that the growth of ε_{Nd} in the upper mantle has been retarded over the past 2,000 Myr, probably because of recycling of continental crust. The general tendency for samples nearer the Archaean craton to have lower initial ε_{Nd} values argues for contamination with small amounts of Archaean crust. Yet these samples—the GMF, Twilight Gneiss and Idaho Springs Formation—all have the same calculated initial $\varepsilon_{Nd} = 3.7 \pm 0.5$, which implies vigorous and thorough regional mixing of a small amount of crust into the mantle near the Archaean plate boundary. The almost ubiquitous disturbance of the Rb-Sr system in crustal rocks makes it difficult to determine a precise $^{87}Sr/^{86}Sr$ evolution curve for depleted mantle. Continuing efforts to establish this curve are, however, extremely important in the context of establishing the history of crust-mantle evolution.

The Nd data confirm earlier conclusions that large segments of the crust in southwestern North America were derived from the mantle ~1,800 Myr ago[32]. Compared with the Mesozoic and Cenozoic patterns, this suggests major contrasts in rates, and possibly styles, of plate tectonics in middle Proterozoic time[33].

We thank Kent Condie and James Robertson for providing well documented samples, Sam Bowring for discussions of regional geology, and P. J. Patchett for reviewing the manuscript. This work was supported by a grant from donors of the Petroleum Research Fund, administered by the American Chemical Society, and NSF grant EAR82-12377.

Received 26 April; accepted 15 August 1984.

