File Copy

LAMONT-DOHERTY GEOLOGICAL OBSERVATORY
OF COLUMBIA UNIVERSITY

TURBIDITY CURRENTS ON THE OCEAN BOTTOM

by

Maurice Ewing, William B. F. Ryan
Hubert D. Needham, and B. Charlotte Schreiber

UNCLASSIFIED

Report Number: CR 70.018

November 1970

Final Report

Contract N62399-69-C-0006

U. S. Naval Civil Engineering Laboratory
Port Hueneme, California 93041

File copy
Folder 6

Lamont-Doherty Geological Observatory
of Columbia University
Palisades, N.Y., 10964

TURBIDITY CURRENTS ON THE OCEAN BOTTOM

by

Maurice Ewing, William B. F. Ryan,
Hubert D. Needham, and B. Charlotte Schreiber

UNCLASSIFIED

Report Number: CR 70.018
November 1970

Final Report
Contract N-62399-69-C-0006

U.S. Naval Civil Engineering Laboratory
Port Hueneme, California 93041

C O N T E N T S

CONTENTS (continued)

Pages

INTRODUCTION

GENERAL BACKGROUND --

The purpose of this report is to contribute information pertinent to the selection of sites for manned and unmanned structures on the ocean floor, and to provide guidance for the prediction of natural hazards to which sites may be exposed during and after the emplacement of such structures.

Emphasis is directed specifically towards that class of disturbance on the ocean floor which is associated with the concept of turbidity currents as understood atpresent from our standpoint as marine geophysicists and geologists. Other, more or less related processes of mass sedimentary transport, including erosion, drifting and episodic deposition, also effect the ocean floor environment. As far as possible, this report seeks to discuss these complex phenomena in fair perspective, since their interactions with ocean floor structures may be analogous to those of turbidity currents.

ANALYSIS OF THE SUBJECT

The report is divided into seven parts:

PART I. Bibliography

PART II. The nature and definition of the major components of intra-oceanic sedimentary processes; the history and meaning of the concept of turbidity currents as episodic phenomena; a general "state of the art" examination of the apparent and possible interdependence of sedimentary processes in the sea; and the recognition of the activity of these processes.

PART III. Discussion of the physical and dynamic properties and depositional characteristics of turbidity currents, both as inferred from full-scale oceanic evidence, from theoretical considerations, and as measured and inferred from model studies.

PART IV. Brief review, with case histories, of the historical record of telephone cable breaks and of other events attributable to turbidity currents or the processes which initiate them; stratigraphic and sedimentological evidence of the activity and effects of turbidity currents; episodic events through time; and turbidity current models.

PART V. Critical appraisal of existing methods for the prediction of the general and specific occurrence of turbidity currents in the ocean; and suggestions for the development of prediction and monitoring techniques, including recommendations for immediate use.

PART VI. A study of the distribution of turbidity current deposits in the ocean in terms of both the environmental factors conducive to their initiation, and the physiographic provinces likely to be effected by their passage across the sea floor and by the deposition of their sedimentary suspensions; and studies of specific areas; and the overall distribution of hazardous and safe areas on the ocean floor.

PART VII General summary, recommendations and ideas.

ACKNOWLEDGEMENTS

This report presents a synthesis of the ideas of many scientists and includes some material heretofore unpublished.

We would like, in particular, to acknowledge helpful discussions with Stephen Eittreim, Robert Embley, David B. Ericson, John Ewing, Bruce C. Heezen, Charles D. Hollister, David Horn, Marcus Langseth, Eric D. Schneider, Edward Thorndike and Marek Truchan.

Hester Haring and the late Leah Miras provided generous help in compilation, editing and typing of the manuscript, and Ellen Miras assisted in the massive task of searching and compiling the references.

1. PREFACE TO SELECTED BIBLIOGRAPHY

We have examined more than 3000 items of published material which are directly or indirectly concerned with turbidity currents or with other dynamic proceses concerned with the question of the construction and occupation of submarine structures.

These 3000 items are a very large portion of the total existing body of pertinent information. Virtually all postdate the commencement of comprehensive scientific investigations of the ocean during the latter part of the nineteenth century, and most postdate the rapid increase of activity in the field during the 1930's, and particularly in the 1950's and 1960's.

We have selected from these 3000 items, about 350 articles in published journals and books, and the selection covers all of the major and most of the minor parts of relevant knowledge in the literature. Separately compiled is a listing of Books, Reprints, Volumes and Charts with designations to show the more important individual chapters. Some of these chapters are listed both in Sections 1 and 2.

Despite the fact that we have listed little more than 10% of the total number of items examined, few of the listed items are specifically pertinent to problems of submarine structural engineering and habitation. We have indicated those which, in the present state of knowledge, seem to be the more relevant.

2. ARTICLES PUBLISHED IN SCIENTIFIC JOURNALS AND BOOKS

References are listed and assigned numbers in alphabetical order, and the numbers are also gathered under the following ten major groups which permit selection of reference by subject matter. The numbers of those items which are either definitive, or appear to be particularly relevant or important, are underlined.

I. Inference of contemporary turbidity current activity on the ocean floor:

19, 32, 46, 47, 48, 58, 64, 71, 72, 75, 82, 84, 85, 87, 98, 99, 104, 108, 109, 112, 113, 115, 119, 122, 126, 133, 141, 142, 143, 144, 145, 149, 153, 154, 155, 156, 158, 159, 160, 161, 162, 163, 164, 165, 166 171, 176, 179, 185, 191, 202, 241, 252, 253, 271, 273, 288, 289, 293, 297, 302, 327, 332.

II. The hydrodynamics of the turbidity current flow regime:

6, 12, 23, 25, 27, 58, 65, 89, 96, 128, 142, 145, 146, 160, 179, 182, 196, 199, 204, 206, 207, 211, 212, 213, 215, 219, 220, 221, 223, 224, 228, 229, 230, 231, 232, 245, 248, 252, 255, 268, 276, 289, 291, 292, 293, 304, 312, 315, 327, 328, 331, 336.

References

III. Slope failure and submarine landslides:

18, 19, 32, 38, 64, 75, 85, 86, 106, 107, 132, 147, 155, 160, 162, 169, 170, 190, 191, 197, 202, 243, 258, 259, 261, 263, 281, 282, 288, 297, 302, 308, 322, 344.

IV. Experimental studies on turbidity currents:

1, 23, 24, 25, 26, 58, 65, 128, 150, 206, 207, 213, 216, 220, 221, 229, 230, 245, 246, 254, 255, 256, 257, 273, 274, 280, 303, 304, 328, 330, 331, 344,

V. General discussion of turbidity current deposits (turbidites):

A. Modern marine turbidites:

48, 72, 82, 108, 109, 110, 115, 149, 183, 185, 194, 222, 228, 268, 269, 271, 287, 289, 293, 301, 302, 332.

B. Ancient flysch deposits:

45, 76, 94, 95, 100, 107, 193, 223, 226, 227, 233, 244, 277, 309, 321, 334.

C. Theoretical or experimental deposits:

2, 3, 11, 17, 199, 205, 208, 264, 265, 290, 292, 336, 337, 339.

VI. Discussion of erosion of deep-sea sediments (drifts, scour, cut and fill):

8, 10, 12, 14, 79, 125, 134, 125, 137, 140, 144, 159, 160, 167, 172, 174, 177, 184, 217, 241, 249.

VII. Processes related to the initiation of turbidity currents (ie. slumping, tsunamis, earthquakes, tectonic activity, oversteepening by deposition):

4, 6, 18, 24, 29, 32, 34, 49, 52, 55, 60, 67, 74, 86, 98, 131, 136, 139, 145, 155, 161, 162, 164, 165, 178, 190, 202, 210, 216, 222, 235, 258, 263, 274, 288, 289, 318, 319, 320, 328.

VIII. On prediction and detection of turbidity currents:

98, 108, 114, 115, 179, 190, 191, 196, 261, 272, 279, 302, 316, 327.

IX. Distribution of oceanic sediments:

20, 22, 36, 70, 73, 99, 110, 113, 115, 117, 118, 119, 122, 123, 124, 126, 131, 140, 144, 149, 187, 250, 285, 289, 301, 302, 326, 332.

X. Physical properties of marine sediments related to slope failure and turbidite deposition:

37, 38, 39, 40, 42, 43, 129, 132, 186, 187, 188, 194, 261, 262, 283, 289, 294.

Bibliography

1. Albertson, M.L., D.B. Simons and E.V. Richardson, 1958. Discussion of the mechanics of ripple formation. Jour. Hydraul. Div. Amer. Soc. Civ. Engrs., 84, (HY 1).

2. Allen, J.R.L., 1962. Asymmetrical ripple marks and the origin of cross-stratification. Nature, 194, 167-169.

3. Allen, J.R.L., 1963. Asymmetrical ripple marks and the origin of waterlaid cosets of cross-strata. Jour. Geol. Lpool. Manchr., 3, 187-236.

4. Allen, J.R.L., 1964. Sedimentation in the modern delta of the River Niger, West Africa. In, Van Straaten, L.M.J.U., (Editor), Deltaic and Shallow Marine Deposits. Elsevier, Amsterdam, 26-34.

5. Allen, J.R.L., 1964. The Nigerian continental margin: bottom sediments, submarine morphology and geological evolution. Marine Geol., 1, 289-332.

6. Allen, J.R.L., 1964. Turbulent boundary layer. Sedimentology, 3, 89-108.

7. Allen, J.R.L., 1965. A review of the origin and characteristics of recent alluvial sediments. Sedimentology, 5(2), 89-191.

8. Allen, J.R.L., 1965. Sedimentation in the lee of small underwater sand waves: an experimental study. Jour. Geol., 73, 95-116.

9. Allen, J.R.L., 1966. On bed forms and palaeocurrents. Sedimentology, 6, 153-190.

10. Allen, J.R.L., 1968. On criteria for the continuance of flute marks, and their implications. Geol. en Mijnbouw, 47(1), 3-16.

11. Allen, J.R.L., 1968. On the character and classification of bed forms. Geol. en Mijnbouw, 47(3), 173-185.

12. Allen, J.R.L., 1968. The extent of erosion beneath sandstones in turbidites as revealed by scour marks. Jour. Sediment. Petrol., 38(4), 1379-1380.

13. Allen, J.R.L., 1968. The nature and origin of bed-form hierarchies. Sedimentology, 10(3), 161-182.

14. Allen, J.R.L., 1969. Erosional current marks of weakly cohesive mud beds. Jour. Sediment. Petrol., 32(2), 607-623.

15. Allen, P., J.R.L. Allen, R. Goldring and I.D. Maycock, 1960. Festoon bedding and "mud-with lenticles" lithology. Geol. Mag. 97, 261-263.

16. Ambraseys, N.N., 1960. The seismic sea wave of July 9, 1956, in the Greek Archipelago. Jour. Geophys. Res., 65, 1257-1265.

17. Anderson, A.G., 1953. The characteristics of sediment waves formed by flow in open channels. Proc. Midwestern Conf. Fluid Mech., 3rd, Minneapolis, Minn., 379-395.

18. Anderson, D.M., R.C. Reynolds and J. Brown, 1969. Bentonite debris flows in Northern Alaska. Science, 164(3876), 173-174.

19. Archanguelsky, A.D., 1930. Slides of sediments on the Black Sea bottom and the importance of this phenomenon for geology. Bull. Soc. Nat. Moscow (Sci. Geol.), 38, 38-80.

20. Arrhenius, G., 1952. Sediment cores from the East Pacific. Rept. Swed. Deep-Sea Exped., 5(1), 227p.

21. Arrhenius, G., 1963. Pelagic sediments. In, Hill, M.N. (Editor), The Sea, 3. Interscience Publishers, New York, 655-727.

22. Arrhenius, G., 1967. Deep sea sedimentation, a critical review of U.S. work 1963-1967. Report to XIV Gen. Assembly of the I.U.G.G.

23. Bagnold, R.A., 1955. Some flume experiments on large grains but little denser than the transporting fluid, and their implications. Proc. Inst. Civ. Engrs., 3, 174-205.

24. Bagnold, R.A., 1956. The flow of cohesionless grains in fluids. Phil. Trans. Roy. Soc. London, (Ser. A), 249, 235-297.

25. Bagnold, R.A., 1962. Autosuspension of transported sediment: turbidity currents. Proc. Roy. Soc., London, (Ser. A), 265, 315-319.

26. Bagnold, R.A., 1963. Beach and near-shore processes. Part 1. Mechanics of marine sedimentation. In, Hill, M.N. (Editor), The Sea, 3. Interscience Publishers, New York, 507-528.

27. Bagnold, R.A., 1968. Deposition in the process of hydraulic transport. Sedimentology, 10(1), 45-56.

28. Bailey, E.B., 1940. Submarine canyons. Nature, 146, 493.

29. Barazangi, M. and J. Dorman, 1969. World seismicity maps compiled from ESSA, Coast and Geodetic Survey, epicenter data. 1961-1967. Bull. Seism. Soc. Amer., 59(1), 369-380.

30. Bates, C.C., 1953. Rational theory of delta formation. Bull. Amer. Assoc. Petrol. Geologists, 37, 2119-2162.

31. Belderson, R.H. and A.H. Stride, 1969. Irish Sea-tidal currents near the sea bed. Nature, 222(5188), 74-75.

32. Bellaiche, G., 1968. Precisions apportées à la connaissance de la Pente Continentale et de la Plaine Abyssale à la suite de trois plongées en bathyscaphe (Archimède). Revue de Géographie Physique et de Géologie Dynamique, 10(2), 137-145.

33. Beneo, E., 1956. Accumuli terziari da risedimentazione (olistostroma) nell'Appennino centrale e frane sottomarine. Boll. Serv. Geol. Italia, 78, 291-321.

34. Benest, H., 1899. Submarine gullies, river outlets and fresh water escapes beneath the sea level. Geog. Jour. 14, 394-413.

35. Berger, W.H. and G.R. Heath, 1968. Vertical mixing in pelagic sediments. Jour. Mar. Res., 26(2), 134-143.

36. Bezrukov, P.L., 1960. Sedimentation in the northwestern Pacific Ocean. Internat'l. Geol. Congress, Copenhagen, Rept. Soviet Geologists, 45-58.

37. Bishop, A.W. and A.K.G. Eldin, 1950. Undrained triaxial tests on saturated sands and their significance in the general theory of shear strength. Geotechniques 2, 14-32.

38. Bishop, A.W. and N. Morgenstern, 1960. Stability coefficients for earth slopes. Geotechniques 10, 129-150.

39. Bjerrum, L., 1961. The effective shear strength parameters of sensitive clays. Proc. 5th Internat'l Conf. Soil Mech. and Found. Eng., Paris 1, 23-28.

40. Bjerrum. L.,S. Kringstad and O. Kummeneje, 1961. The shear strength of fine sand. Proc. 5th Internat'l. Conf. Soil Mech. and Found. Eng., Paris, 1, 29-38.

41. Blanc, J., 1954. Erosion et Sedimentation littorale actuelle dans le Détroit de Messine. Bull. de l'Inst. Océanog., 51(1051), 1-12.

42. Bogarin, J., 1961. Some aspects of the application of the theory of sediment transportation to engineering problems. Jour. Geophys. Res., 66, 3337-3346.

43. Boswell, P.G.H., 1951. The determination of the thixotropic limits of sediments. Jour. Geol. Lpool. and Manchr. 1(1), 1-22.

44. Bouma, A.H., 1964. Ancient and recent turbidites. Geol. en Mijnbouw, 43, 375-379.

45. Bouma, A.H., 1964. Turbidites. In, Bouma, A.H. and A. Brouwer (Editors), Turbidites. Elsevier, Amsterdam, 247-256.

46. Bourcart, J., 1950. Les hypothèses sur la genêse des gorges sous-marines (submarine canyons). Bull. Inform. Comité Central Océanog. Etude Côtes, 2(9), 317-330.

47. Bourcart, J., 1955. Les sables profonds de la Méditerranée. Arch. Sci. (Genève), 8(1), 5-13.

48. Bourcart, J., 1964. Les sables profonds de la Méditerranée occidentale. In, Bouma, A.H. and A. Brouwer (Editors), Turbidites. Elsevier, Amsterdam, 148-155.

49. Bourcart, J. et L.Glangeaud, 1958. Perturbations sous-marines et courants de turbidité résultant du tremblement de terre d'Orléansville. Bull. Inform. Comité Central Océanog. Etude Côtes, 10, 642-654.

50. Bourcart, J., M. Gennesseaux et E. Klimek, 1960. Ecoulements profonds de sables et de galêts dans la grande vallée sous-marine de Nice. Compt. Rend., 250, 3761-3765.

51. Bourcart, Jl, M. Gennesseaux, E. Klimek et Y. Le Calvez, 1960. Les sédiments des vallées sous-marines au large dans le Golfe de Gênes. Compt. Rend., 251, 1443-1445.

52. Braddock, R.D., 1969. On tsunami propagation. Jour. Geophys. Res., 74(8), 1952-1957.

53. Bradley, J.S., 1956. An application of the theory of sediment transport by turbidity currents to the dredging of navigable channels. Jour. Sediment. Petrol., 26, 119-124.

54. Bramlette, M.N. and W.H. Bradley, 1940. Geology and biology of North Atlantic deep-sea cores. U.S. Geol. Surv., Profess. Papers, 196-A, 1-34.

55. Brune, J.N., 1968. Seismic moment, seismicity and rate of slip along major fault zones. Jour. Geophys. Res., 73(2), 777-784.

56. Bryant, W.R., J. Antione, M. Ewing and B. Jones, 1968. Structure of Mexican Continental Shelf and Slope, Gulf of Mexico. Bull. Amer. Assoc. Petrol. Geologists, 52(7), 1204-1228.

57. Buffington, E.C., 1952. Submarine natural levees. Jour. Geol., 60, 473-479.

58. Buffington, E.C., 1961. Experimental turbidity currents on the sea floor. Bull. Amer. Assoc. Petrol. Geologists, 45, 1392-1400.

59. Campbell, A.U., 1967. Lamina, laminaset and bed set. Sedimentology, 8, 7-26.

60. Cann, J.R., 1968. Geological processes at mid-ocean crests. Geophysical Journal of the Royal Astro. Soc., 15(3), 331-341.

61. Cann, J.R., 1969. Spilites from the Carlsberg Ridge, Indian Ocean. Jour. Petrol., 10(1), 1-19.

62. Carsola, A.J., 1954. Submarine canyons on the arctic slope. Jour. Geol., 62, 606-610.

63. Carson, B. and D.A. McManus, 1969. Seismic reflection profiles across Juan de Fuca Canyon. Jour. Geophys. Res., 74(4), 1052-1060.

64. Chamberlain, T.K., 1964. Mass transport of sediment in the heads of Scripps submarine canyon, California. In, Miller, R.L. (Editor), Papers in Marine Geology, Macmillan, New York, 42-64.

65. Chang, Y.L., 1939. Laboratory investigation of flume traction and transportation. Trans. Amer. Soc. Civ. Engrs., Hydraulics Div. 83, Paper 1331.

66. Charnock, H., 1959. Tidal friction from currents near the sea bed. Jour. Geophys., 2, 215-221.

67. Chase, R.L. and E.T. Bunce, 1969. Underthrusting of the Eastern Margin of the Antilles by the floor of the western North Atlantic Ocean, and origin of the Barbados Ridge. Jour. Geophys. Res., 74(6), 1413-1420.

68. Clay, C.S. and P.A. Rona, 1964. On the existence of bottom corrugations in the Blake-Bahama Basin. Jour. Geophys. Res., 69, 231-234.

69. Cloet, R.L., 1954. Sandwaves in the southern North Sea and Persian Gulf. Jour. Inst. Navig., 7, 272.

70. Conolly, J.R., H.D. Needham and B.C. Heezen, 1967. Late Pleistocene and Holocene sedimentation in the Laurentian Channel. Jour. Geol., 75(2), 131-147.

71. Conolly, J.R. and C.C. Von Der Borch, 1967. Sedimentation and physiography of the sea floor south of Australia. Sediment. Geol., 1(2), 181-220.

72. Conolly, J.R. and M. Ewing, 1967. Sediments in the Puerto Rico Tranch. Jour. Sediment. Petrol., 37, 44-59.

73. Correns, C.W., 1939. Pelagic sediments of the North Atlantic Ocean. In, Trask, P.D. (Editor), Recent Marine Sediments, Bull. Amer. Assoc. Petrol. Geologists, 373-395.

74. Coulter, H.W. and R.R. Migliacco, 1966. Effects of the earthquake of March 27, 1964, at Valdez, Alaska. U.S. Geol. Soc. Prof. Paper 542-C.

75. Cousteau, J.Y., 1955. Diving through an undersea avalanche. Nat'l. Geograph. Mag., 107, 538-542.

76. Crook, K.A.W., 1959. Unconformities in turbidite sequences. Jour. Geol., 67, 710-713.

77. Crowell, J.C., 1952. Submarine canyons bordering central and southern California. Jour. Geol., 60, 58-83.

78. Crowell, J.C., 1957. Origin of pebbly mudstones. Bull. Geol. Soc. Amer., 68, 993-1009.

79. Daly, R.A., 1936. Origin of submarine canyons. Amer. Jour. Science, 31, 401-420.

80. Dangeard, L., 1960. Glissements de vase sous-marine et phenomenès de compaction. Observations faites en bathyscaphe. Compt. Rend. 251, 2224-2225.

81. Dangeard, L., 1961. A propos des phenomènes sous-marine profonds de glissement et de resédimentation. Cahiers Océanog., 13, 68-72.

82. Davies, D.K., 1968. Carbonate turbidites, Gulf of Mexico. Jour. Sediment. Petrol., 38(4), 1100-1109.

83. Diessell, C.F.K., 1966. The determination of the direction of transport of fluviatile arenites by orientation analysis of the detrital mica. Sedimentology 7, 167-178.

84. Dietz, R.S., 1953. Possible deep-sea turbidity-current channels in the Indian Ocean. Bull. Geol. Soc. Amer. 64, 375-378.

85. Dill, R.F., 1964. Sedimentation and erosion in Scripps submarine canyon head. In, Miller, R.L. (Editor), Papers in Marine Geology, Macmillan, New York, 23-41.

86. Dill, R.F., 1969. Earthquake effects on fill of Scripps submarine canyon. Bull. Geol. Soc. Amer., 80(2), 321-328.

87. Dill, R.F., R.S. Dietz and H. Stewart, 1954. Deep-sea channels and delta of the Monterey submarine canyon. Bull. Geol. Soc. Amer., 65, 191-194.

88. Dill, R.F. and L.A. Shumaker, 1961. Geological features of La Jolla canyon as revealed by dive number 83 of the bathyscaphe Trieste. U.S. Navy Electron. Lab., Tech. Mem., 516, 1-26.

89. Dott, Jr., R.H., 1963. Dynamics of subaqueous gravity depositional processes. Bull. Amer. Assoc. Petrol. Geologists 47, 104-129.

90. Dott, Jr., R.H. and J.K. Howard, 1962. Convolute lamination in non-graded sequences. Jour. Geol., 70, 114-121.

91. Drake, C.L., M. Ewing and G.H. Sutton, 1959. Continental margins and geosynclines: the east coast of North America north of Cape Hatteras. In, Ahrens, L.H., F. Press, K. Rankama and S.K. Runcorn (Editors), Physics and Chemistry of the Earth, 3, 110-198.

92. Duplaix, S., et A. Cailleux, 1960. La sédimentation dans la fosse de la Romanche. Geol. Rundschau, 49, 502-504.

93. Duplaix, M.S., W.D. Nesteroff and B.C. Heezen, 1965. Mineralogie comparée des sédiments du Tâge (Portugal) et de quelques sables profonds de la plaine abyssale correspondante. Deep-Sea Research, 12, 211-217.

94. Dzulynski, S., 1963. Directional structures in flysch. Studia Geol. Polon., 12, 136pp.

95. Dzulynski, S., M. Ksiazkiewicz and Ph. H. Kuenen, 1959. Turbidites in flysch of the Polish Carpathian Mountains. Bull. Geol. Soc. Amer., 70, 1089-1118.

96. Dzulynski, S. and A.J. Smith, 1963. Convolute lamination, its origin, preservation, and directional significance. Jour. Sediment. Petrol., 33, 616-627.

97. Dzulynski, S. and J.E. Sanders, 1962. Current marks on firm mud bottoms. Trans. Conn. Acad. Arts. Sci., 42, 57-96.

98. Elmendorf, C.H. and B.C. Heezen, 1957. Oceanographic information for engineering submarine cable systems. Bell System Tech. Jour., 36(5), 1047-1093.

99. Emery, K.O., 1960. Basin plains and aprons off southern California. Jour. Geol., 68, 464-479.

100. Emery, K.O., 1964. Turbidites-Precambrian to present. Studies Oceanog., Tokyo, 486-495.

101. Emery, K.O., 1965. Geology of the continental margin off eastern United States. In, Whittard, W.F. and R. Bradshaw (Editors), Submarine Geology and Geophysics, Colston Paper 17, 1-20.

102. Emery, K.O., 1968. Relict sediments of continental shelves of world. Bull. Amer. Assoc. Petrol. Geologists, 52(3), 445-464.

103. Emery, K.O. and R.D. Terry, 1956. A submarine slope off southern California. Jour. Geol., 64, 271-280.

104. Emery, K.O., J. Hülsemann and K.S. Rodolfo, 1962. Influence of turbidity currents upon basin waters. Limnol. Oceanog., 7, 439-446.

