstaff Limestone, Paleocene–Eocene, Utah	22–110	2.75×10^6	Feth, 1964; Picard & High, 1972
Todilto Limestone, L. Jurassic, New Mexico	380	20,000	Feth, 1964; Picard & High, 1972
Lockatong Formation, L. Triassic, New Jersey	225	5.1×10^6	Van Houten, 1964; Picard & High, 1972
Deltaic environments			
Delta topsets, average	15,000–20,000		Kukal, 1971
Mississippi	2,740,000	4 d	Kukal, 1971
Mississippi, channel-mouth bar	500,000	100	Schindel, 1980 (Coleman, 1976)
Mississippi, channel-mouth bar	340,000	195	Schindel, 1980 (Gould, 1970)
Mississippi, delta front	300,000–450,000	1	Kukal, 1971
Mississippi, prodelta	60,000–300,000	1	Kukal, 1971
Mississippi, subaerial average	170,000–200,000	600	Schindel, 1980 (Coleman, 1976)
Mississippi, offshore	200,000	100	Schindel, 1980 (Coleman, 1976)
Mississippi, crevasse splay	30,000–100,000	150	Schindel, 1980 (Coleman, 1976)
Mississippi, adjacent shelf	45,000	1	Kukal, 1971

Area	Rate (B)[b]	Period (yr)[c]	Reference
Mississippi, premodern lobes, average	20,000–25,000	1,000	Schindel, 1980 (Coleman, 1976)
Mississippi, interdistributary bay	19,600	120	Elliott, 1978 (Gagliano & Van Beck, 1970)
Mississippi, Sale Sypremort lobe	8,300–12,500	1,200	Schindel, 1980 (Coleman, 1976)
Mississippi, "maximum"	10,000	11,000	Lisitzin, 1972
Mississippi, submarine, nearshore	8,200	<100	Schindel, 1980 (Shokes & Presley, 1976)
Mississippi, submarine, cont. shelf	6,100	<100	Schindel, 1980 (Shokes & Presley, 1976)
Mississippi, submarine, cont. slope	400	<100	Schindel, 1980 (Shokes & Presley, 1976)
Mississippi River (Miocene)	325	20×10^6	Rainwater, 1966
Mississippi River (Miocene), prodelta	220	20×10^6	Rainwater, 1966
Colorado, Texas	4,064,000	6	Kanes, 1970
Po, Italy, at shoreline, average	465,000	25	Nelson, 1970
Po, at shoreline, range	268,000–653,000	19–45	Nelson, 1970
Rhône	400,000	1	Kukal, 1971
Rhône	700		Schindel, 1980 (Schwarzacher, 1975)
Rhône, river mouth, 50 m depth	350,000	1	Oomkens, 1970
Rhône, mouth of Grand Rhône	14,000	≈5,000	Oomkens, 1970
Rhône, mouth of Petit Rhône	7,600	≈5,000	Oomkens, 1970
Rhône, shoreline	2,000	11,000	Lisitzin, 1972
Rhône, 45 km offshore	6,000	11,000	Lisitzin, 1972
Rhône, 75 km offshore	1,000	11,000	Lisitzin, 1972
Rhine delta, Lake Constance (Bodensee)	2,500,000	10	Müller, 1966
Rhine delta, Lake Constance, average	262,800	50	Müller, 1966
Nile, subaerial portion	10,000	1	Kukal, 1971
Nile	660		Schindel, 1980 (Schwarzacher, 1975)
Fraser, Canada	50,000–300,000		Kukal, 1971
Volga, Caspian Sea	5,000–70,000		Kukal, 1971
Tana River, Japan	30,000–70,000	10	Schindel, 1980 (Ambe, 1972)
Alamo River, Salton Sea, USA	50,000	33	Schindel, 1980 (Arnal, 1961)
Amu Darya River, Aral Sea, USSR	25,000		Kukal, 1971
Orinoco, Venezuela	5,000–6,000	11,000	Lisitzin, 1972
Sabine, Texas	2,930	5,200	Nelson & Bray, 1970
Guadelupe, Texas	2,100	2,000	Donaldson et al., 1970
Rud Hilla, Persian Gulf	800–5,000	<6,000	Schindel, 1980 (Melguen, 1973)
Columbia River, Washington shelf	1,300–3,900	≈100	Schindel, 1980 (Nittrouer et al., 1979)
Huang-He, PRC	1,500		Kukal, 1971
Don, Sea of Azov, USSR, subaerial portion	1,220	1	Kukal, 1971
Malaysia, tide-dominated delta	1,000	100	Galloway & Hobday, 1983 (Coleman et al., 1970)
Bengal cone (Ganges prodelta)	62	10.2×10^6	Moore et al., 1974
Tidal flats, coastal wetlands, and beaches			
Tidal flats			
Jade Busen, Germany	1,450,000	8 d	Kukal, 1971 (Reineck, 1960)
Jade Busen, Germany	11,500	4	Kukal, 1971 (Reineck, 1960)
Jade Busen, Germany	2,200	1,900	Kukal, 1971 (Reineck, 1960)
The Wash, UK	16,000–80,000	9 mo	Schindel, 1980 (Evans, 1965)
Netherlands	10,000–20,000	1	Kukal, 1971
Laguna Madre, Texas	250–5,000	2,500	Schindel, 1980 (Miller, 1975)
Boundary Bay, British Columbia	5,000	20	Schindel, 1980 (Kellerhals & Murray, 1969)
Boundary Bay, British Columbia	420	4,500	Schindel, 1980 (Kellerhals & Murray, 1969)
Beaches			
Chenier beaches, SW Louisiana	6,300–21,200	400–2,200	Reineck & Singh, 1975 (Gould andMcFarlan, 1959)
Padre Island, Texas, barrier beach	2,100	4,000	Reineck & Singh, 1975 (Fisk, 1959)
Galveston Island, Texas, barrier beach	2,860	3,500	Bernard et al., 1962
Fire Island Inlet, New York	103,000	115	Kumar & Sanders, 1974
Wangerooge Inlet, North Sea	88,000	68	Reineck & Singh, 1975 (Reineck, 1958)

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Tidal flats, coastal wetlands, and beaches (cont.)			
Beaches (cont.)			
Nayarit, Mexico	44,000	205	Reineck & Singh, 1975 (Curray et al., 1969)
Christchurch Formation (Holocene), New Zealand, offshore sand	2,700–3,500	5,535	Suggate, 1968
Wetlands, salt marshes			
Long Island Sound, USA	4,700–6,300	<100	Schindel, 1980 (Armentano & Woodwell, 1975)
Denmark	3,600	30	Schindel, 1980 (Schou, 1967)
Connecticut	2,000–6,500	6	Schindel, 1980 (Harrison & Bloom, 1974)
Farm River, Connecticut	1,600	200	Schindel, 1980 (McCaffrey, 1977)
Klang River (Malaysia)	1,000		Schindel, 1980 (Coleman, 1976)
SW Louisiana, salt marsh & lagoon	5,500–27,500	400–1,800	Reinick & Singh, 1975 (Gould and McFarlan, 1959)
Wetlands, peat deposits			
UK, Littoral	9,800		Kukal, 1971
Schwabia, S. Germany, high moors	1,500–1,800	1	Kukal, 1971
North American	550		Kukal, 1971
Borneo (Kalimantan), coastal swamps	4,250	4,000	Galloway & Hobday, 1983 (Stach et al., 1975)
Everglades, Holocene coastal swamp	1,200	3,460	Spackman et al., 1964
Olof Jone Damm, Sweden, fresh water	5,300		Kukal, 1971
Bays, lagoons, and estuaries			
Texas, lagoon	14,300	290	Schindel, 1980 (Moore, 1955)
Texas, lagoon	9,100	68	Schindel, 1980 (Shepard, 1953)
Texas, lagoon, clay and eolian sand	3,800		Kukal, 1971
San Antonio Bay, Texas	3,750	100	Donaldson et al., 1970 (Shepard & Moore, 1960)
Texas, lagoon	2,300	9,300	Schindel, 1980 (Shepard & Moore, 1955)
Padre Island, Texas, lagoon	1,900	4,000	Reineck & Singh, 1975 (Fisk, 1959)
Great Bay, USA, estuary	1,600–7,800	≈100	Schindel, 1980 (Capuzzo & Anderson, 1973)
Long Island Sound, USA	6,000	30	Schindel, 1980 (Thomson & Turekian, 1973)
Long Island Sound	1,000–7,000	<100	Schindel, 1980 (Benninger et al., 1977)
Long Island Sound	500–1,000	10,000	Schindel, 1980 (Benninger et al., 1977)
Mobile Bay, Alabama	5,600	115	Schindel, 1980 (Ryan & Goodell, 1972)
Mobile Bay, Alabama	1,640	6,000	Schindel, 1980 (Ryan & Goodell, 1972)
Firth of Clyde, Scotland	5,000	12,000	Schindel, 1980 (Kuenen, 1950)
Firth of Clyde, Scotland, clay	2,400–3,000		Kukal, 1971
James River, Virginia, estuary	1,500–3,000	75	Schindel, 1980 (Nichols, 1972)
Sea of Azov, USSR, estuary	900–2,400	11,000	Lisitzin, 1972
Hampton, New Hampshire, estuary	1,000–2,300	11,000	Schindel, 1980 (Keene, 1970)
Kiel Bay, Germany, sand & silt	1,500–2,000		Kukal, 1971
Drammens Fjord, Sweden	1,500	12,000	Schindel, 1980 (Kuenen, 1950)
San Francisco Bay, USA	300–1,300	≈2,500	Schindel, 1980 (Story et al., 1966)
Gulf of California, Mexico	1,000	12,000	Schindel, 1980 (Kuenen, 1950)
Gulf of California, clay, diatomite	600–1,000		Kukal, 1971
Gulf of Paria, Venezuela, clay	0–10,000	700	Kukal, 1971 (van Andel & Postma, 1954)
Kara–Bougas–Gol, Caspian Sea, clay & salt	500–700		Kukal, 1971
Inland seas			
Black Sea (Messinian), carbonates, some pebbly mudstone	1,030	800,000	Hsü, 1978
Black Sea, terrigenous clastics	1,000	15,000	Schindel, 1980 (Stoffers et al., 1978)
Black Sea, dolomitic varves	900	81	Schindel, 1980 (Ross et al., 1978)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Black Sea, terrigenous & diatom mud	600	125,000	Hsü, 1978
Black Sea, terrigenous & diatom mud	500	500,000	Hsü, 1978
Black Sea, lacustrine carbonates	310	1.1×10^6	Hsü, 1978
Black Sea, terrigenous mud	100–400	11,000	Lisitzin, 1972
Black Sea	50–400	7,000	Schindel, 1980 (Ross et al., 1970)
Black Sea, coccolith ooze	100–300	3,000	Schindel, 1980 (Stoffers et al., 1978)
Black Sea	200		Kukal, 1971
Black Sea	200	12,000	Schindel, 1980 (Kuenen, 1950)
Black Sea, brackish sapropel	100	4,000	Schindel, 1980 (Stoffers et al., 1978)
Black Sea, marine, terrigenous mud	100	11,000	Hsü, 1978
Black Sea (Pliocene) lacustrine chalk	54	3.45×10^6	Hsü, 1978
Black Sea (pelagic)	10–40	11,000	Lisitzin, 1972
Black Sea (Miocene) black shale	26	4×10^6	Hsü, 1978
Mediterranean, Baleric abyssal plain; hemipelagic & turbidites	160–520	20,000	Rupke, 1975
Mediterranean, Tyrrhenian Sea, calcareous and diatom clay	100–500	12,000	Schindel, 1980 (Kuenen, 1950)
Mediterranean, terrigenous turbidites	300	≈10,000	Cita et al., 1978
Mediterranean, ooze & eolian	200		Kukal, 1971
Mediterranean, deep basin	150	1.9×10^6	Cita et al., 1978
Mediterranean, calcareous ooze	100		Kukal, 1971
Mediterranean, average	25–90	3×10^6	Cita et al., 1978
Mediterranean, pelagic oozes	50	≈10,000	Cita et al., 1978
Mediterranean, ridges & basin margins	25–50	1.9×10^6	Cita et al., 1978
Mediterranean, Adriatic Sea, shells (lag)	10		Kukal, 1971
Mediterranean, pelagic (condensed)	1	1.9×10^6	Cita et al., 1978
Caspian Sea, mouth of Kura River	6,000	7,000	Lisitzin, 1972
Caspian Sea, pelagic	200–600	7,000	Lisitzin, 1972
Caspian Sea, calcareous clays	100–180		Kukal, 1971
Baltic Sea	200–2,000	10,000	Schindel, 1980 (Alhonen, 1966)
Baltic Sea, black organic clays	300		Kukal, 1971
Persian Gulf (eastern basin), terrigenous	410	9,000	Schindel, 1980 (Seibold et al., 1973)
Persian Gulf (central basin), terrigenous and carbonate	70	9,000	Schindel, 1980 (Seibold et al., 1973)
Sea of Okhotsk, W. Pacific, shelf depression and base of slope	90–250	11,000	Lisitzin, 1972
Sea of Okhotsk, central shelf	9–45	11,000	Lisitzin, 1972
Gulf of Mexico, upper slope, sandy, silty clays	70		Kukal, 1971
Gulf of Mexico, lower slope, silty clays	50		Kukal, 1971
Gulf of Mexico, basin floor, calcareous clays	40		Kukal, 1971
Milford Sound (New Zealand), sandy silts	12.5		Kukal, 1971
Terrigenous shelf deposits			
North American shelf	0–400		Kukal, 1971
New Jersey shelf, USA, sand	950–1,300	8,000	Swift et al., 1984
Barents Sea, Arctic Ocean, clays	8–40		Kukal, 1971
North Sea, gale 'Adolph–Bermpohl', sand	4,200,000	36 h	Gadow & Reineck, 1969
North Sea, storm, March 1967, sand	1,100,000	84 h	Gadow & Reineck, 1969
Nompho, Korea, mud	1,500,000	4	Lisitzin, 1972
Antarctic shelf, sand	20–60	10,000	Lisitzin, 1972
Antarctic shelf, mud	200–300	10,000	Lisitzin, 1972
Indochinese shelf	50–200		Kukal, 1971
North Sea sand waves	2,200–4,300	10,000	Houbolt, 1968
High Island, Gulf of Mexico, nearshore sands	1,050–1,760	5,200	Nelson & Bray, 1970
High Island, Gulf of Mexico, nearshore muds	470	5,200	Nelson & Bray, 1970

