CHAPTER FOUR

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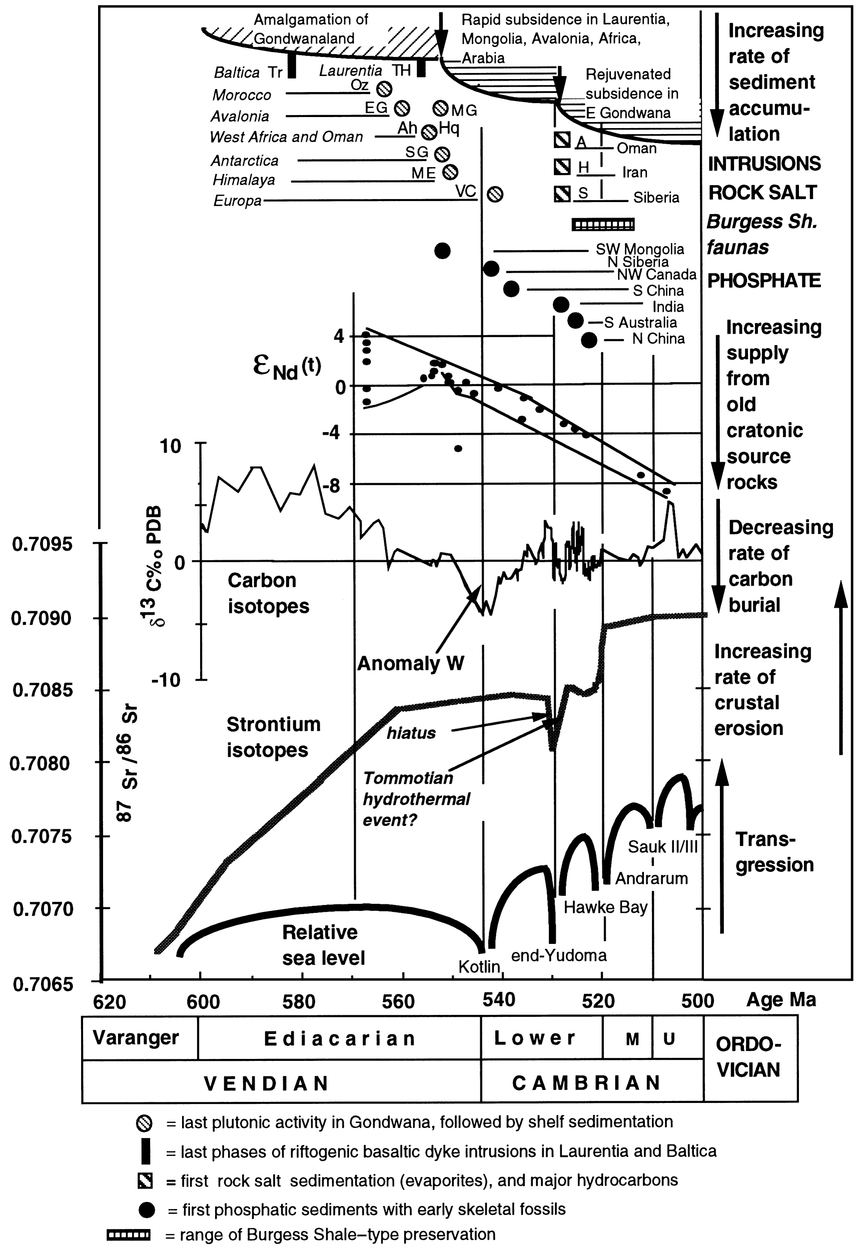
Did Supercontinental Amalgamation Trigger the "Cambrian Explosion"?

A global overview of sediment patterns and accumulation rates, and carbon, strontium, and neodymium isotopes confirms that increasing rates of subsidence and uplift accompanied the dramatic radiation of animal life through the Neoproterozoic-Cambrian interval (ca. 600 to 500 Ma). Peritidal carbonate platforms were drowned, to be followed in places by phosphorites and black shales, while thick evaporites accumulated in interior basins. This drowning of cratons during the latest Neoproterozoic-Cambrian could have brought about major taphonomic changes. The shoreward spread of oxygen-depleted and nutrient-enriched waters favored the preservation of thin skeletons by secondary phosphate and chert in peritidal carbonates and, later, the occurrence of Burgess Shale–type preservation in deeper-water shales. The burial of event sands in rapidly subsiding basins also allowed the paradoxical preservation of deep-water Nereites *ichnofacies in shallow-water sediments.*

THIS CHAPTER ATTEMPTS to put the "Cambrian explosion" into the wider context of events in the lithosphere. The formation and later rapid extensional subsidence of supercontinents in the Neoproterozoic have recently become apparent from a wide range of disciplines, including paleomagnetism, facies and fossil distributions, subsidence curves, and isotopic studies (e.g., Bond et al. 1984; Lindsay et al. 1987; Dalziel 1991; McKerrow et al. 1992; Derry et al. 1992, 1994). At some time before ca. 900 Ma B.P., Antarctica, Australia, Laurentia, Baltica, and Siberia appear to have been united in a Neoproterozoic supercontinent called Rodinia or Kanatia (Torsvik et al. 1996). It is possible that this may have begun to rift apart as early as 800 Ma (e.g., Lindsay and Korsch 1991; Lindsay and Leven 1996); certainly early rift successions can preserve deposits of the older, Rapitan-Sturtian glaciations (ca. 750–700 Ma; Young 1995). At some point after 725 Ma, the western margins of Laurentia and Antarctica-Australia were certainly separated and moving apart (Dalziel 1992a,b; Powell et al. 1993). By ca. 600–550 Ma, Laurentia, Baltica, and Siberia were also in



the process of rifting apart (McKerrow et al. 1992; Torsvik et al. 1996), and here the rift sequences may preserve deposits of the younger, Varangerian (or Marinoan) glaciations (ca. 620–590 Ma; e.g., Young 1995).