105. Emery, K.O. and D.A. Ross, 1968. Topography and sediments of a small area of the continental slope south of Marthas Vinyard. Deep-Sea Research, 15(4), 415-422.

106. Enos, P., 1965. Anatomy of a flysch-Middle Ordovician Cloridorme Formation, northern Gaspé Peninsula. Ph. D. Thesis, Yale Univ., 145pp.

107. Enos, P., 1969. Anatomy of a flysch. Jour. Sediment. Petrol., 39(2), 680-723

108. Ericson, D.B., M. Ewing and B.C. Heezen, 1951. Deep-sea sands and submarine canyons. Bull. Geol. Soc. Amer., 62, 961-965.

109. Ericson, D.B., M. Ewing and B.C. Heezen, 1952. Turbidity currents and sediments in North Atlantic. Bull. Amer. Assoc. Petrol. Geologists, 36, 489-511.

110. Ericson, D.B., M. Ewing, G. Wollin and B.C. Heezen, 1961. Atlantic deep-sea sediment cores. Bull. Geol. Soc. Amer., 72, 193-286.

111. Ericson, D.B. and G. Wollin, 1968. Pleistocene climates and chronology in deep-sea sediments, Science, 162, 1227-1234.

112. Ewing, M., 1953. Turbidity-current transportation of sediments in the Atlantic. Bull. Geol. Soc. Amer., 64, 1419-1420.

113. Ewing, M., 1965. The sediments of the Argentine Basin. Roy. Astron. Soc., 6, 10-27.

114. Ewing, M., B.C. Heezen, D.B. Ericson, J. Northrop and J. Dorman, 1953. Exploration of the Northwest Atlantic Mid-Ocean Canyon. Bull. Geol. Soc. Amer., 64, 865-868.

115. Ewing, M., D.B. Ericson and B.C. Heezen, 1958. Sediments and topography of the Gulf of Mexico. In, Weeks. L.G. (Editor), Habitat of Oil. Amer. Assoc. Petrol. Geologists, Tulsa, 995-1053.

116. Ewing, M. and J.I. Ewing, 1962. Rate of Salt-Dome growth. Bull. Amer. Assoc. Petrol. Geologists, 46(5), 708-709.

117. Ewing, M. and J.I. Ewing, 1964. Distribution of oceanic sediment. In, Yoshida, K. (Editor), Studies in Oceanography, Univ. of Tokyo Press, Tokyo, 525-537.

118. Ewing, M., J.I. Ewing and M. Talwani, 1964. Sediment distribution in the oceans: The Mid-Atlantic Ridge: Bull. Geol. Soc. Amer., 75(1), 17-36.

119. Ewing, M., W.J. Ludwig and J.I. Ewing, 1964. Sediment distribution in the oceans: The Argentine Basin. Jour. Geophys. Res., 69, 2003-2032.

120. Ewing, M. and E.M. Thorndike, 1965. Suspended matter in deep ocean water. Science, 147, 1291-1294.

121. Ewing, M., X. LePichon and J.I. Ewing, 1966. Crustal structure of the mid-ocean ridges. 4. Sediment distribution in the South Atlantic Ocean and the Cenozoic history of the Mid-Atlantic Ridge. Jour. Geophys. Res., 71, 1611-1636.

122. Ewing, M., J.I. Ewing, R.E. Houtz and R. Leyden, 1968. Sediment distribution in the Bellingshausen Basin. Symp. Antarctic Oceanog. 1966, 89-100.

123. Ewing, M., R. Houtz and J.I. Ewing, 1969. South Pacific sediment distribution. Jour. Geophys. Res., 74(10), 2477-2493.

124. Ewing, M., S. Eittriem, M. Truchan and J.I. Ewing, 1969. Sediment distribution in the Indian Ocean. Deep-Sea Research, 16(3), 231-248.

125. Ewing, M., S. Eittriem, J. Ewing and X. Le Pichon (in press) Sediment transport and distribution in the Argentine basin. 3. Nepheloid layer and processes of sedimentation. Physics and Chemistry of the Earth, Vol. 8, Pergamon Press Ltd., Oxford

126. Ewing, J. I., J. L. Worzel and M. Ewing, 1962. Sediments and oceanic structural history of the Gulf of Mexico. Jour. Geophys. Res., 67(6), 2509-2527.

127. Ewing, J. I., X. LePichon and M. Ewing, 1963. Upper stratification of Hudson apron region. Jour. Geophys. Res., 68, 6303-6316.

128. Fahnestock, R.K. and W.L. Haushild, 1962. Flume studies of the transport of pebbles and cobbles on a sand bed. Bull. Geol. Soc. Amer., 73, 1431-1436.

129. Fairbridge, R.W., 1967. Phases of diagenesis and authigenesis. In, Larsen, G. and G.V. Chilingar (Editors), Diagenesis in Sediments: Developments in Sedimentology, 8, 19-89.

130. Feth, J.H., 1954. Convoluted bedding suggests turbidity-current deposition in Pleistocene Lake Bonneville, Utah (abstract). Bull. Geol. Soc. Amer., 65, 1251.

131. Fisk, H.N., E. McFarlan, C.R. Kolb and L.J. Wilbert 1954. Sedimentary framework of the modern Mississippi delta. Jour. Sediment. Petrol., 24, 76-99.

132. Fisk, H.N. and B. McClelland, 1959. Geology of the continental shelf off Louisiana; its influence on offshore foundation design. Bull. Geol. Soc. Amer. 70, 1369-1394.

133. Forel, F.A., 1887. Le ravin sous-lacustre du Rhône dans le lac Léman. Bull. Soc. Vaudoise Sci. Nat., 23, 85-107.

134. Fox, P.J. and B.C. Heezen, 1965. Sands of the Mid-Atlantic Ridge. Science, 149, 1367-1370.

135. Fox, P.J., B.C. Heezen and A. Harian, 1968. Abyssal anti-dunes. Nature, 220, 270-272.

136. Francis, T.J.G., 1968. The detailed seismicity of Mid-Oceanic ridges. Earth and Planetary Science Letters, 4(1), 39-46.

137. Gabriel, V.G., 1957. Turbidity current and its importance in erosion and deposition of sediments. World Oil 144(2), 81.

138. Gennesseux, M., 1962. Les canyons de la baie des Anges, leur remplissage sédimentaire et leur rôle dans la sédimentation profonde. Compt Rend., 254, 2409-2411.

139. Gibson, R.E., 1958. The progress of consolidation in a clay layer increasing in thickness with time. Geotechnique 8, 171-182.

140. Goodell, H.G., 1965. The marine geology of the Southern ocean. 1. Pacific-Antarctic and Scotia Basins. Contribution 11, Sedimentology Res. Lab., Florida State Univ.

141. Gorsline, D.S. and K.O. Emery, 1959. Turbidity current deposits in San Pedro and Santa Monica basins off southern California. Bull. Geol. Soc. Amer., 70, 279-290.

142. Gould, H.R., 1951. Some quantitative aspects of Lake Mead turbidity currents. Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 2, 34-52.

143. Gould, H.R., 1960. Turbidity currents. In, Smith, W.O., C.P. Vetter, G.B. Cummings (and others, Editors), Comprehensive Survey of Sedimentation in Lake Mead 1948-1949. U.S. Geol. Surv., Profess. Papers, 295, 201-207.

144. Griggs, G.B., A.G. Carey, Jr. and L.D. Kulm, 1969. Deep-sea sedimentation and sediment-fauna interaction in Cascadia Channel and on Cascadia Abyssal Plain. Deep-Sea Research, 16(2), 157-170.

145. Grover, N.C. and C.S. Howard, 1938. The passage of turbid water through Lake Mead. Trans. Amer. Soc. Civ. Engrs., 103, 720-790.

146. Gwinn, V.E., 1964. Deduction of flow regime from bedding character in conglomerates and sandstones. Jour. Sediment. Petrol., 34, 656-658.

147. Hadding, A., 1931. On subaqueous slides. Geol. Fören. Stockh. Förh., 53, 377

148. Hamilton, E.L., 1956. Sunken islands of the Mid-Pacific Mountains. Geol. Soc. Amer. Memoir, 64, 97pp.

149. Hamilton, E.L., 1967. Marine geology of abyssal plains in the Gulf of Alaska. Jour. Geophys. Res., 72, 4189-4213.

150. Hampton, Monty A. and I.P. Colburn, 1967. Directional features of an experimental turbidite fan, A report of progress. Jour. Sediment. Petrol. 37(2), 509-513.

151. Harms, J.C. and R.K. Fahnstock, 1965. Stratification, bed forms and flow phenomenon (with examples from the Rio Grande). In, Middleton, B.V. (Editor), Primary Sedimentary Structures and their Hydrodynamic Interpretation. Amer. Assoc. Petrol. Geologists, Spec. Publ. 12, 84-115.

152. Harvey, J.G., 1966. Large sand waves in the Irish Sea. Marine Geol., 4, 49-56.

153. Hayter, P.J.D., 1960. The Ganges and Indus submarine canyons, Deep-sea Research, 6, 184-186.

154. Heezen, B.C., 1955. Turbidity currents from the Magdalena River, Colombia (abstract), Bull. Geol. Soc. Amer., 66, 1572.

155. Heezen, B.C., 1956. Corrientes de turbidez del Rio Magdalena. Bull. Soc. Geog., Colombia, 51/52, 135-143.

156. Heezen, B.C., 1957. 1908 Messina earthquake, tsunami, and turbidity current (abstract). Bull. Geol. Soc. Amer., 68, 1743.

157. Heezen, B.C., 1959. Deep-sea erosion and unconformities. Jour. Geol., 67, 713-714.

158. Heezen, B.C., 1959.[b] Modern turbidity currents. Intern. Oceanog. Congr., 1st Amer. Assoc. Advan. Sci, Preprints, 458-460.

159. Heezen, B.C., 1959.[c] Modern turbidity currents. Eclogae Geol. Helv., 51(3), 521-522.

160. Heezen, B.C., 1959.[d] Dynamic processes of abyssal sedimentation: erosion, transportation, and redeposition on the deep-sea floor. Jour. Geophys., 2, 142-163.

161. Heezen, B.C., 1963. Turbidity currents. In, Hill, M.N. (Editor), The Sea. 3 The Earth Beneath the Sea. Interscience, New York, 742-775.

162. Heezen, B.C. and M. Ewing, 1952. Turbidity currents and submarine slumps, and the 1929 Grand Banks earthquake. Amer. Jour. Science, 250, 849-873.

163. Heezen, B.C., M. Ewing and D.B. Ericson, 1954. Reconnaissance survey of the abyssal plain south of Newfoundland. Deep-Sea Research, 2, 122-133.

164. Heezen, B.C., D.B. Ericson and M. Ewing, 1954. Further evidence for a turbidity current following the 1929 Grand Banks earthquake. Deep-Sea Research, 1, 193-202.

165. Heezen, B.C. and M. Ewing, 1955. Orleansville earthquake and turbidity currents. Bull. Amer. Assoc. Petrol. Geologists, 39, 2505-2514.

166. Heezen, B.C. M. Ewing and R.J. Menzies, 1955. The influence of submarine turbidity currents on abyssal productivity. Oikos, 6, 170-182.

167. Heezen, B.C., R. Coughlin and W.C. Beckmann, 1960.[a] Equatorial Atlantic mid-ocean canyon. Geol. Soc. Amer., Prog. Ann. Meeting, 118.

168. Heezen, B.C., R. Coughlin and W.C. Beckmann, 1960.[b] Equatorial Atlantic mid-ocean canyon. Bull. Geol. Soc. Amer. (Abstract), 71, 1886.

169. Heezen, B.C. and C.L. Drake, 1963. Gravity tectonics, turbidity currents and geosynclinal accumulations in the continental margin of eastern North America. In, Carey, S.W. (Editor), Syntaphral Tectonics Symposium, Univ. of Tasmania, Dept. of Geol. Publications, D1-D10.

170. Heezen, B.C. and C.L.Drake, 1964. Grand Banks Slump. Bull. Amer. Assoc. Petrol. Geologists, 48, 221-225.

171. Heezen, B.C., R.J. Menzies, E.D. Schneider, M. Ewing and N.C.L. Granelli, 1964. Congo submarine canyon. Bull. Amer. Assoc. Petrol. Geologists, 48, 1126-1149.

172. Heezen, B.C. and C.D. Hollister, 1964. Deep-sea current evidence from abyssal sediments. Mar. Geol., 1, 141-174.

173. Heezen, B.C. and T.L. Holcombe, 1965. Patterns of relative relief, slope, and topographic texture in North Atlantic Ocean. Navships N00024-67-C-1186, Technical Report #8

174. Heezen, B.C., C.D. Hollister and W.F. Ruddiman, 1966. Shaping of the continental rise by deep geostrophic contour currents. Science, 152, 502-508.

175. Heezen, B.C., E.D. Schneider and O.H. Pilkey, 1966. Sediment transport by the Antarctic Bottom Current on the Bermuda Rise. Nature, 211, 611-612.

176. Heezen, B.C., M.Ewing and G.L. Johnson, 1966. The Gulf of Corinth floor. Deep-Sea Research, k3, 381-411.

177. Heezen, B.C. and G.L. Johnson, 1969. Mediterranean undercurrent and microtopography west of Gibraltar. Bull. Inst. Oceano., 67, 95pp.

178. Heezen, B.C. and G.L. Johnson, 1969. Alaskan submarine cables: A struggle with a harsh environment. Arctic (in press).

179. Hersey, J.B., 1965. Sediment ponding in the deep sea. Bull. Geol. Soc. Amer. 76(11), 1251-1260.

180. Hesse, R., 1961. Das Flyschgebiet des Zwiesel westlich von Bad Tölz. Z. Deut. Geol. Ges., 113, 293-304.

181. Hills, E.S., 1958. Load casts and flame structures. Geol. Mag., 95, 171-172 (letter).

182. Hinze, J.O., 1960. On the hydrodynamics of turbidity currents. Geol. Mijnbouw, 39, 18-25.

183. Hollister, C.D. and B.C. Heezen, 1964. Modern graywacke-type sands. Science, 146, 1573-1574.

184. Hollister, C.D. and B.C. Heezen, 1967. The floor of the Bellingshausen Sea. In, Hersey, J.B. (Editor), Underwater Photography, Johns Hopkins Univ. Press, 177-190.

185. Holtedahl, H., 1965. Recent Turbidites in the Hardangerfjord, Norway. In, Whittard, W.F. and R. Bradshaw (Editors), Submarine Geol. and Geophys., The Colston Papers 17, Butterworths, London, 107-142.

186. Horn, D.R., B.M. Horn and M.N. Delach, 1968. Correlation between acoustical and other physical properties of deep-sea cores. Jour. Geophys. Res., 73(6), 1939-1957.

187. Horn, D.R., N.N. Delach and B.M. Horn, 1969. Distribution of volcanic ash layers and turbidites in the North Pacific. Bull. Geol. Soc. Amer., 80, 1715-1724.

188. Horn, D.R., B.M. Horn and M.N. Delach, 1970. Sedimentary provinces of the North Pacific. Geol. Soc. America Mem. 126 (in press).

189. Horton, R.E., 1941. Sheet erosion-past and present. Trans. Amer. Geophys. Union, 22, 299-305.

190. Houtz, R.E., 1962. The 1953 Suva earthquake and tsunami. Bull. Seism. Soc. Amer. 52, 1-12.

191. Houtz, R.E. and H.W.Welman, 1962. Turbidity current at Kadavu Passage, Fiji. Geol. Mag., 99, 57-62.

192. Hoyt, J.H., B.L. Oostdam and D.D. Smith, 1969. Offshore sediments and valleys of the Orange River (South and Southwest Africa).Marine Geology.7(1),69-84

193. Hsu, K.J., 1964. Cross-laminations in graded bed sequences. Jour. Sediment. Petrol., 34, 379-388.

194. Hubert, J.F., 1964. Textural evidence for deposition of many western North Atlantic deep-sea sands by ocean bottom currents rather than turbidity currents. Jour. Geol., 72, 757-785.

195. Hunt, J.M., 1967. Red Sea: detailed survey of hot-brine areas. Science, 156 (3774), 514-516.

196. Hurley, R.J., 1960. The geomorphology of abyssal plains in the Northeast Pacific Ocean. Thesis, Univ. of California, Los Angeles, 173pp.

197. Hutchinson, J.N., 1961. A landslide on a thin layer of quick clay at Furre, central Norway. Geotechnique,11, 69-94.

198. Il'in, A.V., K.D. Sabinin and V.A. Shulepov, 1968. Some methods and results of the statistical processing of bottom relief data. Oceanology 8(3), 344-352.

199. Inman, D.L., 1949. Sorting of sediments in the light of fluid mechanics. Jour. Sediment. Petrol., 19, 51-70.

200. Inman, D.L., 1963. Physical properties and mechanics of sedimentation. In, Shepard, F.P., Submarine Geology. 2nd. edn., Harper, New York, 101-151.

201. Jerlov, N.G., 1953. Particle distribution in the ocean. Repts. Swed. Deep-Sea Exped., 3, 73-97.

202. Johnson, G.L. and B.C. Heezen, 1969. Natural hazards to submarine cables. Ocean Engring. (in press).

203. Johnson, M.A., 1962. Turbidity currents. In, Barnes, H. (Editor), Oceanography and Marine Biology, 2, Allen and Unwin , London.

204. Johnson, M.A., 1966. Application of theory to an Atlantic turbidity-current path. Sedimentology, 7(2), 117-130.

205. Jopling, A.V., 1962. Origin of regressive ripples explained in terms of fluid mechanics processes. U.S. Geol. Surv., Profess. Papers, 429D,(15/17), 408pp.

206. Jopling, A.V., 1963. Hydraulic studies on the origin of bedding. Sedimentology, 2, 115-121.

207. Jopling, A.V., 1964. Laboratory study of sorting processes related to flow separation. Jour. Geophys. Res. 69, 3403-3418.

208. Jopling, A.V., 1966. Origin of cross-laminas in a laboratory experiment, Jour. Geophys. Res., 71, 1123-1134.

209. Jordan, G.F., 1962. Large submarine sand waves. Science, 136, 839-848.

210. Julien, A., 1955. Influence du séisme d'Orléansville sur les cables de la Mediterranée (9 et 10 septembre, 1954). Bull. Inform. Comité Central Oceanog. Etude Côtes, 7, 196-200.

211. Kao, T.W., 1965a. Stability of two-layer viscous stratified flow down an inclined plane. The Physics of Fluids, 8(5), 812-820.

212. Kao, T.W., 1965b. The phenomenon of blocking in stratified flows. Jour. Geophys. Research, 70(4), 815-822.

213. Kennedy, J.F., 1963. The mechanics of dunes and antidunes in erodible-bed channels. Jour. of Fluid Mech. 16(4), 521-544.

214. Klein, G. de V., 1967. Paleocurrent analysis in relation to modern marine sediment dispersal patterns. Bull. Amer. Assoc. Petrol. Geologists, 51(3), 366-382.

215. Knapp, R.T., 1943. Density currents: their mixing characteristics and their effect on the turbulence structure of the associated flow. Proc. Hydraul. Conf., 2nd, 1942. Bull. Univ. Iowa Studies Eng., 27, 289-306.

216. Knauss, J., 1965. A technique for measuring deep ocean currents close to the bottom with an unattached current meter, and some preliminary results. Jour. Mar. Res., 23, 237-245.

217. Kuenen, Ph. H., 1937. Experiments in connection with Daly's hypothesis on the formation of submarine canyons. Leidse Geol. Mededel., 8 , 327-335.

218. Kuenen, Ph. H., 1950a. Marine Geology, Wiley, New York, 568pp.

219. Kuenen, Ph. H., 1950b. Turbidity currents of high density. Intern. Geol. Congr., 18th, London, 1948. Rept., 8, 44-52.

220. Kuenen, Ph. H., 1951. Properties of turbidity currents of high density. In, Hough, J.L. (Editor), Turbidity currents. Soc. Econ. Paleont. Miner., Spec. Publ., 2, 14-33.

221. Kuenen, Ph. H., 1952. Estimated size of the Grand Banks turbidity current. Amer. Jour. Science, 250, 874-884.

222. Kuenen, Ph. H., 1953a. Origin and classification of submarine canyons. Bull. Geol. Soc. Amer., 64, 1295-1314.

223. Kuenen, Ph. H., 1953[b] Significant features of graded bedding. Bull. Amer. Assoc. Petrol. Geologists, 37, 1044-1066.

224. Kuenen, Ph. H., 1956. The difference between sliding and turbidity flow. Deep-Sea Research, 3, 134-139.

225. Kuenen, Ph. H., 1957. Sole markings of graded greywacke beds. Jour. Geol., 65, 231-258.

226. Kuenen, Ph. H., 1960. Sand-its origin, transportation, abrasion and accumulation. (Du Toit Mem. Lecture 6, 1959), Ann. Geol. Soc. So. Africa,62,1-33

227. Kuenen, Ph. H., 1964. Deep-sea sands and ancient turbidites. In, Bouma, A.H. and A. Brouwer (Editors), Turbidites. Elsevier, Amsterdam, 3-33.

228. Kuenen, Ph. H., 1965[a] The shell pavement below oceanic turbidites. Marine Geol., 2, 236-246.

229. Kuenen, Ph. H., 1965[b] Experiments in connection with turbidity currents and clay suspension. In, Whittard, W.F. and R. Bradwhaw (Editors), Submarine Geology and Geophysics, Colston Papers 17, 47-74.

230. Kuenen, Ph. H., 1966. Experimental Turbidite lamination in a circular flume. Jour. Geol., 74, 523-545.

231. Kuenen, Ph. H., 1967. Emplacement of flysch-type sand beds. Sedimentology, 9(3), 203-244.

232. Kuenen, Ph. H. and C.I. Migliorini, 1950. Turbidity currents as a cause of graded bedding. Jour. Geol., 58, 91-127.

233. Kuenen, Ph. H. and H.W. Menard, 1952. Turbidity currents graded and non-graded deposits. Jour. Sediment. Petrol., 22, 83-96.

234. Kuenen, Ph. H. and F.L. Humbert, 1964. Bibliography of turbidity currents and turbidites. In, Bouma, A.H and A. Brouwer (Editors), Turbidites. Elsevier, Amsterdam, 222-246.

235. Lachenbruch, A.H. and B.V. Marshall, 1968. Heat flow and water temperature fluctuations in the Denmark Strait. Jour. Geophys. Res., 73(18), 5829-5847.

236. Lamont, A., 1957. Slow anti-dunes and flow marks. Geol. Mag., 94, 472-480.

237. Latham, G.V., R.S. Anderson and M. Ewing, 1967. Pressure variations produced at the ocean bottom by hurricanes. Jour. Geophys. Res., 72(22), 5693-5704.

238. Laughton, A.S., 1960. An interplain deep-sea channel system. Deep-Sea Research, 7, 75-88.

239. Laughton, A.S., 1962. Discrete hyperbolic echoes from an otherwise smooth deep-sea floor. Deep-Sea Research, 9, 218-219.

240. Laughton, A.S., 1963. Microtopography. In, Hill, M.N.(Editor), The Sea, 3. Interscience, New York, 437-473.

241. Laughton, A.S., 1968. New evidence of erosion in the deep ocean floor.Deep Sea Research, 15(1), 21-29.

242. Lundgren, H. and I.G. Jonsson, 1964. Shear and velocity distribution in shallow channels. Jour. of Hydrol. Div., ASCE, 90, (HY6), 257-264.

243. Maxwell, J.C., 1959. Turbidite, tectonic and gravity transport, northern Apennine Mountains, Italy. Bull. Amer. Assoc. Petrol. Geol., 43(11), 2701-2719.

244. McKee, E.D., 1957. Primary structures of some recent sediments. Bull. Amer. Assoc. Petrol. Geologists, 41, 1704-1747.

245. McKee, E.D., 1965. Experiments on ripple lamination. In, Middleton, G.V. (Editor), Primary sedimentary structures and their hydrodynamic interpretation. Soc. Econ. Paleontologists Mineralogists,Spec.Pub.,12,66-83

246. McKee, E.D. and M. Goldberg, 1969. Experiments on formation of contorted structure in mud. Bull. Geol. Soc. Amer., 80(2), 231-243.

247. McGregor, B.A., 1968. Features at the Sohm Abyssal Plain terminus. Marine Geol. 6(5), 401-414.

248. Menard, H.W., 1950. Sediment movement in relation to current velocity. Jour. Sediment. Petrol., 20, 148-160.

249. Menard, H.W., 1952. Deep ripple marks in the sea. Jour. Sediment. Petrol., 22, 3-9.

250. Menard, H.W., 1955. Deep-sea channels, topography and sedimentation. Bull. Amer. Assoc. Petrol. Geologists, 39, 236-255.

251. Menard, H.W., 1956. Archipelagic aprons. Bull. Amer. Assoc. Petrol. Geol., 40, 2195-2210.

252. Menard, H.W. and J.C. Ludwick, 1951. Applications of hydraulics to the study of marine turbidity currents. In, Hough, J.L. (Editor), Turbidity currents and the transportation of coarse sediments to deep water, a Symposium. Soc. Econ. Paleontologists, Mineralogists, Spec. Publ., 2, 2-13.

253. Menard, H.W., S.M. Smith and R.M. Pratt, 1965. The Rhône deep-sea fan. In, Whittard, W.F. and R. Bradshaw (Editors), Submarine Geology and Geophysics, Colston Papers, 17, Butterworths, London, 271-285.