1. Schilling, J. -G. & Winchester, J. W. in *Mantles of the Earth and Terrestrial Planets* (ed. Runcorn, S. K.) 267-283 (Interscience, London, 1967).
2. Armstrong, R. L. *Phil. Trans. R. Soc.* **A301**, 443-472 (1981).
3. DePaolo, D. J. *Geophys. Res. Lett.* **10**, 705-708 (1983).
4. DePaolo, D. J. *Geochim. cosmochim. Acta* **44**, 1185-1196 (1980).
5. Jacobsen, S. B. & Wasserburg, G. J. *J. geophys. Res.* **84**, 7411-7427 (1979).
6. Allègre, C. J., Dupré, B., Richard, P., Rousseau, D. & Brooks, C. *Earth planet. Sci. Lett.* **57**, 25-34 (1982).
7. Taylor, S. R. in *Island Arcs, Deep Sea Trenches and Back-Arc Basins* (eds Talwani, M. & Pitman, W. C.) 325-336 (American Geophysical Union, Washington DC, 1977).
8. Condie, K. C. *Resource Map 13* (New Mexico Bureau of Mines and Mineral Resources, 1981).
9. Condie, K. C. & Nuter, J. A. *Precambr. Res.* **15**, 131-155 (1981).
10. Condie, K. C. & McCrink, T. P. *Precambr. Res.* **19**, 141-166 (1982).
11. DePaolo, D. J. *Nature* **291**, 193-196 (1981).
12. Patchett, P. J., Kouvo, O., Hedge, C. E. & Tatsumoto, M. *Contr. Miner. Petrol.* **78**, 279-297 (1981).
13. Premo, W. R. & VanSchmus, W. R. *Geol. Soc. Am. Abstr. Prog.* **14**, 346 (1982).
14. Bickford, M. E., Bowring, S. A. & Gray, J. E. *Geol. Soc. Am. Abstr. Prog.* **14**, 303 (1982).
15. Bowring, S. A. & Condie, K. C. *Geol. Soc. Am. Abstr. Prog.* **14**, 304 (1982).
16. Condie, K. C. & Shadel, C. A. *Can. J. Earth Sci.* (in the press).
17. Houston, R. S. & McCallum, M. E. *Geol. soc. Am. Spec. Pap.* **68**, 91 (1961).
18. Condie, K. C. *Geology* **10**, 37-42 (1982).
19. DePaolo, D. J. *J. geophys. Res.* **86**, 10470-10488 (1981).
20. Papanastassiou, D. A., DePaolo, D. J. & Wasserburg, G. J. *Proc. Lunar Sci. Conf. 8th* 1639-1672 (1977).
21. Zindler, A. H. in *Komatiites* (eds Arndt, N. T. & Nisbet, E. G.) 399-420 (Allen & Unwin, London, 1982).
22. Patchett, P. J. *Trans. Am. geophys. Un.* **64**, 905-906 (1983).
23. Cohen, R. S., Evensen, N. M., Hamilton, P. J. & O'Nions, R. K. *Nature* **283**, 149-153 (1980).
24. Patchett, P. J. & Chauvel, C. *Geophys. Res. Lett.* **11**, 151-153 (1984).
25. Allègre, C. J. & Rousseau, D. *Earth planet. Sci. Lett.* **67**, 19-34 (1984).
26. DePaolo, D. J. & Wasserburg, G. J. *Geophys. Res. Lett.* **3**, 743-746 (1976).
27. O'Nions, R. K., Hamilton, P. J. & Evensen, N. M. *Earth planet. Sci. Lett.* **34**, 13-22 (1977).
28. DePaolo, D. J. & Wasserburg, G. J. *Geophys. Res. Lett.* **4**, 465-468 (1977).
29. Jacobsen, S. B. & Wasserburg, G. J. *J. geophys. Res.* **84**, 7429-7445 (1979).
30. DePaolo, D. J. *Eath planet. Sci. Lett.* **43**, 201-211 (1979).
31. Jacobsen, S. B. & Heier, K. *Lithos* **11**, 257-276 (1978).
32. Nelson, B. K. & DePaolo, D. J. *Bull. geol. Soc. Am.* (in the press).
33. Hoffman, P. F. & Bowring, S. A. *Geology* **12**, 68-72 (1984).
34. Condie, K. C. *J. Geol.* **88**, 603-609 (1980). Taylor, S. R. in *Island Arcs, Deep Sea Trenches and Back-Arc Basins* (eds Talwani, M. & Pitman, W. C.) 325-336 (American Geophysical Union, Washington DC, 1977).
35. Robertson, J. M. *Geol. Soc. Am. Abstr. Prog.* **39**, 103 (1981).
36. DePaolo, D. J. & Wasserburg, G. J. *Geophys. Res. Lett.* **3**, 249-252 (1976).
37. Wooden, J. L., Mueller, P. A., Hunt, D. K. & Bowes, D. R. *Spec. Publ. Montana Bur. Mines. Geol.* **84**, 45-55 (1982).
38. DePaolo, D. J. thesis, Calif. Inst. Technol. (1978).
39. Claoué-Long, J. C., Thirlwall, M. F. & Nesbitt, R. W. *Nature* **307**, 697-701 (1984).
40. Bokhari, F. Y. & Kramers, J. D. *Earth planet. Sci. Lett.* **54**, 409-422 (1981).
41. McCulloch, M. T., Gregory, R. T., Wasserburg, G. J. & Taylor, H. P. *Earth planet. Sci. Lett.* **46**, 201-211 (1980).
42. Morris, J. D. & Hart, S. R. *Geochim. cosmochim. Acta* **47**, 2015-2030 (1983).
43. McCulloch, M. T. & Perfit, M. R. *Earth planet. Sci. Lett.* **56**, 167-179 (1981).
44. Nohda, S. & Wasserburg, G. J. *Earth planet. Sci. Lett.* **52**, 264-276 (1981).
45. Hickey, R. L. & Frey, F. A. *Geochim. cosmochim. Acta* **46**, 2099-2115 (1982).
46. DePaolo, D. J. & Johnson, R. W. *Contr. Miner. Petrol.* **70**, 367-380 (1979).
47. Hawkesworth, C. J., O'Nions, R. K., Pankhurst, R. J., Hamilton, P. J. & Evensen, N. M. *Earth planet. Sci. Lett.* **36**, 253-262 (1977).
48. Cohen, R. S. & O'Nions, R. K. *Earth planet. Sci. Lett.* **61**, 73-84 (1982).
49. Hawkesworth, C. J. & Powell, M. *Earth planet. Sci. Lett.* **51**, 297-308 (1980).
50. Hawkesworth, C. J., O'Nions, R. K. & Arculus, R. J. *Earth planet. Sci. Lett.* **45**, 237-248 (1979).