The assembly of another supercontinent, Gondwana, also took place during the Ediacarian to Early Cambrian interval. (*Ediacarian* is here used to indicate that period of the Late Neoproterozoic between the Marinoan glaciation at ca. 600 Ma and the base of the Cambrian at ca. 543 Ma). This involved the amalgamation of the separate

Figure 4.1 Isotopes, sea level, fossil taphonomy, and global tectonic changes during the Vendian-Cambrian interval. Basic dykes in Baltica and Laurentia indicate a final phase of rifting: *Tr* = Troms, Norway (582 ± 30 Ma; Torsvik et al. 1996); *TH* = Tibbit Hill, Quebec (554 Ma; Kumarapeli et al. 1989). Latest Pan-African plutonic events may indicate the final phases of amalgamation in West Gondwana: *EG* = Ercall Granophyre, England (560 ± 1 Ma, U/Pb zircon; Tucker and Pharaoh 1991); *Ah* = Ahaggar plutons, West Africa (556 ± 12 Ma, U/Pb zircon; Betrand-Sarfati et al. 1995); *Hq* = granite and ignimbrite below Huqf Group, Oman (556 ± 10 Ma, Rb/Sr; Burns et al. 1994); *ME* = granites from the Mount Everest region, Nepal, Himalaya (550 ± 16 Ma, Rb/Sr; Ferrara et al. 1983); *MG* = Marystown Group volcanics, southeastern Newfoundland (552 ± 3 Ma, U/Pb zircon; Myrow and Hiscott 1993); *Oz* = Ourzazate volcanics, Morocco (563 ± 2.5 Ma, U/Pb zircon; Odin et al. 1983); *SG* = postorogenic quartz syenite, Skelton Group, Antarctica (551 ± 4 Ma, U/Pb zircon; Rowell et al. 1993); *VC* = Vires-Carolles granite, Brioverian France (540 ± 10 Ma, U/Pb monazite; Dupret et al. 1990). Thick rock salt accumulated during rapid subsidence of extensional basins: *A* = Ara Salt Formation, Oman (Burns and Matter 1993; Loosveld et al. 1996); *H* = Hormuz Salt Formation, Iran (Brasier et al. 1990; Husseini and Husseini 1990). Burgess Shale–type faunas are confined to the medial Lower to Middle Cambrian (Butterfield 1996). Phosphatic sediments with early skeletal fossils first appear in the transition to more rapid subsidence and/or flooding of the platforms (sources cited in figures 4.2 and 4.3). ε_{Nd}(t) data recalculated from Thorogood 1990, using revised ages. The carbon isotope curve is composite, compiled from the Vendian of southwestern Mongolia (Brasier et al. 1996), Early to Middle Cambrian of Siberian Platform (Brasier et al. 1994), and Middle to Upper Cambrian of the Great Basin, USA (Brasier 1992b). The strontium isotope curve is based on least-altered samples (compiled from Burke et al. 1982; Keto and Jacobsen 1987; Donnelly et al. 1988, 1990; Derry et al. 1989, 1992, 1994; Narbonne et al. 1994; Nicholas 1994, 1996; Smith et al. 1994; Brasier et al. 1996). The sea level curve is based on data in Brasier 1980, 1982, and 1995; Notholt and Brasier 1986; Palmer 1981; and Bond et al. 1988.

crustal blocks of Avalonia, Europa, Arabia, Africa, Madagascar, South America, and Antarctica (together forming West Gondwana) and resulted in the compressional Pan-African orogeny, which culminated between ca. 560 and 530 Ma. Orogenic closure of the Pan-African compressional basins was accompanied in many places by igneous intrusions. In figure 4.1, we have plotted some of the youngest dated phases of igneous activity, as well as the riftogenic dyke swarms of Laurentia. Although geologic evidence indicates that East Gondwana (India, South China, North China, Australia) collided with West Gondwana along the Mozambique suture between ca. 600 and 550 Ma, recent paleomagnetic evidence has also suggested that final amalgamation did not take place until the Early Cambrian (Kirschvink 1992; Powell et al. 1993).

Pan-African amalgamation of Gondwana appears to have been accompanied by the widespread development of subsiding foreland basins, as documented in figures 4.1–4.3. Sediments of "rift cycle 1" (sensu Loosveld et al. 1996) begin with the Sturtian Ghadir Mangil glaciation in Arabia, dated to ca. 723 Ma (Brasier et al. 2000). The development of thick salts in the Ara Formation, once thought to be rift deposits of Tommotian age (Loosveld et al. 1996; Brasier et al. 1997), now appear to be foreland basin deposits of late Ediacarian age (Millson et al. 1996; Brasier et al. 2000).

Subductive margins were also developed along the borders of eastern Australia and Antarctica (e.g., Millar and Storey 1995; Chen and Liu 1996) and Mongolia (e.g., Şengör et al. 1993; but see also Ruzhentsev and Mossakovsky 1995) in the Early to