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Shallow-water carbonates			
Individual coral growth rates			
Range, coral growth	850–150,000		Schlager, 1981
Massive coral	4,000		Kukal, 1971
Pristatophyllum, Devonian	2,000–6,200	33	Faul, 1943
Atlantic corals			
Corals, Florida Bay, leeward	570	528	Kukal, 1971
Montastrea annularis			
inshore, patch reef, <6 m depth, Florida	8,200	50	Shinn, Lidz et al., 1989 (Hudson, 1981)
offshore, >6 m depth, Florida	6,300	50	Shinn, Lidz et al., 1989 (Hudson, 1981)
platform margin, <3 m depth, Florida, windward	11,200	50	Shinn, Lidz et al., 1989 (Hudson, 1981)
Key West, Florida	2,800–5,800	15	Weber & White, 1977
Florida, 0–5 m	6,000	3	Huston, 1985 (Vaughan, 1915)
Virgin Islands, 0 m	9,170 ± 1,330		Baker & Weber, 1975
Virgin Islands, 5 m	9,950 ± 1,430		Baker & Weber, 1975
Virgin Islands, 9 m	10,410 ± 1,240		Baker & Weber, 1975
Virgin Islands, 13.5 m	9,690 ± 1,360		Baker & Weber, 1975
Virgin Islands, 18 m	6,540 ± 3,560		Baker & Weber, 1975
Virgin Islands, 22.5 m	2,060 ± 540		Baker & Weber, 1975
Virgin Islands, 27 m	1,560 ± 200		Baker & Weber, 1975
Virgin Islands, 2 m, leeward	7,600 ± 330	4 m	Gladfelter et al., 1978
Virgin Islands, forereef, 10 m, windward	7,600 ± 820	4 m	Gladfelter et al., 1978
Jamaica, 0–1 m	6,950	≥ 3	Huston, 1985
Jamaica, 5 m	7,400 ± 3,100	≥ 3	Huston, 1985
Jamaica, 10 m	7,430 ± 1,920	≥ 3	Huston, 1985
Jamaica, 10 m	6,680 ± 2,000	2	Dustan, 1979
Jamaica, 20 m	1,770 ± 700	≥ 3	Huston, 1985
Jamaica, 28 m	1,700 ± 400	2	Dustan, 1979
Jamaica, 30 m	1,750 ± 330	≥ 3	Huston, 1985
Jamaica, 45 m	1,630 ± 1,200	2	Dustan, 1979
Curacao, 6–15 m	6,300–7,800		Huston, 1985 (Bak, 1976)
Belize	12,000		Weber & White, 1977
Caribbean	3,000–12,000		Weber & White, 1977
Caribbean	6,000		Ghiold & Enos, 1982 (Macintyre & Smith, 1974)
Pleistocene, Florida	5,000	400	Shinn, Lidz et al., 1989
Pleistocene, Florida	5,200	31	Landon, 1975
Montastrea cavernosa			
Jamaica, 10 m	3,600 ± 1,900	≥ 3	Huston, 1985
Jamaica, 20 m	6,840 ± 2,670	≥ 3	Huston, 1985
Jamaica, 30 m	4,100 ± 1,390	≥ 3	Huston, 1985
Key West	2,250–4,050	22	Weber & White, 1977
Florida	3,200–5,700	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Acropora palmata			
Caribbean	100,000–120,000	1	Shinn, Lidz et al., 1989 (Lewis et al., 1968)
Florida, <5 m	25,000–40,000	3	Huston, 1985 (Vaughan, 1915)
Curacao, <5 m	88,000		Huston, 1985 (Bak, 1976)
Virgin Islands, 1/2 m, leeward	56,900 ± 4,100	2 m	Gladfelter et al., 1978
Virgin Islands, 1/2 m , windward	68,500 ± 6,900	2 m	Gladfelter et al., 1978
Virgin Islands, 9 m , windward	77,000 ± 6,900	2 m	Gladfelter et al., 1978
Acropora cervicornis			
Jamaica, windward	266,000 ± 129,000	1	Buddemeier & Kinzie, 1976 (Lewis et al., 1968)
Jamaica, <5 m, windward	109,000–159,000	1	Tunnicliffe, 1983
Jamaica, 6–15 m, windward	80,000–140,000	1	Tunnicliffe, 1983
Jamaica, 25 m, windward	92,000–148,000	1	Tunnicliffe, 1983

Area	Rate (B)[b]	Period (yr)[c]	Reference
Virgin Islands, 10m, windward	71,000 ± 6,500	2 m	Gladfelter et al., 1978
Florida, <5 m, windward	105,100 ± 16,500	1	Shinn, 1966
Florida, 1 m , leeward	45,700 ± 18,400	9 m	Shinn, 1966
Florida, <5 m	40,000–45,000	3	Huston, 1985 (Vaughan, 1915)
Barbados, leeward	145,000 ± 559,000	1	Buddemeier & Kinzie, 1976 (Lewis et al., 1968)
Colpophyllia natans			
Jamaica, 5 m	9,000 ± 1,100	≥ 3	Huston, 1985
Jamaica, 10 m	8,100 ± 1,400	≥ 3	Huston, 1985
Jamaica, 20 m	4,200 ± 850	≥ 3	Huston, 1985
Diploria spp.	3,500–10,000	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Diploria stringosa, Bermuda	3,300–3,500	50	Dodge & Vaisnys, 1980
D. labyrinthiformis, Florida	3,500 ± 600	3–27	Ghiold & Enos, 1982
Solenastrea bournoni, Florida Bay, leeward	8,900	100	Shinn, Lidz et al., 1989
Porites porites			
Florida	14,000–17,000	6	Landon, 1975
Caribbean	21,000–36,000		Landon, 1975 (Lewis et al., 1968)
Pleistocene, Florida	10,500	6	Landon, 1975
Porites furcata, Florida & Bahamas	9,000–23,000	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Porites astreoides			
Jamaica, 0–1 m	5,030 ± 560	≥ 3	Huston, 1985
Jamaica, 5 m	5,000 ± 1,500	≥ 3	Huston, 1985
Jamaica, 10 m	3,300 ± 770	≥ 3	Huston, 1985
Jamaica, 20 m	2,700 ± 220	≥ 3	Huston, 1985
Jamaica, 30 m	2,300 ± 250	≥ 3	Huston, 1985
Virgin Islands, 2 m	3,450 ± 320	8 m	Gladfelter et al., 1978
Virgin Islands, 10 m	3,000 ± 120	3 m	Gladfelter et al., 1978
Florida	4,300		Ghiold & Enos, 1982 (Kissling, 1977)
Florida, Bahamas	5,700–13,000	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Dendrogyra cylindrus, Florida	5,000	1	Shinn, Lidz et al., 1989
Siderastrea sidera			
Jamaica, 10 m	7,150	≥ 3	Huston, 1985
Jamaica, 20 m	3,000 ± 800	≥ 3	Huston, 1985
Jamaica, 30 m	3,100	≥ 3	Huston, 1985
Florida	2,200–2,700	22	Landon, 1975
Pleistocene, Florida	1,500	19	Landon, 1975
Favia fragum, Florida	2,900–3,800	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Manicina sp. Florida	2,500–8,700	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Agaricia argaricites, Jamaica, 0–30m	1,110 ± 270	≥3	Huston, 1985
Agaricia sp, Florida	3,800	3	Ghiold & Enos, 1982 (Vaughn, 1915)
Pacific corals			
Acropora spp	85,000–226,000		Buddemeier & Kinzie, 1976
Acropora spp, Samoa, 0–5 m	4,000–185,000		Huston, 1985 (Mayor, 1924)
A. abrantoides, Yapp	125,000–130,000		Huston, 1985 (Tamura & Hada, 1932)
A. pulchra	101,000–172,000		Huston, 1985 (Tamura & Hada, 1932)
Astreapora myriophythalma			
Enewetak, 6–15 m	7,500–13,000		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 16–25 m	5,000–5,500		Huston, 1985 (Buddemeier et al., 1974)
Pocillopora spp., Samoa, 0–5 m	7,000–35,000		Huston, 1985 (Mayor, 1924))
P. damicornis	13,900–27,800		Buddemeier & Kinzie, 1976
Guam, 0–5 m	33,300		Huston, 1985 (Neudecker, 1977)
Guam, 6–15 m	36,700		Huston, 1985 (Neudecker, 1977)
Guam, >25 m	18,100		Huston, 1985 (Neudecker, 1977)
Panama, 3 m	39,600 ± 1,500		Glynn, 1976
Panama, 6 m	33,600 ± 2,100		Glynn, 1976
Panama, 0–15 m	44,300–59,300		Huston, 1985 (Wellington, 1982)
P. eydouxi, Enewetak, 0–5 m	50,000		Huston, 1985 (Buddemeier et al., 1974)

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Shallow-water carbonates (cont.)			
Pacific corals (cont.)			
Psammocora sp., Enewetak, 0–5 m	30,000		Huston, 1985 (Buddemeier et al., 1974)
Pavona sp., Samoa, 0–5 m	32,000		Huston, 1985 (Mayor, 1924)
P. clavus, Panama, 0–5 m	15,500–23,000		Huston, 1985 (Wellington, 1982)
Panama, 6–15 m	12,000–19,000		Huston, 1985 (Wellington, 1982)
P. gigantea, Panama, 0–5 m	10,000–19,500		Huston, 1985 (Wellington, 1982)
Panama 6–15 m	8,000–17,000		Huston, 1985 (Wellington, 1982)
Fungia fungites, Enewetak, 6–15 m	10,000–12,000		Huston, 1985 (Buddemeier et al., 1974)
Porites spp, 0–5 m	7,000–48,500		Huston, 1985 (various)
P. lutea			
Enewetak, 0–5 m	5,000–13,500		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 6–15 m	5,000–11,000		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 16–25 m	3,000–9,500		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, >25 m	5,000–6,000		Huston, 1985 (Buddemeier et al., 1974)
P. lobata, Enewetak, 6–15 m	10,000–11,500		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 16–25 m	5,000–6,000		Huston, 1985 (Buddemeier et al., 1974)
Favia pallida			
Enewetak, 0–5 m	5,500–7,500		Huston, 1985 (Highsmith, 1979)
Enewetak, 6–15 m	5,000–7,000		Huston, 1985 (Highsmith, 1979)
Enewetak, 15–25 m	4,000–7,000		Huston, 1985 (Highsmith, 1979)
Enewetak, 25–30 m	4,000–6,500		Huston, 1985 (Highsmith, 1979)
F. speciosa			
Enewetak, 0–5 m	4,600		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 6–15 m	5,600–8,500		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 16–25 m	6,500–7,000		Huston, 1985 (Buddemeier et al., 1974)
Goniastrea retiformis			
Enewetak, 0–5 m	8,000–10,000		Huston, 1985 (Buddemeier et al., 1974)
Enewetak, 6–15 m	5,000–9,500		Huston, 1985 (Highsmith, 1979)
Enewetak, 16–25 m	6,000		Huston, 1985 (Highsmith, 1979)
G. parvistella, 0–5 m	1,300–12,500		Huston, 1985 (Buddemeier et al., 1974)
Platygyra laminella, 6–15 m	6,700–8,000		Huston, 1985 (Buddemeier et al., 1974)
Reefs			
Coral, atolls, windward	14,000		Kukal, 1971
Atolls, lagoon reefs, leeward	3,800		Kukal, 1971
Range, <5 m depth	1,100–20,000		Schlager, 1981
Range, 10–20 m depth	500–2,000		Schlager, 1981
Range, atolls, outer reefs	330–910		Kukal, 1971
Atlantic reefs			
Alcaran Reef, Mexico, *Acropora cervicornis* reef	12,000	775	Macintyre et al., 1977
Alcaran Reef, Mexico, head-coral reefs	6,000	1,175	Macintyre et al., 1977
Grecian Rocks, Florida, windward	1,300	6,000	Shinn, 1980
Holocene, Miami, Florida	6,500	2,000	Lighty et al., 1978
Miami, Florida	740	4,900	Shinn, Hudson et al., 1977
Bal Harbor, Florida	380	6,300	Shinn, Hudson et al., 1977
Long Reef, Florida, windward	650	5,600	Shinn, Hudson et al., 1977
Carysfort Reef, Florida, windward	860–4,850	700–5,250	Shinn, Hudson et al., 1977
Big Pine Key, Florida, leeward	490–1,510	1,000–7,200	Shinn, Hudson et al., 1977
Dry Tortugas, Florida	1,910–4,470	130–6,000	Shinn, Hudson et al., 1977
St. Croix	15,200		Adey et al., 1977
St. Croix, algal ridge	6,000		Adey, 1977
Hess Channel, St. Croix	2,300	3,400	Adey et al., 1977
Galeta Point, Panama, *Acropora palmata* reef			
range	1,290–10,810	390–4,400	Macintyre & Glynn, 1976
average	3,900		