254. Middleton, G.V., 1965. Antidune cross-bedding in a large flume. Jour. Sediment. Petrol., 35(4), 922-927.

255. Middleton, G.V., 1966a. Small-scale models of turbidity currents and the criterion for autosuspension. Jour. Sediment. Petrol., 36, 202-208.

256. Middleton, G.V., 1966b. Experiments on density and turbidity currents. I. Motion of the head. Can. Jour. Earth Sciences, 3, 523-546.

257. Middleton, G.V., 1966c. Experiments on density and turbidity currents. II. Uniform flow of density currents, Can. Jour. Earth Sciences, 3, 627-637.

258. Monroe, J.N., 1969. Slumping structures caused by organically derived gases in sediments. Science, 164(3886), 1394-1395.

259. Moore, D.G., 1961. Submarine slumps. Jour. Sediment. Petrol, 31, 343-357.

260. Moore, D.G., 1962. Bearing strength and other physical properties of some shallow and deep-sea sediments from the North Pacific. Bull. Geol. Soc. Amer., 73, 1163-1166.

261. Moore, J.C., 1964. Giant submarine landslides on the Hawaiian Ridge. U.S. Geol. Survey Proj. Paper 501-D, 95-98.

262. Moore, D.G. and G. Shumway,1959. Sediment thickness and physical properties: Pigeon Pt. shelf, Cal. J.Geophys.Res.,64,367-374.

263. Morgenstern,N.R., 1967. Submarine slumping & the initiation to turbidity currents. In, Richards,A.F.(Ed.),Marine Geotechnique, Univ. of Illinois Press, Urbana, 189-220.

264. Moss, A.J.,1962. The physical nature of common sandy & pebbly deposits, I. Amer. J. Science, 260, 337-373.

265. Moss, A.J., 1963. The physical nature of common sandy & pebbly deposits, II. Amer. J. Science 261, 297-343.

266. Mutti,E.,1965. Submarine flood tuffs(ignimbrites) associated turbidites in Oligocene deposits of Rhodes Island(Greece). Sedimentology, 5, 65-88.

267. Needham,H.D.,D. Habib & B.C. Heezen,1969. Upper carboniferous palynomorphs as a tracer of red sediment dispersal patterns in the northwest Atlantic. J. Geol. 77, 113-120.

268. Nesteroff,W.D.,1961. La"séquence-type" dans les turbidites terrigènes modernes. Révue Géogr. Phys.Géol.Dyn.,4,263-268.

269. Nesteroff,W.D. & B.C. Heezen,1963. Essais de comparison entre les turbidites modernes et le flysch. Révue de Géogr. Phys. et de Géol. Dyn. 5(2), 115-127.

270. Nesteroff,W.D.,B.C. Heezen & J.R. Conolly,1966. Decouverte de turbidites heterogenes dans le bassin Austral.Acad.Sci.,Paris,263, 1052-1054.

271. Norin,E.,1958. The sediments of the Central Tyrrhenian Sea. Rept. Swed. Deep-Sea Exped., 8, 136pp.

272. Nota,D.J.G.,1958. Sediments of the Western Guiana shelf. Rpts. of the Orinoco Shelf Exped. 2,Med. Landbouwhogeschool Wageningen, 58, 1-98.

273. Nowroozi,A.A.,M. Ewing,J.E. Kafe & M. Fliegel,1968. Deep ocean current & its correlation with the ocean tide off the coast of northern Calif. J. Geophys. Res., 73(6), 1921-1932.

274. Oulianoff,N.,1958. Effet des vibrations experimentales sur la sédimentation. Compt. Rend. 247, 2404-2407.

275. Passega,R.,1957. Texture as characteristic of clastic deposition. Bull.Amer.Assoc.Petrol.Geol.,41,1952-1984.

276. Plapp, J.E. & J.P. Mitchell, 1960. A hydrodynamic theory of turbidity currents. J. Geophys. Res.,65,983-992.

277. Potter,P.E.,1967. Sand bodies & sedimentary environments: A Review. Bull. Amer. Assoc. Petrol. Geol.,51(3),337.

278. Rast,N.,1962. Textural evidence for the origin of ignimbrites. L. & M. J. Geol., 3(1), 97-108.

279. Reid,J., Jr., H. Stommel, E.D. Stroup & B.A. Warren,1968. Detection of a deep boundary current in the western south Pacific. Nature, 217(5132), 937.

280. Rees, A.I.,1966. Some flume experiments with a fine silt. Sedimentology, 6, 208-240.

281. Reimnitz,E. & N.F. Marshall,1965. Effects of the Alaska earthquake & tsunami on recent deltaic sediments. J. Geophys. Res. 70, 2363-2376.

282. Renz,O., R. Lakeman & E. van der Meulen,1955. Submarine sliding in Western Venezuela. Bull. Amer. Assoc. Petrol. Geol., 39, 2053-2067.

283. Richards,A.F., 1962. Investigations of deep-sea sediment cores, II: Mass physical properties. U.S. Navy Hydrographic Office Tech. Rept. 102.

284. Richardson,P.D.,1968. The generation of scour marks near obstacles. J. Sediment. Petrol., 38(4), 965-970.

285. Rodolfo,K.S.,1969. Bathymetry & marine geology of the Andaman basin, & tectonic implications for southeast Asia. Bull. Geol. Soc. Amer.,80(7), 1203-1230.

286. Rowe,G.T. & R.J. Menzies,1968. Deep bottom currents off the coast of North Carolina. Deep-Sea Research,15,711-719.

287. Rusnak,G.A. & W. Nesteroff,1964. Modern turbidites; terrigenous abyssal plain versus bioclastic basin. In, Miller, R.L.(Ed.) Papers in Marine Geology,Macmillan,N.Y., 488-507.

288. Ryan,W.B.F. & B.C. Heezen,1965. Ionian Sea submarine canyons & the 1908 Messina turbidity current.Bull. Geol. Soc. Amer. 76, 915-932.

289. Ryan,W.B.F., F. Workum, Jr. & J.B. Hersey, 1965. Sediments on the Tyrrhenian abyssal plain. Bull. Geol. Soc. Amer., 76(11), 1261-1282.

290. Sanders,J.E.,1960. Origin of convolute laminae. Geol. Mag., 97, 409-421.

291. Sanders,J.E., 1963. Concepts of fluid mechanics provided by primary sedimentary structures. J. Sediment. Petrol., 33, 173-179.

292. Sanders,J.E.,1965. Primary sedimentary structures formed by turbidity currents & related resedimentation mechanism. In, Middleton,G.V.(Ed.), Primary Sedimentary Structures and their Hydrodynamic Environment.Am.Assoc.Pet.Geol.,Spec. Publ.12, 192-219.

293. Schneider,E.D. & B.C. Heezen,1966. Sediments of the Caicos outer ridge,the Bahamas. Bull.Geol.Soc.Amer.,77,1381-1398.

294. Schreiber,B.C.,1968. Sound velocity in deep sea sediments. J. Geophys. Res., 73(4), 1259-1268.

295. Scholl,D.W.,E.C. Buffington & D.M. Hopkins,1968. Geologic history of the continental margin of North America in the Bering Sea. Marine Geol., 6(4), 297-330.

296. Seilacher,A.,1962. Paleontological studies on turbidite sedimentation and erosion. J. Geol., 70, 227-234.

297. Shepard,F.P.,1951. Mass movements in submarine canyon heads. Trans. Amer. Geophys. Un., 32, 405-418.

298. Shepard,F.P.,1955. Delta front valleys bordering the Mississippi distributaries. Bull.Geol.Soc.Amer.,66, 1489-1498.

299. Shepard,F.P.,1964. Criteria in modern sediments useful in recognizing ancient sedimentary environments. <u>In</u>, Van Straaten,L.M.J.V. (Ed.),Deltaic & shallow marine deposits. Elsevier,Amsterdam,1-25.

300. Shepard,F.P.,1965. Ir[illegible]tance of submarine valleys in funneling sediments to the de[illegible] sea. <u>In</u>,Sears,M.(Ed.),Progress in Oceanography, 3, Pe[illegible]on Press, 321-332.

301. Shepard,F.P. & G. Einsele,1962. Sedimentation in San Diego Trough and contributing submarine canyons. Sedimentology,1, 81-133.

302. Shepard,F.P.,R.F. Dill & U. Von Rad,1969. Physiography and Sedimentary processes of La Jolla submarine fan and fan valley,Calif. Bull. Amer. Assoc. Petrol. Geol., 53(2), 290-298.

303. Simons, D.B.,E.V. Richardson & M.L. Albertson,1961. Flume studies using medium sand (0.45 mm). U.S. Geol. Survey, Water Supply Papers, 1498 A, 76pp.

304. Simons, D.B.,E.V. Richardson & W.L. Haushild,1963. Some effects of fine sediment on flow phenomena. U.S. Geol. Survey,Water Supply Paper 1498 G, 47pp.

305. Simons,D.B.,E.V. Richardson & G.F. Nordin,Jr.,1965. Sedimentary structures generated by flow in alluvial channels. <u>In</u>, Middleton, G.V.(Ed.), Primary Sedimentary Structures and their Hydrodynamic Interpretation. Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 12, 34-52.

306. Starke,G.W. & A.D. Howard,1968. Polygenetic origin of Monterey Submarine Canyon. Bull. Geol. Soc. Amer., 79, 813-826.

307. Sternberg,R.W. & J.S. Graeger,1965. An instrument system to measure boundary-layer conditions at the sea floor. Marine Geol., 3, 475-482.

308. Stanley,D.J., 196 . Large mudstone nucleus sandstone spheroids in subma[illegible]ine channel deposits. J.Sediment Pet[illegible]ol.,34(3),672-676.

309. Stanley,D.J.,1967. Comparing patterns of sedimentation in some modern & ancient submarine canyons. Earth & Planetary Science Letters, 3, 371-380.

310. Stanley,D.J. & A.H. Bouma,1964. Methodology & paleogeographic interpretation of flysch formations: a summary of studies in the Maritime Alps. In, Bouma,A.H. & A. Brouwer(Ed.),Turbidites. Elsevier,Amsterdam, 34-64.

311. Sternberg,R.W.,1967. Measurements of sediment movement & ripple migration in shallow marine environment. Marine Geol,5(3), 195-205.

312. Stoneley,R.,1956. On Turbidity Currents. Nat'l Bur. Standards (U.S.),Washington,D.C.,16pp.

313. Stride,A.H.,1963. Current-swept sea floors near the southern half of the British Isles. Quart.Geol.Soc. London,119,175-199.

314. Sullwold,Jr.,H.H.,1961. Turbidites in oil exploration. In, Peterson,J.A. & J.C. Osmund(Ed.), Geometry of sandstone bodies. Bull. Amer. Assoc. Petrol. Geol., 63-81.

315. Sutherland,A.J.,1967. Proposed mechanism for sediment entrainment by turbulent flows. J. Geophys. Res.,72(24),6183-6194.

316. Swallow,J.C. & L.V. Worthington,1961. An observation of a deep counter-current in the western North Atlantic. Deep-Sea Research, 8, 1-19.

317. Swallow,J.C. & L.V. Worthington,1969. Deep currents in the Labrador Sea. Deep-Sea Research, 16(1), 77-84.

318. Sykes,L.R.,1966. The seismicity & deep structure of island arcs. J. Geophys. Res., 71(12), 2981-3006.

319. Sykes,L.R.,1967. Mechanism of earthquakes & nature of faulting on the mid-oceanic ridges. J. Geophys. Res., 72(8), 2131-2153.

320. Talwani,M.,1964. A review of marine geophysics. Marine Geol., 2, 29-80.

321. Ten Haff,E.,1959. Properties & occurrence of turbidites. Geol. en Mijnbouw, 21, 217-222 .

322. Terzaghi,K.,1956. Varieties of submarine slope failures. Proc. 8th Texas Conf. Soil Mech. Found. Eng., 52, 41pp.

323. Thorndike,E.M., & M. Ewing,1969. Photographic determination of ocean-bottom current velocity. Marine Tech. Soc. J.,3(1)45-50.

324. U.S. Bureau of Reclamation,1941-1947. Lake Mead Density Current investigations. 1-2(1941), 1-453; 3(1947), 454-904.

325. Van Andel,Tj. H.,1960. Sources & dispersion of Holocene sediments northern Gulf of Mexico. In, Shepard, F.P.,F.B. Phleger, & Tj. H. Van Andel(Ed.), Recent sediments, Northwest Gulf of Mexico. Amer.Assoc.Petrol.Geol.,Tulsa, 34-55.

326. Van Andel,Tj.H.,1964. Recent marine sediments of the Gulf of Calif. In, Van Andel,Tj. H. & G.G. Shor(Ed.), Marine geology of the Gulf of Calif.,Amer.Assoc.Petrol.Geol.Mem.3,216-310.

327. Van Andel, Tj. H. & P.D. Komar, 1969. Ponded sediments of the Mid-Atlantic Ridge between 22° and 23° north latitude. Bull. Geol. Soc. Amer., 80(7), 1163-1190.

328. Van Der Knapp, W. and R. Eijpe, 1968. Some experiments on the genesis of turbidity currents. Sedimentology, 11(1/2), 115-124.

329. Van Der Lingen, G.J., 1969. The turbidite problem. N.Z. J. Geol. Geophys., 12, 7-50.

330. Vanoni, V.A., 1963. Sediment transportation mechnics. Suspension of sediment. J. Hydrol. Div., ASCE, 89, (HY5), 45-76.

331. Vanoni, V.A., 1966. Sediment transportation mechanics: Initiation of motion. J. Hydrol. Div., ASCE, 92, (HY2), 291-314.

332. Van Straaten, L.M.J.U., 1964. Turbidite sediments in the south-eastern Adriatic Sea. In, Bouma, A.H. & A. Brouwer (Eds.), Turbidites, Elsevier, Amsterdam, 142-147.

333. Vogt, P.R., E.D. Schneider & G.L. Johnson, 1969. The crust & upper mantle beneath the sea. In, Hart, P.J. (Ed.), The Earth's crust and upper mantle. Amer. Geophys. Union, Geophys. mono. 13, 556-617.

334. Von Rad, U., 1968. Comparison of sedimentation in the Bavarian flysch (Cretaceous) and Recent San Diego Trough (Calif.). J. Sediment. Petrol., 38(4), 1120-1154.

335. Volkmann, G., 1962. Deep current observations in the western North Atlantic. Deep-Sea Research, 9, 493-500.

336. Walker, R.G., 1963. Distinctive types of ripple-drift cross-lamination. Sedimentology, 2, 173-188.

337. Walker, R.G., 1965. The origin and significance of the internal sedimentary structures of turbidites. Proc. Yorkshire Geol. Soc., 35, 1-32.

338. Weber, J.N., 1958. Recent grooving in lake bottom sediments at Great Slave Lake, Northwest Territories. J. Sediment. Petrol., 28, 333-341.

339. Williams, E., 1960. Intra-stratal flow and convolute folding. Geol. Mag. 97, 208-214.

340. Windisch, C.C., R.J. Leyden, J.L. Worzel, T. Saito & J. Ewing, 1968. Investigation of horizon beta. Science, 162, 1473-1479.

341. Worzel, J.L., R. Leyden & M. Ewing, 1968. Newly discovered diapirs in Gulf of Mexico. Bull. Amer. Assoc. Petrol. Geol., 52(7), 1194-1228.

342. Wüst, G., 1955. Stromgeschwindigkeiten im Tiefen-und Bodenwasser des Atlantischen ozeans. Deep-Sea Research, 3(Suppl.), 373-397.

343. Yalin, M.S., 1964. Geometrical properties of sand waves. Proc. Amer. Soc. Civ. Engrs., J. Hydraulics Div., 90, (HY5), 105-119.

344. Yano, K. and A. Daido, 1965. Fundamental study on mud-flow. Bull. Disaster Prev. Res. Inst., Kyoto Univ., 14, 69-83.

1.3 BOOKS, REPORTS, OTHER VOLUMES AND CHARTS

I. GENERAL MARINE GEOLOGY: The following have summaries of general interest concerned with the eology of the ocean floor and with oceanic currents.

Books:

Open-channel Hydraulics (1959). Chow, V.T., McGraw-Hill, New York, 680p.

Problèmes de Géologie Sous-marine (1958). Bourcart, J., Masson et C^{ie}, Paris, 125p.

The Floor of the Ocean (1942). Daly, R.A., Univ. North Carolina Press, Chapel Hill, 177p.

The floor of the Oceans, 1. The North Atlantic (1959). Heezen, B.C., M. Tharp and M. Ewing. Geol. Soc. Amer., Spec. Paper 65, 122p.*

The Gulf Stream: A physical and dynamical description (1960). Stommel, H., Univ. of Calif. Press, Berkeley, 202p.

The Oceans: Their physics, chemistry and general biology (1942). Sverdrup, H.U., M.W. Johnson and R.H. Fleming, Prentice-Hall, New York, 1087p.

Symposium and collected volumes:

Deep-sea Photography (1967). Hersey, J.B. (Editor), Johns Hopkins Oceanographic Studies, no. 3, 310p.

Search and serendipity. W.L. Brundage, Jr., C.L. Buchanan and R.B. Patterson, 75-88.

Geological applications of sea-floor photography. C.O. Bowin, R.L. Chase and J.B. Hersey, 117-140.

Photography of seamounts. R.M. Pratt, 145-150.

Current markings on the continental slope. D.M. Owen and K.O. Emery, 167-172.

The floor of the Bellingshausen Sea. C.D. Hollister and B.C. Heezen, 177-190.

Underwater photography of the Carlsberg Ridge. A.S. Laughton, 191-206.

Marine Geology of the Gulf of California (A Symposium) (1964). Van Andel, Tj. H., and G.G. Shor, Jr. (Editors), Amer. Assoc. Petrol. Geologists, Memoir 3, 408p.

Recent marine sediments of Gulf of California. Tj. H. Van Andel

Sea floor valleys of Gulf of California. F.P. Shepard.

* Asterisks designate volumes and/or chapters which are either definitive or are of particular relevance and importance.

Diagenesis in Sediments. Developments in Sedimentology, 8. Larsen, G. and
G.V. Chilingar (Editors), Elsevier, Amsterdam, 551p.

Phases of diagenesis and authigenesis. R.W. Fairbridge, 19-90.

Diagenesis of sandstone. E.I. Dapples, 91-126.

Recent Sediments, Northwest Gulf of Mexico, (A Symposium)(1960). Shepard,F.P.,
F.B. Phleger, and Tj. H. Van Andel (Editors). Amer. Assoc. Petrol.
Geologists, Tulsa, 394p.

Sources and dispersion of Holocene sediments, Northern Gulf of Mexico.
Tj. H. Van Andel, 34-55.

Sediments and history of Holocene transgression, continental shelf,
Northwest Gulf of Mexico. J.R. Curray, 221-266.

Regional aspects of modern sedimentation in Northern Gulf of Mexico and
similar basins, and paleogeographic significance. Tj. H. Van Andel and
J.R. Curray, 345-364.

The Encyclopedia of Oceanography. Encyclopedia of Earth Sciences, I(1966).
Fairbridge, R.W. (Editor), Reinhold Publ. Co., New York, 1020p.*
See index listing on: marine topography, turbidites, currents and
regional sediment studies.

The Sea. 3 (1963). Hill, M.N. (Editor), Wiley, New York, 963p.*

Topography of the deep-sea floor. B.C. Heezen and H.W. Menard, 233-280.

II. PHYSIOGRAPHIC DIAGRAMS: The following charts and accompanying explanations provide a primary basis for interpretations of dynamic processes which affect the ocean floor.

Physiographic diagram: Atlantic Ocean, sheet 1. (1957). Heezen, B.C. and
M. Tharp. Geol. Soc. Amer.

Physiographic diagram: Atlantic Ocean (revised)(1968). Heezen, B.C. and
M. Tharp. Geol. Soc. Amer.

Atlantic Ocean Floor (1968). Heezen, B.C., M. Tharp and H.C. Berann.
Nat'l. Geogr. Soc., 133(6).

Physiographic diagram of the South Atlantic (1961). Heezen, B.C. and
M. Tharp, Geol. Soc. Amer.

Physiographic diagram of the Indian Ocean (1965). Heezen, B.C. and M. Tharp.
Geol. Soc. Amer.

Physiographic Map of the North West Pacific Ocean (1967). [illegible]rova, N.A.,
Academy of Sciences, Inst. of Oceanology, Moscow, 710-71[illegible].

Physiographic Map of the North Eastern Pacific (1964). In, Menard, H.W., Marine Geology of the Pacific, McGraw-Hill, New York.

The Indian Ocean Floor (1967). Heezen, B.C., M.Tharp and H.C.Berann. Nat'l. Geogr. Soc., 132(4).

The Pacific Ocean Floor (1969). Heezen, B.C., M.Tharp and H.C.Berann. Nat'l. Geogr. Soc., 136(4).

III. TURBIDITY CURRENTS: The following include discussions regarding the effects of turbidity currents on the ocean floor, their flow regimes, places of occurrence and mechanisms of initiation.

Books:

Marine Geology of the Pacific (1964). Menard, H.W., McGraw-Hill, New York, 271p.

Submarine Canyons and Other Sea Valleys (1966). Shepard, F.P. and R.F. Dill, Rand McNally, Chicago, 381p.

Submarine Geology, 2nd edn. (1963). Shepard, F.P., Harper, New York, 557p.

The floors of the Oceans I. The North Atlantic (1959). Heezen,B.C.,M.Tharp and M. Ewing, Geol. Soc. Amer., Spec. Paper 65, 122p.

The Sea off Southern California. A Modern Habitat of Petroleum (1960). Emery, K.O., Wiley, New York, 366p.

Symposium and Collected Volumes:

Submarine Geology and Geophysics, Colston Papers (no. 17) (1965). Whittard, W. and R. Bradshaw (Editors), Butterworths, London, 464p.

Experiments in connection with turbidity currents and clay-suspensions. Ph. H. Kuenen, 47-74.

Recent turbidites in the Hardangerfjord, Norway. H. Holtedahl, 107-142.

The Rhône deep-sea fan. H.W. Menard, S.M.Smith and R.M.Pratt, 271-286.

Papers in Marine Geology (Shepard commemorative volume) (1964). Miller,R.L. (Editor), Macmillan, New York, 531p.

Sedimentation and erosion in Scripps Submarine Canyon Head. R.F. Dill, 23-41.

Mass transport of sediment in the heads of Scripps Submarine Canyon, California. T.K. Chamberlain, 42-64.

Analysis of flow in Cascadia Deep-Sea Channel. P.J. Hurley, 117-13[illegible].

The Sea, 3(1963). Hill,M.N.(Ed.),Wiley,New York, 963p.*

Abyssal Plains. B.C. Heezen & A.S. Laughton, 312-364.*

Submarine Canyons. F.P. Shepard, 480-506.*

Sedimentation. R.A. Bagnold and D.L. Inman. 507-553.

Turbidity currents. B.C. Heezen, 742-775.*

Turbidity Currents and the transportation of coarse sediments to deep water. (1951). A Symposium. Soc. Econ. Paleontologists Mineralogists, Spec. Publ. 2, 107p.

Applications of hydraulics to the study of marine turbidity currents. H.W. Menard and J.C. Ludwick, 2-13.

Properties of turbidity currents of high density. Ph. H. Kuenen. 14-33.

Some quantitative aspects of Lake Mead turbidity currents. H R. Gould, 34-52.

Transportation of sand into deep water. F.P. Shepard, 53-65.

Displaced foraminifera faunas. F.B. Phleger, 66-75.

Sedimentary history of the Ventura Basin, California, & the action of turbidity currents. M.L. Natland & Ph. H. Kuenen, 76-105.

IV. TURBIDITY CURRENT DEPOSITS: The following deal primarily with the description of the facies of near bottom transported sediments (including turbidites). Proper interpretation of the sediments is necessary to appreciate the dynamics of sedimentary regimes in the deep-sea.

Books:

Atlas & Glossary of Primary Sedimentary Structures(1964). Pettijohn, F.P. & P.E. Potter. Springer-Verlag, Berlin, 370p.

Current Ripples(1968). Allen,J.R.L. North Holland,Amsterdam, 447p.

Cyclic Sedimentation. Developments in Sedimentology 10,(1967). Duff, P. McL.D., A. Hallam & E.K. Walton. Elsevier, Amsterdam, 280p.

Methods for the study of sedimentary structures(1969). Bouma, A.H. Wiley - Intersciences, New York, 458p.

Paleocurrents & basin analysis(1963). Potter, P.E. & F.J. Pettijohn. Springer-Verlag, Berlin, 296p.

Particle size: Measurement, interpretation, & application (1963). Irani, R.R. and C.F. Callis. John Wiley and Sons, New York, 165p.

Petrology of Sedimentary Rocks (1959). Folk, R.L. Hemphills Book Store, Austin, 154p.

Sedimentary Features of Flysch and Graywackes (1965). Dzulynski, S. and E.K. Walton. Elsevier, Amsterdam, 274p.

Sedimentology of Some Flysch Deposits. A Graphic Approach to Facies Interpretation (1962). Bouma, A.H., Elsevier, Amsterdam, 168p.

Scientific Method in Analysis of Sediments (1967). Griffiths, J.C., McGraw-Hill Book Company, New York, 508p.

Symposia or Collected Volumes:

Deltaic and Shallow Marine Deposits (1964). Van Straaten, L.M.J.U. (Editor), Elsevier, Amsterdam, 464p.

Criteria in modern sediments useful in recognizing ancient sedimentary environments. F.P. Shepard, 1-25.