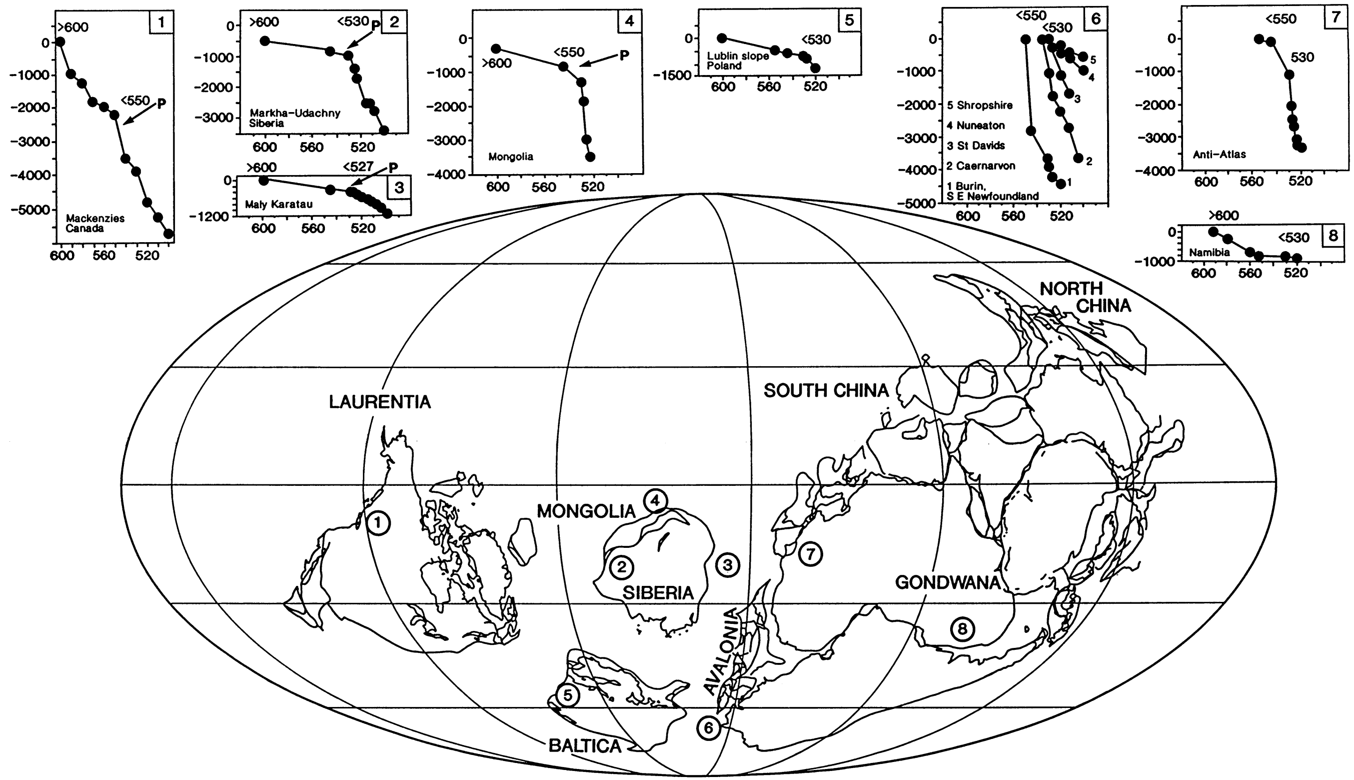


Figure 4.2 Sediment accumulation rates for the Ediacarian-Cambrian interval placed against latest Vendian to Nemakit-Daldynian paleogeography of McKerrow et al. (1992). Based on data in the following sources: Mackenzies, Canada (Narbonne and Aitken 1990); Markha-Udachny area of Siberian Platform (Astashkin et al. 1991); Maly Karatau in Kazakhstan (Cook et al. 1991); southwestern Mongolia (Brasier et al. 1996); Lublin Slope, Poland, Baltica (Moczydłowska 1991); Avalonian, Burin Peninsula, Newfoundland (Landing 1992): Caernarvon, North Wales; St. Davids, South Wales, Nuneaton, English Midlands; Shropshire, Welsh borderlands (Rushton 1974; Brasier 1989); Anti-Atlas, Morocco (Sdzuy and Geyer 1988); Namibia, southwestern Africa (Kaufman et al. 1994). The time scale is adapted from sources cited in Bowring et al. 1993, Tucker and McKerrow 1995, and Brasier 1995: base of Vendian = 610 Ma; Varangerian glacials = 610–600 Ma; base of Ediacarian = 600 Ma; main Ediacaran faunal interval = 580–555 Ma; late Ediacarian (Kotlin) interval = 555–545 Ma; base of Nemakit-Daldynian = base of Cambrian herein = 545 Ma; base of Tommotian = 530 Ma; base of Atdabanian = 528 Ma; base of Botoman = 526 Ma; base of Toyonian = 523 Ma; base of Middle Cambrian = 520 Ma; base of Upper Cambrian = 510 Ma; base of Ordovician = 500 Ma. *P* marks the first phosphatic sediments with early skeletal fossils. The numbers (e.g., <550) give the suggested timing of renewed rift/drift in millions of years ago (Ma).