Area	Rate (B)[b]	Period (yr)[c]	Reference
Galeta Point, reef-flat rubble, windward	600	5,500	Macintyre & Glynn, 1976
Boo Bee patch reef, Belize lagoon, leeward	1,600	8,800	Halley et al., 1977
Pacific reefs			
Tarawa Atoll , windward	8,200	550	Marshall & Jacobson, 1985
Tarawa Atoll, windward	5,000–5,400	2,070	Marshall & Jacobson, 1985
Nishimezaki Reef, Ryukyus	3,900	7,400	Takahashi et al., 1988
Kikai-jima, Ryukyus	2,900–4,000	8,720	Takahashi et al., 1988 (Konishi et al., 1978)
Hanauma Reef, Hawaii	2,900	6,970	Takahashi et al., 1988 (Easton & Olson, 1976)
Great Barrier Reef, Holocene	200–600	9,000	Davies & Marshall, 1979
Great Barrier Reef, algal pavement	2,800	alkalinity meas.	Davies & Marshall, 1979
Great Barrier Reef, reef, flat, coral zone, windward	3,100	alkalinity, 24 h	Davies & Marshall, 1979
Great Barrier Reef, reef flat, windward	5,000	alkalinity, 24 h	Davies & Marshall, 1979
Great Barrier Reef, leeward margin, leeward	6,000	alkalinity, 24 h	Davies & Marshall, 1979
Carter Reef, Great Barrier Reef	260–2,200	320–1,940	Hopley, 1977
Orpheus Island, Great Barrier Reef	4,000	7,300	Takahashi et al., 1988 (Hopley & Barnes, 1985)
Northern Great Barrier Reef	67–100	15×10^6	Davies, 1988
Central Great Barrier Reef	60–75	4×10^6	Davies, 1988
Southern Great Barrier Reef			
(Heron Island, Wreck Reefs)	50–75	$2–3 \times 10^6$	Davies, 1988
Enewetak, Holocene	320	13,000	Saller, 1984
Chiriqui Gulf, Panama, reef flat, average	2,640	2,825	Glynn & Mcintyre, 1977
Chiriqui Gulf, Panama, reef flat, range	1,100–4,800	5,585–210	Glynn & Mcintyre, 1977
Panama Bay, Panama, reef flat	1,300	4,150	Glynn & Mcintyre, 1977
Other carbonate environments			
Andros Island, storm-tide layers	320,000–6,750,000	1.3–16.5 h	Hardie & Ginsburg, 1977
Stromatolites	1,460,000	24 h	Kukal, 1971
Bermuda stromatolites, subtidal	365,000	1–6 d	Gebelein, 1969
Algal ridge, St. Croix, windward	6,000		Adey, 1977
Calcareous algae	2,000–7,000		Kukal, 1971
Calcareous algae, summer	12,000	1 m	Kukal, 1971
Ooid shoals (range), windward	550–2,000	≈3,000	Schlager, 1981
Great Bahama Bank, leeward	800–1,100	2,500	Schindel, 1980 (Cloud, 1962)
Great Bahama Bank, Andros lobe, leeward	200–850	≈7,000	Enos, 1974
Little Bahama Bank	1,200	10,000	Sarg, 1988 (Hine et al., 1981)
Little Bahama Bank, Bight of Abaco (ave.), leeward	120	5,500	Neumann & Land, 1975
Little Bahama Bank, Bight of Abaco (core dates), leeward	200–300	≈1,000	Neumann & Land, 1975
Florida, forereef slope	710	10,000	Enos, 1977
Florida outer shelf margin, windward	490–1,010	6,000	Enos, 1977
Florida inner shelf margin	180–610	5,000	Enos, 1977
Florida inner shelf margin	220	10,200	Stockman et al., 1967
Rodriguez bank, Florida inner shelf margin	1,000	5,000	Wilson, 1975 (Turmel & Swanson, 1972)
Ceasar Creek Bank, Florida inner shelf margin	1,390	4,000	Warzeski, 1976
Florida Bay, mud banks, leeward	330	3,000	Stockman et al., 1967
Florida Bay, interbank, leeward	53	3,000	Stockman et al., 1967
Florida Bay, western mud bank, leeward	460	4,000	Schindel, 1980 (Scholl, 1966)
Florida Bay, western mud banks, leeward	620	4,000	Wanless & Tagett, 1989; K.K. Mukerji, unpub., 1987

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Shallow- water carbonates (cont.)			
Other carbonate environments (cont.)			
Florida Bay, Cross Bank, leeward	1,585	1,700	E.A. Shinn & P.R. Rose, unpub., 1963
Florida Bay, Crane Key, leeward	1,000	3,000	Wilson, 1975 (Stockman et al., 1967)
Florida Bay, Cape Sable	4,000	50	Gebelein, 1977
Belize lagoon, adjacent patch reef	400–500	8,800	Halley et al., 1977
Galeta Point, Panama, forereef talus	2,500	2,500	Macintyre & Glynn, 1976
Galeta Point, fore-reef pavement, windward	600	3,750	Macintyre & Glynn, 1976
Galeta Point, back reef, windward	670	3,000	Macintyre & Glynn, 1976
Northeast Yucatan, lagoon, leeward	1,000	5,000	Wilson, 1975 (Brady, 1971)
Alcaran Reef, Mexico, coral rubble & sand, windward	2,500	4,500	Macintyre et al., 1977
Alcaran Reef, Mexico, reef flat, windward	2,000	3,500	Macintyre et al., 1977
Gulf of Mexico (oyster reef)	730	2,100	Schindel, 1980 (Shepard & Moore, 1955)
Persian Gulf	5–50	10,000	Schindel, 1980 (Sarntheim, 1971)
Trucial Coast, Persian Gulf, lagoon, leeward	140	5,000	Schindel, 1980 (Kinsman, 1969)
Great Barrier Reef, sand flats	200	alkalinity meas.	Davies & Marshall, 1979
Great Barrier Reef, reticulated lagoon, leeward	1,000	alkalinity, 24 h	Davies & Marshall, 1979
Great Barrier Reef, deep lagoon	300	alkalinity, 24 h	Davies & Marshall, 1979
Chiriqui Gulf, Panama, forereef slope, ave.	7,500	430	Glynn & Macintyre, 1977
Chiriqui Gulf, Panama, forereef slope, range	500–20,800	1,065–130	Glynn & Macintyre, 1977
Tidal flats, range	400–950	≈3,000	Schlager, 1981
Cape Sable, FLorida, tidal flats	11,000	19 m	Gebelein, 1977
Cape Sable, Florida, tidal flats	2,000–5,500	50	Gebelein, 1977
Andros Island, Bahamas, tidal flats	490	5,000	Hardie & Ginsburg, 1977
Andros Island, Bahamas, NW tidal flats	700	2,200	Wilson, 1975 (Shinn et al., 1965)
Andros Island, Bahamas, SW tidal flats	800	≈5,000	Gebelein, 1975
Trucial Coast, Persian Gulf (intertidal)	400	5,000	Schindel, 1980 (Kinsman, 1969)
Sabkha, Trucial Coast, Persian Gulf	500	5,000	Wilson, 1975 (Kinsman, 1969)
Sabkha, Persian Gulf	29–94	3,000	Schindel, 1980 (Illing et al., 1965)
Sabkha Faishak, Persian Gulf	1,000	4,000	Wilson, 1975 (Illing et al., 1965)
Bathyal and abyssal deposits			
North Atlantic, Holocene, average	88.5	11,000	Ericson et al., 1961
North Atlantic, Hologene, range	5–636	11,000	Ericson et al., 1961
North Atlantic, L. Pleistocene, glacial, average	63	50,000	Ericson et al., 1961
North Atlantic, L. Pleistocene, glacial, range	10–>203	50,000	Ericson et al., 1961
California Borderlands, diatomaceous mud	880		Kukal, 1971
Yellow Sea, Sea of Japan, diatomaceous clay	50–200		Kukal, 1971
East India Sea, calcareous clay	850		Kukal, 1971
Atlantic and Pacific	10–150	≈100,000	Schindel, 1980 (Ku et al., 1968)
North Pacific Ocean	11	7×10^6	Schindel, 1980 (Dymond, 1966)
Barbados (airborne)	0.6	9 m	Schindel, 1980 (Delany et al., 1967)
Hemipelagic deposits			
Continental slope, "blue mud", average	17.8		Kukal, 1971
Atlantic Ocean average	50–100		Kukal, 1971
Canadian slope, Atlantic	30–300	10,000	Lisitzin, 1972
European basin, Atlantic	485	11,000	Lisitzin, 1972
Senegal continental slope, Atlantic	200–418	11,000	Lisitzin, 1972
North American basin, west Atlantic	90–480	11,000	Lisitzin, 1972
Cuban slope, Caribbean	245–300	11,000	Lisitzin, 1972
Argentine Basin, S. Atlantic	3–45	11,000	Lisitzin, 1972
Demerara & Ceara Rise, W. equatorial Atlantic	3–400	15,000	Scholle et al., 1983 (Damuth, 1977)
Demerara & Ceara abyssal plain	15–40		Lisitzin, 1972
Ceara abyssal plain, equatorial Atlantic	200		Berger, 1974 (Hayes et al., 1972)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Nares abyssal plain (turbidites)	25–30	10,000	Kuijpers et al., 1987
W. Greater Antilles Outer Ridge, N. Atlantic	200–300		Kuijpers et al., 1987
E. Greater Antilles Outer Ridge, N. Atlantic	30		Kuijpers et al., 1987
Baleric abyssal plain, Mediterranean	160–520	20,000	Rupke, 1975
North American rise, N. Atlantic	34–68	12,000	Schindel, 1980 (Emery et al., 1970)
North American abyssal plain, N. Atlantic	20	12,000	Schindel, 1980 (Emery et al., 1970)
Nares abyssal plain, N. Atlantic	2.5–10	10,000	Lisitzin, 1972
Sohm abyssal plain, N. Atlantic	20–360	10,000	Lisitzin, 1972
Cape Verde abyssal plain, mid-Atlantic	10–40	10,000	Lisitzin, 1972
Pernambuco abyssal plain, S. Atlantic	0.8	10,000	Lisitzin, 1972
Guinea abyssal plain, E. equatorial Atlantic	25–50	10,000	Lisitzin, 1972
Niger fan, E. equatorial Atlantic	25–65	10,000	Lisitzin, 1972
Angola abyssal plain, S. Atlantic	7.5–23	10,000	Lisitzin, 1972
Cape abyssal plain, S. Atlantic	1.5–12	10,000	Lisitzin, 1972
Bengal cone, Indian Ocean	64	10.2×10^6	Moore et al., 1974
North Pacific Ocean, blue mud	10		Kukal, 1971
Bering Straits, gray mud	80–4,500		Kukal, 1971
Bering Sea, basins	70–360	10,000	Lisitzin, 1972
California borderland	50–2,000		Berger, 1974 (Bandy, 1968)
California borderland	3–400	15,000	Scholle et al., 1983 (Prensky, 1973)
California borderland, ridges	50		Lisitzin, 1972 (Emery & Bray, 1962)
California borderland, proximal basin	1,800		Lisitzin, 1972 (Emery & Bray, 1962)
California borderland, distal basins	200–400		Lisitzin, 1972 (Emery & Bray, 1962)
California continental slope	80		Lisitzin, 1972
Kuril–Kamchatka trench	20–30	10,000	Lisitzin, 1972
Andean trench, E. Pacific	18–36	11,000	Schindel, 1980 (Lisitzin, 1972)
Antarctic slope, gray, silty clay	10–160		Kukal, 1971
Red clay			
Oceanic average	7–13	12,000	Schindel, 1980 (Kuenen, 1950)
North Pacific Ocean	1–2		Opdyke & Foster, 1971
North Pacific (muddy)	10–15		Opdyke & Foster, 1971
North Pacific Ocean	0.2–6	10,000	Lisitzin, 1972
Tropical North Pacific	0–1		Berger, 1974
Indian Ocean	0.5–4.6	100,000	Schindel, 1980 (Kuznetsov, 1969)
North and South Pacific	2		Van Andel et al., 1975
Nares abyssal plain, N. Atlantic	5–10		Kuijpers et al., 1987
Brown clay			
Oceanic average	2–10		Kukal, 1971
Nares abyssal plain, N. Atlantic	13–24	$10–24.8 \times 10^3$	Kuijpers et al., 1987
Pelagic carbonate			
Oceanic average, *Globigerina* ooze	10–80		Kukal, 1971
Oceanic average, *Globigerina* ooze	8–14	12,000	Schindel, 1980 (Kuenen, 1950)
Pacific Ocean, average	5.5	3×10^6	Davies & Worsley, 1981
Indian Ocean	10–40		Scholle et al., 1983 (Goldberg & Koide, 1963)
Indian Ocean average	11.9	3×10^6	Davies & Worsley, 1981
Atlantic Ocean, average	17.3	3×10^6	Davies & Worsley, 1981
Mid-Atlantic Ridge (crest)	1.7–185	11,000	Lisitzin, 1972
Mid-Atlantic Ocean	29	8,000	Schindel, 1980 (Nozaki et al., 1977)
North Atlantic Ocean	10–80		Scholle et al., 1983 (Ericson et al., 1961)
North Atlantic, Rockall Bank	25–75	10,000	Lisitzin, 1972
North Atlantic (40–50°N)	35–60		Berger, 1974 (Mcintyre et al., 1972)
North Atlantic (5–20°N)	14–40		Berger, 1974 (Schott, 1935)
Equatorial Atlantic Ocean	20–40		Berger, 1974 (Schott, 1935; Ericson et al., 1956)
South Atlantic Ocean	20–50		Scholle et al., 1983 (Ericson et al., 1961)
South Atlantic, Brazilian slope	30–50	10,000	Lisitzin, 1972
Caribbean Sea	24	116,000	Schindel, 1980 (Broecker & van Donk, 1970)

Table 1 (cont.)

Area	Rate (B)[b]	Period (yr)[c]	Reference
Bathyal and abyssal deposits (cont.)			
Pelagic carbonates (cont.)			
Caribbean Sea	12		Kukal, 1971
Caribbean Sea	20–110		Lisitzin, 1972
Caribbean Sea	10–60		Scholle et al., 1983 (Prell & Hay, 1976)
Caribbean Sea	28		Berger, 1974 (Emiliani, 1966)
Atlantic, Mediterranean	20–100		Kukal, 1971
Menorca Rise, Mediterranean, Quaternary nanofossil marls	108	1.8×10^6	Hsü, Montadert et al., 1978
Black Sea, nannofossil ooze	100–300	3,000	Schindel, 1980 (Stoffers et al., 1978)
Mediterranean, pelagic ooze	50	10,000	Cita et al., 1978
Panama Basin, eastern equatorial Pacific	9–100		Scholle et al., 1983 (Swift, 1977)
Equatorial Pacific Ocean	10–25	10,000	Lisitzin, 1972
Western equatorial Pacific Ocean	11–50		Scholle et al., 1983 (Berger et al., 1978
Equatorial Pacific Ocean	5–18	10^6	Berger, 1974 (Hays et al., 1969)
Central equatorial Pacific Ocean	10–20	10^6	van Andel et al., 1975
Eastern equatorial Pacific Ocean	30		Berger, 1974 (Blackman, 1966)
East Pacific Rise (0–20°S)	20–40		Berger, 1974 (Blackman, 1966)
East Pacific Rise (30°S)	3–10		Berger, 1974 (Blackman, 1966)
East Pacific Rise (40–50°S)	10–60		Berger, 1974 (Blackman, 1966)
Northwest Providence Channel, Bahamas	15–22	$183–93 \times 10^2$	Boardman and Neumann, 1984
Northwest Providence Channel, periplatform ooze	69–43	1,400–6,550	Boardman and Neumann, 1984
Biogenic siliceous sediments			
Oceanic average, radiolarian ooze	5		Kukal, 1971
North & equatorial Atlantic Ocean	2–7		Berger, 1974 (Turekian, 1965)
South Atlantic Ocean	3–18	10,000	Lisitzin, 1972
South Atlantic Ocean	2–3		Berger, 1974 (Maxwell et al., 1970)
Pacific Ocean, diatom ooze	5–50		Kukal, 1971
Equatorial Pacific, siliceous ooze	4–5	10^6	van Andel et al., 1975
Equatorial Pacific, siliceous ooze	2–5		Berger, 1974
Equatorial Pacific Ocean	2–25	10,000	Lisitzin, 1972
Antarctic Ocean, radiolarian ooze	11–140		Scholle et al., 1983 (Hays, 1965)
Antarctic Ocean	0.7–32	10,000	Lisitzin, 1972
Indian Ocean, diatom ooze	5–20	100,000	Schindel, 1980 (Kuznetsov, 1969)
Gulf of California, varved diatomites	4,700–5,400		Lowe, 1976 (Calvert, 1966)
Vancouver Island fiord, varved diatomaceous sediment	4,000		Lowe, 1976 (Gross et al., 1963)
Freshwater diatomites (average)	300–1,000		Kukal, 1971

a. General format of the table is after Schindel (1980), as are many of the data. Other major sources are Kukal (1971), who does not generally indicate his sources, methods, or duration of observation, and Lisitzin (1972). The reference given in parentheses is the primary data source; those not listed among the references may be found from the secondary source cited.

b. Derived from many different types of measurement and from observations spanning vastly different time intervals. No corrections have been made for compaction. Longer periods of observation include some compaction, as well as more lacuna, than observations of shorter duration, a point emphasized by Schindel (1980). Rates are in Bubnoff ($mm/10^3$ yr = $m/10^6$ yr).

c. Time interval of observation is years, unless indicated otherwise.