A mechanism for magmatic accretion under spreading centres

John A. Whitehead Jr, Henry J. B. Dick & Hans Schouten

Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543, USA

The most active volcanic zone on Earth occurs along the global mid-ocean ridge system where the lithosphere is pulled apart and new oceanic crust is created at rates between 10 and 200 km Myr^{-1}. This volcanism is segmented, and is concentrated along the axial valleys and highs of slow- and fast-spreading ridges respectively, with transform fault zones separating, and generally offsetting, each active ridge segment. Volcanic activity may be nearly absent at the ridge transform intersection where a slow-spreading ridge is offset, and greatly reduced where a fast-spreading ridge is offset, implying thinner crust in these regions[1]. The source of magmas erupted along ocean ridges is the underlying mantle, which undergoes decompression melting when rising to fill the gap between the spreading plates. The resulting magma aggregates in the upper mantle and ascends to the crust due to its low density compared with the parent mantle rock. The melt is believed to pool in crustal magma chambers—whence it periodically erupts to the surface. Typically, the formation, ascent and aggregation of magmas along the mid-ocean ridges is regarded two-dimensionally and modelled in a section through the crust and mantle across the strike of the ridges (see ref. 2). Here, however, we consider the problem with a three-dimensional model for which we provide supporting geological evidence, in which the magmas rise as a result of a gravitational instability (modified Rayleigh–Taylor) in the underlying partial melt zone.

It is reasonable, to a first approximation, to assume that the mantle is behaving as a viscous newtonian fluid, so we can expect the upwelling to be two dimensional with a maximum under the smoothed location of the spreading centre. If this is the case, we also postulate that a linear region of high-melt content exists in the mantle below the ridge where melt aggregates from the rising asthenosphere. This region could be approximated as a horizontal cylindrical body with relatively low viscosity and density compared with the overlying mantle. In such circumstances, it will develop a gravitational instability leading to regularly spaced vertical protrusions. We have conducted some simple experiments in which a water–glycerine mixture was quickly injected into glycerine along a horizontal line. Although this line will gradually rise because the water–glycerine mixture is less dense than the pure glycerine, an instability will always also develop as shown in Fig. 1 and lead to semi-spherical pockets. It is reasonable to expect that a linear

Table 1 Composition of postulated parental mantle, abyssal peridotites and a primitive abyssal basalt

	Parental mantle models Pyrolite*	Pyrolite†	Xenoliths‡	Fracture zone peridotites§	Primitive abyssal basalt tholeiite‖
SiO_2	46.1	45.0	42.9	43.6	50.9
TiO_2	0.02	0.17	0.2	0.02	0.89
Al_2O_3	4.3	4.4	5.8	1.18	16.0
FeO	8.2	7.6	9.2	8.20	8.83
MgO	37.6	38.8	37.2	45.2	9.02
CaO	3.1	3.4	3.7	1.13	12.4
Na_2O	0.04	0.40	0.7	0.02	2.08
Cr_2O_3	—	0.45	0.2	0.22	—
NiO	—	0.26	0.2	0.22	—
Mg/(Mg+Fe)	0.891	0.901	0.878	0.908	—

* Pyrolite model[21].
† Pyrolite model[22].
‡ Upper mantle model[23].
§ Average abyssal peridotite[9].
‖ Primitive abyssal tholeiite (TR154–1802)[24].