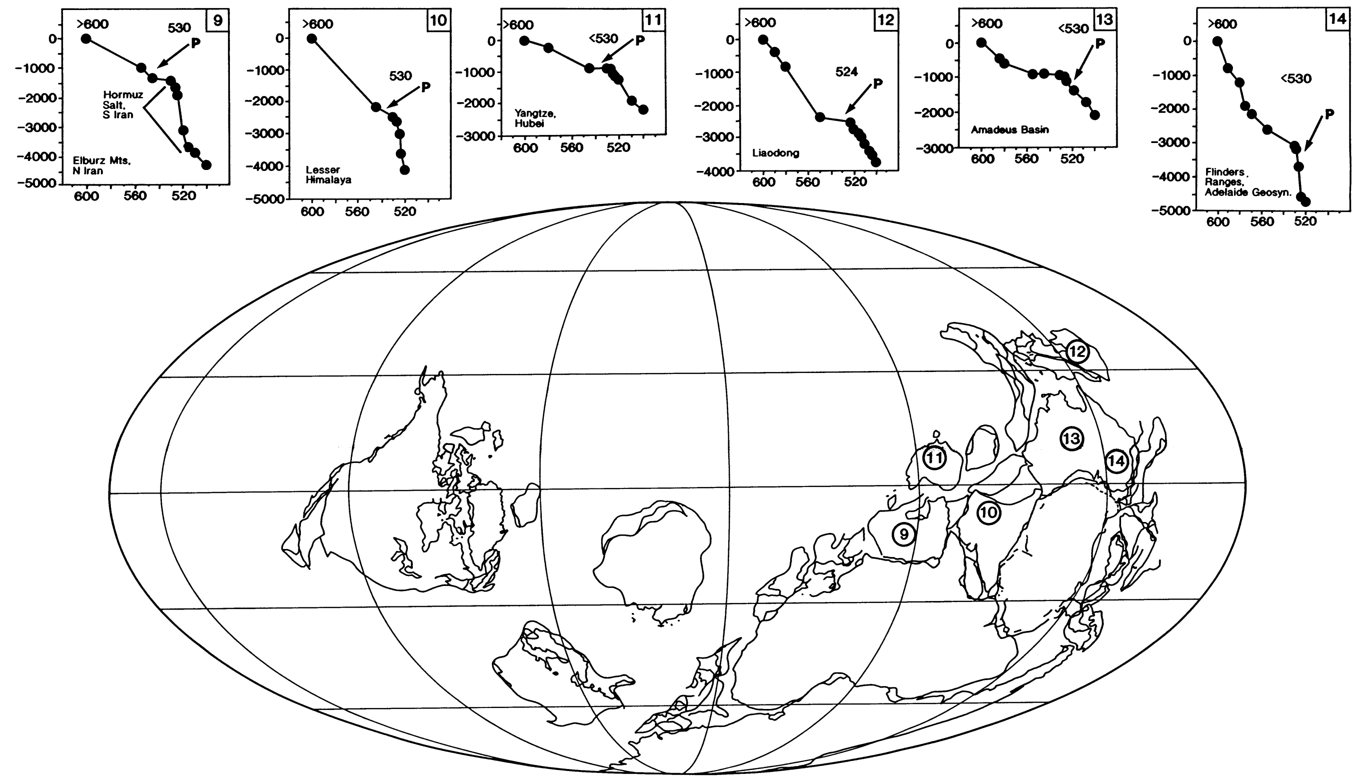


Figure 4.3 Sediment accumulation rates for the Ediacarian-Cambrian placed against the Atdabanian-Toyonian paleogeography of McKerrow et al. (1992). Based on data in the following sources: Elburz Mountains, Iran (Hamdi et al. 1989); Lesser Himalaya, India (Shanker and Mathur 1992); Yangtze Gorges, Hubei, South China, and Liaodong, North China (Wang 1986; Chang 1988; Lindsay 1993); Amadeus Basin, central Australia, and Flinders Ranges, Adelaide Geosyncline, southern Australia (Lindsay et al. 1987; Jenkins et al. 1993; Lindsay 1993). Key as for figure 4.2.

Middle Cambrian. Below, we explore the possibility that the amalgamation of Gondwana between ca. 555 and 510 Ma helped to bring about dramatic changes in the rate of sediment accumulation and in the biosphere over the Precambrian-Cambrian transition.

SEDIMENT ACCUMULATION RATES

Plots of sediment thickness against time can give an impression of the changing rate of sediment accumulation (figures 4.2 and 4.3). Such curves may, however, be skewed by the effects of compaction, which is greatest in siliciclastic sediments (especially argillites) and least in early-cemented carbonates. Rather than make assumptions about the degree of compaction and cementation, we here plot the raw data. Sediment accumulation rates are therefore likely to be underestimates in the case of finer clay-rich clastic lithologies. Inspection of the data, however, suggests that changes in sediment accumulation rate cannot be explained by changes in lithology and compaction alone.

In order to portray the tectonic component, data on the sediment accumulation rate should be "backstripped" by making corrections not only for the assumed effects of cementation and compaction but also for the isostatic effects of sediment loading; and further corrections should be made for the effects of water depth, the isostatic effect of seawater loading, and the stretch factor due to crustal extension (e.g., Watts 1982). If the sediments are mainly shallow-water deposits, as in this study, then backstripping tends to change the amplitude but not the general shape of the curves. In this study, we have found that selection of different time scales has relatively little effect on the shapes of the curves.

Backstripped tectonic subsidence curves have been used to track the thermal relaxation of the crust following rifting events, such as those during the Neoproterozoic-Cambrian. As rift turned to drift and ocean basins widened, extension on the margins of cratons is believed to have encouraged rapid rates of subsidence that diminished with time, in general accordance with geophysical models (e.g., Bond et al. 1985, 1988; Lindsay et al. 1987). The latter authors, by backtracking post-rift tectonic subsidence curves from the Middle-Late Cambrian, have estimated that a major phase of continental breakup took place in the Neoproterozoic–Early Cambrian (then dated at 625–555 Ma).

In figures 4.2 and 4.3 we have plotted sediment accumulation data against a time scale adapted from sources in Bowring et al. 1993, Tucker and McKerrow 1995, and Brasier 1995. We note that the rifting cratons of "Rodinia" are widely believed to have resulted from the breakup of Rodinia before ca. 720 Ma (Laurentia, Baltica, Siberia; figure 4.2), and show relatively low average rates of sediment accumulation during the early Ediacarian (ca. 600–550 Ma), followed by more rapid rates in the latest Ediacarian (after ca. 550 Ma, Mackenzies, Mongolia) to Early Cambrian (after ca. 530 Ma, Siberia, Kazakhstan, Baltica). These patterns may be attributed to a progressive attenuation in the thermal relaxation of the crust following the initial rifting of Rodinia in the Riphean, followed either by renewed phases of rifting (Laurentia, Baltica) or by the development of foreland basins (Siberia, Mongolia) across the Precambrian-Cambrian transition.

A similar pattern is seen in East Gondwana (Iran to Australia; figure 4.3), where an initial phase of rifting also seems to have been Riphean-Varangerian (ca. 725–600 Ma). There the rates of sediment accumulation in the Ediacarian interval (ca. 600–543 Ma) appear to have been relatively low, with some evidence for condensation and hiatus in the earliest Cambrian. A sharp change in the estimated rate of sediment accumulation coincides with major facies changes that suggest a renewed phase of subsidence close to the Precambrian-Cambrian boundary (ca. 545–530 Ma).

In West Gondwana (e.g., Avalonia, Morocco), the Ediacarian was characterized by rapid rates of sediment accumulation in compressive settings, which concluded with igneous intrusions, uplift, and cratonic amalgamation by ca. 550 Ma (figures 4.1 and 4.2). This phase was rapidly followed by the formation of extensional strike-slip basins that began to accumulate thick volumes of sediment.