*Thickness of individual flood deposits are reported as yearly rates on the assumption of annual flooding. These are more reasonable figures for modeling than calculated "instantaneous" sedimentation rates; moreover, the actual duration of flooding is rarely reported. It must be noted that most studies of floods are of exceptional events rather than of typical annual floods. For this reason some of the multiyear averages reported may be the most reasonable for simulations.

sediment is input from a point source, such as a river mouth, or a line source, such as a carbonate platform margin, or is uniformly distributed, as in pelagic sedimentation. A related consideration is that of throughput. As illustrated by a deltaic environment, some of the sediment input from the river accumulates in various subenvironments of the delta, but some is redistributed as hemipelagic or turbidite input to deeper environments. Menard (1961) provides some insight into apportionment to different environments in his analysis of diverse drainage basins (table 2).

In the carbonate realm sedimentation input from outside sources is typically minor or negligible compared with in situ production. The production rate is thus of prime importance. Carbonate production rates vary so greatly in magnitude and in response to various controlling factors that for some purposes it is desirable to model these responses. Lerche et al. (1987) explored the impact of some major controls on carbonate production rates and modeled their influence on the configuration of carbonate bodies. They introduced depth- and distance-dependent functions for food supply, light ("photosynthetically active radiation"), temperature, salinity, and oxygen concentration. These variables illustrate that, in general, climate can affect production rates, in addition to the character, of carbonate sediments profoundly. There are exceptions to the generalization that carbonate sediments are tropical (Teichert, 1958; Milliman, 1975, p. 204; Lees, 1975; Leonard et al., 1981; Rao, 1981), but data on these production rates are lacking. Carbonate production rates are generally greater in windward settings than in leeward ones, producing an inherent asymmetry in carbonate platforms. The rather sparse data to substantiate this (see table 1, Shallow-Water Carbonates) suggest that rates differ by factors of 2–4. Only the individual coral growth rates and carbonate fixation estimated from alkalinity measurements are truly production rates; other values are accumulation rates in a strict sense, although they may approximate production rates. It is normally more expedient to use accumulation rates; most of the data are in these terms (table 1, Shallow-Water Carbonates, Bathyal and Abyssal Deposits), and accumulation constitutes the sedimentary record (table 3).

Production rates exclude transported sediment, whereas this sediment is inherently incorporated in the accumulation rate. This difference leads to examination of the well-established dogma in carbonate sedimentology that most sediment is produced in situ and that lateral transport is minor or negligible [cf. Wilson (1975, p. 7)]. Scale must again be considered. Lateral transport may be appreciable on the scale of a reef, the nearest approximation of a point source in most carbonate realms. It is generally considered negligible on the scale of a basin or a platform, but it is clear that significant transport is necessary for lateral progradation of the slopes of carbonate platforms (Bosellini, 1984; Playford et al., 1989; Eberli and Ginsburg, 1989). Transport must likewise control sedimentation in tidal flats where in situ production is negligible. Periplatform carbonate ooze (Schlager and James, 1978), an important component of highstand accumulations in proximal portions of basins (Boardman and Neumann, 1984; Droxler and Schlager, 1985; Shinn, Steinen et al., 1989), demonstrates lateral transport of carbonates in suspension. The possibility of some lateral transport must therefore be considered to realistically model carbonate accumulation in two or three dimensions [cf. Spencer and Demicco (1989)].

One-dimensional models, focused on simulation of sequences by vertical accretion, generally ignore lateral transport (Read et al., 1986). This essentially denies the possibility of autocyclic sequences that are controlled by lateral progradation of sediment (Ginsburg, 1971). Current two-dimensional models generally deal with progradation in carbonates in essentially the same way as progradation in terrigenous clastics is treated (Demicco and Spencer, 1989; Lawrence et al., 1990; Bosence and Waltham, 1990). Accumulation produces vertical aggradation until the available space is filled; surplus sediment is then redistributed into adjoining areas.

More data exist for pelagic sedimentation rates in both carbonate and noncarbonate sediments than for any other environment, in part because of the Deep Sea Drilling Project, which calculates accumulation rates for each datable sedimentary interval. Time spans are typically a few million years. Only a reasonable sampling of these data is presented in tables 1 and 3. Impetus to systematically glean rates from the 100-plus volumes of the Deep Sea Drilling Project is reduced by the fact that pelagic sedimentation rates are generally the lowest encountered and the most stable. In some settings, however, basinal sedimentation rates may be of prime importance. Harris (1989) demonstrated the influence of basinal accumulation rates on progradation of Middle Triassic platform margins in the Dolomites of northern Italy. Progradation of platform margins is typically in response to increased shallow-water production rates or reduced accommodation space, but increases in basinal accumulation rates, especially through shifts to siliciclastics, volcanoclastics, or evaporites, also can dramatically increase progradation rates (Harris, 1989).

Mixed carbonate and terrigenous environments are not yet fully integrated into most models, even those capable of dealing with either terrigenous or carbonate sources [cf. Lawrence et al. (1990)]. Several new considerations are introduced. The cumulative sedimentation from both sources influences the overall sedimentation rate. When some threshold in terrigenous input is reached, carbonate productivity and therefore accumulation rate are apparently suppressed (Mount, 1984; Walker et al., 1983). Neither the threshold nor the rate of suppression can be quantitatively defined at present.

It is likely that the type of impinging terrigenous sediment must be considered in addition to its volume. Organisms can probably tolerate accumulation of sand and coarser sediments better than they can tolerate mud. Sand creates un-

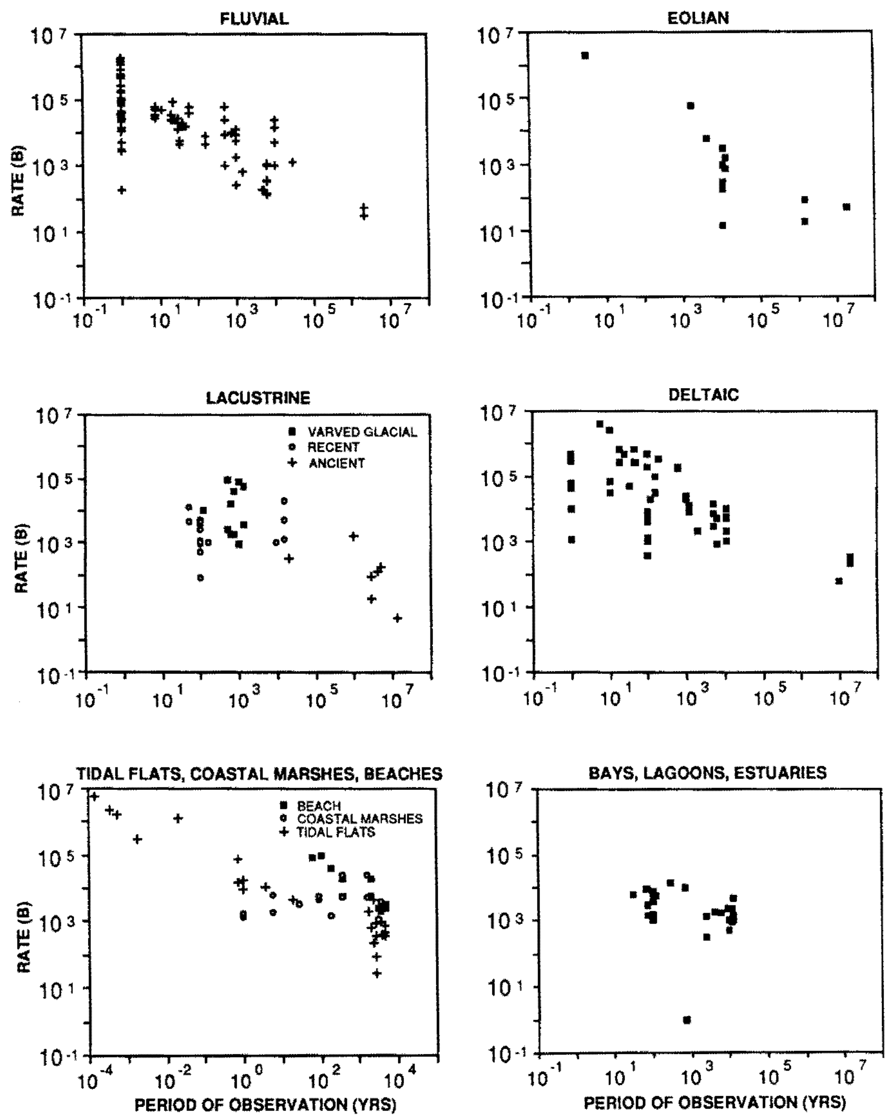

Figure 1. Rates of sedimentation versus period of observation. The strong inverse relationship on most plots illustrates the gaps in the geologic record, the "long periods of boredom and short periods of terror" (Ager, 1981, p. 107). High "instantaneous" rates, for example, those generated by floods on floodplains and in deltas, are moderated by extended hiatuses. Note the contrast with more stable abyssal-plain environments. Even lakes show an inverse trend when ancient lacustrine environments are considered. Unfortunately, there seem to be no instantaneous rates on turbidity current deposition. Longer periods of observation include some apparent rate reduction because of compaction. No corrections for compaction have been made. Units of sedimentation rate are Bubnoffs (1 B = 1 mm/10^3 yr = 1 m/10^6 yr). Plots are logarithmic. Data from table 1. Expanded from Schindel (1980, fig. 1).

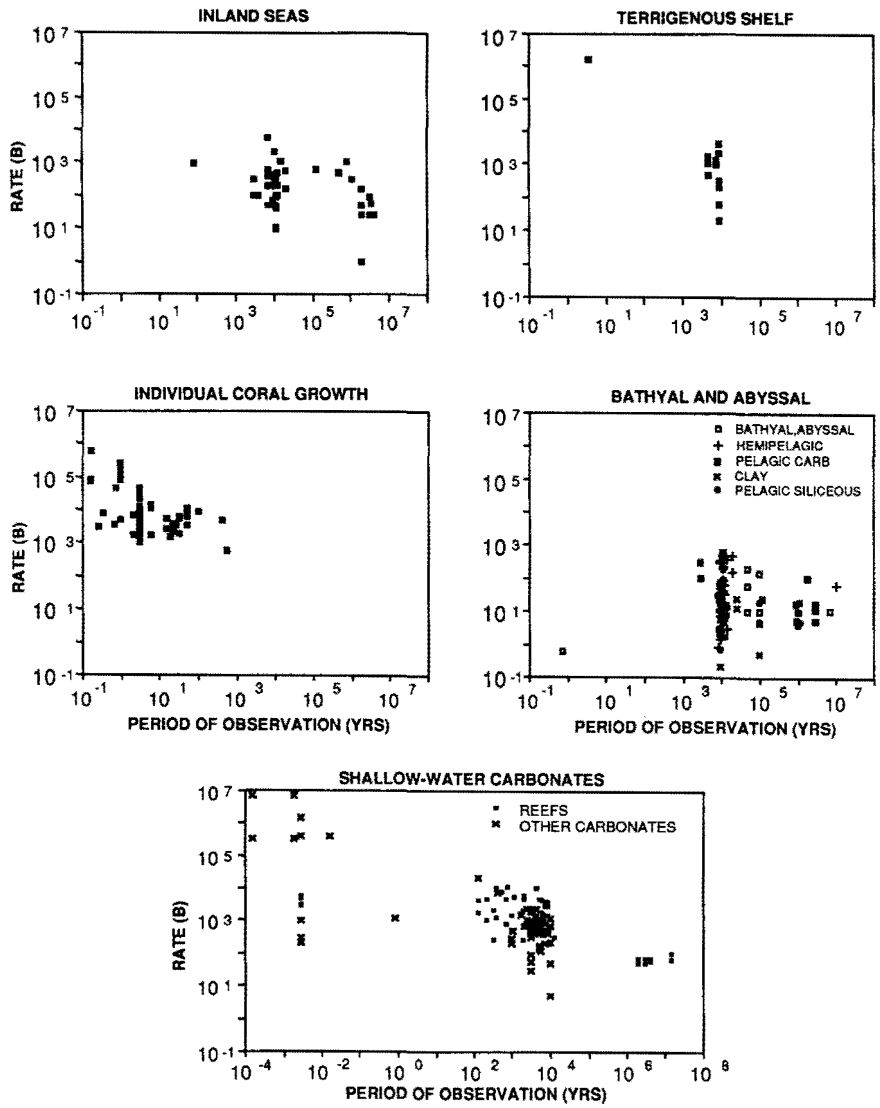

Figure 1 (cont.)

Table 2. Partitioning of sediment by depositional environment

Source Area	Volume of Sediment (10^6 km^3)	Depositional Sites (% Volume) Continental & Shelf	Continental Rise	Abyssal Plain
Appalachian Mountains	7.8	29	54	17
Mississippi drainage basin	11.1	81	11	8
Himalaya Mountains	8.5	49	1	50

Data from Menard (1961, p. 159).