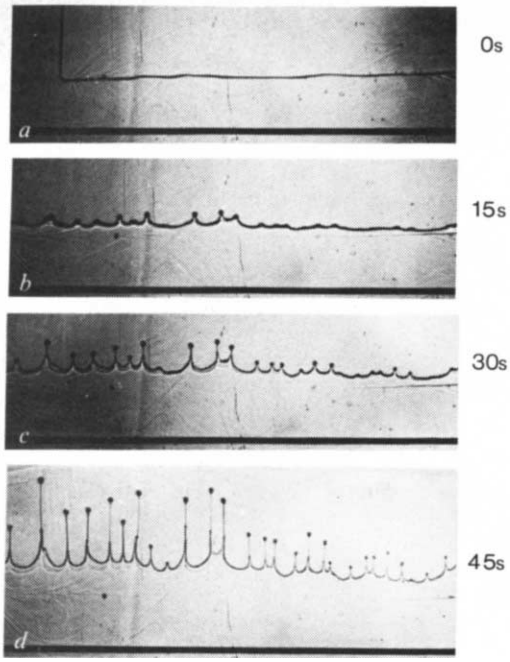

Fig. 1 Gravitational instability of a horizontal line of water/glycerine mixture in a bath of pure glycerine. Kinematic viscosity of the glycerine is estimated as $5.0\ \mathrm{cm^2\ s^{-1}}$, that of the water/glycerine mixture as $0.015\ \mathrm{cm^2\ s^{-1}}$, and reduced gravity $g^* = g\Delta\rho/\rho = 260\ \mathrm{cm\ s^{-2}}$. The dark stripe at the bottom is 0.64 cm diameter tubing, and the initial diameter of the conduit is estimated with a precision optical micrometer to range from 0.16 to 0.24 cm. The spacing of the 25 diapirs in *d* is 39.5 cm for an average of 1.6-cm spacing, giving a ratio of spacing to conduit diameter of 7–10. Marsh argues that the formula for spacing is $d/D = 2\pi/2.15\ (\mu_1/\mu_2)^{1/3}$, where μ_1 is viscosity of the upper fluid and μ_2 is viscosity of the conduit fluid with perhaps an order one correction in the coefficient[4]. Our parameter group $2\pi/2.15\ (\mu_1/\mu_2)^{1/3}$ is ~20, roughly twice as big, but well within the acceptable range given.

region of partially molten mantle in the Earth will behave in a similar manner and will lead to fairly regularly spaced protrusions from which the melt will ascend to form magma chambers at spreading ridges.

The instability is similar to the classical Rayleigh–Taylor instability in fluid mechanics. It occurs when there is a density inversion, that is a layer of heavy fluid over light. Whitehead and Luther[3] reported some experiments and theory on this instability which were relevant when both viscosities were large. They found that the ratio of the wavelength of fastest-growing disturbances to the thickness of the unstable layer was large as the viscosity contrast became large. Thus, a thin layer of relatively low-viscosity fluid can be expected to develop very widely spaced protrusions. Marsh[4,5] isolated a similar problem, the instability of a ribbon-like, or cylindrical (with its axis horizontal), body of low-viscosity material. The ribbon-like region was produced in the laboratory by allowing black oil or glycerine to rise from a slit in the bottom of a container of glycerine. Marsh suggests that this diapir-genesis occurs under island arcs and is the mechanism that leads to the formation of the magma chambers for the island-arc volcanoes. Noting the regular spacing of magmatic centres across Iceland, Sigurdsson and Sparks[6] suggest that a similar instability exists at the base of the lithosphere in Iceland.

The suggestion that there is convection at right angles to the direction of relative plate movement is not new (see ref. 7). However, we suggest that an instability due to density change through melting in the mantle beneath the ridge produces upwellings with length scales of the spacing between the midpoints of spreading centre segments. This also implies that axial magma chambers are localized at the midpoints of individual ridge segments.