Sedimentation in the modern delta of the Niger, West Africa. J.R.L. Allen, 26-34.

Ancient deltaic sedimentation in eugeosynclinal belts. R.H.Dott, Jr., 105-113.

Le Flysch: definition; depot de faible profondeur. M. Rech-Frollo, 347-355

Distribution and lateral variability of heavy minerals in the Annot Sandstones. D.J. Stanley, 388-398.

Geometry of Sandstone Bodies (A Symposium) (1961). Peterson, J.A. and J.C. Osmond (Editors), Amer. Soc. Petrol. Geologists, 45th Annual Meeting, 240p.

Problems and principles of sandstone-body classification. G. Rittenhouse, 3-12.

Turbidites in oil exploration. H.H.Sullwold, Jr., 63-81.

Primary Sedimentary Structures and Their Hydrodynamic Interpretation (1965). Middleton, G.V. (Editor), Amer. Assoc. Petrol. Geol., Spec. Publ. 12, 265p.

Hydro-mechanical principles of sediment structure formation. L.I. Briggs and G.V. Middleton, 5-16.*

Experimental work on primary sedimentary structures, L.M. Brush, Jr., 17-24

Laboratory study of the distribution of grain sizes in cross-bedded deposits. A.V. Jopling, 53-65.

Experiments on ripple lamination. E.D. McKee, 66-83.

Stratification, bed forms and flow phenomena (with an example from the Rio Grande). J.C. Harms and R.K. Fahnestock, 84-115.

Sedimentary structures: Mississippi River deltaic plain. J.M. Coleman and S.M. Gagliano, 133-148.

Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms. J.E. Sanders, 192-219.*

Progress in Oceanography 4. The Quaternary History of the Ocean Basins. A Symposium (1965, INQUA), (1967) Sears, M. (Editor), Pergamon Press, Oxford, 344p.

Some deep-sea sediments in the Western Mediterranean Sea. K.G. Eriksson, 267-282.*

Turbidites (1964). Bouma, A.H. and A. Brouwer (Editors), Elsevier, Amsterdam, 264p.

Deep-sea sands and ancient turbidites. Ph. H. Kuenen, 3-33.*

Methodology and paleogeographic interpretation of flysch formations: a summary of studies in the Maritime Alps. D.J. Stanley, 34-64.

Turbidite sediments in the southwestern Adriatic Sea. L.M.J.V. Van Straaten, 142-147.*

Le sables profonds de la Méditerranée occidentale. J. Bourcart, 148-155.

Allodapische Kalke, Turbidite in Riff-Mahen Sedimentations-Becken. K-D Meischner, 156-191.

Bibliography of turbidity currents and turbidites. Ph. H. Kuenen and F.L. Humbert, 222-246.*

Turbidites. A.H. Bouma, 247-256.

V. ENGINEERING PROPERTIES OF MARINE SOILS: Direct measurements of the physical properties of marine soils are reported in the following documents and provide the engineer with data on the particle size distribution, bearing strength, shear strength, porosity, density, acoustic properties and composition of the sediments.

Acoustic and other physical properties of shallow-water sediments off San Diego (1956). Hamilton, E.L., G. Shumway, H.W. Menard and C.J. Shipek. Jour. Acoust. Soc. Amer., 28 (1), 1-15.

A summary of engineering properties, sediment size and composition analysis of cores for NAVADO III; December 1964 - September 1965 (1966). E. Achsetter, D. Hill and R. Stewart. Laboratory report 283, Geol. Res. Branch, Ocean Surveys Div., Ocean Surveys Dept., 75p.

Correlation between acoustical and other physical properties of deep-sea cores Norwegian Basin; Tech. Rept. 1, Task Area 1. (1967). Horn, D.R., U.S. Naval Oceanographic Office, Marine Geophysical Survey Project (unpublished report, prepared by Texas Instruments, Inc.) 88p.

Correlation between acoustical and other physical properties of Mediterranean deep-sea cores; Tech. Rept. 2.,Task Area 6.(1967). Horn,D.R., B.M. Horn & M.N Delach. U.S. Naval Oceanographic Office, Marine Geophysical Survey Project (unpublished report, prepared by Texas Instruments, Inc.),152p.

Correlation between acoustical and other physical properties of deep-sea cores, Northeast Atlantic: Tech. Rept. 3, Task Area 2.(1967). Horn, D.R., M.N. Delach and B.M. Horn. U.S. Naval Oceanographic Office, Marine Geophysical Survey Project (unpublished report, prepared by Texas Instruments, Inc.), 115p.

Correlation between acoustical and other physical properties of deep-sea cores, West European and Iberian Basins, Northeast Atlantic: Tech. Rept. 4, Task Area 3 (1968). Horn, D.R., M.N. Delach and B.M. Horn. U.S. Naval Oceanographic Office, Marine Geophysical Survey Project (unpublished report, prepared by Texas Instruments, Inc.), 100p.

Correlation between acoustical and other physical properties of deep-sea cores, Northwest Atlantic; Tech. Rept. 5, Task Areas 7 and 4 (1968). Horn, D.R., M.N. Delach and B. M. Horn. U.S. Naval Oceanographic Office, Marine Geophysical Survey Project (unpublished report prepared by Texas Instruments, Inc.), 103p.

Sonic properties of deep-sea cores from the North Pacific Basin and their bearing on the acoustic provinces of the North Pacific: Tech. Rept. 10, (1968). Horn, D.R., B.M. Horn and M.N. Delach. Lamont Geol. Observatory, CU-10-68, NAVSHIPS N00024-67-C-1186, 357p.

Core, sound velocimeter, hydrographic, and bottom ocean stations, Marine Geophysical Survey Program 65-67, Western North Atlantic and Eastern and Central North Pacific Oceans, Area I(Regional Sohm Abyssal Plain, SP-96-I-8), 8(1966). Schreiber,B.C.(unpublished report prepared for NAVOCEANO by Alpine Geophysical Assoc., Inc.), 27p.

Core, sound velocimeter, hydrographic and bottom ocean stations, Marine Geophysical Survey Program 65-67, Western North Atlantic and Eastern and Central North Pacific Oceans, Area II (Region of the Hatteras Abyssal Plain, SP-96-II-8), 8 (1967). Schreiber, B.C. (unpublished paper prepared for NAVOCEANO by Alpine Geophysical Assoc.,Inc.),22p.

Core, sound velocimeter, hydrographic, and bottom ocean stations, Marine Geophysical Survey Program 65-67, Western North Atlantic and Eastern and Central North Pacific Oceans, Area I supplement (Region of the Sohm and Hatteras Abyssal plains, SP-96-1S-8B) 8,(1967). Schreiber, B.C. (unpublished report prepared for NAVOCEANO by Alpine Geophysical Assoc.,Inc.

Core, sound velocimeter, hydrographic, and bottom ocean stations, Marine Geophysical Survey Program 65-67, Western North Atlantic and Eastern and Central North Pacific Oceans, Area V (Region of Hawaiian Ridge SP-96-V-8) 8, (1968). Schreiber, B.C. (unpublished report, prepared for NAVOCEANO by Alpine Geophysical Assoc., Inc.), 42p.

Investigations of deep-sea sediment cores, II: Mass physical properties, (1962). Richards, A.F., U.S. Navy Hydrographic Office Tech. Rept. 102.

Physical analysis of deep-sea sediments, (1957). Sutton, G.H., Berckheimer, G.H. and J.E. Nafe, Geophysics, 22, 719-812.

Physical properties of marine sediments, (1963). Nafe, J. and C.L. Drake. In, Hill, M.N.(Editor), The Sea. 3. Wiley, New York, 794-815.

Sound speed and absorption studies of marine sediments by a resonance method, 1. (1960). Shumway, G., Geophysics, 25(2), 541-467.

Sound speed and absorption studies of marine sediments by a resonance method, II, (1960). Shumway, G., Geophysics, 25(3), 659-682.

Sound speed and related physical properties of sediments from experimental Mohole (Guadaloupe Site), (1965). Hamilton, E.L., Geophysics 30(2), 257-261.

Variation with depth in shallow and deep water marine sediments of porosity, density and the velocities of compressional and shear waves, (1957). Nafe, J.E and C.L. Drake, Geophysics, 22 (3), 523-552.

PART II

SUBMARINE SEDIMENTARY PROCESSES

2.1 GENERAL

The dynamics of submarine sedimentation are visualized by deduction from geophysical observations of the ocean floor and from direct study of sediment samples obtained through coring and drilling. Instrumental monitoring of ocean floor currents and internal waves, using current meters and sea floor cameras, provides a few actual measurements of the hydrodynamic realm, but investigations conducted from submersibles have produced the only direct visual data available on the movements of sediment. Theoretical considerations help to confine speculation.

The major components of intra-oceanic sedimentary processes involve *in situ* production of sediment on or beneath the ocean floor (post-depositional diagenesis), settling of sediment through the water column, and horizontal movement of sediment along or near the bottom. More than one of these components is commonly important in the sedimentary regime at any site on the ocean floor; and generally there are few areas where any single component is completely non-operative.

Under special circumstances, these sedimentary processes might be active at a locality on the ocean floor in such a manner as to be directly hazardous to man-made structures. These possibilities must be exp,ored. Consequently, it will be necessary to review briefly here the basic mechanisms of sediment production, transport, deposition and diagenesis.

2.2 IN SITU PRODUCTION OF SEDIMENT

Principal types of authigenic sediment are manganese oxides, cherts and phosphorites, the mineral glauconite, zeolite, and possibly some poorly crystalline clays. In anaerobic environments, minerals such as pyrite may form; and in restricted basins, salts may be directly precipitated. The development of all these sediment types involves crystallization near the sediment-water interface (or at deeper depths in the sediment layers, e.g., with respect to chert, which would, nevertheless, be within reach of drilling and thus of potential usefulness in foundation work, as discussed in Part VII of this report).

With regard to environmental hazards, time spans involved in the production of authigenic sediment are probably too long to be relevant.

The residence time of manganese, for example, is about 1500 years and accretion rates are of the order of a few millimeters per thousand years. The precipitation of manganese crusts on structures would be minimal. However, some areas of the sea floor contain authigenic sediments which might provide specific settings for engineering activities. For instance, in certain explored areas, there are layers beneath the sea floor which, through diagenetic processes, have become consolidated into brittle rock; yet there are zones between these layers in which the sediments are still moist and soft. In regions of compaction the flow of pore water is restricted to the softer, permeable horizons and the flowage may cause anomalously high pore water pressures along some of the bedding planes. These high pressures may facilitate the mass downslope transport of the superficial sediment layers, by reducing the effective coefficient of sliding friction. Mass transport of the sea floor is sometimes observed on deep sea fans where rapid local accumulations of sediment near deep sea channels load that region to the extent of generating downslope slumping along the channel levees and even generating geysers of pore water through faults in the sea floor. Examples of authigenic sediments and related phenomena are shown in Fig. II-1.

The formation of layers of salt (halite, gypsum, anhydrite) in the distant past can result in a peculiar kind of subsequent earth movements. Salt layers consist of light low-density materials which, after sufficient burial, remobilize and flow upward into the overlaying sediment. Salt tectonics produces such features as piercement domes and elongate ridges. Vertical movement of these domes is approximately 1-2 mm/yr (Ewing and Ewing, 1962), and can cause related faulting and slope instabilities. Active diapiric structures are found in large regions of the Gulf of Mexico, Mediterranean and Red Seas, and may even occur in the open oceans.

2.3 SETTLING OF SEDIMENT THROUGH THE WATER COLUMN

Settling of sediment through the water column is usually termed pelagic sedimentation, as is the rafting of debris from melting icebergs. The process of pelagic sedimentation can be visualized as approximately vertical deposition.

The main types of material thus deposited are: (1) shells and tests of plankton (with depositional rates, typically a few centimeters per 1000 yr.), of which carbonate constituents are limited to depths of less than approximately 2000 fm.; (2) wind-blown detritus, including tephra from explosive volcanic eruptions; (3) cosmic particles; (4) rafted material, which may include boulders weighing many tons; (5) water-borne particles from rivers.

In almost every case, this type of sedimentation is slow, considering the expected life-span of sea floor structures. However, even slow accumulation might adversely affect windows and light-

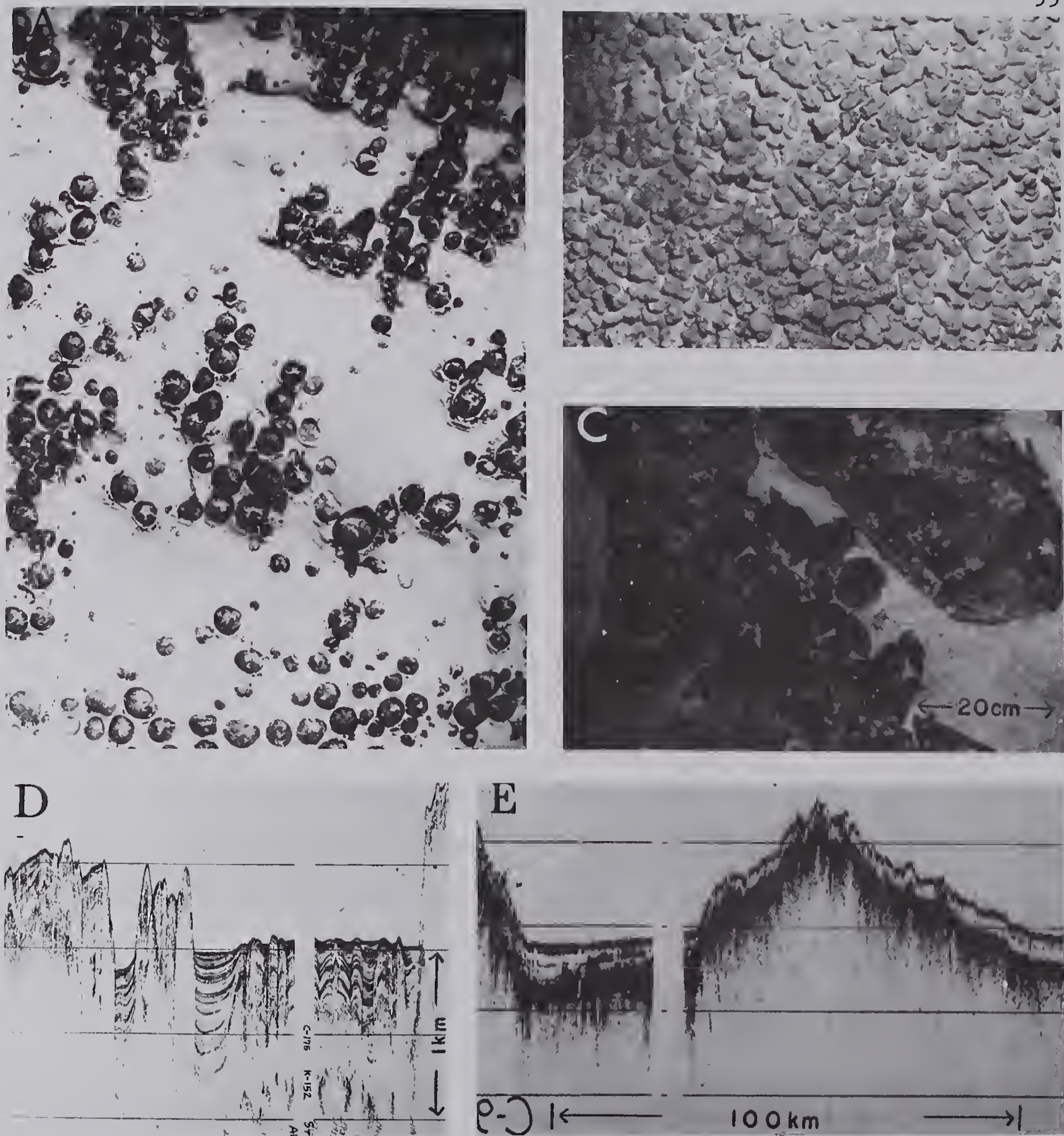


FIGURE II-1 Examples of authigenic sediments and related phenomena

A. Spherical manganese nodules are formed by slow precipitation of manganese oxides on a central nucleus. (RC9-K69-15)

B. Vast pavements of nodules with diameters of about 3 cm. (RC9-K68-5)

C. Accreted manganese crust on soft sediment. (V18-K23-2)

D. Diapiric intrusions (piercement domes) in a sedimentary basin are produced by vertical migration of salt layers. (RC9: 15-16 July 1965)

E. Post depositional diagensis has turned a subbottom sedimentation layer (strong reflector) into hard brittle rock. The processes of cementation can produce siliceous cherts of calcereous limestones, depending upon the composition of the original sedimentary layers. (RC9: 22 July 1965)

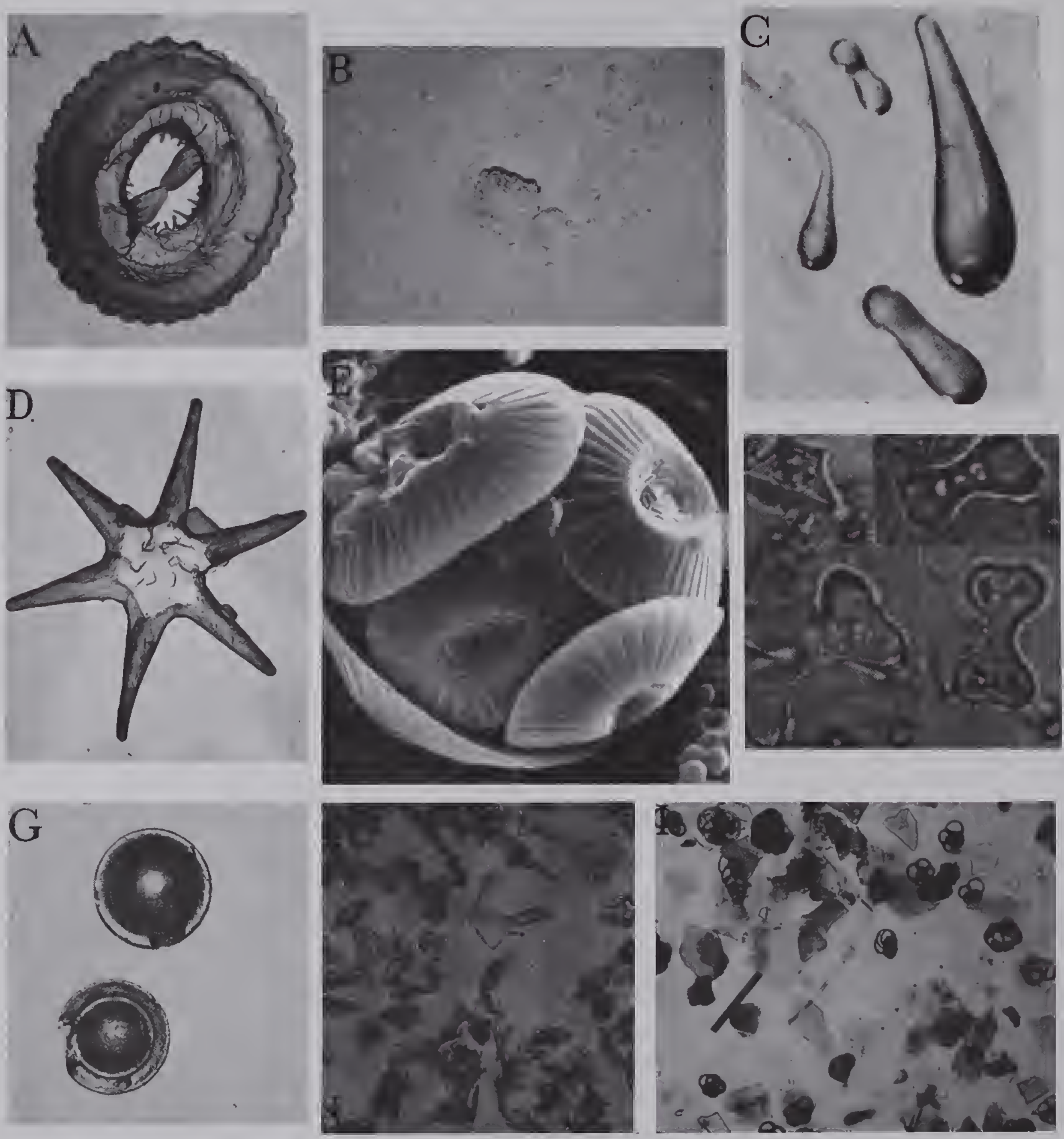


FIGURE II-2 - <u>Examples of sedimentary constituents in pelagic sediments</u>

A. Fragment of a modern coccolithophoridae, diameter 5 microns. (A. McIntyre)
B. An ice-rafted boulder, diameter 10 cm.
C. Cosmic tektite, diameter 50 microns. (Glass and Heezen, 1967)
D. Discoaster, diameter 5 microns. (A. McIntyre)
E. Living coccolithophoridae. (A.McIntyre)
F. Wind-blown opal phytoliths, diameter 10 microns. (Folger *et al.*, 1967)
G. Cosmic spherules.(Glass and Heezen, 1967)
H. Volcanic tephra, shard diameters are 70 microns. (Ryan *et al.*, 1963)
I. Wind-blown dust and fresh water diatoms mixed with foraminifera. Diameters are 100 microns. (Ryan *et al.*, 1963)

transmitting apertures. In shallow water, the biological productivity of the plankton and nekton might hinder visibility through water, but in the deeper oceans this slow particle-by-particle settling produces negligible light scattering or attenuation.

Examples of the sedimentary materials found in pelagic sediments are shown in Fig. 11-2

2.4 BOTTOM AND NEAR-BOTTOM MOVEMENT OF SEDIMENT

There are four principal types of bottom and near-bottom movement of sediment in the oceans: (1) relatively low velocity mass movements (creep) of materials of all dimensions which are relatively free of included sea water; (2) avalanches, or relatively high velocity mass movements; (3) relatively low velocity near-bottom transport of silt and clay (and possibly sand-sized particles) by bottom currents, and possibly by other mechanisms analagous to so-called low density turbidity currents, (these, however, are not well understood and may involve secular flow.); (4) powerful, high-velocity transportation of coarse sand (and possibly pebbles) and finer sediment by turbidity currents.

2.4-1 Low velocity mass movements

Low velocity slippage and creep are probably the major types of movement in this category. Gentle drifting of sand in submarine canyons may be regarded either as mass movement or as low-velocity near-bottom transport. The creeping of sediment can occur even on very gentle slopes of less than a few degrees, such as those on the continental rise or deep-sea aprons. Movement on such low slopes is invariably aided by mechanisms such as those previously discussed, i.e. anomalous increases in entrained pore-water pressure as the result of over-burden on sediment bodies with impermeable inter-beds (Hubbert and Rubey, 1959; Rona and Clay, 1967).

Seismic reflection profiler records, which display in cross-section the sub-bottom sedimentary strata, have not revealed as yet many occurrences of significant creep or overthrusts, except in the very limited regions of oceanic trenches. An example of large-scale gravity sliding is shown in Fig. II-3A, where a large terrain, many tens of km in length and over 300 m in thickness, has glided northward (i.e. toward the left) into a deep sea trench south of Turkey in the Mediterranean Sea .

A case for downslope <u>decollement</u> may be argued for this profile, where the folded sediment at (d) may have slid on an incompetent horizon below reflector "M" from the rise at (c). In fact, thinning and rifting of the superficial cover at (c) may indicate the place of departure of this large slide. The net distance travelled is a few kilometers, although the glide plane is over 80 km long. In Fig. 11-3B, a slump block has slid down a continental slope a distance of about 7 km and has travelled up over the contemporary sea floor. The site of detachment is clearly seen upslope, and the slide block itself is only moderately deformed. From geophysical evidence such as this reflection profile it is very difficult to estimate the time span required for this translation.

However, photographs of the sea floor (Fig. 11-3D) sometimes show very fresh and sharp-looking scars which indicate quite sudden and very recent movement.

2.4-2 Avalanches

Submarine avalanches encompass the phenomena of the falling, tumbling and rapidly sliding downslope of parts of the sea floor under the pull of gravity. On steep slopes the avalanches might include boulder-sized fragments or hugh blocks many kilometers in length and many hundreds of meters in thickness. Such large masses, which move as more or less cohesive units, are called "allochthonous terrains." They may creep downslope over extended periods of time on basal units which act as lubricating horizons, or they may glide in rapid motion at express train velocity. Motions of allochthonous units which result in contortions within the unit, such as fracturing, folding , fluid flowage and squeezing without failure, are generally termed slumping. Conversely, submarine flows are those phenomena which encompass mostly fluid motion of wet viscous sea-floor sediments.

Avalanches can be recognized in the stratigraphic superposition of older sediments on younger terrains, and this superposition can be seen on seismic reflection profiles, and in cores from outcrop exposures. Similarly, part of the exposed gliding horizon may reveal grooves and characteristic texture called "slickensides" (Fig. II-3C). Coring and sea floor photography may reveal terminal slump blocks which, as sedimentation deposits, are called breccias (Fig. II-3E). Breccia units range from a few meters across and a few centimeters thick to immense emplacements in depressions at the base of steep escarpments. Coring can also reveal overturned beds which indicate folding as the result of creep or rapid displacement. In sediments with extremely homogeneous lithologies, slumping may be recognized by mixing of faunal assemblages of different ages or environments, although such mixing can also occur from the winnowing action of bottom currents. Generally, however, associations of benthic fauna (bottom dwellers) from markedly different depth do indicate downhill transport by slumping or flow.