Table 3. Sedimentation rates of "chemical" rocks in the geologic record

Age/Location	Rate (B)[a]	Period (m.y.)[b]	Reference[c]
Platform carbonates			
Late Cambrian Whipple Cave Fm., Nevada	60	≈9	Cook & Taylor, 1977
Late Cambrian tidal flats, Appalachians	25	18	Laporte, 1971
Late Cambrian subtidal, Appalachians	34	18	Laporte, 1971
Early Ordovician Ellenburger Group, Texas	15	27	Sarg, 1988 (Loucks & Anderson, 1980)
Early Ordovician Arbuckle Group, Oklahoma	110	27	after Wilson, 1975
Silurian pinnacle reefs, Michigan	13	14	Sarg, 1988 (Mesolella et al., 1974)
Late Silurian, Appalachians	100	≈6	Laporte, 1971
Late Silurian, Midcontinent, USA	25	≈6	Laporte, 1971
Early Devonian (Gedinnian) Helderberg Group, New York	15	7	Laporte, 1971
Middle Devonian Keg River platform, Alberta, Canada	14	11	Sarg, 1988 (Schmidt et al., 1980)
Late Devonian Swan Hills, Alberta, Canada	122	1	Sarg, 1988
Devonian (Givetian–Famenian), Canning basin	30	20	Schlager, 1981 (Playford & Lowrie, 1966)
Mississippian (Kinderbook–Meramec), Rocky Mountains	50–80	15	Schlager, 1981 (Rose, 1976)
Mississippian (Osage), Indiana	15	2	Brown et al., 1990
Mississippian (Osage), Indiana, tidal flats	350,000	1×10^{-6}	Brown et al., 1990
Mississippian (Meramec–Chester), Rocky Mountains	100–150	8	Schlager, 1981 (Rose, 1976)
Pennsylvanian–Permian Nansen Fm. Sverdrup basin, Canada	37	52	Davies, 1977
Early Permian Wichita Fm., Texas	50	11	Sarg, 1988 (Silver & Todd, 1969)
Early Permian (Longyinian), Yangtze platform China	33–135	7	Enos, 1992
Early Permian (Qixian), Yangtze platform China	5–150	6	Enos, 1992
Early Permian (Maokouan), Yangtze platform China	3–67	15	Enos, 1992
Late Permian (Longtan/Changxing), Yangtze platform, China	7–110	15	Enos, 1992
Permian Clear Fork Fm., Texas	365	1	Sarg, 1988 (Sarg & Lehmann, 1986)
Permian Grayburg Fm., Delaware basin, USA	160	1	Sarg, 1988 (Sarg & Lehmann, 1986)
Permian Capitan Fm., Delaware basin, USA	75	3	Schlager, 1981 (Harms, 1974)
Permian Capitan reef, Delaware basin, USA	55–83	9	Sarg, 1988 (Silver & Todd, 1969)
Permian San Andres Fm., Delaware basin	180	1	Sarg, 1988 (Sarg & Lehmann, 1986)
Triassic (late Anisian–Ladinian) Northern Calcareous Alps	100	7	Schlager, 1981 (Ott, 1967)
Triassic (Early Carnian) Dolomites	300–500	4	Schlager, 1981
Late Triassic, Tethys (Alps, Apennines)	100		Bernoulli, 1972
Early Jurassic, Tethys (Alps, Apennines)	15–40	≈20	Bernoulli, 1972
Jurassic Haynesville Fm., Texas	95	2	Sarg, 1988
Late Jurassic Smackover Fm., Arkansas	83	4	Sarg, 1988
LateJurassic Friuli platform, southern Alps	30–45	20	Schlager, 1981 (Winterer & Bosellini, 1981)
Early Cretaceous Shuaiba, Middle East	155	1	Sarg, 1988
Mid-Cretaceous (Albanian–Cenomanian) Golden	100	15	Enos, 1977
Lane, Mexico	80	15	Wilson, 1975 (Coogan et al., 1972)
Cretaceous–Cenozoic, Andros well, Bahamas	35	120	Wilson, 1975 (Goodell & Garman, 1969)
Cretaceous–Cenozoic, Sunniland field, Florida	30	120	Wilson, 1975
Mesozoic–Cenozoic, Persian Gulf (maximum)	30	200	Wilson, 1975
Late Eocene, Enewetak	170	3.4	Saller, 1984
Early Miocene, Enewetak	76	7.1	Saller, 1984
Late Miocene Terumbu Fm., S. China Sea	80–286	0.8–5.2	Sarg, 1988 (Rudolph & Lehmann, 1987)
Quaternary, Enewetak	11.5	0.59	Saller, 1984
Middle Miocene–Holocene, northern Great Barrier Reef	67–100	15	Davies, 1988

Age/Location	Rate (B)[a]	Period (m.y.)[b]	Reference[c]
Pelagic and deep-water carbonates			
Late Cambrian Hales Lst. (lower) Nevada	14	≈9	Cook and Taylor, 1977
Late Cambrian Frederick Lst., Maryland	50	16	Reinhardt, 1977
Late Pennsylvanian–Early Permian Hare Fiord Fm. (lower), Sverdrup basin, Canada	16	28	Davies, 1977
Late Cretaceous Marne a Fucoidi (argillaceous)	15	≈15	Bernoulli, 1972
Early Jurassic (Pliensbachian), High Atlas, Morocco	63–100	5–8	Evans & Kendall, 1977
Early Jurassic Monte Sant'Angelo Lst., Apennines (including platform debris)	13	20	Bernoulli, 1972
Early Jurassic, Corniola Fm., Apennines, & Sihiais Lst, Greece, (pelagic & calc. turbidites)	15–25	≈20	Bernoulli, 1972
Middle Jurassic Lamellibranch Lst, Apennines, S. Alps, Greece	3–8	19	Bernoulli, 1972
Late Jurassic Oberalm Beds, Austrian Alps	17–51	5–15	Garrison & Fischer, 1969
Late Jurassic Cat Gap Fm., N. Atlantic	8–14	16	Jansa et al., 1979
Late Jurassic, Early Cret. Maiolica, Apennines, s. Alps	10	≈23	Bernoulli, 1972
Late Cretaceous Chalk, UK			
range	3–60	32	Scholle et al., 1983 (Hancock, 1975)
average	15	32	
Late Cretaceous Chalk, Danish trough, North Sea	100		Scholle et al., 1983
Late Cretaceous, Tongue of the Ocean	8	≈30	Bernoulli, 1972
Cretaceous, Italy			
range	7–50		Scholle et al., 1983 (Arthur, 1979)
average	12		
Late Cretaceous chalks, Western Interior, USA			
range	6.5–50		Scholle et al., 1983 (Kauffman, 1977)
average	35		
Early and middle Miocene, nannofossil marls, Menorca Rise, Mediterranean	103	7	Hsü, Montadert et al., 1978
Miocene Great Abaco Fm, N. Atlantic (intraclastic debris)	9–43	4.6–6.3	Jansa et al., 1979
DSDP cores, to site 335, 3 my averages	0.6–17	3	Davies and Worsley, 1981
Pacific Oceanic Plateaus			
Aptian–Quaternary, Ontong Java Plateau	11.1	113	Jenkyns, 1978 (Moberly & Larsen, 1975)
Berriasian–Quaternary, Magellan Rise	8.9	131	Jenkyns, 1978 (Moberly & Larsen, 1975)
Barremian–Quaternary, Manihiki Plateau	7.8	116	Jenkyns, 1978 (Moberly & Larsen, 1975)
Berriasian–Quaternary, Shatsky Rise	4.9	131	Jenkyns, 1978 (Moberly & Larsen, 1975)
Cenomanian–Quaternary, Hess Rise	3.6	96	Jenkyns, 1978 (Moberly & Larsen, 1975)
Condensed Sequences			
Late Devonian Cephalopodenkalk, Germany	1.5–2	14	Tucker, 1974
Late Devonian Griotte, France	≈7	7	Tucker, 1974
Late Triassic Hallstatt Lst., Austrian Alps	0.5–1.5	20	Garrison & Fischer, 1969
Early Jurassic Adnet Beds, Austrian Alps	0.6–1.0	15–25	Garrison & Fischer, 1969
Early–Middle Jurassic, Ammonitico Rosso, Apennines, s. Alps, Greece	2.5–6.5		Bernoulli, 1972
Early Pliocene nannofossil marls, Cretan Basin, Mediterranean	9	2.4	Hsü, Montadert et al., 1978
Pliocene nannofossil marls, Menorca Rise, Mediterranean	16	3.4	Hsü, Montadert et al., 1978
Pleistocene, Mediterranean	1	1.9	Cita et al., 1978
Siliceous rocks			
Silurian–Mississippian or Devonian Caballos Novaculite, Texas	0.3–4.5	105–48	Folk & McBride, 1976
Devonian Arkansas Novaculite, Arkansas and Oklahoma (varved)			
range	1,000–2,500	0.1	Lowe, 1976
average	1,250	0.1	

Table 3 (cont.)

Age/Location	Rate (B)[a]	Period (m.y.)[b]	Reference[c]
Siliceous rocks (cont.)			
Jurassic Radiolarite, Apennines	3–9		Schlager, 1974
Middle Jurassic Ruhpolding Radiolarite, Austria	0.7–1	20–30	Garrison & Fischer, 1969
Tithonian–Barremian Radiolarite Group, s. Alps	5.4	16	Bernoulli, 1972
Tithonian–Barremian Scisti ad Aptici, Apennines	5.8	16	Bernoulli, 1972
Tithonian–Barremian U. Posidonia Beds, Greece	3.1	16	Bernoulli, 1972
Eocene Bermuda Rise Fm, N. Atlantic, chert and siliceous mudstone	5–8	≈10	Jansa et al., 1979
Miocene Monterey Fm, California, diatomite	8–200		Scholle et al., 1983 (Garrison & Douglas, 1981)
Miscellaneous pelagic rocks			
Anhydrite, L. Permian Castile Fm, W. Texas–New Mexico	1,825	0.3	Dunham, 1972 (Udden, 1924)
Carbonaceous clays, Early Cretaceous Hatteras Fm., N. Atlantic	3–19	15–25	Jansa et al., 1979
Variegated clays, Late Cretaceous, N. Atlantic	1–3	27	Jansa et al., 1979
Hemipelagic mud, Eocene–Pleistocene Blake Ridge Fm., N. Atlantic	3–200	1.7–2.5	Jansa et al., 1979
Hemipelagic mud, Pleistocene, Nares Abyssal Plain, N. Atlantic	400–500	≈0.05	Kuijpers et al., 1987
Evaporites			
Late Silurian, Salina Group, Michigan basin	180	≈6	Alling & Briggs, 1961
Late Silurian, Salina Group, Appalachians	150	≈6	Alling & Briggs, 1961
Late Permian Castile Anhydrite, Texas–New Mexico (varved)	1,825	0.3	Dunham, 1972 (Udden, 1924)
Messinian, Sicily	160	1.2	Decima & Wezel, 1973
Messinian, DSDP Site 124, Baleric basin	67	1.2	Decima & Wezel, 1973
Messinian, DSDP Site 132, Tyrrhenian Sea	30	1.2	Decima & Wezel, 1973

a. Bubnoff, 1 B = 1 mm/10^3 yr = 1 m/10^6 yr.

b. Time intervals for stratigraphic units of Mesozoic and Cenozoic age are from Haq et al. (1987). Paleozoic intervals are from Palmer (1983).

c. Not all primary sources are listed in references; see secondary source for original reference.

stable, shifting substrates; it probably has little direct impact on the organisms' metabolism. Finer suspended sediment, however, has the more direct influence of fouling the feeding mechanisms of many carbonate-producing organisms or of smothering them (Ginsburg and Shinn, 1964; Wilson, 1975, pp. 1–3), although it may also produce fluid substrates inimical to epifauna. It is nevertheless probable that the tolerance of carbonate organisms to mud is higher than generally recognized. Many carbonate rocks include a high percentage of mud, carbonate or terrigenous, that was introduced over a long period of time. Some carbonate producers were excluded, but others survived or even flourished, and carbonate sedimentation continued [cf. Laporte (1969, p. 115)]. There is no indication that the inhibiting effects of terrigenous mud are any different from those of carbonate mud; carbonate-producing organisms cannot be expected to be mineralogists. Terrigenous mud influxes can, however, include land-derived excess nutrients that could further suppress carbonate productivity (Hallock and Schlager, 1986, p. 394). Productivity suppression from sediments or pollutants introduced by human activity (Weiss and Goddard, 1977; Smith et al., 1981) offers the best possibilities for quantification, an example of Nietzschean serendipity.

Lag time

Inundation of carbonate platforms during transgression apparently does not lead to the immediate onset of rapid production of carbonate sediment. Stated another way, carbonate production does not reach its full potential for a finite period (Schlager, 1981). Carbonate sediment accumulation therefore tends to lag the relative rate of sea-level rise, resulting in a deepening sequence (Read et al., 1986). The interval between initial inundation and onset of rapid sediment accumulation is the lag time. The formation of shoaling-upward platform cycles so common in the geologic record requires a lag time, according to current concepts of sedimentation (Read et al., 1986; Ginsburg, 1971). Otherwise, rapid carbonate sedimentation would maintain the sediment surface at sea level, and accumulation rates less than the relative rate of sea-level rise would form a continuously deepening sequence. Lag time is also essential to Ginsburg's (1971) autogenic cycles in which carbonate sediment builds up to sea level and progrades toward the platform edge, reducing the area of carbonate production until progradation ceases. To produce a transgression and begin a new cycle rather than maintain a steady-state aggradation, sediment accumulation must drop appreciably below the relative rate of subsidence for a finite period, the lag time.

Lag time has not been considered in terrigenous siliciclastic cycles because the capacity for the in situ production is lacking; sediment input does not necessarily change with submergence. It has been shown by analysis and by simulation that asymmetric shoaling-upward cycles can be produced by symmetric (e.g., sine wave) eustatic oscillations in sea level superposed on constant subsidence and sedimentation rates [cf. Jervey (1988)]. The rate of subsidence plus sea-level fall must exceed the rate of sediment supply near the inflection point of the sea-level curve, the point of maximum rate of fall. Such cycles could also be produced in carbonate sedimentation, of course, if the rate of sediment production were less than the combined maximum rates of sea-level fall and subsidence. Such solutions appear rather contrived, given the demonstrable rapid rates of carbonate production in shallow water (tables 1, Shallow-Water Carbonates; table 3, Platform Carbonates; Schlager, 1981). Moreover, the resulting cycles should invariably terminate with subaerial exposure of the upper part of the cycle and should commonly show a deepening portion of the cycle. Such elements are not rare in shoaling-upward platform cycles, but they do not appear to be the general case.