LITHOFACIES CHANGES

Lithofacies changes provide further evidence for the rapid flooding of carbonate platforms between ca. 550 and 530 Ma B.P. The replacement of peritidal carbonates, especially "primary" dolomite, by neritic limestones and/or siliciclastic units above the Precambrian-Cambrian boundary (Tucker 1992; Brasier 1992a) broadly coincides in places (e.g., Mongolia; Lindsay et al. 1996) with the change from slower to more rapid rates of sediment accumulation. Hence, the mineralogic shift from dolomite to calcite/aragonite can be explained, in part, by the "drowning" of peritidal platforms, brought about by increased subsidence and relative sea level rise.

The widespread occurrence of phosphorites and cherts across the Precambrian-Cambrian boundary interval has for many years been related to the explosion of skeletal fossils in the Early Cambrian (e.g., Brasier 1980; Cook and Shergold 1984), but the connection has remained somewhat enigmatic. Brasier (1989, 1990, 1992a,b) has summarized evidence for the widespread development of "nutrient-enriched waters" during this interval and has argued that their incursion dramatically enhanced the preservation potential of early, thin-shelled skeletal fossils that herald the Cambrian period. These phosphatic sediments typically lie within the upper parts of dolomitic facies or rapidly succeed them. In figures 4.1–4.3 it can be seen that the first appearance of phosphatic beds with early skeletal fossils tends to coincide with the switch from slow to more rapid sediment accumulation. This may be explained by the interaction between phosphorus-rich oceanic waters and calcium-rich platformal waters under relatively low rates of sediment accumulation. Such conditions appear to have been widespread in the late Ediacarian to Tommotian (ca. 555–530 Ma). At first, the peritidal carbonate banks discussed above may have acted as barriers.

Later drowning of these barriers allowed incursions of nutrient-enriched water masses from the outer shelf and open sea. This drowning of barriers was made possible by the interrelated factors of increased subsidence and relative sea level rise.

Many Asiatic successions also show abrupt transitions from a restricted carbonate platform to organic-rich black shales over this interval, as, for example, in the latest Ediacarian of southwestern Mongolia (ca. 550–543 Ma, Brasier et al. 1996; Lindsay et al. 1996), and between the latest Ediacarian and mid-Atdabanian of southern Kazakhstan, Oman, Iran, Pakistan, India, and South China (ca. 545–527 Ma). These laminated black shales have many distinctive features: (1) they are basin-wide; (2) they follow a well-defined sequence boundary indicated by a major break in deposition, often with evidence for karstic solution of underlying peritidal carbonates; (3) they coexist with or overlie phosphatic dolostone beds and bedded cherts; (4) they contain high levels of organic matter with distinctively negative $\delta^{13}C$ values and positive $\delta^{34}S$ values; (5) they are highly metalliferous, with high concentrations of vanadium, molybdenum, cobalt, and barium; (6) in India, Oman, and China, they are accompanied by carbonates yielding a large negative carbon isotope anomaly (e.g., Hsu et al. 1985; Brasier et al. 1990, 2000), which is consistent with the turnover of aged, nutrient-enriched, and poorly oxygenated bottom waters (Brasier 1992a).

These anoxic marker events appear to lie in the interval between slower and more rapid rates of sediment accumulation. Drowning of the platform is indicated by the abrupt change in facies, from dolomites and peritidal phosphorites beneath. It therefore appears that a change in sedimentary regime took place, from one in which sediment accumulation rates were “space limited” (in the carbonate platform) to one in which they were “supply limited” (in the black shales).

Although gypsum, anhydrite, and evaporitic fabrics are not uncommon within the peritidal dolomite facies discussed above, thick layers of rock salt (halite) became widespread in the latest Ediacarian to the Early Cambrian. Indeed, some of the world’s thickest successions of rock salt were laid down from ca. 545 Ma onward (e.g., figures 4.1 and 4.3). These include the Hormuz Salt of Iran, the Ara Salt of Oman (both thought to be latest Ediacarian), the Salt Range salt of Pakistan (Atdabanian-Botoman), and the Usolka and contemporaneous salts of Siberia (Tommotian-Atdabanian; see Husseini and Husseini 1990; Kontorovitch et al. 1990; Burns and Matter 1993; Brasier et al. 2000). The preservation of thick halite implies interior basins with low siliciclastic supply, restricted by major barriers. The Hormuz and Oman salt horizons are also associated with volcanic rocks (e.g., Husseini and Husseini 1990; Brasier et al. 2000), which are taken to indicate an extensional tectonic setting. These salt deposits are therefore thought to have accumulated within interior barred basins formed by renewed subsidence of the basement (e.g., Loosveld et al. 1996). Poor bottom-water ventilation also led to anoxic conditions, so that associated sediments can be important as hydrocarbon source rocks (e.g., Gurova and Chernova 1988; Husseini and Husseini 1990; Mattes and Conway Morris 1990; Korsch et al. 1991).

THE EDIACARIAN-CAMBRIAN Sr AND Nd ISOTOPE RECORD

Figure 4.1 shows that least-altered values of $^{87}Sr/^{86}Sr$ rose almost continuously from ca. 0.7072 in the Varangerian to 0.7090 in the Late Cambrian, punctuated by a fall in values during the Tommotian (Derry et al. 1994; Brasier et al. 1996; Nicholas 1996). The low Riphean-Varangerian values have been attributed to the influence of hydrothermal flux on new ocean floors during rifting of the Rodinia (e.g., Veizer et al. 1983; Asmerom et al. 1991). The rise in Vendian $^{87}Sr/^{86}Sr$ ratios has been explained by accelerating rates of uplift and erosion associated with the Pan-African orogeny (e.g., Derry et al. 1989, 1994; Asmerom et al. 1991; Kaufman et al. 1994) and late Precambrian glaciations (Burns et al. 1994). The decline in seawater $^{87}Sr/^{86}Sr$ values in the Tommotian perhaps reflects a drop in the rate of erosion and subsidence, a decrease in silicate weathering rate, and/or the influence of rift-related hydrothermal activity (Derry et al. 1994; Nicholas 1996). It is interesting to note that this $^{87}Sr/^{86}Sr$ shift and the preceding hiatus found across much of the Siberian Platform and possibly beyond (Corsetti and Kaufman 1994; Ripperdan 1994; Knoll et al. 1995; Brasier et al. 1996) (figure 4.1) are both broadly coincident with the inferred shift from slower to more rapid sediment accumulation on many separate cratons (figures 4.2 and 4.3).