Rapid movements of soft, unconsolidated marine sediments will invariably result in a certain amount of material being carried into suspension as a mud cloud. Indeed, many breccias found in the axes of depressions are capped with a thin layer of residual material which has settled out locally from such clouds and accumulated on top of the slump debris.

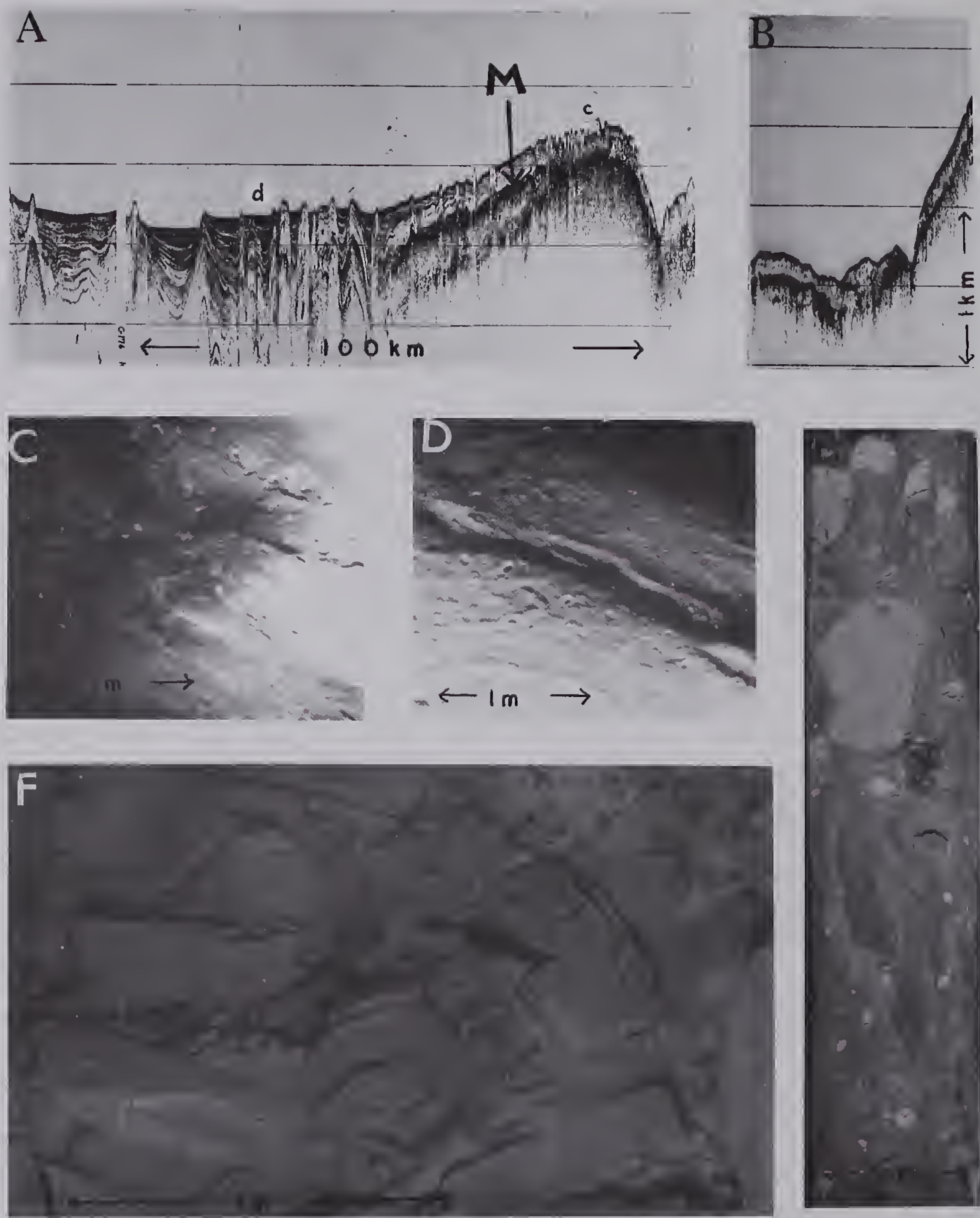


FIGURE II-3- Examples of low-velocity mass movements of sedimentary terraines and avalanches involving rocks and sediment

A. Gravitational gliding of superficial sedimentary layers in the Mediterranean Sea. The downslope creep has produced folds at (d) and has left tension cracks at (c). Vertical scale is 500 m between lines. RC9:16 July 1965.(Ryan et al. in press). B. A slope block which has ridden up onto the basin floor. RC9: 20 July 1965. C. Photograph of sea-floor slickensides. RC9-K164. D. Slump scars. *Chain* 61-12. E. Microbreccia containing slump debris in a deep-sea core. RC9-188. F. Avalanche talus of siltstone blocks. *Eastward* 6-66.

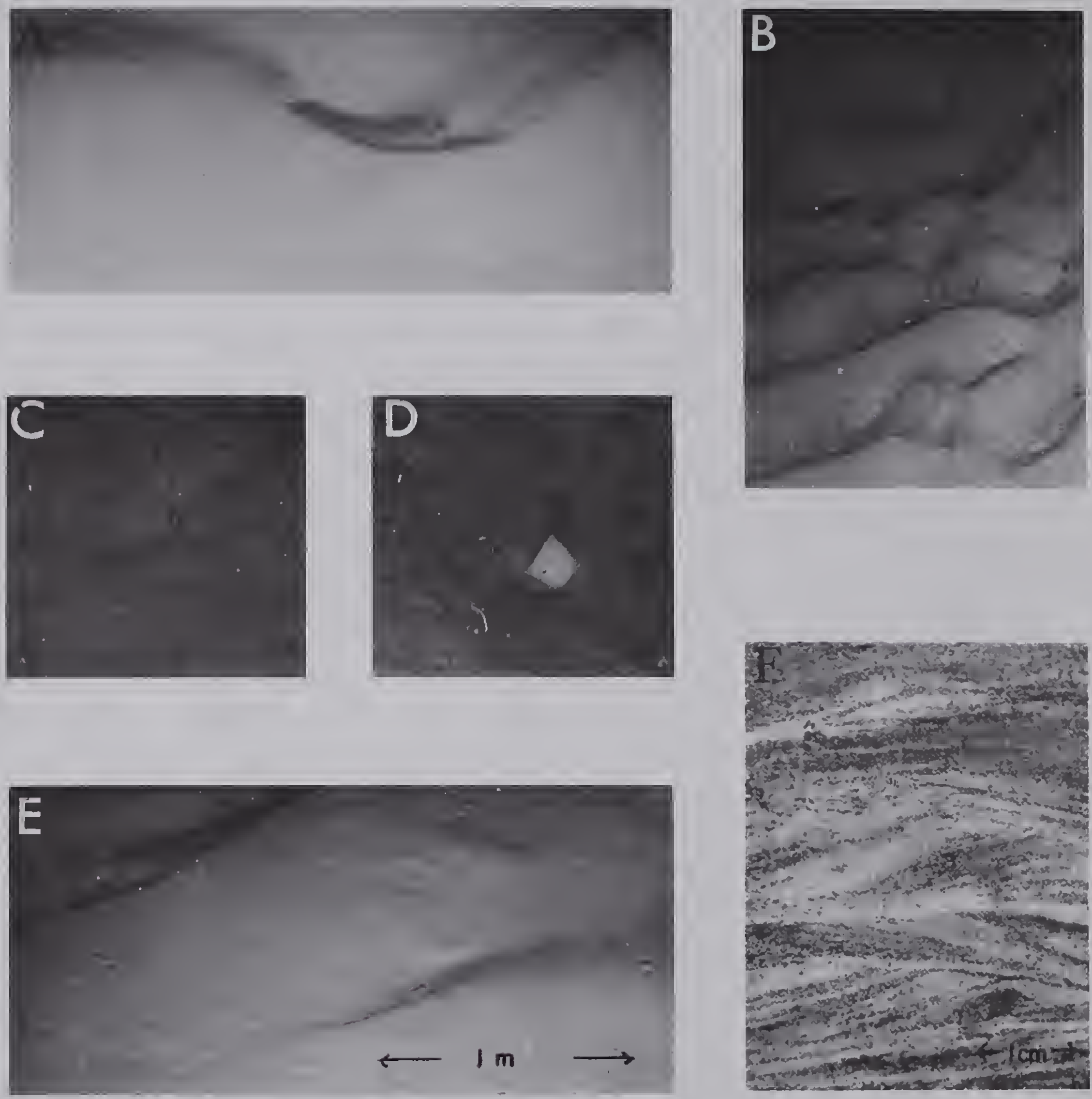


FIGURE II-4 - <u>Examples of near-bottom current activity - scour erosion and sedimentary drifts</u>

A. Sedimentary drifts in fine-grained lutite. *Eastward* 8-67-8.
B. Ripple marks in sand. *Eastward* 9-66-106.
C. Scour around pebbles. *Eastward* 41-49-8.
D. Sponge bent over in current.
E. Large dunes. *Eastward* 8-67-8.
F. Cross-bedded winnowed deep-sea sand in a core sample. *Chain*.

2.4-3 Low velocity near-bottom transport

Near-bottom deep sea currents involve stream-type flow where, on a given slope, velocity is a function of stream-flow and is independent of concentration of suspended material. This appears to be a major low-velocity transporting mechanism along some continental margins. Possibly some process of an indeterminate nature, having a downslope gravity-controlled component related to density-current dynamics, may also operate near the sea floor. We are unsure.

The near-bottom currents, which effectively transport sediment horizontally along the sea floor, generally flow across-slope, parallel to regional bathymetric contours and often are of sufficient velocity to cause local corrugations on the bottom, such as lineations, scour marks, ripple marks, and tool marks (Ewing and Davis, 1967). Evidence from bottom photographs (see Fig. II-4) indicates that, in this environment, the drifting and deposition of sediment may be affected by obstructions such as man-made structures. Turbidity currents could, by drifting, cause the slow burial of structures or, by scour and erosion, undermine structures. A sea floor structure, depending on its configuration, may be considered analogous to a snow fence in drifting snow (produces scour).

In regions of near-bottom current activity, the absence of old tracks and trails of bottom crawling animals is evidence that the reworking of the sediments by the currents is active continually. Thermal gradients in the uppermost meter of the sea floor, measured in regions of current-produced lineations on the Blake - Bahama Outer Ridge in 1967 by thermoprobes with multiple thermisters, can be interpreted to indicate that here is a region where the net deposition is about 75-150 cm/1000 yrs. The sea floor apparently is reworked to a depth of 50 cm in less than 6 months. Thus the net deposition rate may have little to do with the amount of local drifting or scour that takes place around a structure. This subject will be treated further in a later section.

Often, in regions of near-bottom currents, sediment is found to be in tenuous suspension in a layer above the bottom that is many hundreds of meters thick. This layer, called a nepheloid layer (Ewing and Thorndike, 1965), gives a useful measurement of the presence of bottom sediment in motion. The concentration of the layer as measured _in situ_ by light-scattering meters is sufficient, in many regions, to affect seriously visibility above the bottom, and to make it impossible to photograph the sea floor with any detail from heights of only a meter or two. M. Ewing and E. Thorndike at Lamont-Doherty have now made nephelometer observations at 1800 stations in the world ocean

The processes discussed here as low-velocity near-bottom transport merge into those which are usually included under the concept of turbidity currents. To most marine geologists turbidity currents have come to imply exclusively high-density suspensions and high-velocity motion. In much the same way, it is not entirely satisfactory to distinguish between avalanches and turbidity currents, as one may be the cause of the other, or may transform into the other.

2.4-4 Turbidity Currents

The downslope flow, on or near the bottom, of sedimentary materials which have been mixed in water and tossed into a turbulent suspension is called a turbidity current. The density of the suspension is greater than that of the water above it. A turbidity current is a specific type of density current caused by different concentrations of suspended sediments.

Turbidity currents originate upslope of a sedimentary basin where, by some trigger mechanism, the sediment fails and starts to slump. As all or parts of the slump are transformed into a suspension of higher density than the surrounding water, a resulting turbidity current flows downslope with increasing velocity (often through channels) until it reaches lower gradients, where it spreads out laterally and decreases in velocity. The current picks up new material in the path of motion if the velocity and turbulence are strong enough to retard the settling of entrained particles.

Turbidity currents can flow for several thousand kilometers, moving out to the extremes of a sedimentary depression. As the current slackens, suspended sediment is deposited vertically and laterally in graded beds known as turbidites. Depending upon the properties of the initial suspension, the grain size of the entrained particles, and the steepness of the downhill slope, the velocities of turbidity currents range from about 10 cm/sec to over 50 knots (2,500 cm/sec).

The basal portion of the average turbidite bed may be deposited on the basin floor in a few hours. The thickness of individual beds ranges from a few millimeters to several meters. Therefore, turbidity currents are potentially destructive to ocean floor structures. They have broken, displaced and buried deep-sea cables.

2.5 THE TURBIDITY CURRENT CONCEPT AND ITS INTERDEPENDENCE WITH OTHER PROCESSES

In 1885 Forel associated sedimentary transport in Swiss lakes to flows of turbid water. During the early 1900's civil engineers had investigated and were aware of types of sedimentary transport in reservoirs, which were essentially what we call turbidity currents. Daly (1936) was responsible for the recognition of possible evidence of turbidity current activity in the ocean, in the form of an agent for the erosion of submarine canyons. Subsequently, many other workers were also involved (Natland, Migliorini, Shepard), but it was not until 1951 that Ericson, Ewing, and Heezen described definite sedimentary evidence of rapid and large-scale bottom transport in the sea. This evidence related to the formation of abyssal plains (Tolstoy and Ewing 1949) and submarine deltas.

Questions regarding the roles of secular and episodic submarine transportation of sediment and the stability of bottom water regimes at any one site have been raised by the relatively high velocities of bottom water flow (up to 15 cm/sec) observed in steady-state abyssal circulation This has been described by Heezen, Hollister and Ruddiman (1966) and was long predicted by Wüst (1936).

Thus we have two, or possibly more, possible transport mechanisms for deep sea sediments, and it is sometimes difficult to distinguish between them. The concept of turbidity currents in the modern ocean is supported by inferences between cause and effect, and not as yet by direct observations of their activity or physical properties, or by general equations precisely describing the physics of their behavior. Hence, any discussion of turbidity currents and their depositional products may be considered controversial. It is our intent, within the limits of todays' "state of the art" to describe the evidence as we interpret it.

There are considerable problems in drawing a line between what is and what is not a turbidity current. Similarly, the difficulties of finding direct, firm evidence for a turbidity currents' responsibility for a particular phenomenon, focus attention on the possible and apparent interdependence and similarity of other processes, such as slumps, slides, avalanches, creep, etc.

2.6 POWERFUL MOVEMENTS OF SEDIMENT OR LARGE-SCALE RAPID DEPOSITION OF SEDIMENT

Direct evidence

Among the more direct indications of powerful sedimentary transport and episodic deposition are (1) deformed or disrupted submarine telegraph cables outside areas of seismic-tectonic-volcanic disturbance, (2) the presence of fresh, green terrestrial organic plant life wedged in frayed submarine cables, (4) winnowing and abrasion seen in bottom photographs, (5) thermal anomalies in the sea floor sediment layer, and (6) the presence of a nepheloid layer.

Although there is no way to prove a unique relationship between an observed event and its depositional results, in many areas the texture, grain size, grading etc. of mapped upper sedimentary layers, are compatible with the high current velocities associated with our concept of the turbidity current regime.

Indirect evidence

An understanding of the relationships of topographic slope and basin geometry to depositional mechanisms provides excellent indications of possible means of transport of sediments which carpet the sea floor. The cutting of canyons and channels, the construction of fan valleys, the growth of channel levees and the filling of sedimentary ponds can all be recognized by a study of the structure of sub-surface sedimentary layering as recorded in seismic reflection profiles (see Fig. II-5).

Sediment cores from the deep sea show structural, textural stratigraphic and compositional properties which constitute proof or give indications of displaced sediment, - for example, beach sands, tests of shallow water benthonic plankton, or organic plant debris found in cores from deep oceanic depths.

There are many unresolved problems, such as the lateral extent of turbidity flows, their thickness above the ocean floor, etc. Some of these problems are discussed at length in part III. For example, the height of the current above the floor relates to dynamic properties and observed sediment distribution and, of course, this height is critical as far as the influence of turbidity currents on ocean floor structures is concerned.

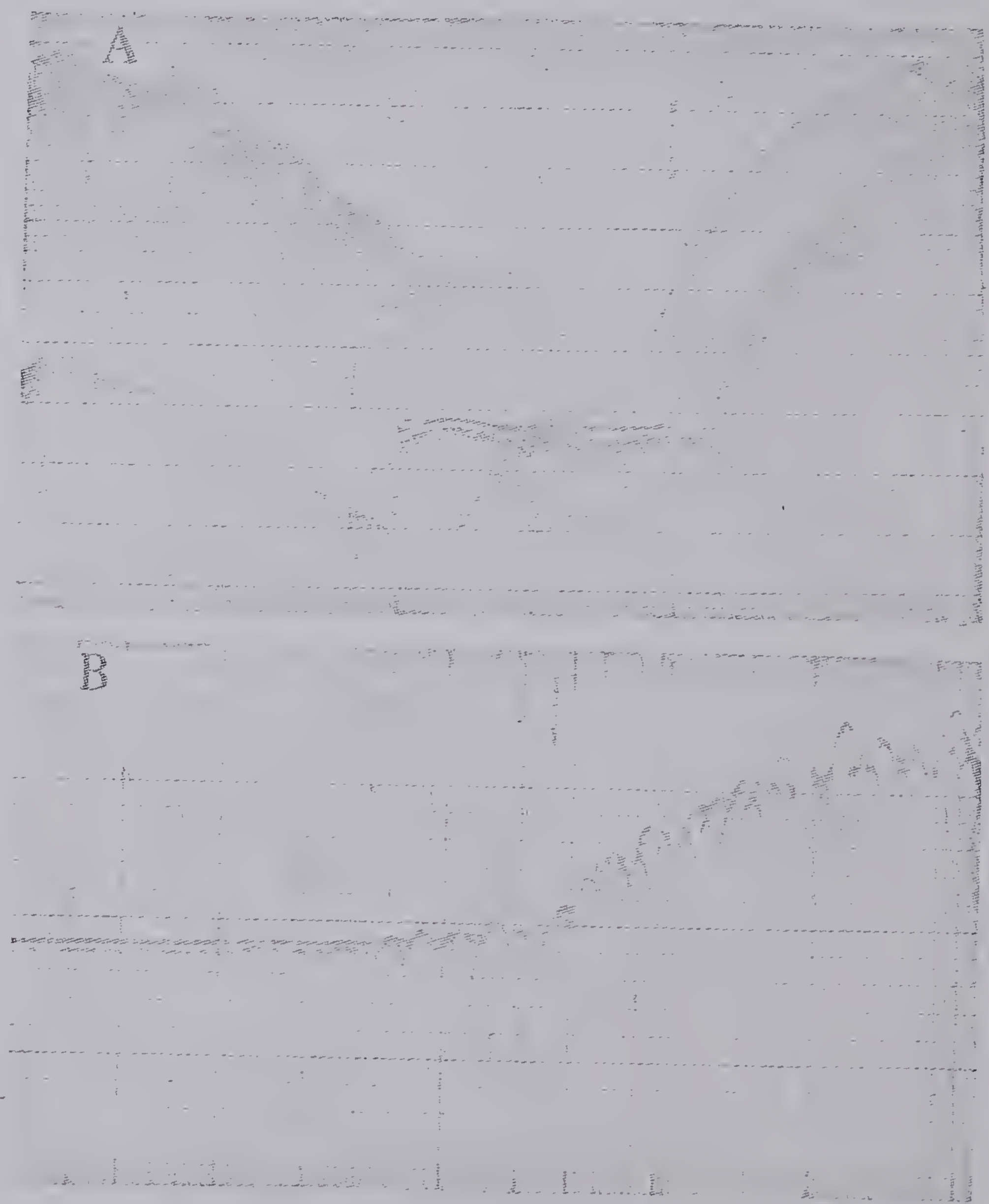


FIGURE II-5 - Examples of inferred evidence for turbidity currents

A. Echogram across a submarine canyon. The erosion of the submarine canyons has been attributed in part to the action of high-velocity turbidity currents (horizontal scale 10 km; vertical scale ≃350 m) from Ryan and Heezen 1965.

B. The sediments on the very flat abyssal plains must have been distributed by fluid processes controlled by gravity, such as density currents. Turbidity currents are a type of density current. (horizontal scale 100 km; vertical scale ≃ 750 m) from Ryan and Heezen 1965.

PART III

PROPERTIES AND DEPOSITIONAL CHARACTERISTICS OF TURBIDITY CURRENTS

3.1 THE HYDROLOGIC REALM

Theoretically, sediment can be transported indefinitely above certain minimal concentrations in turbulent suspension, as long as the gravity gradient is sufficient to override the inhibiting forces generated by internal friction. On the other hand, the physics of turbulent suspension is complex and does not provide exact limiting values for precise conditions of flow. Similarly, the lack of definite information about the circulation and behavior of turbidity currents makes it not yet possible to provide the physicists or hydrodynamicists with the parameters for rigorous theoretical calculations. Among other, Menard and Ludwick (1951), Kuenen (1952), Bagnold (1956, 1968), Stoneley (1957), Allen (1964), and Ichiye (1966) have given theoretical treatment of turbulent suspensions involving their generation, maintenance and decay.

The aspects of the hydrologic nature of turbidity currents most pertinent to the emplacement of structures on the ocean floor are: (1) the mechanism of generation (i.e., transfer of slumps into suspension); (2) the downslope velocities; (3) the turbulence of the flows; (4) the thickness of the turbulent layer above the bottom; (5) the density of the sediment suspensions; and (6) the time span of a turbidity current event. Any consideration of the mode of turbidity current flow would suggest that the parameters of velocity, density, and thickness would vary along the path of the flow depending on such variables as topographic slope, proximity to source, channel width (for confined flows), and general basin geometry (the capability of a flow to disperse laterally). Conversely, the ability of the turbulent flow to erode the sea bed, and/or deposit sediment, would also vary along the path, and would be controlled by the instantaneous nature of the local hydrologic realm.

It is important to understand the variations in the hydrologic realm along the path of a turbidity current. Ocean - floor structures conceivably could be placed in areas of the oceans susceptible to moderate turbidity currents if it could be predicted that the flow would be small enough to cause no damage.

3.2 INDICATORS OF THE BEHAVIOR OF THE HYDROLOGIC REALM

There are two lines of evidence that are the most reliable indicators of the nature of hydrologic real at any one site. The first is the inference that can be drawn from details of sediment deposits (i.e., the contact features, particle-size, textural parameters, geometry of current-induced structures such as cross-bedding and sole-markings, and vertical/lateral extent of beds and other features). The second line of evidence comes from observations on the timing of sequential downslope breakage of submarine cables (which permit calculations of velocity), the locations of such breaks (which show whether flows are channelized or not, i.e., height of flows), and the thermal anomalies in the recent deposits (which permit the calculation of time span of a turbidity current erosional and depositional cycle).

In Section III aspects of the first line of evidence will be discussed; and in Section IV, the second. Inference may be drawn from the details of modern oceanic sediments examined in deep-sea cores and from seismic reflection profiles; and, sometimes in greater detail, from ancient sediments now exposed as outcrops on the continents.

3.2-1 Oceanic sediments

Density current transport and deposition is inferred to be the reason for the extreme flatness of the layers of sediment found on abyssal plains (gradients $<1:1000$). the process of selective deposition in the deepest depression of a sedimentary basin is called "sediment ponding" (Hersey, 1965). Menard (1955) pointed out the control exercised by ocean floor relief on the distribution of clastic sediments derived from erosion on the continents, and showed that the morphology of the northeast Pacific basin was controlled somewhat by the locations of sediment traps. In the North Atlantic, the topography is highly irregular where deep-sea areas are separated from the continent by basins and troughs; while in other areas there is a gradual and continuous slope from the continent to the ocean basin. In such areas, plains slope out from the continent; although in some places there are long ridges, which apparently act as dams to the outward flow of sediments.

Sometimes rough-bottomed basins are found near the flat abyssal plains, but these regions are always ringed by hills or lie on the "lee" side of ridges relative to the continuous downslope transport of sediment material. Menard interpreted this physiographic evidence as an indication that sediment moved by a gravity-controlled transport mechanism (i.e. fluid) would always seek the lowest level of a basin. Sediment could be deposited all along the downslope course of transport, but could not be carried up onto or over topographic elevations.

It was apparent to Menard that turbidity currents could account for the observed sea floor morphology, including the fan-shaped aprons built landward of the abyssal plains and radiating from the mouths of submarine canyons; and the small box-shaped and leveed channels meandering across the fans, connecting the plains and canyons.

Figure III-1A is a reflection profile across an abyssal plain which illustrates the ponded nature of the sedimentary layers. Figure III-1D shows in cross-section a deep-sea channel with its natural levees. Figure III-2 shows various types of topography resulting from sedimentation by turbidity currents and by pelagic processes (from Menard, 1955).

Studies by Ericson et al. (1951), on the sediments from the abyssal plains and the aprons leading to them, revealed the presence of coarse-grained terrigenous sands, many containing tests of displaced shallow-water benthonic foraminifera, and many showing a vertical decrease in grain size from the base of a bed to the top (grading). Sediments on hills rising from the plain have quite different characteristics. They have no sand, even when the cores are taken on slopes only a few tens of meters above the level of the plain.