Although lag time has a profound effect on the character of simulated shoaling-upward cycles [cf. Read et al. (1986, p. 108), Goldhammer et al. (1987), and Koerschner and Read (1989)], processes that may cause sediment accumulation to temporarily lag subsidence are not recognized. It is clear that aggradation of sediment into the supratidal zone terminates carbonate production because of subaerial exposure. It is not clear why sediment production does not recommence immediately upon submergence. One possibility is that extremely shallow water results in temperature fluctuations or periodic exposure that inhibits carbonate production. The well-established increase in carbonate production rates with decreasing depth (fig. 2) may have an upper limit somewhat below sea level.

Another possibility is that lag time has a physical basis that reflects lack of accumulation above some profile of equilibrium, such as the wave base (Enos, 1989). Accumulation of modern carbonate sediments in south Florida is confined below a threshold depth that appears to be a function of fetch, suggesting that fairweather or storm wave base is the control. Critical depths are approximately 3 m (10 ft) in the open shelf margin of south Florida and 2 m (7 ft) in the protected inner shelf of Florida Bay. In the more open Atlantic setting of Antigua, West Indies, the threshold appears to be approximately 5 m (16 ft) (Weiss and Multer, 1988).

If either control, decreased productivity or wave base, is valid, the appropriate parameter is lag depth rather than lag time. The corresponding time is determined by the rate of change in relative sea level, which is a function of eustasy, subsidence, compaction, and accumulation rate. Goldhammer et al. (1987) used a constant lag depth of 1 m (3 ft) in simulations of shoaling-upward cycles in the Middle Triassic of the Dolomites. In contrast, Read et al. (1986) assigned various durations to lag time and observed how this influenced the cycles generated. It is obviously desirable that the lag parameter be an empirical input rather than an unknown. If the lag parameter is closely related to water depth, then the threshold will not be reached simultaneously across a sloping surface. Deeper areas would begin accumulating sediment while shoal areas still lie above the threshold depth. If wave energy is the ultimate control, then sheltered areas would have a shallower lag depth than more exposed areas.

In summary, appropriate lag depth or time cannot be satisfactorily specified at present. Specification will be possible only when the processes are better understood. If the threshold is energy related, it should become possible to make reasonable assumptions if enough is known about the regional setting. It would also be clear what parameters must be studied in modern environments to obtain better definition for modeling.

Accommodation space

Accommodation space (or accommodation potential) is the increment of room available for sediment accumulation. The upper limit of the space is the level above which net erosion will occur. In nearshore settings this is generally taken as sea level, although a profile of equilibrium with a basinward slope, a well-established concept in subaerial settings, probably also applies below sea level (Enos, 1977, pp. 106–107). The lower limit is the depositional interface, so the instantaneous accommodation space is approximately water depth. Because this increment varies with sediment accumulation, some arbitrary fixed datum, such as basement, is used to define accommodation space. Accommodation space then increases in response to eustatic rises in sea level, subsidence, and erosion. For some purposes these components can be lumped together in a single accommodation parameter. In general, it is essential to isolate the components and thereby illustrate their influence on accumulation patterns.

Sea-level fluctuations The changes in sea level used for simulations are based on simple mathematical models or empirical sea-level curves. Some short-term simulations use a constant sea level or a uniform rate of change. This is realistic only for time spans less that 10^3–10^4 years; in fact, small-scale high-frequency oscillations may have durations

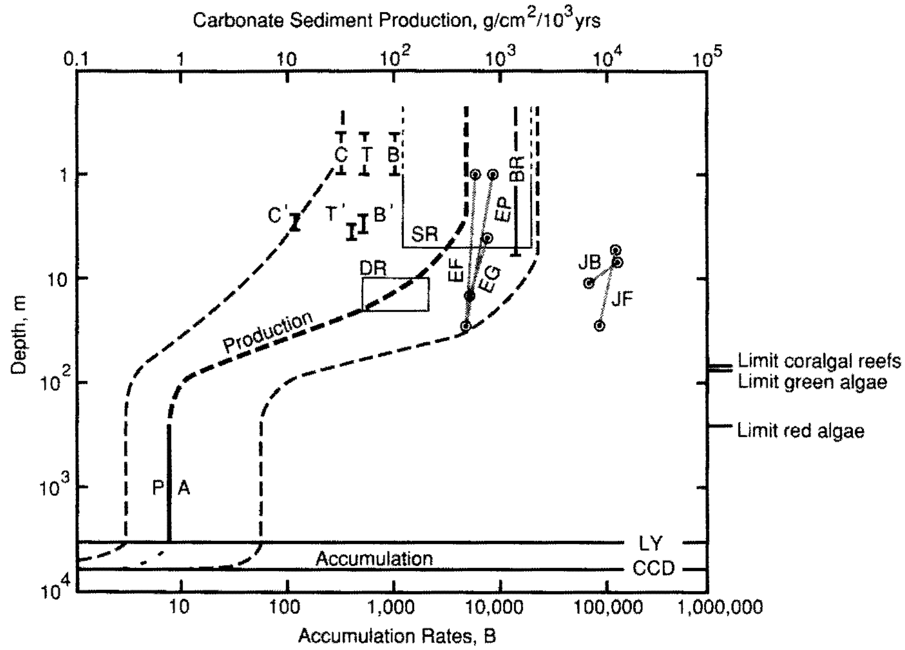

Figure 2. Carbonate sediment production as a function of depth. This curve is a hybrid, as were the original attempts to illustrate depth-related variations (Garrison and Fischer, 1969, fig. 22; Wilson, 1975, fig. I–2); those curves have nevertheless proved fertile. This attempt to quantify the curve encountered a paucity of data on in situ production, especially at intermediate depths. These are needed to document the thresholds related to algal productivity, stressed by R. N. Ginsburg [cf. Wilson (1975) and Schlager (1981)], and the lower limit of coral-algal reef growth (≈70 m; James, 1977). In contrast, data on sedimentation rates, as opposed to production rates, in shallow-water and abyssal settings are abundant (table 1). Data used in constraining the curve are: BR, productivity of a Barbados reef in a sheltered setting (Stearn et al., 1977). SR, compilation of accretion rates (mm/10^3 yr) of reefs at less than 5 m (15 ft) depths (Schlager, 1981). DR, compilation of accretion rates of reefs at 10–20 m (30–65 ft) depth (Schlager, 1981). EF, EG, and EP, regressions of growth rate versus depth for three species of corals in Enewetak Atoll (Highsmith, 1979). JF and JB, regressions of growth rate versus depth of *Acropora cervicornis* on the forereef and backreef, respectively, in Jamaica (Tunnicliffe, 1983); these are linear growth rates of a branching coral, so they do not constrain accretion rates, but they should help define the variation with depth. C, T, and B, total productivity of carbonate mudbanks with varying degrees of restriction in south Florida [Upper Cross Bank, Tavernier Key, and Buchanan Banks, respectively (Bosence, 1988, 1989)], paired with productivity from adjacent interbank areas (C′, T′, B′). PA, accumulation rates of pelagic carbonates from the equatorial Pacific; average for last million years and range for past 45 m.y. (van Andel et al., 1975). LY, the lysocline, a threshold of accelerated carbonate dissolution with depth, currently about 3,500–4,800 m (11,000–16,000 ft) (Scholle et al., 1983). CCD, the carbonate compensation depth, which ranges from <4,000 m (<13,000 ft) to >5,000 m (>16,000 ft) in the present oceans (Berger and Winterer, 1974). Benthic production is assumed to approach 0 with the disappearance of red algae at depths of approximately 250 m (800 ft). Pelagic production is a function of surface conditions, so it is shown constant with depth. Pelagic accumulation rates are affected primarily by dissolution on the seafloor. They decrease sharply below the lysocline and drop to 0 at the carbonate compensation depth. The conversion from production rates in grams per square centimeter per 1,000 years to sedimentation rates in Bubnoffs (mm/10^3 yr) varies with the porosity and mineralogy of the sediment; for example, 1 g/cm^2/10^3 yr would equate to a sedimentation rate of 14.7 B for calcitic sediment with 75% porosity (mud), but only 8.5 B of aragonitic sediment with 40% porosity (sand). The conversion used, 10 B = 1 g/cm^2/10^3 yr, would apply to aragonitic sediment with about 65% porosity.

of only a few hundred years (Dominquez et al., 1987). Sine or cosine functions of various amplitudes and wavelengths, which may be superposed on other functions, are also used (Bice, 1988; Jervey, 1988). Empirical sea-level functions include extrapolations of Quaternary sea-level curves (Watney et al., 1989), inferred from the coupling of glacial ice volume with $\delta^{18}O$ content of pelagic foraminifers, reflecting sea-surface temperatures (Matthews, 1984). These data have been extended through the Cenozoic (Prentice and Matthews, 1988; Matthews, 1984). For longer-term simulations sea-level curves derived by sequence stratigraphy (Vail et al., 1977; Haq et al., 1987) have been used, although these also have their critics (Miall, 1986; Burton et al., 1987; Matthews, 1988; Christie–Blick et al., 1988; Gradstein et al., 1988).

Curiously, no one seems to have gone back to the roots and directly applied the complex wave functions resulting from the periodicities in orbital parameters (Milankovich, 1941) to pre-Pleistocene fluctuations in sea level. This would seem particularly appropriate in view of the current overwhelming acceptance of the Milankovich band of orbital fluctuations as a cause of short-term changes in sea level, as inferred from their postulated control on climate (Milankovich, 1941) and thereby on the waxing and waning of glaciers (Denton and Hughes, 1983).

Subsidence Subsidence also has at least three components: tectonics, isostasy, and compaction. Some models simply assume a constant rate of subsidence at a given point, which translates into a linear boundary (constant slope) in two dimensions [cf. Jervey (1988)]. This simple model essentially incorporates all three components.

Tectonic subsidence Commonly used functions for tectonic subsidence are based on crustal stretching and cooling (McKenzie, 1978; Steckler and Watts, 1978; Sclater et al., 1980; Watts et al., 1982). These functions accurately model the subsidence of passive continental margins as they move away from spreading centers. Subsidence rate depends on the spreading rate and time or distance from the spreading center. Such functions also have been applied to continental-interior and cratonic basins and to oceanic islands and trenches (Watts et al., 1982). A different model for the tectonic component is provided by the theory of flexure of an elastic plate (Turcotte and Schubert, 1982, pp. 125ff). This model, which predicts differential subsidence (including local uplift) across a basin, seems most appropriate to foreland basins subjected to rapid tectonic and sediment loading from adjacent mountain belts.

Differential warping is generally superposed on the stately submergence of passive margins, expressed as local arches and basins. The Cape Hatteras and Cape Fear arches and the Baltimore Canyon basin of the Atlantic margin of North America are examples. Warping, local faulting, and regional subsidence can be empirically adjusted from burial history, which gives temporal changes in depth to basement. On active margins these three factors are typically the dominant components of subsidence. This empirical approach incorporates a second component of subsidence, isostatic adjustment.

Isostatic subsidence The crust subsides isostatically in response to sediment loading. The amount of subsidence can be approximated as the thickness of mantle material with a density of approximately 3.2 g/cm^3 that would be displaced by an increment of unconsolidated sediment with the appropriate density. For short spans of time, $<10^4$ years, it may be necessary to consider the viscous delay in response to loading.

Isostatic response to loading by water is not normally considered in current modeling programs. It is unlikely to be important in areas that are submerged to depths greater than the increment of sea-level rise so that loading would be uniform. In shallow-water areas, however, water loading is differential and may become a significant parameter as resolution improves through more sophisticated modeling and improved parameter definition.

Compaction The magnitude of compaction-induced subsidence can be three-quarters of the total thickness of muddy sediments because initial porosities are 70–80% (Hedberg, 1936; Rieke and Chilingarian, 1974; Enos and Sawatsky, 1981). Compaction-related subsidence can vary drastically with grain size and with degree of grain support, cementation, and sorting within sediments of the same composition. Variations related to composition are generally much less. Terrigenous muds are probably the most homogeneous sediment type in their compactional behavior; they generally show an exponential decline in porosity with depth. Absolute values vary appreciably among different basins, however (fig. 3). Far fewer empirical data are available on sandstones. The most comprehensive data set is that of G. I. Atwater and E. E. Miller (Blatt, 1979); these data indicate a linear rather than an exponential decrease in porosity with depth (fig. 4).

Compaction of carbonates has generated considerable controversy (Weller, 1959; Pray, 1960; Shinn, Halley et al., 1977; Shinn and Robbin, 1983). For pelagic carbonate muds a tremendous volume of data is available from the porosity versus depth curves generated by the Deep Sea Drilling Project. Available summaries are selective and out of date (fig. 5). The most comprehensive data set for shallow-water carbonates (fig. 6) is that of Schmoker and Halley (1982). Their results do not differentiate porosity loss by cementation from that by physical compaction. This point is important for modeling because changes in bulk volume, not in porosity, determine subsidence. Cement from external sources reduces porosity without reducing bulk volume; thus porosity curves converted to bulk volume loss may exaggerate compaction. This also applies to muddy carbonate rocks and to siliciclastics, although imported cement is probably less common. Conversely, secondary porosity produced by dis-

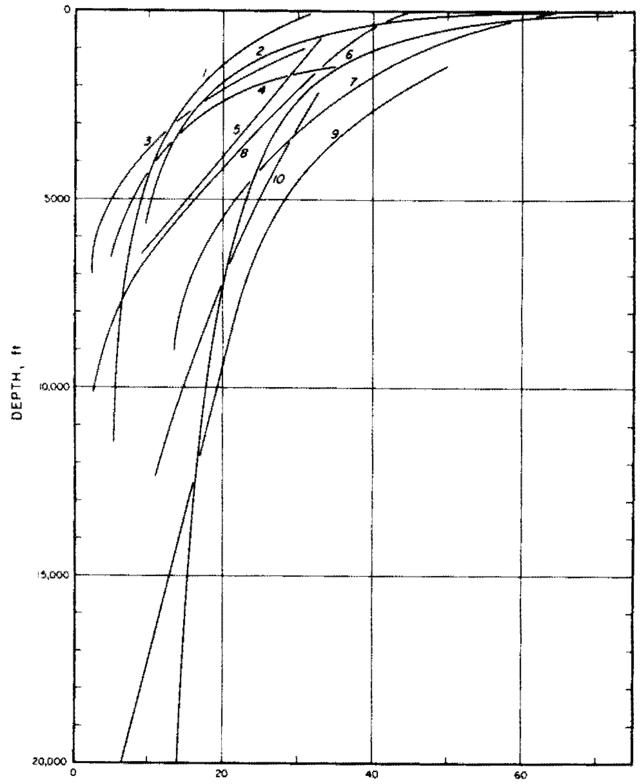

Figure 3. Representative fitted curves of porosity versus depth in mudrock. Locations: (1) Ciscaucasus, USSR. (2) Compilation, Tertiary and Quaternary. (3) Oklahoma, USA, Pennsylvanian and Permian. (4) Japan, Tertiary. (5) Venezuela, Tertiary. (6) Gulf Coast, USA, Tertiary. (7) Japan. (8) Compilation, mainly of locations 3 and 5. (9) Gulf Coast, USA, Tertiary. (10) Gulf Coast, USA, Tertiary. From Rieke and Chilingarian (1974, p. 42); references are given therein.