High crustal erosion rates have been inferred from late Tommotian to Late Cambrian $^{87}Sr/^{86}Sr$ values (Derry et al. 1994). This suggests that uplift and erosion of Pan-African orogenic belts (Avalonia and the Damara-Gariep belt of Namibia, for example) may have provided a source of radiogenic ^{87}Sr through the Cambrian. This interpretation is supported by studies of $\varepsilon_{Nd}(t)$ values in Ediacarian to Cambrian clastics from the Avalonian terranes of England and Wales. These sediments show a progressive reduction in the signal left by juvenile igneous rocks and an increase in the radiogenic component, between ca. 563 and 500 Ma (Thorogood 1990). Such a change in sediment supply suggests that younger accretionary margins became progressively submerged while older, interior crystalline rocks of the craton were uplifted and eroded, presumably as bulging of the crust and transgression of the platform proceeded. Comparison of the $^{87}Sr/^{86}Sr$ record of the Ediacarian-Cambrian with that of the Cenozoic (Derry et al. 1994) suggests that the inferred uplifted regions of Gondwana could even have experienced major montane glaciations through the latest Ediacarian-Cambrian interval.

THE EDIACARIAN-CAMBRIAN CARBON ISOTOPIC RECORD

Carbon isotopes show a long-term trend of falling values, from maxima of +11‰ $\delta^{13}C_{PDB}$ in the post-Sturtian interval (ca. 730–600 Ma B.P.) to +8 in the Ediacarian and +5.5 in the Cambrian (Brasier et al. 1996, 2000). On this broad-scale trend are superimposed a series of second-order cycles, which in the Cambrian appear to have been about 1 to 5 m.y. long, some of which can be correlated globally (e.g., Brasier et al. 1990; Kirschvink et al. 1991; Ripperdan 1994; Brasier et al. 1996; Calver and Lindsay 1998).

Above, we have argued for increasing rates of sediment accumulation through this time interval, which might be expected to have increased the global rates of carbon burial (cf. Berner and Canfield 1989). The long-term trend for carbon burial, however, is for falling values through the Neoproterozoic-Cambrian (figure 4.1). This means that increases in carbon burial due to raised rates of sediment accumulation must have been offset by raised rates of organic carbon oxidation. Such oxidation could have been brought about by a range of factors, including uplift and erosion of sedimentary carbon, greater ocean-atmosphere mixing (e.g., glacial climates, Knoll et al. 1996) and innovations in the biosphere (e.g., fecal pellets, Logan et al. 1995; bioturbation, Bottjer and Droser 1994, Brasier and McIlroy 1998).

The second order, 1–5 m.y. cycles in $\delta^{13}C$ may contain signals that relate to subsidence and sea level. Such a connection has been argued at higher levels in the geological column, as, for example, in the Late Cambrian (Ripperdan et al. 1992) and in the Jurassic-Cretaceous (e.g., Jenkyns et al. 1994). This has led to the suggestion that positive $\delta^{13}C$ excursions may record an increase in the burial of organic matter connected with the rapid areal expansion of marine depositional basins during "transgressions." Conversely, the negative $\delta^{13}C$ excursions may record reduced rates of carbon burial and increased rates of carbon oxidation during "regressions."

It is difficult to test for a connection between $\delta^{13}C$ and sea level in the Ediacarian-Cambrian interval without access to a set of rigorously derived sea level curves. Figure 4.1 shows a notional global sea level curve that depicts the major Cambrian transgression divided into major transgressive pulses. It is notable that several of the carbon isotopic maxima can be traced to these pulses; e.g., the appearance of laminated black limestones of the Sinsk Formation in Siberia coincided with the Botoman $\delta^{13}C$ maximum (Brasier et al. 1994; Zhuravlev and Wood 1996), and the influx of flaggy, phosphatic "outer detrital belt" carbonates of the Candland Shales in the Great Basin coincided with the Upper Cambrian sea level maximum (Bond et al. 1988; Brasier 1992c). Negative excursions can also, in several cases, be connected with evidence for emergence and omission surfaces. These are named in figure 4.1 and include the *Kotlin regression* prior to negative anomaly "W"; the *end-Yudoma regression* at the top of the Nemakit-Daldynian in Siberia (e.g., Khomentovsky and Karlova 1993; correlated with the top of the Dahai Member in South China, according to Brasier et al. 1990); the *Hawke Bay regression* across the Lower-Middle Cambrian boundary interval (i.e., the Sauk I-II boundary of Laurentia, according to Palmer 1981; with similar breaks in Baltica and Avalonia, according to Notholt and Brasier 1986); the *Andrarum regression* associated with the *Lejopyge laevigata* Zone of the Middle Cambrian in Scandinavia (correlated into Avalonia by Notholt and Brasier [1986] and possibly into Laurentia); and the Sauk II-III regression of Laurentia (Sauk II-III boundary of Palmer 1981 and Bond et al. 1985).

Of particular interest is the negative $\delta^{13}C$ interval of anomaly W, here taken to correlate the Precambrian-Cambrian boundary. Major sedimentary breaks occur close to this anomaly across the globe, which could be taken to indicate a synchronous regression during which sediments were removed by erosion (Brasier et al. 1997). At earlier times in the Neoproterozoic, negative $\delta^{13}C$ anomalies of this amplitude are associated with glacial/deglacial carbonates (Knoll et al. 1996). Although no glacial sediments are known from the Precambrian-Cambrian boundary interval, a glacially driven overturn of a stratified water column provides a possible explanation for the advection of light ^{12}C into surface waters (cf. Aharon and Liew 1992; Knoll et al. 1996). A similar explanation has also been put forward to explain falling $\delta^{13}C$ in the Tommotian (e.g., Ripperdan 1994).