Figure III-3A is a 12kHz echogram across a small abyssal plain, showing shallow sub-bottom reflecting horizons which by coring can be demonstrated to correspond to levels where graded terrigenous beds are found (Ryan, 1970). These layers can be seen to pinch out at both margins of the small basin, thus demonstrating that these sediments are not carried up over small hills. Figure III-3B shows examples of graded sand layers recovered in the cores. Figures III-4A and B are examples of pelagic sediments deposited on the topographic highs inaccessible to gravity-transported sediment flows from the continent.

Significance of grading

Many of the layers on abyssal plains show obvious grading; most reveal at least some slight grading. Inversely graded beds are very rare, although some beds in deep-sea channels have small inversely graded units in their basal parts and repeated grading within a single bed. Graded bedding (Kuenen and Migliorini, 1950: Kuenen, 1953, 1964, 1967; and Bouma 1962) significantly attests to the deposition of sediment from a gradually waning current carrying suspended material, as do internal sedimentary structures within the graded beds. A facies model, synthesised by Bouma (1962), has been used by many authors, and will be discussed here. We recognize that this model has many exceptions, but it is useful.

According to the model, a complete bed, as found on abyssal plains and inferred to be deposited by turbidity currents (i.e., a turbidite), consists of five intervals, each characterized by one dominant type of sedimentary structure. Beds displaying all five intervals are rare. However, the order of superposition of intervals is always the same.

First there is (1) a graded interval which is structureless and homogeneous, apart from grading. This interval exists only in the most coarse materials of the beds. We sometimes observe poorly-developed laminations in this lowermost unit (see Figure III-5), and often there is much mud matrix present in the basal sand. This unit is followed in vertical sequence by (2) an interval of parallel lamination, usually well-developed with good sorting of medium-to-fine sand. Mud may be concentrated along laminations, giving some a characteristic dark color and enhancing their visual appearance. In the next unit of (3) current-ripple bedding and convolute bedding, the appearance of the bedding is generally enhanced not by contributions of mud, but by sorting of minerals in the size fractions of sand to silt. These units are followed by (4) an upper interval of parallel lamination in coarse-to-fine silt, passing upwards into (5) a pelitic interval of structureless silt and clay with an extremely wet, homogeneous appearance. Each such layered turbidite bed is separated by intervals of pelagic sedimentation from later turbidity current events. This will be discussed later.

Many authors, including Nesteroff and Heezen (1963), Dzulynski and Walton (1965), Schneider and Heezen (1966), Von Rad (1968), and van der Lingen (1969) have provided interpretations of the origin of these intervals in the turbidity facies model. We shall discuss only those aspects from which we may infer the nature of the hydrologic regime that was active on the sea floor at the time of deposition of the respective interval. To avoid excessive detailed argument, generalizations will be made.

A thin "pavement" of shells and tests of forminifera, assumed to be deposited in the pelagic interval, is sometimes found at the base of the graded interval. This concentration of tests indicates a winnowing of finer clay size materials from the pelagic sediments and strongly suggests a period of erosion on the local sea floor by the turbidity current before deposition began.

An important question in considering ocean floor structure sites, and a difficult one to answer, is: how much erosion has occurred? Other ways in which erosion can be inferred are cited below.

Seismic reflection profiles show that sub-bottom horizons appear to outcrop on the walls of submarine canyons and deep-sea channel walls (Dill, 1964; Ryan et al. 1965; and Laughton, 1968) show evidence of slumping, eroded cliffs, and slabs of semi-consolidated sediment scoured by currents. Some channels, particularly in the Indian Ocean on the Idus and Ganges Cones, or in the Gulf of Alaska are clearly the result of net deposition rather than erosion. This is affirmed by evidence in several areas along the course of the channel that the flat floor contained within levees is 20 to 150 meters above the general level of the fan or cone (Fig. III-6A). However, even in these cases, many dominantly depositional channels meander in tight curves back and forth across the fan (Ryan et al. 1967). The erosion must occur locally on the outer walls at the bends. The sediment removed from there is replaced on the inside terrace or carried further down stream (Fig. III-6D), as is often the case in deltaic distributaries on land.

Cores taken near the axis of the deep-sea channels often show incomplete turbidite sequences. These incomplete sequences suggest that erosion at the base of the turbidity current flow removes the upper part of the sea-floor and parts of previously deposited turbidites. Bedding structures cut-off abruptly in the middle of an interval and replaced by the lower graded interval of the next overlying bed are said to be truncated. This is best illustrated in core photographs. Other indications come from graphs of mean grain-size and various textural parameters (see Fig. III-7A) which show incomplete graded sequences replaced by new overlying sequences (Horn et al. in press). Further more, foraminiferal and detrital pavements are common along the basal contacts and often

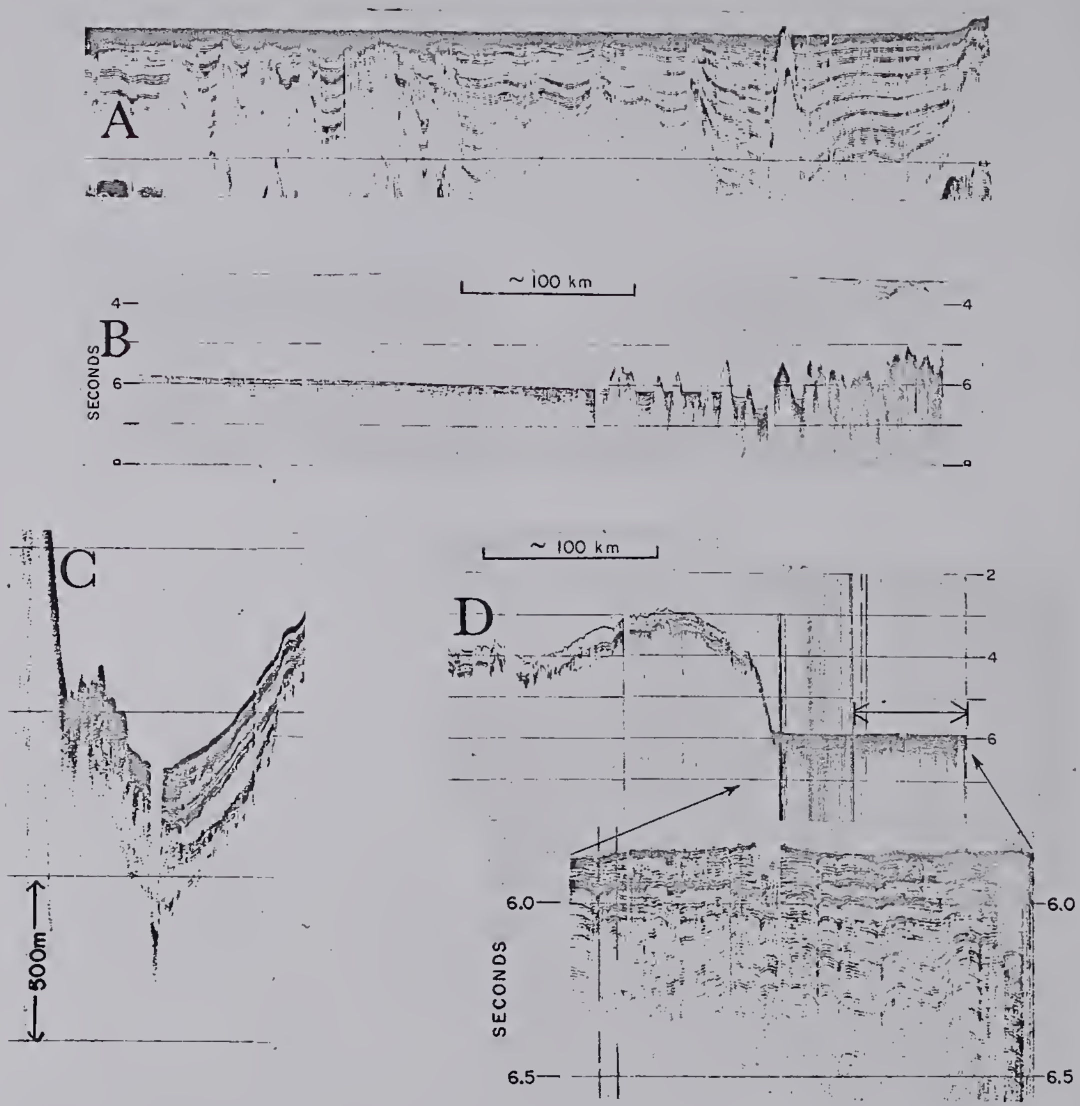


FIGURE III-1 - Sedimentary layering produced by "fluid" sediment flows i.e. turbidity currents

A. "Sediment ponding" occurs in small abyssal plains where contemporary influxes of sediment always fill the deepest depressions so as to maintain a level sea floor, despite active folding of the sub-strata. RC9, Balearic Abyssal Plain.

B. Hills at the seaward margin of a submarine fan act as dams and prevent the transport of continentally derived clastics into isolated depressions on the flank of the mid-oceanic ridge. (Ewing, Eittreim, Truchan and Ewing 1969)

C. The fluid flows carve deep canyons into the continental slope.(Ryan et al. in press)

D. Flows across deep-sea fans are confined to channels; however, the spillage part of the sediment suspension from the channel builds up natural levees above the level of the fan itself. (Ewing, Eittreim,Truchan and Ewing 1969)

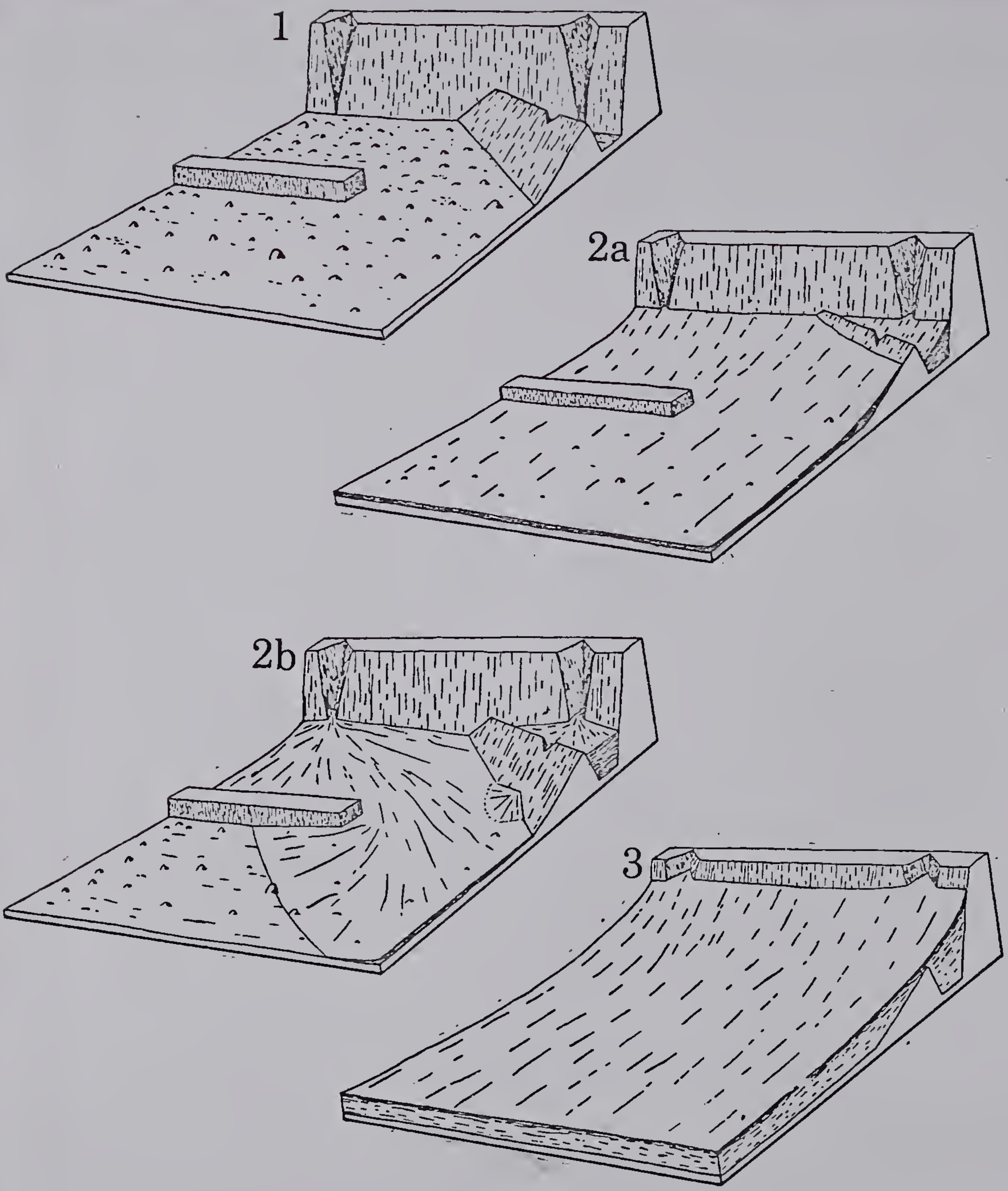


Figure III-2 - Different topography resulting from sedimentation by turbidity currents and by volume suspension in ocean. Relief about 2 miles, vertical exaggeration x 100 (1) Continental slope with submarine canyons (origin not relevant), and irregular deep-sea floor without sedimentary cover. (2a) Deposition of sedimentary blanket by volume suspension. Blanket thins from shore outward but is otherwise virtually uninfluenced by relief. (2b) Deposition of flood sediment by turbidity currents.Submarine canyons are at apices of low flat deep-sea fans. Sediment from one canyon is ponded by tilted fault block, and comparatively thick deposit is formed. Sediment from other canyon flows around low fault block, but some alee parts of deep-sea floor remain uncovered. (3) More mature stage of deposition in which initial topography is completely covered. Deep-sea fans have coalesced, alee areas have been filled and topography gives few obvious clues about dominant type of deposition. From Menard (1955).

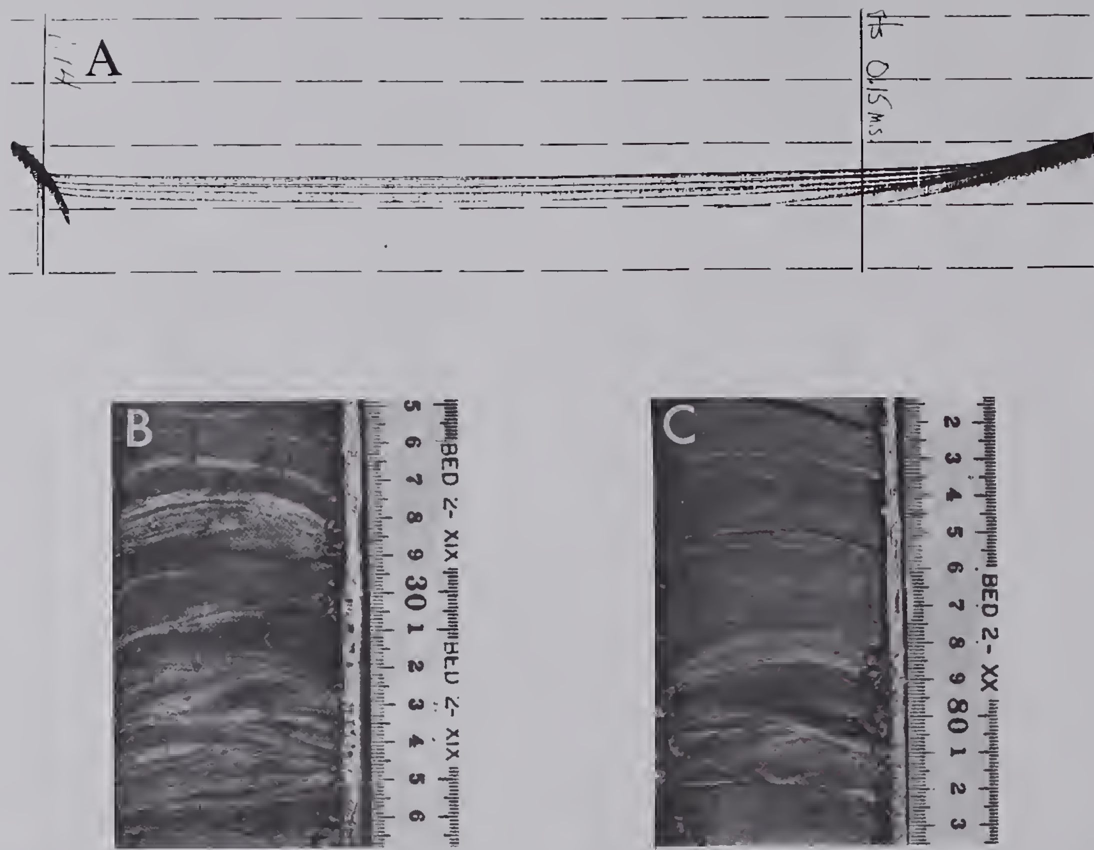


FIGURE III-3 - The interrelation of abyssal plains and deep-sea sands

A. Sub-bottom reflecting horizons beneath small abyssal plains are continuous from margin to margin and pinch out against all protruding relief. The depth of the reflectors correspond exactly to the levels of layers of deep-sea sands interbedded in lutite. Vertical scale is 40 meters between scale lines. (Ryan et al., in press)

B & C. The sand layers recovered in cores from this abyssal plain show vertical grading, have sharp base contacts on the lutite, exhibit current-bedding structures such as parallel lamination and cross-bedding, and range in thickness from 2 to 15 cm. (Ryan, Workum and Hersey, 1965)

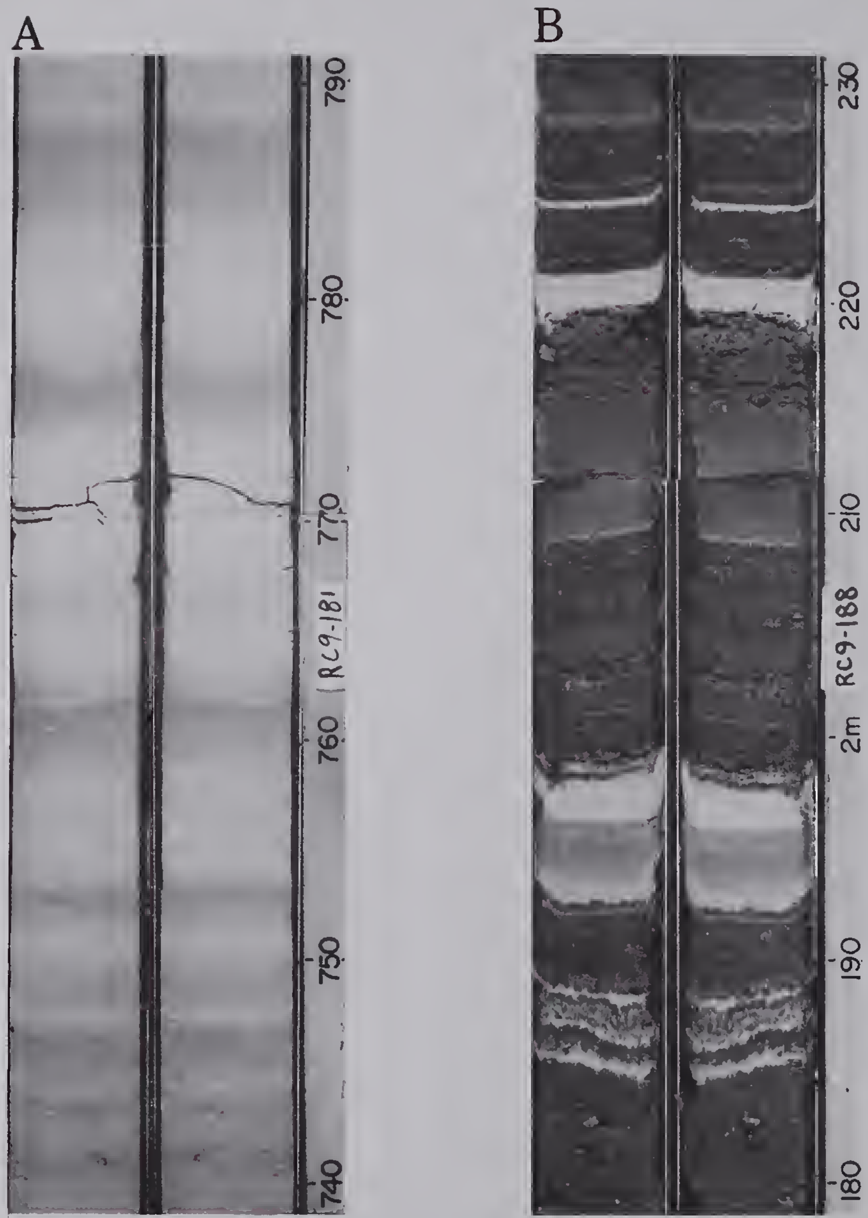


Figure III-4 - Pelagic sedimentation

A. & B. These core photographs show the accumulation of continuous pelagic deposition layer by layer as the result of vertical settling of detrital and biogenic materials. The change in colors and texture represent changes in sediment source, climate , productivity, and oxygen level of the bottom water. No current activity is indicated. These sediments are found interbedded between turbidites, or on isolated topographic elevations in abyssal plains.

resedimented, partly imbricated mud pebbles are interbedded in the sand of the overlying turbidite. The materials in the pavements and mud pebbles are derived from the local channel floor or wall by current erosion (as opposed to slumping) by the process of winnowing and scour.

Another indication of erosion at the base of a turbidity current flow is the development of scour marks (i.e., grooves, flutes and tool marks) in the subadjacent lutite directly beneath. Examples of this are shown in Fig. III-8, A and B. The scour depressions are subsequently filled with basal sands of the turbidites, with a marked appearance of bedding unconformity.

In conclusion, evidence has been presented by many authors that certain structures of the deep-sea sand beds indicate deposition in hours or days, rather than over a period of years or hundreds of years.

3.2-2 Flysch-type sandstones in outcrops on the continent

For many years, the deep-sea sands have been compared to the flysch-type sand beds deposited in ancient sedimentary troughs, now uplifted onto land (Gorsline and Emery, 1959; Nesteroff and Heezen, 1963; Kuenen, 1964). (For an excellent summary, see Von Rad, 1968 and Tables III-1 and III-2 from his test.) As is reflected in the literature, far more turbidites have been studied on land than in the oceans - and outcrops enable a more comprehensive view of sedimentary facies. The beds and internal structures can be traced over much greater distances and the bedding contacts can be examined in planar view as well as in cross-section (something impossible in small-diameter, deep-sea cores). Much of the discussion of the previous section on deep-sea sands applies to flysch-type sands as well.

Several facts about the nature of the hydrologic regime during deposition are clearly revealed in continental exposures. The erosional marks at the base of the turbidite layers beneath the graded interval can be examined in terms of the direction of current flow, type of marks, i.e., grooves, flutes, ripple marks (which tell about local current dynamics), and vertical extent (depth) of erosion (see Fig. III-10). The strike of the sole marks is a preserved indication of the direction of the flow at the time of erosion and has, in many areas, been used to give an indication of paleoslopes in the ancient basin.

The facies changes that take place over a period of time at the same location are shown by the vertical succession of turbidity layers in outcrops on large escarpments. One may extrapolate the affects of changes in sources or changes in basin geometry as the basin fills with sediment.

3.2-3 Length of time of the turbidity current depositional cycle

Core samples give only slight evidence of the length of time required to deposit the various intervals of the turbidity unit. Thus, direct sampling cannot yet settle the question of whether the turbidity unit is really (1) a product of high-speed episodic flows, or (2) steady-state continuous deposition from normal traction bottom currents. (For summaries of these arguments, see Hubert, 1964; Kuenen, 1967; Von Rad, 1968; van der Lingen, 1969.)

(1) From a sedimentological examination of flysch deposits, Kuenen (1967) has developed a line of reasoning which argues that turbidite deposition occurs primarily from episodic flows; i.e., as if "in a single casting". He further states that the observed bedding structures can only have resulted from deposition by currents containing sediment in suspension. This does not deny the existence of sedimentary deposition from normal near-bottom currents in abyssal environment, but infers that currents move their sand load by rolling and saltation and flow, regardless of the presence of sediment in suspension or even of bottom slopes. Turbidity currents, on the other hand, carry their total load in suspension, apart from a possible minor traction carpet (Sanders, 1965). For normal currents to continuously deposit repeated sequences of vertically graded beds over large lateral areas, they must wax and wane in a very short period of time and yet at intervals of hundreds or thousands of years. The upward fining of the grain size in a sand bed deposited by bottom traction cannot be explained as being the result of a gradual shifting of the current towards a finer source material; hence, grading must be attributed to gradual waning of the current itself.

The deposition of a bed by a turbidity current is an entirely different process from that of shifting a bottom load by traction. The latter erodes or deposits according to its competency for bottom traction, whereas the deposition by a turbidity current is governed by capacity for suspension (Kuenen, 1967). In treating this aspect of the capacity for suspension, and comparing the structures seen in cores to those created in experimental flume studies, arguments can be proposed for a short time interval to explain: (1) the admixture of lutum in the lower graded interval; (2) the climbing ripples, from which the angle of climb is a direct function of deposition rate; (3) the very homogeneous and slick-looking pelitic interval with no signs of post-depositional winnowing; (4) the restriction of benthic animal burrowing to only the upper part of the pelitic interval and the overlying pelagic interval, - never in the lower intervals of the turbidity sequence; and most important; (5) the development of convolute lamination during deposition (in the interval of current-ripple lamination). Convolute lamination may be produced by current drag (Sanders, 1960), loading (Dzulynski and Walton, 1965) or by a quick-sand condition, - liquification (Kuenen, 1967), each suggesting a very swift accumulation of sediment with high porosity rendering the deposit expremely hydroplastic. In flume studies, such a condition deteriorates in a few hours as water is extruded by compaction.