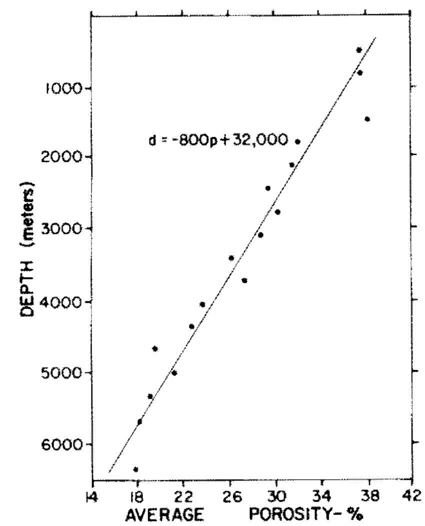

Figure 4. Porosity versus depth of late Tertiary sandstones from Louisiana, US Gulf Coast. The 17,367 analyses have been averaged in thousand-foot intervals for calculation of the least-squares fit. Unfortunately, no error estimates or petrology are available. Unpublished data of G. I. Atwater and E. E. Miller (1965); from Blatt (1979, p. 146).

solution at depth (Schmidt and McDonald, 1979) might lead to underestimates of compaction. Reef carbonates are typically considered noncompactible in simulations and burial history routines, but even in these lithologies considerable compaction may ensue through pressure solution at depth [Mossop (1972) documents 13% compaction; Anderson and Franseen (1991) document 30%].

Compaction curves for mixed terrigenous and carbonate sediments have not generally been isolated, although data are certainly available from the Deep Sea Drilling Project, where carbonate content is measured for the same intervals as porosity. Compaction curves of muddy terrigenous and carbonate sediments (figs. 3 and 5) generally do not differ sufficiently to preclude the use of the curve for the dominant sediment type or, preferably, a weighted average of the two curves for mixed sediments. A further consideration in the compaction of mixed sediments, not incorporated into existing models, is the possible effect on pressure solution. Conventional wisdom is that terrigenous clay in excess of approximately 10% enhances the susceptibility of carbonate sediments to pressure solution (Wanless, 1979; Bathurst, 1975; Dunnington, 1967). Documentation is inadequate and the mechanisms are not understood, but good empirical evidence that terrigenous clays do enhance pressure solution has been provided by McNeice (1987). It would be possible to incorporate this threshold in modeling, but there are scant quantitative data on pressure solution versus depth and even less on the relationship between pressure solution and terrigenous content of sediments.

In rapidly accumulating muddy sediments the possible generation of overpressure and the consequent retardation of compaction should be considered (Bradley, 1975; Plumley, 1980; Carstens and Dypvik, 1981; Shi and Wang, 1986). Overpressure resulting from sedimentation in excess of the rate at which pore fluids can escape would be amenable to modeling (Mudford and Best, 1989).

Erosion Two environments of erosion must be considered in sedimentary modeling. Subaerial erosion contributes to sediment input and modifies the final configuration of the eroded area. Submarine erosion involves redistribution of sediment and corresponding changes in the final configuration.

Rates of subaerial erosion are complex functions of such factors as elevation, lithology, climate, and vegetation. The first two parameters may evolve from the simulation, and the last two, largely independent of parameters being modeled, are either ignored or indirectly specified by user-selected

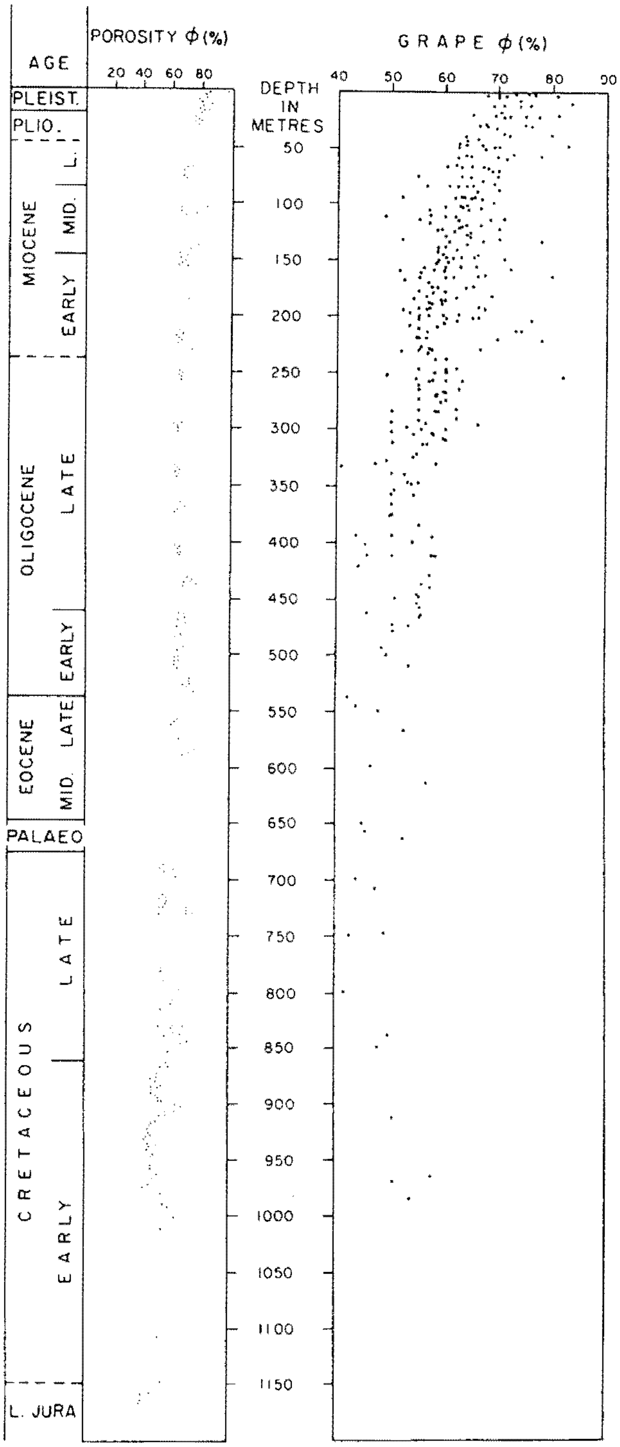

Figure 5. Porosity and depth in pelagic carbonates from the Pacific and Indian oceans. The left column is from Deep Sea Drilling Project (DSDP) site 167, Magellan Rise, central North Pacific. The right column has samples with 80% or more calcium carbonate from 11 other DSDP sites. Note that the ages apply only to site 167 (left-hand column). From Schlanger and Douglas (1974, p. 119).

values. If the factors causing erosion are not a focus of the study, the typically complex interrelations of erosion are best treated by using empirical net rates of erosion. Estimates of erosion rates in various climates, terrains, and lithologies (tables 4–8) are provided by Corbel (1959a,b), Menard (1961), Ritter (1967), and Meybeck (1976). An empirical relationship potentially useful in modeling was derived by Ahnert (1970):

$$d = 0.1535h, \tag{1}$$

where d is the rate of denudation in Bubnoff units (mm/10^3 yr) and h is relief in meters. The relationship proposed by Schumm (1963) for relatively small drainage basins [<1,500 mi^2 (4,000 km^2)] in semi-arid areas of the western United States may be more convenient in modeling:

$$\log D = 26.9H - 1.7, \tag{2}$$

where D is the denudation rate in feet per 1,000 years (305 B) and H is the ratio of relief to the length of a drainage basin. The relief to length ratio is dimensionless and can be represented by the slope of a surface tilted above sea level in a simulation.

Erosion in subaerially exposed carbonate sediments is typically ignored. For short time spans (perhaps 10^4–10^5 yr) this is probably justified, because the initial stage of carbonate erosion is typically dissolution, which produces secondary porosity, not a lowering of the land surface. Over more extended periods, however, the collapse of larger pore spaces (e.g., caverns) can result in significant lowering of the landscape, which may be differential and partially predictable (Purdy, 1974). Most of the data on erosion in carbonate terrains are values for net transport of ions by rivers. With simplifying assumptions, these can be expressed as changes in elevation in the landscape (table 9).

The erosion of subaerially exposed terrigenous terrains has different implications for modeling than erosion of carbonate terrains in both the erosional and depositional realms. Carbonate material is removed primarily by dissolution and does not directly produce new sediment (Bosence and Waltham, 1990). Eroded terrigenous material must be redistributed and adjusted for any changes in bulk density between the original material and the unconsolidated sediment.

Submarine erosion by waves or by slumping and sliding can be significant. Some modeling programs test for slope stability after each increment of sediment is added and remove material where the simulated slopes are in excess of what is considered stable (Lawrence et al., 1990), a value commonly specified by the user. Marine engineering practice provides some empirical data (Roberts et al., 1980). Angles of large-scale slopes in carbonates and siliciclastics have been summarized by Schlager and Camber (1986); slopes of carbonate platforms tend to be much steeper and to covary with height (fig. 7). Few data are available on submarine

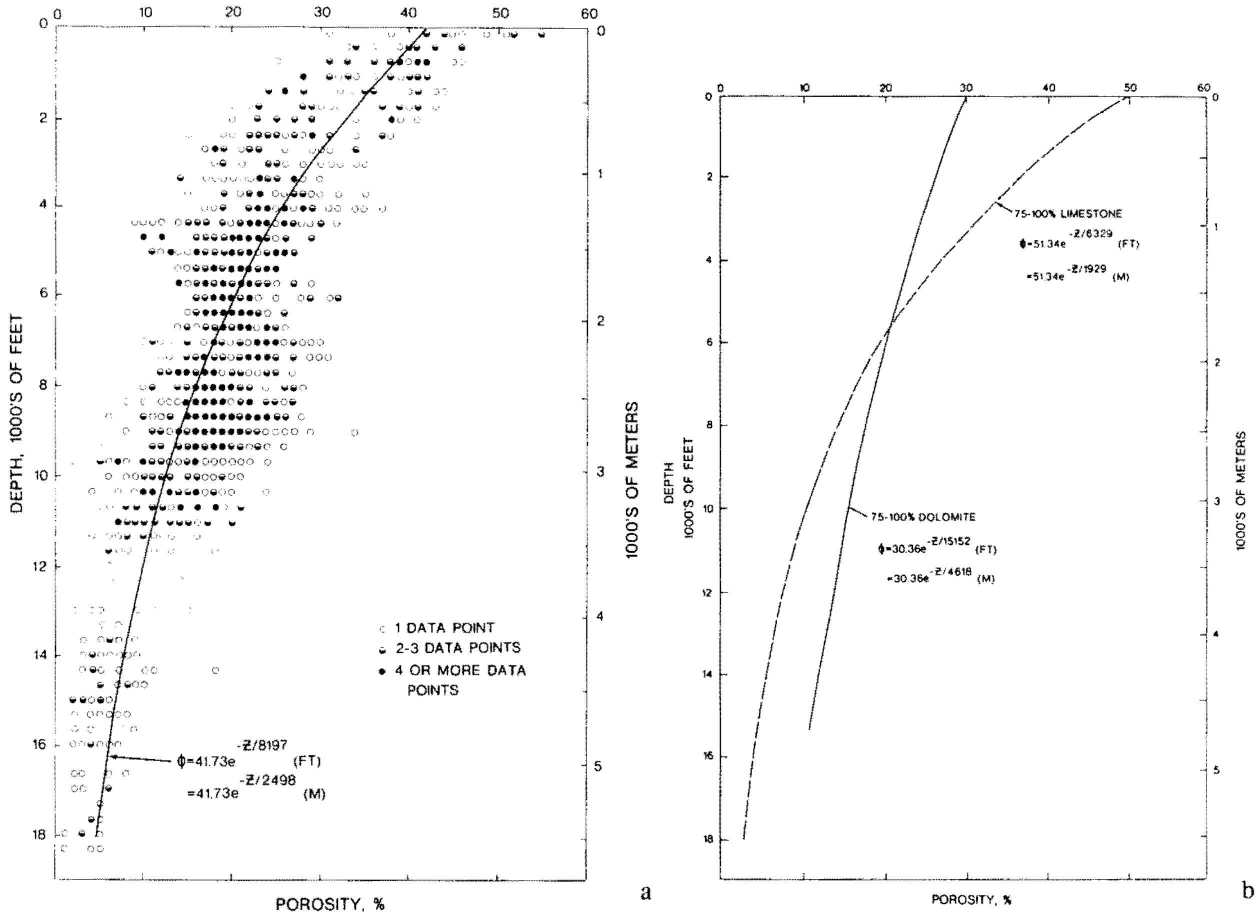

Figure 6. (a) Porosity versus depth from shallow-water carbonates, Pleistocene to Early Cretaceous in age, from south Florida. $N = 1{,}302$; 167 data points are gravity data from boreholes at shallow depths; the remainder are data points from porosity logs. From Schmoker and Halley (1982, p. 2,566). (b) Data from part a separated into dolomites (75–100% dolomite; $N = 336$) and limestones (75–100% $CaCO_3$; $N = 489$). From Schmoker and Halley (1982, p. 2,569).

erosion rates by waves and other shear forces. Rates of intertidal erosion of carbonates (table 10) (Trudgill, 1985) are apparently the nearest approximations available.

Summary and conclusions

This review was undertaken to summarize optimum sources of input data for sedimentation simulations and to point out where additional data or refinement of concepts is needed from fieldwork. The trinity of parameters—accumulation rates, lag time, and accommodation space—are judged to be the most fundamental. Available data are summarized in the figures and tables. Urgent needs are an elucidation of lag time, quantification of compaction by pressure solution, and evaluation of the effects of siliciclastic influx on both carbonate sedimentation and pressure solution.

Acknowledgments Lynn Watney persuaded me to undertake this summary, and Evan Franseen shepherded the later stages of preparation. Hal Wanless made many perceptive criticisms that helped to improve the manuscript, and Tony Simo smoothed some rough edges. Randy Farr was a skilled and patient coach at computer drafting. Special thanks are due to Nuria Wells for processing countless drafts, including the horrendous table 1.