Some positive $\delta^{13}C$ anomalies in shallow-water carbonates are accompanied by very positive $\delta^{34}S$ values in anhydrite (e.g., Mattes and Conway Morris 1990). Oxygen released by the burial of organic carbon may not, therefore, have been counterbalanced by the oxidation of sulfides to sulfates in the oceans (cf. Veizer et al. 1980). This has raised the possibility that the partial pressure of atmospheric oxygen could have risen over this interval, favoring the radiation of large, oxygen-hungry metazoans (cf. Knoll 1992; Sochava 1992). More-detailed $\delta^{34}S$ records are, however, needed to test this hypothesis.

IMPLICATIONS FOR THE BIOSPHERE

In figure 4.4 the ways in which raised rates of sediment accumulation/subsidence could have affected the hydrosphere, lithosphere, biosphere, and fossil record across the Precambrian-Cambrian transition are summarized. Submergence of shallow shelves inevitably led to an expansion of habitat area and, as we have argued, also caused phosphorus- and silica-rich waters to invade platform interiors. It may be argued that these environmental changes had a major ecologic impact upon the biota, encouraging blooms of eutrophic plankton, which in turn may have favored the development of a wide range of suspension feeders and the migration of pandemic phosphatic and siliceous taxa (figure 4.4). The reciprocal uplift of hinterland margins, indicated by the strontium isotope curve and by the thick succession of siliciclastic sediments, may well have delivered yet more phosphorus and iron into the oceans, thereby sustaining or raising its productivity (Derry et al. 1994).

This evidence for drowning of platforms also helps to explain some peculiar aspects of Cambrian fossil preservation. The development of secondary phosphatization of thin $CaCO_3$ or organic-walled skeletons during the latest Ediacarian to Atdabanian (e.g., Brasier 1980) is closely related to the timing of subsidence of carbonate platforms (figures 4.1–4.3). In Mongolia, for example, Cambrian-type siliceous sponge spicules and phosphatized early skeletal fossils first appear in the latest Ediacarian (Brasier et al. 1996, 1997) (figure 4.1). In India and North China, flood-

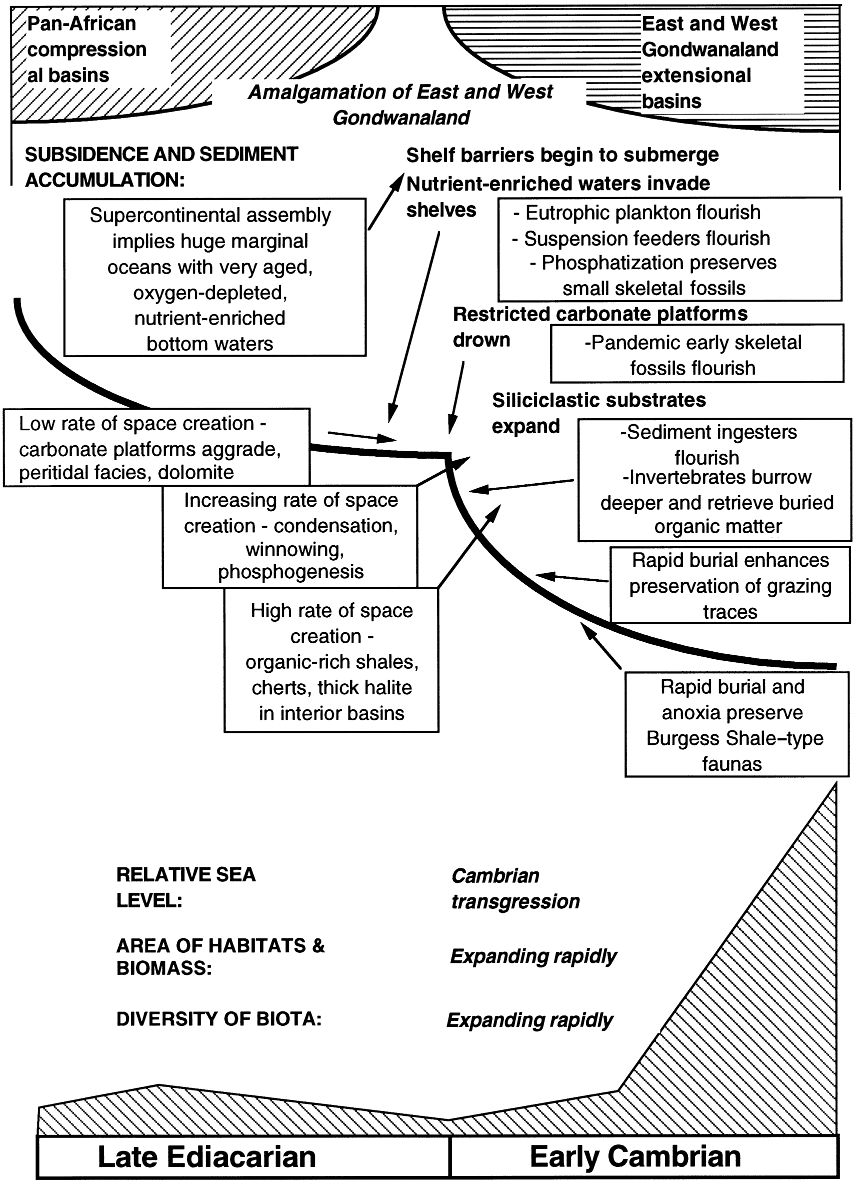


Figure 4.4 Model showing the inferred influence of global tectonic changes upon subsidence, sediment accumulation rate, sea level, nutrients, fossil preservation, and the adaptive radiation of the Cambrian fauna.

ing of the carbonate platforms brought phosphatic sediments with early skeletal fossils that were a little younger (figure 4.1).