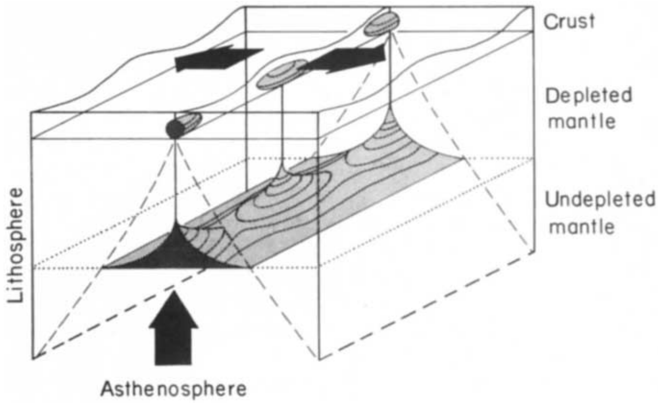

Fig. 2 Viscous asthenosphere rises two-dimensionally beneath the boundary between two spreading lithospheric plates. The lithosphere thickens away from the spreading boundary (dashed lines). Above a certain level (dotted lines), the rising asthenosphere passes through a zone in which partial melt can form which collects at some level below the base of the lithosphere. Due to its lower viscosity and density, the partial-melt zone develops a gravitational instability leading to regularly spaced concentrations of melt which rise to the surface to form crustal magma chambers. The asthenosphere, effectively depleted by this process, must continue to rise viscously and on cooling become lithosphere. The wavelength of the gravitational instability (and the subsequent spacing of spreading centres or transform zones) should also depend on the width of the partial melt zone which is expected to be proportional to the spreading rate.

Evidence for a partially molten zone beneath the base of the lithosphere is seen from regional magnetotelluric measurements in Iceland. These measurements indicate an anomalous conducting layer at a depth of 8–15 km beneath the Neovolcanic Zone which is interpreted as a partially molten zone near the top of the mantle[8]. This zone may correspond to the partially molten region we require in the mantle beneath mid-ocean ridges.

Direct evidence for melt accumulation from a partially molten zone beneath the mid-ocean ridges can be seen in plagioclase-bearing peridotites from oceanic fracture zones. These rocks constitute ~30% of dredged abyssal peridotites[9] and generally contain <0.5% interstitial plagioclase which is believed to have largely crystallized from trapped melt[10–12]. Given the proportion of plagioclase in primitive abyssal tholeiite, it represents <1% trapped melt in the peridotites, although in rare cases, as locally along the Romanche Fracture Zone, peridotites from one or more dredge haul contain an average of ⩾6% plagioclase[9] corresponding to ~13% trapped melt. This is direct evidence for local accumulation of melt in the mantle, which we believe is the result of a Rayleigh–Taylor instability in the partially molten zone and a rapid migration of melt towards the resulting protrusions into the overlying mantle. Green and Hibberson[13] studied the stability of plagioclase in peridotites at high pressures and found that it was unstable at pressures >8.5 kbar. Based on these phase relations, the zone of melt aggregation represented by the plagioclase-peridotites must lie at <25 km in the mantle.

Abyssal peridotites dredged largely from oceanic fracture zones are generally accepted as the residues of mantle melting which generated abyssal basalts. Almost 70% of these peridotites are plagioclase-free spinel peridotites[9]. Table 1 compares their average composition with independent estimates of undepleted upper mantle composition and a typical primitive abyssal basalt. This shows that melting of the upper mantle would rapidly deplete the solid residue in those elements which occur in far greater abundance in basalt than in the postulated source rock. From the relative concentrations of CaO and Al_2O_3 in Table 1, somewhere between 15 and 20% melt would have to be removed from the postulated parents to produce the observed concentrations of these oxides in the fracture zone peridotites.

Taking all abyssal peridotites together gives an average of 0.9% plagioclase which would correspond to ~1–2% trapped melt. Thus most of the melt which was formed in peridotites dredged from fracture zones must have been drained off. Most of these rocks, however, were dredged from regions in fracture zones where only peridotites were exposed for areas of hundreds, if not thousands, of square kilometres and where little, if any, basalt exists. Accordingly, the melts were not emplaced in the same region as their parental peridotites. We propose that melt formed at depth near fracture zones migrates laterally through the mantle towards an upwelling beneath the centre of the ridge segment rather than that the depleted peridotites were emplaced laterally into the fracture zone after melting beneath the ridge.