Table 4. Erosion rates by relief and climate[a]

Relief and Climate[b]	Rate of Chemical Erosion (B)	Total Rate of Erosion (B)
Lowlands		
Periglacial, permafrost (15/yr); Bear Lake, Canada	13	15
Continental, cold (28–75/yr); E. Canada, Sweden	17	19
Maritime, temperate (33/yr); NE Europe	24 ± 12	32 ± 11
Continental, temperate, Mississippi basin	15	59
Missouri basin (4.4/yr)	6	55
Mississippi basin (Ritter, 1967)	14	46.4
Mississippi basin, Quaternary (Menard, 1961)		42
Continental, arid (0.7/yr); New Mexico	1	12
Tropical desert, central Sahara		1?
Hot, seasonally wet and dry; Paraguay	11	32
Tropical, humid (38/yr); Congo basin	15	22
Mountains		
Periglacial (20/yr); Brooks Range, Alaska	12	300
Periglacial, humid (250/yr); Norway	325	580
Maritime, humid (482 cm); Juneau, Alaska	192	800
Maritime, cold-temperate (118/yr); glaciated Alps	99 ± 59	203 ± 181
Maritime, temperate; Swiss Alps (Blatt et al., 1980)		70–910
Alps, long-term (volume Rhone fan; Blatt et al., 1980)		400
Maritime, temperate; Mt. Ranier (Blatt et al., 1980)		3,000–8,000
Maritime, temperate; Appalachians (Menard, 1961)		8
Appalachians, southern (Blatt et al., 1980)		41
Long-term (Cenozoic detrital sed. vol.; Matthews, 1975)		27
Appalachians, northern (Blatt et al., 1980)		48
Long-term (Cenozoic detrital sed. vol.; Matthews, 1975		5
Mediterranean, high mountains (62/yr); Italy, France	78	449
Mediterranean, semi-arid (60/yr); Italy	40	100
Continental, temperate (90/yr); central Europe	50 ± 25	102 ± 61
Continental, Himalayas (Menard, 1961)		1,000
Himalayas, Kosi R. (100/yr)		1,145
Himalayas (Blatt et al., 1980)		720
Hot and humid (76/yr); Usumacinta, Mexico & Guatemala	30	92
Hot and arid (4–10/yr); southeast USA	7	200
Hot and arid (0.6/yr), Tunisia	8	130

After Corbel (1959b), except as noted. Corbel assumed rock densities of 2.5 g/cm^3 to convert weights of material transported to surface lowering.

a. Units of erosion are Bubnoff (mm/10^3 yr = m/10^6 yr).

b. Figures in parentheses are runoff (precipitation minus evapotranspiration) in centimeters per year.

Table 5. Regional erosion rates in the United States

Region	Mechanical (B)	Chemical (B)	Total Rate of Erosion (B)	Period (yrs)
North Atlantic	23.7	18.5 ± 2.2	42.2	4–10
South Atlantic & eastern Gulf of Mexico	16.8	15.9 ± 5.3	32.7	4–8
Western Gulf of Mexico	34.6	8.1 ± 5.3	42.7	6–9
Mississippi River basin	32.4	14.0 ± 0.6	46.4	12
Colorado River basin	142.8	5.9 ± 2.7	148.7	32
Pacific drainage in California	71.8	15.9 ± 5.5	87.7	3–13
Columbia River basin	15.0	16.5 ± 3.6	31.5	2–4

After Ritter (1967).

Suspended and dissolved loads only; bed loads not included. Assumed rock densities are 2.64 g/cm^3. Erosion rates given in Bubnoff (mm/10^3 yr = m/10^6 yr).

Table 6. Rates of erosion from major drainages of the world

River	Drainage Area (10^3 km^2)	Runoff (cm/yr)	Mechanical Erosion (B)	Chemical Erosion (B)	Total Erosion (B)
Amazon (Brazil)	6,300	88	39.9	17.6	47.5
Congo (Zaire)	4,000	31	5.0	4.4	9.4
Congo (Corbel, 1959b)*		38	6.3	14.4	20.7
Mississippi (USA)	3,267	17.8	36	15	51
Mississippi (Corbel, 1959b)*			42	14.2	56.2
Mississippi (Ritter, 1967)			32.4	14.0	46.4
Nile (Egypt)	3,000	2.8	14	2.2	16
Parana (Argentina)	2,800	20	15	7.6	23
Parana (Corbel, 1959b)*			20	10	30
Yenisey (Siberia, USSR)	2,600	21	1.9	11	13
Ob (Siberia, USSR)	2,500	15	2.4	7.6	10
Lena (Siberia, USSR)	2,430	21	2.4	14	16
Yangtze (PRC)	1,950	35	186	–	>186
Amur (Heilong) (Siberia)	1,850	19	5.2	4.1	9.3
Mackenzie (Canada)	1,800	17	25	15	39
Madeira (Brazil)	1,380	73	59	16	75
Hsi (Pearl) (PRC)	1,350	18	34	–	>34
Volga (Russia, USSR)	1,350	20	7.2	22	29
Zambesi (Mozambique)	1,340	17	28	4.4	33
Niger (Nigeria)	1,125	17	23	3.4	26
Murray (Australia)	1,070	2.1	11	3.1	14
St. Lawrence (Canada)	1,025	32.8	1.9	20	22
Orange (South Africa)	1,000	9.1	57	4.5	61
Ganges (Bangladesh)	975	37.8	203	30	233
Indus (Pakistan)	950	22	189	25	214
Orinoco (Venezuela)	950	100	34	20	54
Danube (Romania)	805	25.2	32	28	60
Mekong (Vietnam)	795	72.5	165	28	193
Negro (Brazil)	755	189	3.8	3.8	7.6
Huang (Yellow) (PRC)	745	10.7	814	–	>814
Columbia (USA)	670	37.5	16	20	36
Columbia (Ritter, 1967)			15.0	16.5	31.5
Kolyma (Siberia, USSR)	645	18.2	3.5	–	>3.5
Colorado (USA)	635	3.2	330	8.7	338
Colorado (Corbel, 1959b)*		4.4	207	10.7	217.8
Colorado (Ritter, 1967)			143	5.9	149
Chari (Chad)	600	6.9	2.5	1.7	4.1
Brahmaputra (Bangladesh)	580	104	519	49	568
Xingu (Brazil)	540	45	0.3	1.1	1.4
Tapajós (Brazil)	500	45	0.5	1.4	1.8
Dnieper (Ukraine, USSR)	500	10	0.8	8.3	9.2
Amu–Darya (Uzbekistan, USSR)	450	10	79	23	102
Irrawady (Burma)	430	98	265	–	>265
Don (Russia, USSR)	420	6.6	5.2	13	18
Tigris–Euphrates (Shatt El Arab) (Iraq)	410	14	95	16	110
Maranon (Peru)	407	85	95	34	129
Ucayali (Peru)	400	76	116	52	169
Uruguay (Uruguay)	350	45	15	8.7	24
Magdalena (Colombia)	240	98	379	44	423
Rhine (Corbel, 1959b)*	225	49	1.9	28.3	30.2

After Meybeck (1976) with some additions and comparison. Estimates are based on suspended and dissolved loads only; bed loads are not included, except as noted by asterisks. Converted from weights assuming rock densities of 2.64 g/cm^3 (Ritter, 1967). Units of erosion are Bubnoff (mm/10^3 yr = m/10^6 yr).

*Estimate includes bed load.

Table 7. Rates of erosion of rivers fed by glaciers

River	Rate of Erosion (B)
Hidden Glacier (Alaska; rapid advance)	30,000
Muir (Alaska)	5,000
Bosson (Chamonix, France)	1,800
Nant Blanc (French Alps)	1,600
Heilstuga (Norway)	1,400
Memurelven (Norway)	1,600
Auserfjötur (Iceland)	2,200
Jokullsá (Iceland)	2,200
Hoffelsjökull (Iceland)	3,200
Hofsjökull (Iceland)	1,800
Isortok (Greenland)	2,500
Saskatchewan (Canada)	2,000

From Corbel (1959b, p. 16). Includes estimates of bed load. Assumed rock densities are 2.5 g/cm^3. B = Bubnoff (mm/10^3 yr).

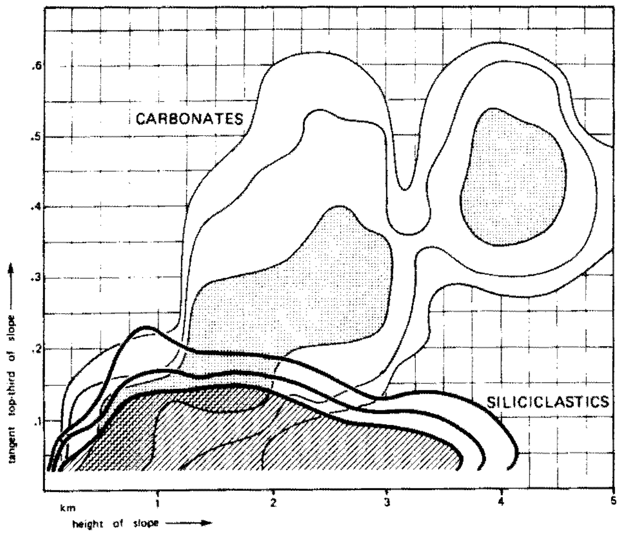

Figure 7. Slope angles on large slopes: angle of upper one-third of slope versus height of slope. Contours indicate concentrations of 0.5%, 1%, and 2% of total sample in unit area of 0.25 km × 0.05 tan *S*, measured as the moving average of 9 unit cells. Carbonate sample includes Bahamas and Marshall Islands (atolls) (N=413). Siliciclastic sample is based on Atlantic continental slopes (N = 72). Carbonate slopes steepen with height up to at least 5,000 m (15,000 ft). Siliciclastics follow this trend only to 500 m (1,500 ft); slope height above 500 m has no influence on steepness of siliciclastic slopes. Data of Schlager and Camber (1986); from Schlager (1988).

Table 8. Average erosion rates by continent

	Mechanical		Chemical		Total	
Continent	L/K[a]	G&M[b]	L/K	G&M	L/K	G&M
North America	27.9	33.0	15.0	13.0	42.8	46.0
South America	35.3	23.2	20.9	11.6	56.2	34.8
Asia	62.8	122.4	16.2	12.6	79.0	134.9
Africa	17.7	6.2	9.6	9.0	27.3	15.2
Europe	16.5	9.8	11.9	18.0	28.4	27.8
Australia	12.2	10.0	4.2	1.0	16.4	10.9
World total	36.8	53.1	14.1	11.4	50.9	64.5
World total (Ritter, 1967)	43–89		9.9		53–99	

Units are Bubnoff: 1 B = 1 mm/10^3 yr = 1 m/10^6 yr. Estimates in tons per square kilometer per year were converted using 2.64 g/cm^3 as the average density of crustal rocks (Ritter, 1967). Continental areas cited by Kukal (1971, p. 30) were used in calculations for uniformity between estimates.

a. Estimates from Lopatin (1952), in Kukal (1971, p. 30).

b. Estimates from Garrels and Mackenzie (1971, p. 120).

Table 9. Rates of chemical erosion of limestones

Climate/Location	Runoff (cm/yr)	Erosion Rate (B)
Arctic, dry		
Svalbard; Tanana, Alaska		40
Victoria Island, Canada		5
Somerset I, Canada (A & S)	10	2
Arctic, humid		
Gold Creek, SE Alaska		530
Capilano, British Columbia		420
Svartisen, northern Norway		400
Maritime, cold		
Vercors, French Alps		240
Lismore, Scotland		150
St. Casimir, Quebec		160
St. Thérésé, Quebec		120
NW England (Sweeting)		40
Maritime, temperate		
Derbyshire, England (A & S)		83
Fergus R., Ireland (A & S)		55
Mendip Hills, England (A & S)	82	63
Thames, England (Sweeting)		104
range		13–288
Lee, Essex, England (Sweeting)		63
range		23–155
Derwent, England (Sweeting)		66–197
Lesse, Belgium		27
Tamis, Yugoslavia		21
Alpine		
Triglav, Yugoslavia (A & S)	280	130
Tolminka, Yugoslavia (A & S)	310	102
Tatry Mtns., Czechoslovakia (A & S)	122	50
range	110–160	33–95
Jura Mtns., Switzerland (A & S)		98
Continental, cold winters		
Jasper, Alberta		40
Whitehorse, Yukon Terr., Canada		32
Fort Simpson, Mackenzie Terr., Canada		40
Gulf of Bothnia, Finland		30
Continental, temperate		
Kentucky (Sweeting)		64
range		3–297
Coolamon, NSW, Australia (A & S)	120	24
Krakow, Poland	25	20
Texas, U.S.A.	4	5
Mediterranean		
Postojna, Yugoslavia (A & S)	160	110
Bosna, Yugoslavia (A & S)	150	90
Trieste, Italy (A & S)	70	48
Senj, Yugoslavia (A & S)	46	28
Podovi, Yugoslavia (A & S)	25	15
Yugoslavia, karst mountains (humid)		60
Marseilles (dry)		10
South Algeria (arid)		6
Tropical, humid		
Jamaica (A & S)	105	73
range	55–135	40–96
Puerto Rico (A & S)	70	41
Florida (A & S)	50	33
Kissimmee, Florida (Sweeting)		27
range		16–63
Usumacinta R., Mexico & Guatemala, mountains		45
Champoton R., Yucatan, lowland		16

Sources: Corbel (1959b, p. 19; 1959a), Sweeting (1964), and Atkinson and Smith (A & S) (1976).
Units of erosion are Bubnoff: 1 B = 1 mm/10^3 yr = 1 m/10^6 yr.

Table 10. Marine erosion: Coastal erosion of limestones

Location	Substrate	Rate
Red Sea	Reef limestone	1,000
Barbados	Beach rock (grazers)	1,000–2,000
Bikini Atoll, Marshalls	Beachrock	300
Southwestern Australia	Beachrock	27,000–67,000
Heron Island, Great Barrier Reef	Beachrock	500
Aldabra Atoll, Indian Ocean	Reef limestone (intertidal)	500–4,000
Aldabra Atoll, Indian Ocean	Reef limestone (subaerial)	260
Aldabra Atoll, Indian Ocean	Reef limestone (with sand)	1,250
Aldabra Atoll, Indian Ocean	Reef limestone (no sand)	1,010

From Trudgill (1985, p. 159). Surface lowering in B (mm/10^3 yr)

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