In clastic sediments, the first main indications of the Cambrian radiation are given by trace fossils. Here, one of the main puzzles has been the preservation of deep-water *Nereites* ichnofacies in shallow waters during the Cambrian (e.g., Crimes 1994). At

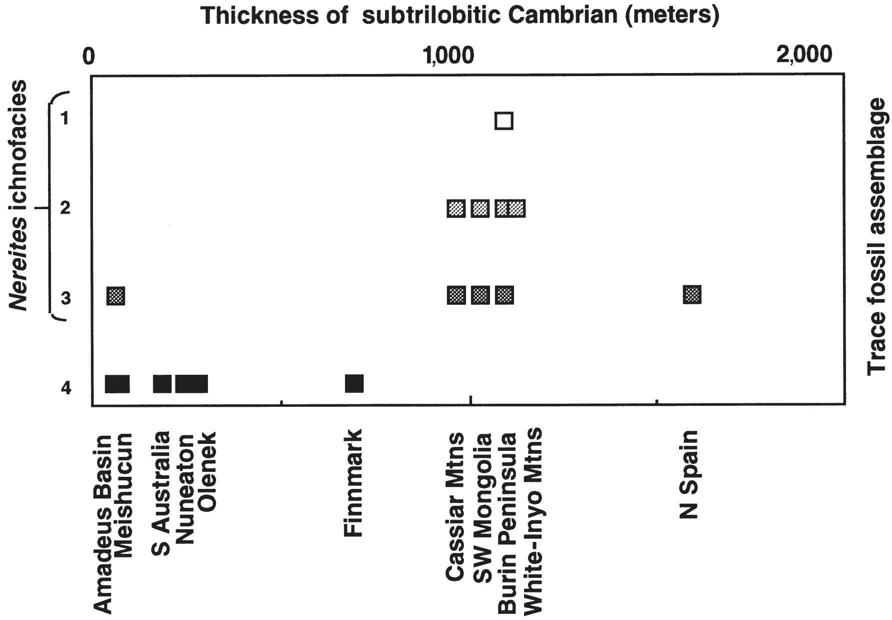


Figure 4.5 The paradox of deeper-water *Nereites* ichnofacies traces (assemblages 1, 2, and 3) in shallow water subtrilobitic Cambrian facies, which may be related to rapid rates of sediment accumulation. *1* = graphoglyptids (e.g., *Palaeodictyon* isp.), *2* = *Helminthopsis* and *Helminthoida* isp., *3* = *Taphrhelminthopsis* isp., *4* = other ichnogenera. Thickness and ichnotaxa from sources in Crimes (1989).

higher levels in the stratigraphic column, the distribution of these grazing traces has been related to the incidence of event sands, such as turbidites, which cast and preserve the delicate top tier of the ichnofauna (Bromley 1990). A review of the literature suggests that these grazing traces tend to be best represented in subtrilobitic Cambrian successions that are relatively thick (figure 4.5). Hence, the paradox of deeper-water *Nereites* ichnofacies traces in shallow-water sediments may well have been enhanced by conditions of rapid deposition, which led to the preservation of a greater number of sand-mud interfaces.

A further paradox of Cambrian fossil preservation concerns the restriction of Burgess Shale–type Lagerstätten to the Early and Middle Cambrian (Conway Morris 1992) (figure 4.1), despite the presence of suitable, anoxic, and poorly bioturbated facies at other times. Butterfield (1995, 1996) has suggested that this paradox could be explained by the restricted temporal distribution of volcanogenic clay minerals with antienzymatic and/or stabilizing effects. Here, we wonder whether Burgess Shale–type preservation was enabled by frequent pulses of fine-grained sedimentation along rapidly subsiding margins, leading to rapid burial and early diagenesis.

One of the most dramatic effects of sea floor subsidence on the Cambrian fossil record was arguably that of sudden "explosive phases" in diversification. The explosion in diversity at the base of the Tommotian certainly coincides with a rapid change in $^{87}Sr/^{86}Sr$ and lies above a major karstic surface (Brasier et al. 1996). In southwestern Mongolia, where rejuvenation of subsidence began in the latest Ediacarian, there is no clear Tommotian explosion in diversity (Brasier et al. 1996). This "Tommotian explosion" can therefore be regarded as an artefact brought about by missing time followed by abrupt facies changes (Lindsay et al. 1996), together caused by a rejuvenation of subsidence along the margin of the Mongolian arc.

CONCLUSION

Evidence is given above for increasing rates of subsidence and sediment accumulation during the Cambrian. This is no longer consistent, however, with the simple hypothesis that a rift-to-drift transition took place over the Precambrian-Cambrian boundary interval (Bond et al. 1984, 1985, 1988); foreland basins related to the amalgamation of Gondwana were also forming at this time. Such subsidence is also in line with the evidence for a major rise in relative sea level, from a low point during the Varangerian glaciation to a high point somewhere in the Late Cambrian (see figure 4.1; e.g., Brasier 1980, 1982, 1995; Bond et al. 1988).

A rise in the rate of sediment accumulation between ca. 550 and 530 Ma suggests that rapid subsidence took place in cratonic margins and interior basins around the globe. The large supplies of clastic sediment that flooded into these basins imply high rates of uplift and erosion of the basin hinterlands, which in turn can provide a plausible explanation for the progressive rise in $^{87}Sr/^{86}Sr$ and change in $\varepsilon_{Nd}(t)$ through the Cambrian.

Carbon isotopic maxima in the Early and Late Cambrian appear to coincide with transgressive pulses, and several carbon isotopic minima (e.g., anomaly W close to the Precambrian-Cambrian boundary, and that at the base of the Tommotian) are associated with widespread breaks in deposition. Further work is needed to test the hypothesis that these negative excursions relate to episodes of cooler (glacial?) climate and oceanic overturn. Recent work on the very complex section preserved in the Nama Basin in Namibia emphasizes the possibility of breaks in sedimentation in other basins (Grotzinger et al. 1995).

A picture emerges of Neoproterozoic to Early Cambrian oceans that were well fed with biolimiting nutrients, derived perhaps from high rates of erosion and runoff and enhanced by montane glaciations. As the platforms extended and subsided across the Neoproterozoic-Cambrian transition, new kinds of fossil preservation became possible: phosphatization and silicification of early skeletal fossils; rapid burial of delicate, grazing trace fossils; and rapid burial and diagenesis of Burgess Shale–type faunas. These changes have amplified and distorted our view of the evolutionary radiation, making the fossil record appear more stepwise, with explosive phases in diversity that we suspect are largely illusory.

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