Figure 2 summarizes our suggested structure and flow field. The asthenosphere rises viscously beneath the spreading centres. Above a cetain level, the rising asthenosphere enters a zone in which partial melt can form. To a first approximation the viscous flow of the asthenosphere is two-dimensional with its axis parallel to the ridge. Thus the rate of partial melt formation is nearly uniform below the ridge axis. The partially molten zone will have relatively low viscosity and density compared with the overlying mantle. The partially molten zone thus develops regularly spaced protrusions centred beneath individual ridge segments. The melt, which can migrate by porous flow, may move more rapidly than the partially molten layer as a whole and concentrate at the tops of the rising protrusions. This would lead to gravitational separation of melt from the mantle and its ascent to form or feed crustal magma chambers. Above the melt aggregation zone, the depleted asthenosphere must continue to rise viscously and, on cooling, become lithosphere.

Young melt concentrations will replenish the older crustal magma chambers. New crust is formed by crystallization of gabbro along the margins of the magma chambers, by the injection of dykes from the magma chambers into the existing crust, both laterally and vertically along the spreading centre axes, and by extrusion of pillow basalts and massive flows from magma chambers and dykes to the top of the crust. Furthest away from the crustal magma chambers, that is, in the transform zones between spreading centres, fewer dykes will be injected and accordingly fewer pillow basalts and massive flows will be extruded, thus producing thinner crust.

This model provides an explanation for the regular segmentation of ocean ridges. Schouten and Klitgord[14] postulated that the sea floor in the slow-spreading North Atlantic (and in other oceans) was formed in a relatively stationary series of spreading centre cells separated by transform-fault zones. From Mesozoic marine magnetic anomalies (155–108 Myr ago) in the western North Atlantic, they estimate a typical separation between spreading centre cells of the order of 50 km, similar to the present distribution of axial valleys on the Mid-Atlantic Ridge. The magnitude of offset between adjacent spreading centres is typically <30 km, can change with time, and can even go through zero and change sign[15]. This is inconsistent with an offset-generating mechanism inferred from the rheology of the plate material alone, as the ridges should lose their 'memory' of the fracture zone at 'zero offset'. The stability of spreading centre cells, then appears independent of offsets between them[14–16]. The model proposed here predicts regular segmentation of ocean ridges, particularly where no large-offset transforms exist. The model also predicts that the crust at small and zero-offset fracture zones should thin, which is consistent with what little is known from seismic studies of these features[15,17–19].

Lateral flow of melt towards regularly-spaced protrusions centred beneath individual ridge segments would inhibit flow and mixing of melt in the mantle between ridge segments. Accordingly, melts erupted at each ridge segment will be derived largely from the mantle beneath that segment. This provides a simple explanation for the geochemical discontinuities observed in ridge basalts across fracture zones (see ref. 20) and strongly supports a laterally heterogeneous oceanic mantle. This effect is heightened at large-offset fracture zones where old-cold lithosphere, juxtaposed against the rising asthenosphere, depresses the partially molten zone[1] and produces a substantial physical barrier to the lateral flow of magmas. We note that our model does not rule out segregation and intrusion of melt to the crust from greater depths in the mantle, nor does it rule out the occurrence of secondary instabilities in the partially molten layer either near fracture zones or laterally away from the ridge axis. We point out, however, that it does predict the presence of geochemical discontinuities in ridge basalts even at zero-offset fracture zones.

This paper developed from a series of interdisciplinary seminars and direct research support sponsored by the Center for the Analysis of Marine Systems of the Woods Hole Oceanographic Institution with funds provided by the Exxon Education Foundation. We thank B. D. Marsh, W. B. Bryan, S. Humphries and J. Karson for constructive criticism. We acknowledge support from the NSF and the Office of Naval Research. Robert Frazel aided in design and construction of the experiment and photography. Anne-Marie Michael, Alice Tricca and Pam Foster processed the manuscript. Woods Hole Oceanographic Institution contribution no. 5668.

Received 2 May; accepted 14 August 1984.