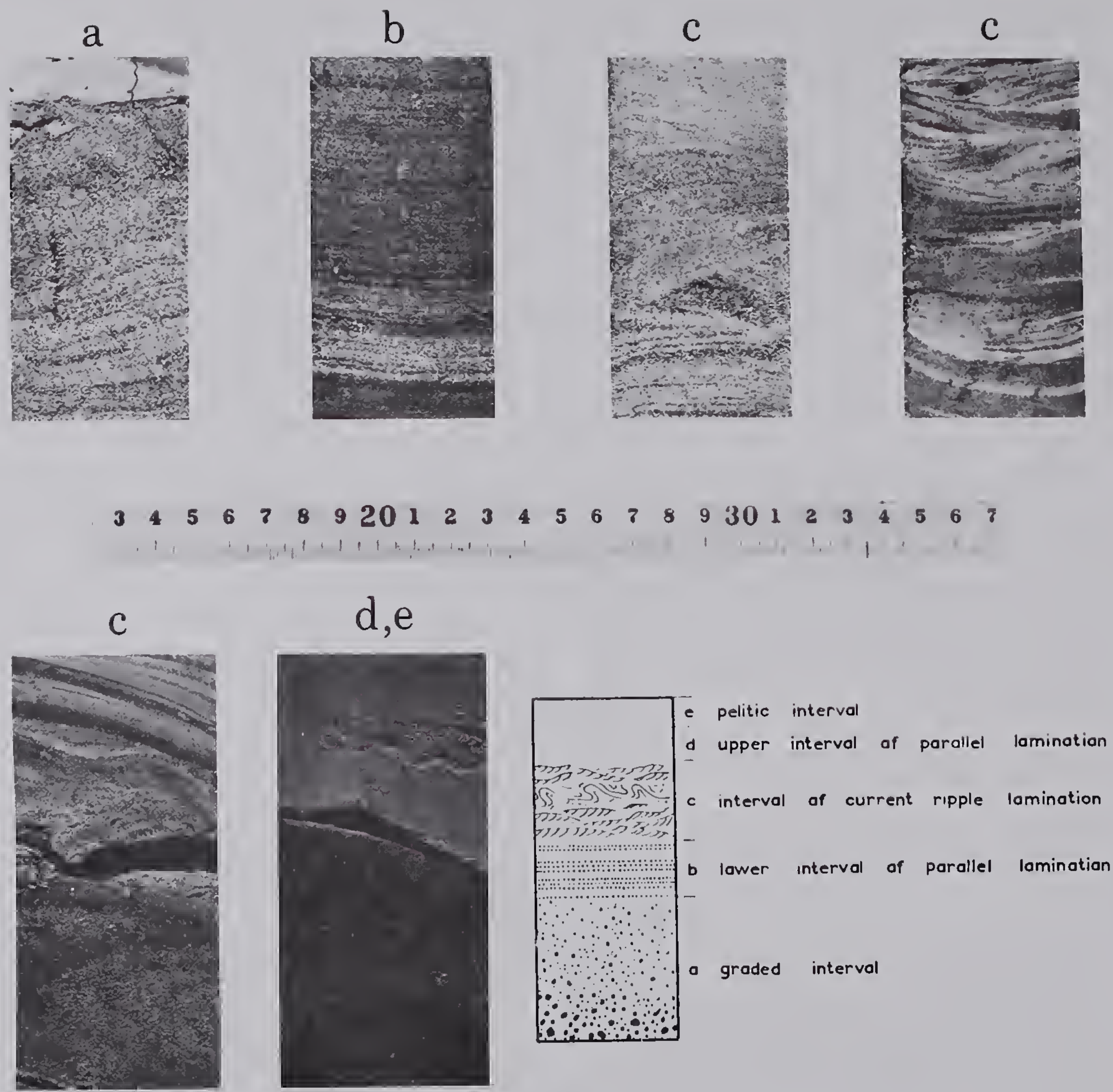


FIGURE III-5 <u>The five major structural intervals of the turbidite model</u>

Each interval is characterized by one dominant type of sedimentary structure as shown in the various core photographs. The interval of current ripple lamination includes climbing ripples, cross-bedding, and convolute laminations. Beds displaying all five intervals are rare, and although all may not be present in one bed, the order of superposition of intervals is always the same (after Bouma, 1962). Knowledge of the process which build these bedding structures permits insight into the hydrologic realm of turbidity current deposition.

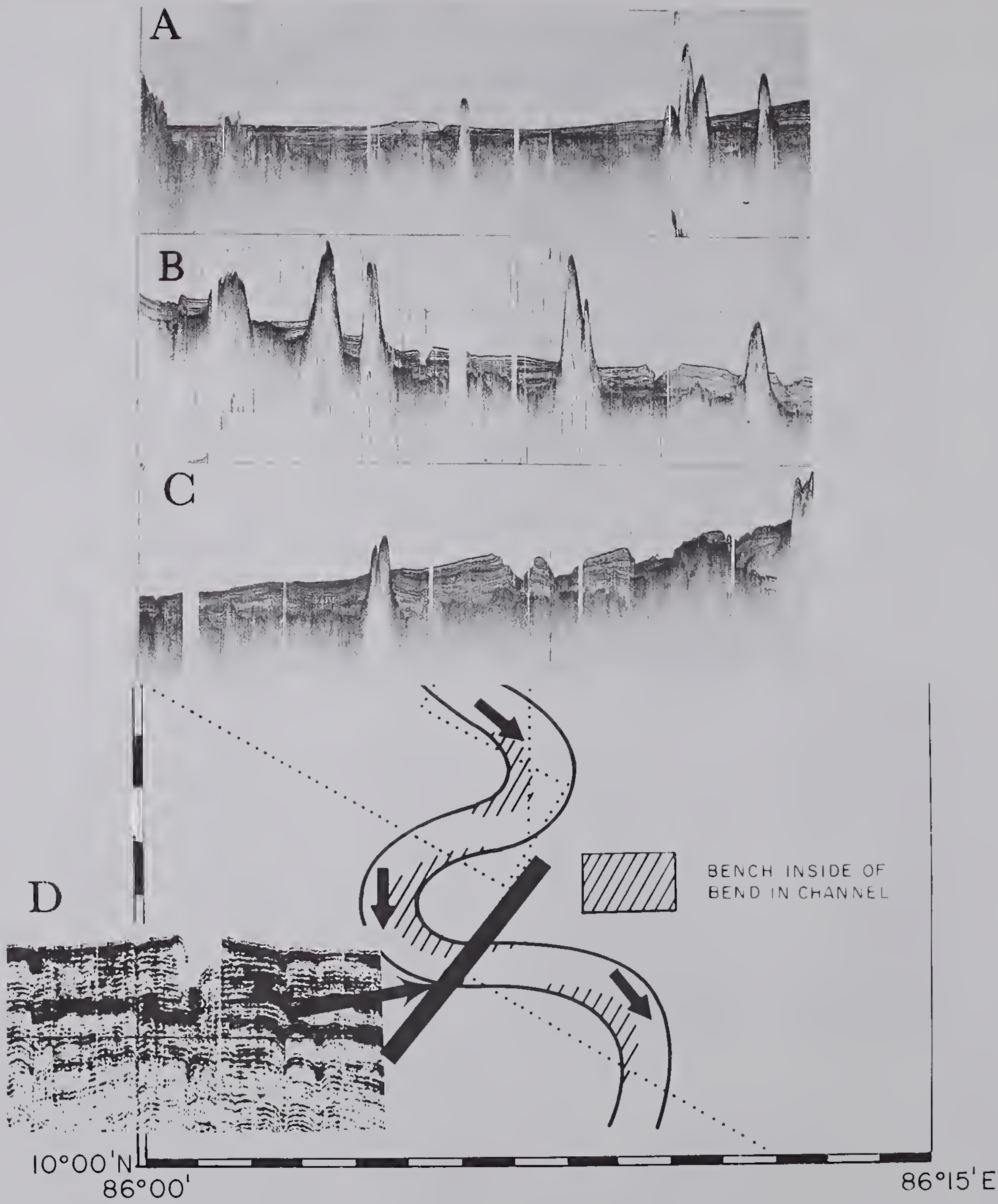


FIGURE III-6 Examples of the structure and configuration of deep-sea channels

A. Note the high natural levee on the channel in the center of the profile. Horizontal scale, 200 km; vertical scale, 1000 m between scale lines.

B. & C. Erosion of channel walls is shown by outcropping of sedimentary strata. The benches on the channel walls are protrusion of sub-bottom layers more resistant to erosion, assumed to be by turbidity currents. Profiles A,B, and C from Ewing, Aitken, and Ludwig, 1968.

D. Meandering deep-sea channels are found on many fans; The cross-section of the channel shows a steep eroded wall on the outside of bends, and depositional terraces on the inside. (Ewing, Eittreim, Truchan and Ewing, 1969)

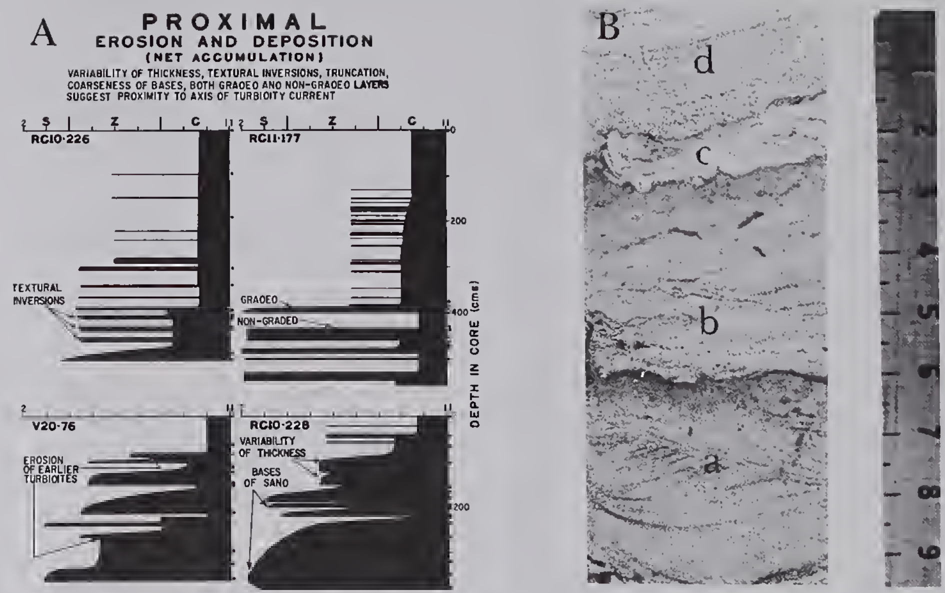


FIGURE III-7 Texture and bedding structures of turbidites from regions within or proximal to deep-sea channels

A. Graphs of mean grain size show incomplete grading, inverse grading, and truncation of turbidite sequences. Note variable thickness of beds. (Horn, Horn, and Delach, 1970)

B. Sands in a core from the proximal zone, showing four different turbidite units (a,b,c, and d), each truncating the upper intervals of the previously deposited unit. The dark contacts between the units are major erosional surfaces and consist of foraminiferal pavements and other winnowed debris. The individual turbidites are recognized here by gross differences in mineralogy and color. (C. Hollister, Columbia Univ. Ph.D. thesis, 1967)

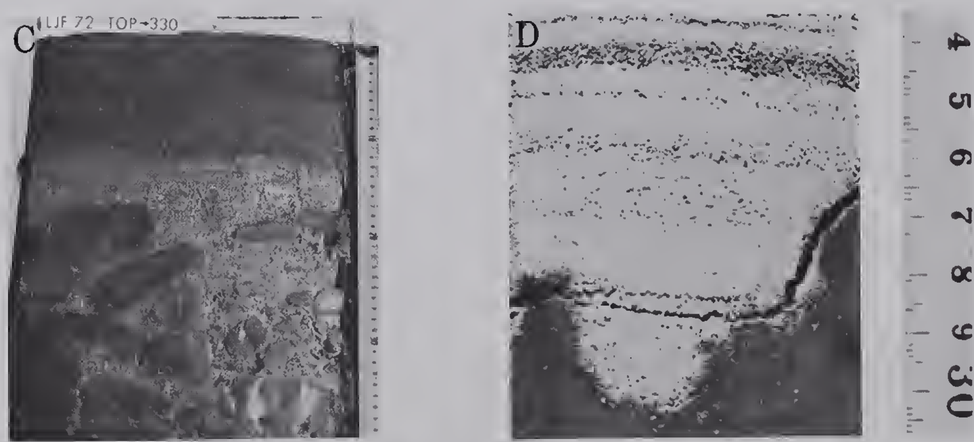


FIGURE III-8 Evidence in cores from fans of erosion of the sea-floor by turbidity currents prior to deposition

C. Cross-section of a scour mark (flute or tool?) which has obviously been cut into the subadjacent clay and has only then been filled with the laminated sands. (Schneider and Heezen, 1966)

D. Resedimented, partly imbricated mud pebbles interbedded in a thick graded coarse sand layer. The mud pebbles are probably derived from erosion of channel walls. This core was retrieved from a small distributary channel of the outermost part of the La Jolla fan valley; depth 1083 meters (from Von Rad, 1968).

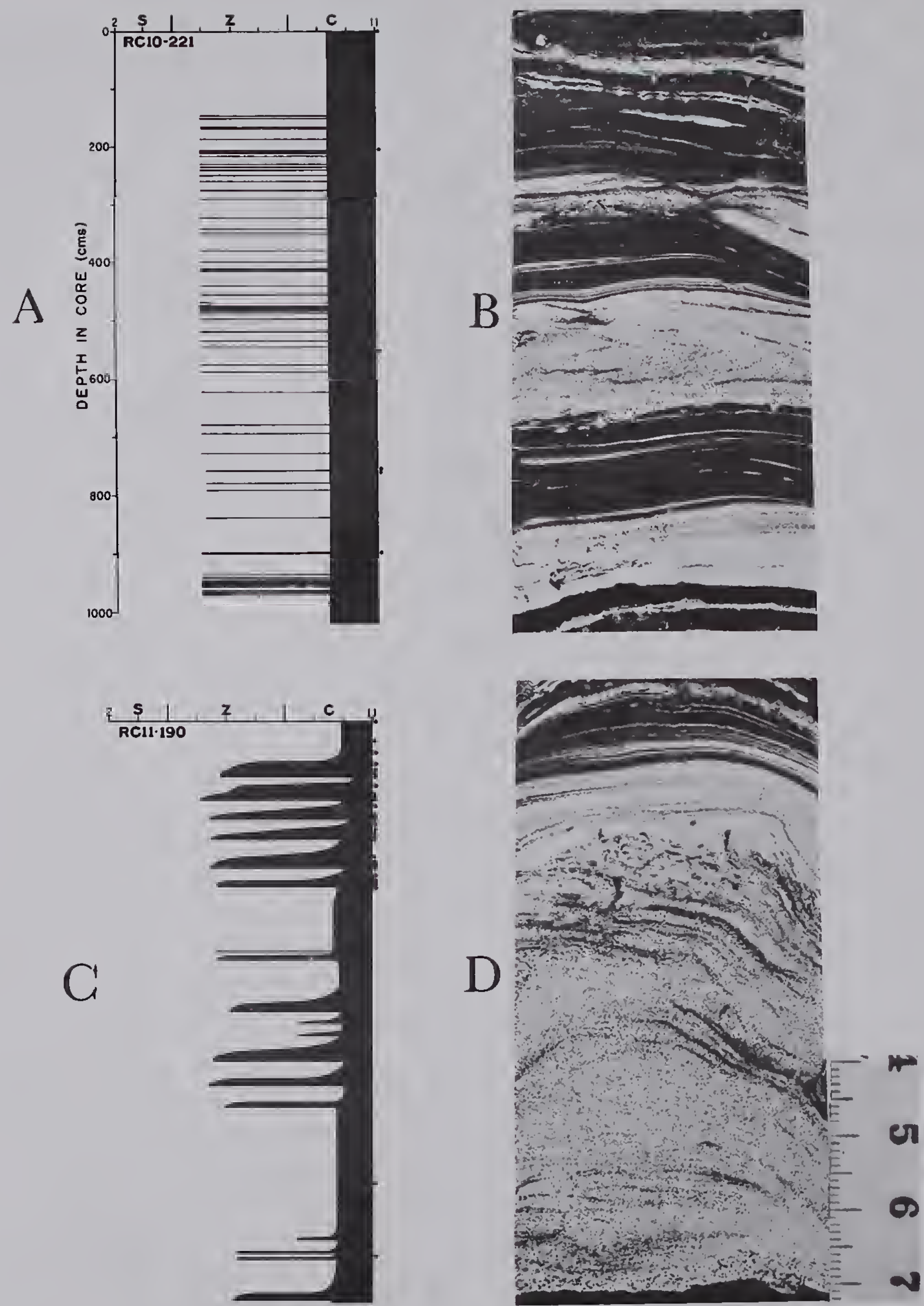


FIGURE III-9 <u>Textures and bedding structure of turbidites on the abyssal plain</u>

A. Plot of mean grain size for Core RC10-221 from a distal fan province. The clastic beds are medium silt, are numerous per meter of core, and have uniform grain size with little or no grading. (Horn <u>et al</u>, 1970)

B. Example of typical coarse-grained layers in cores from the distal fan (or continental rise) province. Note abrupt bedding contacts at both top and bottom of the silt beds, and the presence of parallel laminations and cross-bedding. Grading is lacking. (Hollister, 1967)

C. Plot of mean grain size for Core RC11-190 from an abyssal plain province well away from the mouths of deep-sea channels. Note the readily apparent grading within each turbidite ranging from about 50 microns at the base to 5 microns in the upper pelitic interval; note also the irregular intervals of pelagic sedimentation between the turbidites. (Horn <u>et al</u>, 1970)

D. Example of a typical turbidite layer from an abyssal plain distal from the channel or fan. Note sharp base contact, grading, and the sequence of the five turbidite intervals. (Hollister, 1967)

In a few carefully documented cases, the amount of erosion of the sea floor prior to deposition from an individual flow can be estimated by measuring the amount of normal pelagic sequence missing. The amount of sequence missing would be judged from stratigraphic correlations with similar pelagic sediments in nearby non-turbidite regions. Studies of stratigraphic sequences in the Mediterranean Sea, Silver Abyssal Plain and Caicos Ridge, Tagus Abyssal Plain, and Sigsbee Abyssal Plain indicate that a single turbidity current can remove by erosion up to 2 m of material from the sea floor in areas within channels, on the lower fans, and in regions of the abyssal plain near the mouths of these channels. (For a case study, see Part IV.)

3.3 THICKNESS OF TURBIDITE BEDS

The amount of material deposited at any one locality along the path of a turbidity current depends on many factors: (1) the magnitude of the current; (2) the bed-load; (3) the grain-size of the sediment suspension; (4) the gradient of the topographic slope; and (5) the distance from the source. Beds can range from less than 1 cm to several meters. Long cores at a given locality recover a series of turbidite beds of more or less the same thickness, suggesting a certain predictability.

Generally, the upper course of a deep-sea channel acts as a bypass route. Deposition here consists of only the most coarse pebbles or boulders, and often is of the same magnitude as pre-depositional erosion (steady-state).

Within the lower course of the channels, near the seaward edge of the fan, beds vary in thickness from a few centimeters to about 10 cm, and consist of mostly the <u>lower graded interval</u> and the lower interval of parallel laminations. The upper intervals are very thin and are never deposited, or are truncated by subsequent flows. In regions far from the source, on the distal margins of the abyssal plain the units are generally much thicker -- 20-50 cm -- but contain primarily the fine silt and clay of the <u>upper interval of parallel lamination</u> and the pelitic interval. Little basal erosion is evidenced in these regions. (See Figure III-9).

Detailed studies of cores from several abyssal plains, particularly in the Mediterranean Sea, and on the floor of the trenches, have allowed individual turbidite beds to be traced laterally across the basin floors from source regions to the distal margins (Ewing <u>et al</u>, 1958; Van Straaten, 1964; Ryan <u>et al</u>. 1965; Schneider and Heezen, 1966. Where a single bed can be traced over a wide area of the sea floor, it is usual to find a very gradual change in total thickness and a decrease to finer grain sizes away from the source.

Many turbidity beds can be traced over the entire abyssal plain in basins with diameters of only a few hundred kilometers. In the larger abyssal plains like those in the North American Basin, individual flows travel many thousands of kilometers along slopes as gentle as 1:2000. However, only the finest material from the largest events actually reaches the distal regions, and cores from these areas contain fewer turbidites per meter of sediment.

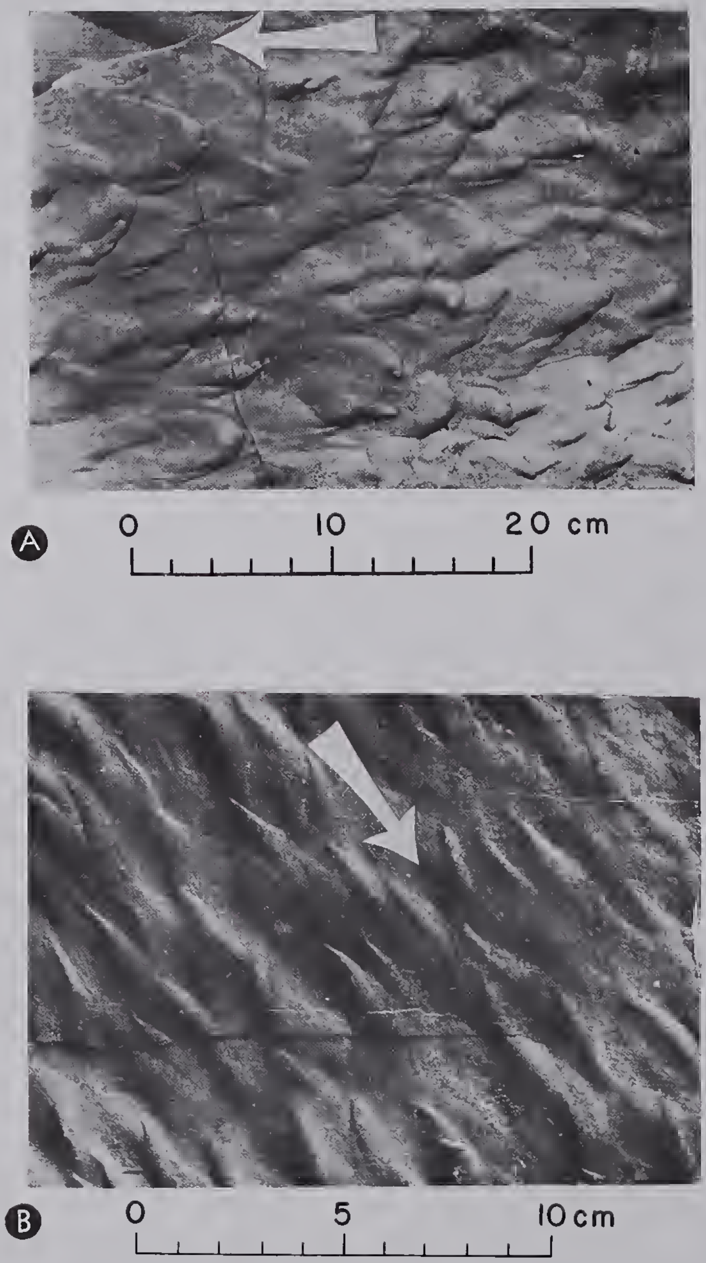


FIGURE III-10 Photographs of scour and tool marks shown as counterparts on basal surfaces of sandstone (flysch) slabs.

A. Longitudinal scour marks (pointed and spirally-twisted flutes). Note indications of only one direction of flow, inferred to be the result of scouring of uniform firm cohesive mud by eddies of turbulent suspension and subsequent deposition of sediment from suspension, which filled bottom depressions and molded an impression of the bottom relief.

B. Longitudinal scour marks (pointed flutes). Arrows show direction of transport of turbulent suspension. A and B from Sanders (1965).

PART IV

HISTORICAL RECORD OF TURBIDITY CURRENTS

4.1 SUBMARINE CABLE BREAKS

For nearly a century and a half, the ocean floors have been criss-crossed with a system of submarine telegraph and telephone cables. By 1858 a cable had been laid between Ireland and Newfoundland. That cable was operational until it broke only three months later. The breaking of the cable provided the first suggestion of activity rather than tranquillity on the deep-sea floor. In fact, until recently, cable breaks provided the only available historical record of strong movements on the sea bed. In more recent years, seismic profiling, deep-sea nephelometry, deep-sea photography, and other techniques have provided additional sources of information.

In some cases, a logical interpretation of the cause of sequential breaking of cables implies downslope movement of sediment masses at high velocities and often over great distances, even on gentle slopes. Furthermore, the precise location of cable breaks and faults has pin-pointed the particular topographic settings where such slides and flows are most likely to occur.

Some case histories of cable breaks on the ocean floor are discussed below.