1. Fox, P. & Gallo, D. G. *Tectonophysics* **104**, 205–242 (1984).
2. Ahern, J. L. & Turcotte, D. L. *Earth planet. Sci. Lett.* **45**, 115–122 (1979).
3. Whitehead, J. A. Jr & Luther, D. S. *J. geophys. Res.* **80**, 705–717 (1975).
4. Marsh, B. D. *J. Geol.* **87**, 687–713 (1979).
5. Marsh, B. D. *Am. Scient.* **62**, 161–171 (1979).
6. Sigurdsson, H. & Sparks, S. *Nature* **274**, 126–130 (1978).
7. Richter, F. M. & Pearsons, B. *J. geophys. Res.* **80**, 2529–2541 (1975).
8. Thayer, R. E., Bjornsson, A., Alvarez, L. & Hermance, J. F. *Geophys. J. R. astr. Soc.* **65**, 423–442 (1981).
9. Dick, H. J. B. & Fisher, R. L. in *The Mantle and Crust-Mantle Relationships, Proc. 3rd. int. Kimberlite Conf.* (ed. Kornprobst, J.) 295–308 (Elsevier, Amsterdam, 1984).
10. Church, W. R. & Stevens, R. K. *J. geophys. Res.* **76**, 1460–1466 (1971).
11. Dick, H. J. B., Fisher, R. L. & Bryan, W. B. *Earth planet. Sci. Lett.* (in the press).
12. Dick, H. J. B. & Bullen, T. *Contr. Miner. Petrol.* **86**, 54–76 (1984).
13. Green, D. H. & Hibberson, W. *Lithos* **3**, 209–221 (1970).
14. Schouten, H. & Klitgord, K. D. *Earth planet. Sci. Lett.* **59**, 255–266 (1982).
15. Schouten, H. & White, R. S. *Geology* **8**, 175–179 (1980).
16. Schouten, H. & Klitgord, K. D. *Nature* **303**, 549–550 (1983).
17. White, R. S. & Matthews, D. H. *Geophys. J. R. astr. Soc.* **61**, 401–435 (1980).
18. Sinha, M. C. & Louden, K. E. *Geophys. J. R. astr. Soc.* **75**, 713–736 (1983).
19. Mutter, J. C., Detrick, R. S. & NAT Study Group, *Geology* (in the press).
20. Machado, A., Ludden, J. N., Brooks, C. & Thompson, G. *Nature* **295**, 226–228 (1981).
21. Ringwood, A. E. *Composition and Petrology of the Earth's Mantle* (McGraw-Hill, New York, 1975).
22. Green, D. H., Hibberson, W. O. & Jaques, A. L. in *The Earth: Its Origin, Structure and Evolution* (ed. McElhinny, M. W.) 165–199 (Academic, New York, 1979).
23. Carter, J. L. *Bull. geol. Soc. Am.* **81**, 2021–2034 (1970).
24. Sigurdsson, H. *J. geophys. Res.* **86**, 9483–9502 (1981).

Anaerobic consumption of organic matter in modern marine sediments

A. Yu. Lein

Institute of Biochemistry and Physiology of Microorganisms, USSR Academy of Sciences, Pushino, Moscow Region, 142292, USSR

Previous estimates of the global organic carbon (C_{org}) budget have generally considered the anaerobic consumption of C_{org} to be of minor importance. Data for the rates of bacterial sulphate reduction in various morphological zones of the oceans indicate, however, that approximately 14% of the total C_{org} reaching the sediment–water interface is consumed by anaerobic processes. Of this, some 70% is taken up in sediments on the continental shelves and slopes in water depths of less than 1,000 m. Although this represents only a small fraction of the global primary carbon productivity, it may have important consequences for models of the carbon and sulphur cycles.

There is some controversy over the rate of primary productivity in the world ocean. Values ranging from 20 Gton C yr^{-1} (refs 1, 2) to 126 Gton C yr^{-1} (ref. 3) have been reported. In view of recent work which questions the validity of the ^{14}C-technique[4] and considers contributions from picoplank-