4.2 THE GRAND BANKS TURBIDITY CURRENT

On November 18, 1929, an earthquake of magnitude 7.2 shook the continental shelf and slope south of Newfoundland. The six submarine cables lying nearest the continental slope were broken instantaneously, and for more than 13 hours after the earthquake, there occurred an orderly sequence of breaks of cables lying in increasingly deeper water for over 300 miles from the earthquake epicenter. Twelve cables were broken in 25 places. De Smitt (1932) showed that (1) the direction of progression of destruction was seaward, (2) the velocity of progression decreased with increase in distance from epicenter, (3) the area of broken cables was restricted to the deepest part of the ocean — narrow in the northern part and broadening to the south, with breaks radiating outward and with not a single break on the shallow continental shelf, and (4) the ocean floor in the neighborhood of the breaks had an average slope of $1^{\circ}50'$ and that about half the breaks were on a slope of less than 1°.

The relationship between earthquakes and submarine telegraph cable breaks was pointed out by Milne, as early as 1897, and by others; but the first strong argument that earthquake-triggered turbidity currents occur in the modern oceans was made by Heezen and Ewing in 1952.

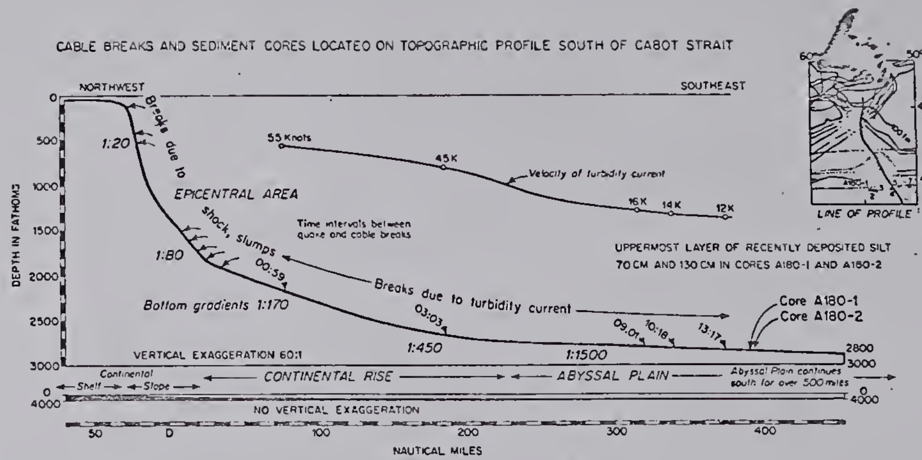
They concluded that the "events associated with the Grand Banks earthquake ... may be considered as a full-scale experiment in erosion, transportation and deposition of marine sediments by a turbidity current in which the submarine cables served to measure its progress and to give tangible evidence of its force."

Following the initial shock, which occurred beneath the contiental slope south of the Laurentian Channel, the trans-Atlantic submarine cables were broken in a north to south sequence, as shown in Fig IV-1A. According to Heezen and Ewing's explanation, each successive cable was broken by a single large dense flow of sediment and water which had originated from a series of slumps in the epicentral area. They inferred that this flow was a turbidity current, and concluded from the points of breakage, that the current had travelled down and across the continental slope and rise, and ocean basin floor, far out onto the Sohm Abyssal Plain, - a distance of well over 450 miles from its source area (Fig. IV-1B). By observing the time of each cable break and the distance from the earthquake epicenter, they calculated that on the continental slope, where the bottom gradient is 1:10 to 1:30, the velocity of the turbidity current exceeded 50 knots (see graph in Fig. IV-1A). By the time the turbidity current caused the seaward-most cable (on the abyssal plain) to snap, the flow was still travelling at a speed in excess of 12 knots along a gradient of less than 1:1500.

Sediment cores taken on the abyssal plain south of the epicentral area revealed an uppermost layer of graded clastic sediment (Heezen et al., 1954). The area of the deposit covered at least 100,000 km^2.

The high velocities theorized for the turbidity current and the association of this event with essentially instantaneous deposition on the ocean floor of the layer of mud with a volume in excess of 10^{11} m^3 was hard to comprehend, and initially was not accepted (Kullenberg, 1954; Shepard, 1954; Jones, 1954). Shepard suggested that the slides developed more slowly on the gentle outer slopes and built up pressure against the cables gradually until they broke, or that the sequence could be explained as at least in part a coincidence and in part to the slower development of slides at an increasing distance from the epicenter. (For a review of statements of the arguments against the concept of the high velocities and instantaneous deposition, the reader is referred to Heezen, 1963, pp. 750-751)

A.



B.

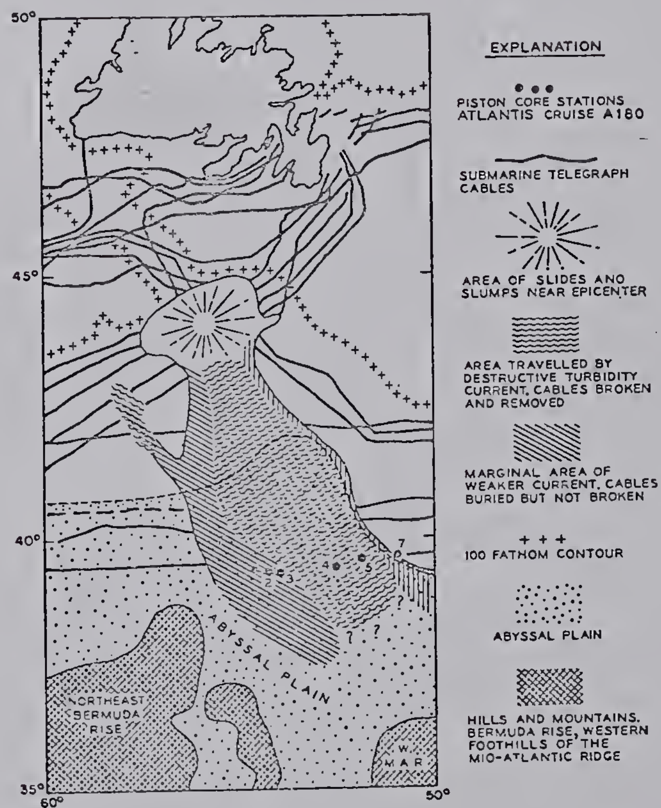


FIGURE IV-1 Path and velocity of the Grand Banks turbidity current (Heezen, 1963)

A. Topographic profile south of the Laurentian Channel with location of cable breaks and sediment cores. Superimposed is a graph of the velocity of the Grand Banks turbidity current.

B. The path of the Grand Banks turbidity current showing locations of cable breaks, region where cables were buried, and area of earthquake epicenter and initial slides and slump.

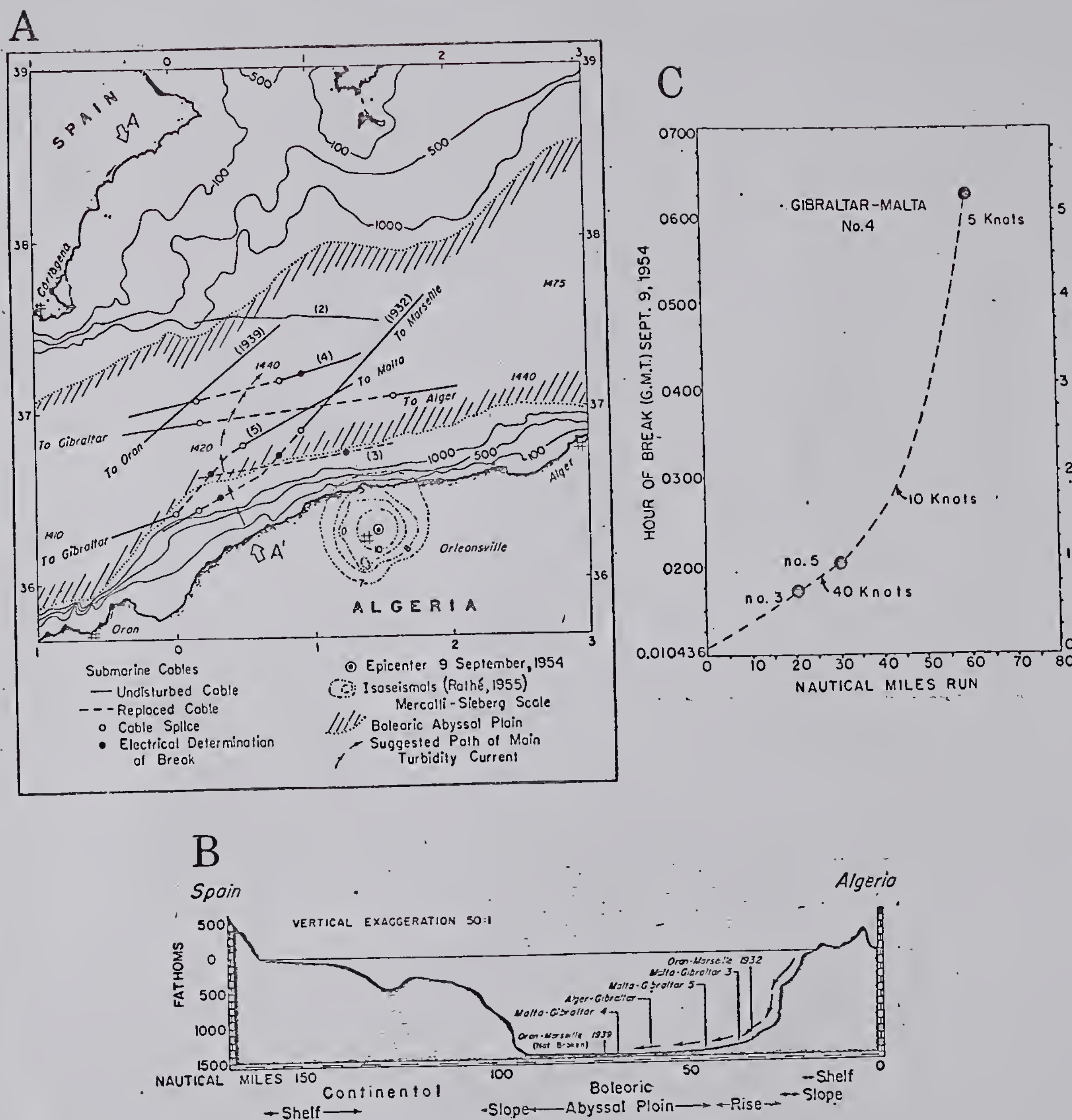


FIGURE IV-2 - Path and velocity of the Orleansville turbidity current (Heezen and Ewing, 1955)

A. Plan view showing distribution of submarine cables on the southern Balearic Abyssal Plain, and the location of the section of the cables destroyed following the 1954 Earthquake near Orleansville. Note that the northernmost cables were not interrupted by the turbidity current.

B. Profile of the basin showing the downslope flow of the earthquake initiated turbidity current.

C. Time-distance graph for the 1954 turbidity current. The velocities determined are of the same sort of magnitude as the 1929 Grand Banks turbidity current.

4.3 ORLEANSVILLE TURBIDITY CURRENTS

Much of the skepticism concerning the high velocities attributed to turbidity currents was dispelled when Heezen and Ewing (1955) reported a similar series of sequential downslope cable breaks on the floor of the western Mediterranean following an earthquake near Orleansville, Algeria, in 1954. Five submarine cables crossing the continental slope and Balearic Abyssal Plain were snapped in an orderly sequence. The timing of the breaks was such that it could again be inferred that if this were a single event initiated from one source, then the speed of the flow of sediment exceeded 40 knots by the time it reached the base of the steep continental slope. The path of the Orleansville turbidity current is shown in Figs. IV-2A and 2B. The velocity of the current, calculated from the time elapsed between the disruptions of service of each cable circuit, and on the assumption that the flow followed the path shown in Figs. IV-2A and 2B, is shown in Fig. IV-2C.

In the case of the Orleansville event, it is possible that several submarine avalanches along the Algerian coast were triggered simultaneously by the earthquake, and that individual turbidity currents coalesced to form the broad path of destruction along the southern boundary of the abyssal plain. It is interesting that these turbidity currents, having reached the axis of the plain at 37°N, did not flow up hill to the north and break the northernmost cables, even though to do this the current would have had to travel up a slope of only 1:1500. In fact, cores from the tops of small knolls that protrude only a few tens of meters above the base level of the abyssal plain contain no layers of graded sands and silts, as do the numerous cores on the flat plain itself (Eriksson, 1965).

4.4 THE TURBIDITY CURRENT AT KADAVU PASSAGE, FIJI

On September 14, 1953, an earthquake near Suva, Fiji, triggered a large submarine avalanche which collapsed into the Kavadu Passage and generated a turbidity current (Heezen, 1959). This current swept southward down the natural channel between the islands (Fig. IV-3A) destroying over 75 miles of submarine cable in its path. An observed displacement of the shallow sea floor near the mouth of the harbor triggered an avalanche of a large mass of mud and coral debris. A tsunami was seen at the reef entrance only seconds after the shock was felt at Suva (Houtz and Wellman, 1962).

The avalanche developed into a turbidity current as it moved down the steep (15°) insular slope for a distance of at least 10 miles. The minimum strain on the cable was in excess of seven tons, its breaking strength. The current probably reached at least 20 knots, judged by the time of the interruption of service on the cable. The cable running southeast along a sill in the strait was undamaged since it was not in the path of the avalanche. Of the cable that broke, over 60 miles was either totally buried or torn up and carried out of range of the subsequent repair survey search. However, a 40 m length of twisted cable which was completely knotted upon itself was recovered. Corrosion had been removed in some areas by an apparent sand-blasting effect, judged to have occurred after the knotting had taken place. Houtz and Wellman (1962) speculated that the extent of sand-blasting (compared with other experimental evidence) indicates that the turbulent flow of sand was travelling as fast as 20 knots as it swept the cable downslope. The recovered strand was found several miles from its original site.

4.5 THE 1908 AND 1909 MESSINA TURBIDITY CURRENTS

In the early morning of December 12, 1908, a seismic disturbance beneath the Strait of Messina (between Sicily and Calabria, Italy) caused a submarine avalanche in the head of a canyon a few miles west of the town of Gallico (Ryan and Heezen, 1965). Because of continuous strong bottom currents passing along the shore northwards through the Strait, a large reservoir of sediment had accumulated in the head of this and other canyons. Shortly after the initial slide, the avalanche destroyed a submarine cable at a depth of 250 fm where it crossed the narrow passage between Calabria and Sicily. This was the first failure of the cable since its emplacement in 1904. It is inferred that the avalanche of mud and sand evolved into a turbid sediment suspension which, under the influence of gravity, propelled itself further down the canyon, because some nine hours and forty-five minutes later another cable (along the Malta-Zanta route) snapped exactly where it crossed this same canyon (see Fig. IV-3B). The average velocity of the turbidity current to this point can be calculated to have been in excess of 12 knots. However, it is likely that the initial velocity down the continental slope was much higher and that at the point in which the current encountered the Malta-Zante cable, in a depth of 1790 fm, it was travelling more slowly, having dispersed a long sediment tail with a subsequent loss in density and capacity. Half an hour later, the flow apparently persisted at the cable, because changes in electrical resistance at the break continued to occur.

On February 26, 1909, following one of the numerous aftershocks of the Messina earthquake, the same Malta-Zante cable was again faulted where it crossed the axis of the second canyon. This indicates that another avalanche took place in a canyon further east on the continental slope of Calabria.

Four cables laid down in 1869 connecting Malta to Alexandria, Egypt, have never experienced a deep water failure. However, these cables pass south of the Messina Abyssal Plain and thus are not in the path of turbidity currents travelling through channels across the Messina Cone. Furthermore, the only deep water failure of the Malta-Zante cable occurred in connection with the 1908 earthquake and 1909 aftershocks.

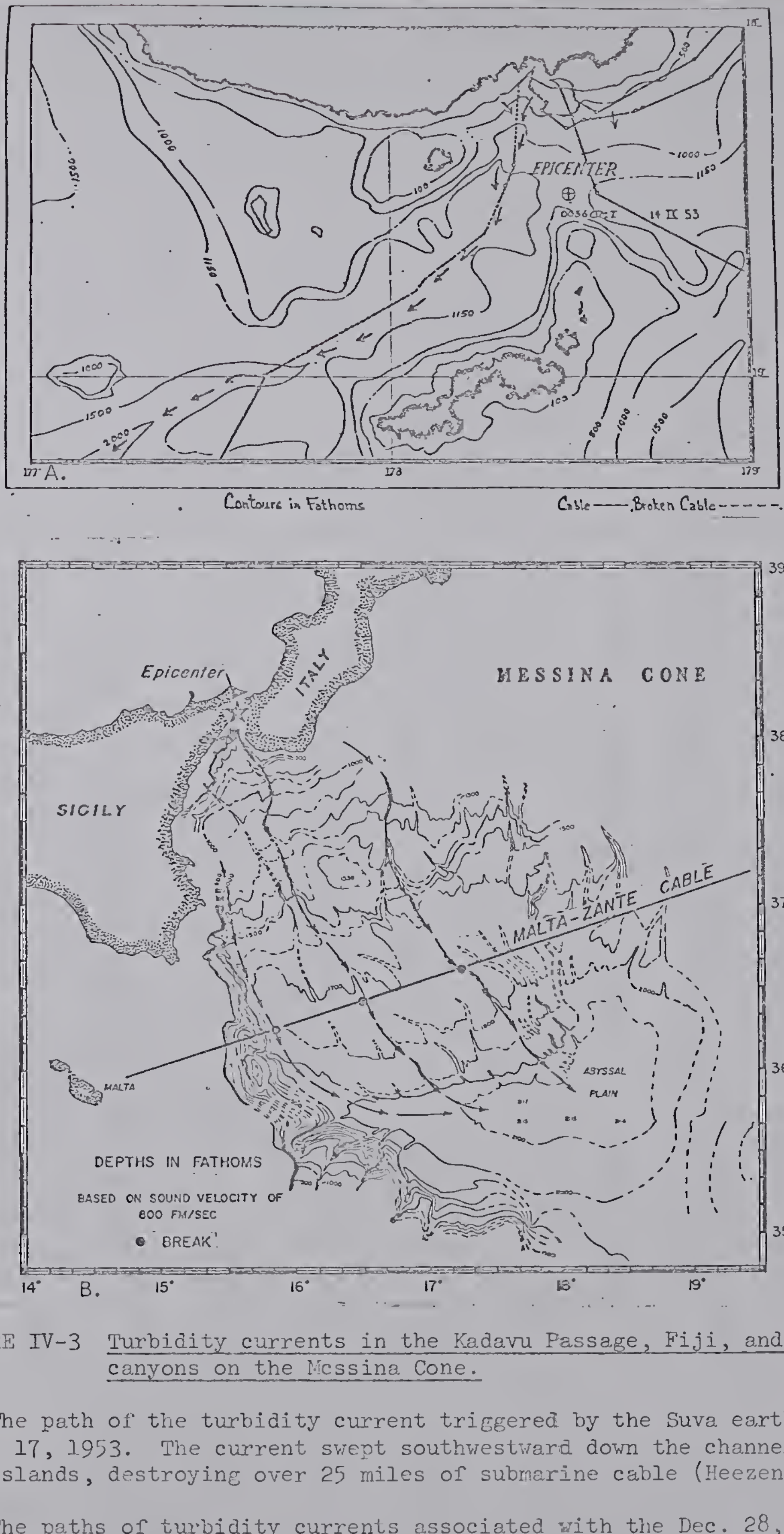


FIGURE IV-3 Turbidity currents in the Kadavu Passage, Fiji, and in canyons on the Messina Cone.

A. The path of the turbidity current triggered by the Suva earthquake of Sept. 17, 1953. The current swept southwestward down the channel between the islands, destroying over 25 miles of submarine cable (Heezen, 1959)

B. The paths of turbidity currents associated with the Dec. 28, 1908 Messina earthquake and the Feb. 26, 1909 aftershock. The turbidity currents were confined to submarine canyons on the Messina Cone and snapped the Malta-Zante cable exactly where it crossed the canyon (Ryan and Heezen, 1965)

Piston cores taken in 1964 on the floor of the submarine canyons where the cables broke contained an upper layer of graded sand with a thickness of more than 3 m. The sand contained fragments of volcanic pumice of identical composition to fragments found wedged in the twisted and pinched ends of the broken cable when it was retrieved for repairs. The cable was apparently destroyed by a sudden action rather than prolonged attrition, for there were no signs of continuous wear by chafing.

4.6 CONGO AND MAGDALENA TURBIDITY CURRENTS

The Congo and Magdalena submarine cones are areas of particular interest because here cables have been laid directly seaward and oblique to submarine distributary systems of major rivers. A study of cable breaks for periods of more than twenty years has shown a marked coincidence between cable failures and abrupt changes in depth and shape of the river mouth and has established that these changes occurred only during periods of peak sediment discharge.

For instance, Heezen et al.(1964) have documented that cable breaks within the Congo Submarine Canyon on the Congo Cone occur with the greatest frequency in that region where the canyon has cut into the continental shelf. This river discharges nearly 10^{10} tons of sediment annually. Much of this sediment is presently trapped in the Stanley Pool; however, the remainder is dropped directly into the head of the Congo Canyon. At periods of peak flood activity cable breaks attest to submarine flows which are either the products of direct low density transport of turbid river suspensions, or, more likely, the products of submarine slumps and turbidity currents generated by instabilities within the extremely rapidly accumulating reservoir of sediment in the canyon head.

Similarly on the Magdalena Cone off Colombia South America, Heezen (1956) found that cable breaks occur every few years in response to visible slumps of the Magdalena river fan. The location of the cable breaks has consistently been confined to those sections of the cable which cross canyons and gullies cut into the cone. An interesting aspect of the 1935 Magdalena cable failure was that grass, still green, was brought up wrapped around the damaged cable, which lay in 750 fm. 12 miles seaward of the river mouth. In fact, cores taken on many submarine cones (i.e. Nile Cone, Amazon Cone, Congo Cone, Indus Cone, Ganges Cone, Cascadia Cone, etc.) contain numerous layers of organic debris of terrestrial origin; sometimes individual beds contain up to 90% plant debris.

The submerged deltas of major rivers have been areas of repeated trubidity current activity throughout most of geologic time. There are several reasons for this great activity. First, the high bedload of the river discharge accumulates directly into large submarine reservoirs. Secondly, organic debris in these sediments, upon decomposition, produces gases which render these high porosity plastic sediment bodies extremely vulnerable to catastrophic failure (Terzaghi, 1956; and Monroe, 1969). These regions are not only potential sources of turbidity sediment, but the trigger effects may be supplied by the river as well.

4.7 EPISODIC EVENTS THROUGH TIME

Although cable breaks give a history of sea floor activity over the last centruy and a half, and the mechanisms and sequences of cable breaks yield evidence about the processes of episodic events, the history revealed by cable breaks is short. The record is entirely too short to give a reasonable prospective concerning the frequency of episodic events. Furthermore, it is difficult to prove from the history of cable failures whether events such as the Grand Banks turbidity current are the principal events to be considered when planning the emplacement of sea floor structures, or whether turbidity current activity of the type and magnitude that occurs on submarine cones such as the Magdalena is more important. The Magdalena-type events are far more predictable than the Grand Banks-type, for the latter depend upon the occurrence of a large earthquake in a region of the slope which has been aseismic for a long time.

A better statistical representation of the frequency and magnitude of turbidity currents can be gathered from the study of deep-sea stratigraphy, i.e. the study of the succession of layers of sediment at the same location on the sea floor, deposited during a time framework of hundreds, thousands and millions of years.

Deep-sea stratigraphy involves data from two major sources, - the succession of layers of sediment seen in seismic reflection profiles and the sequences of sedimentary materials recovered in deep-sea cores. The recovery of sediment layers is accomplished by piston coring in the upper 20 m of soft unconsolidated material, or by drilling, for sampling to greater depths (National Science Foundation 1969, Initial Reports of the Deep Sea Drilling Project, Vol. 1, NSFSP-1, U. S. Government Printing Office, Washington, D. C.).

4.7-1 Events in acoustic records

Distinct sedimentary events of the past can be seen in the attitude and relationship of certain sub-bottom interfaces which are revealed as reflectors of acoustic energy and which are interpreted as equivalent to layers of sediment, or structural boundaries within layers. For instance, Figure IV-4A shows a section of a 3.5 khz seismic reflection profile where the sub-bottom reflecting horizons are interpreted to be equivalent to interfaces between layers of sediment of different acoustic properties and deposited sequentially in time. In this profile the deeper horizons are draped uniformly over the hills and across the valley floor; and the thickness of the layers remains uniform. However, of interest is an upper transparent layer (noted as 'a' in the profile) which is much thicker in the valley than on the hills. This type of non-uniform layering with greater sediment thickness in troughs is called "differential sedimentation." The sediment comprising layer 'a' in Fig. IV-4A has been transported into this region and has been deposited by a process apparently different than the process or processes which deposited all the older layers. This unique change in sedimentation process is believed by us to represent a depositional event. We do not necessarily know the length of time of this event from analyses of the acoustic record alone, for it might have been very rapid (a few days) or much longer (thousands of years). The lithology of the layers of sediment sampled in cores, would permit one to deduce (1) whether the material of layer 'a' in the valley is the product of a local slump, driven by gravity to the bottom of the topographic depression (i.e. rapid deposition) or (2) whether the material is a residual accumulation of sediments swept off the high and settled in the lows, during an intermittent period of strong bottom current activity. If the differential deposition was due to bottom currents, the current activity may have lasted for thousands of years.

Another example of deposition events recorded in acoustic records is illustrated in the reflection profile in Fig. IV-4B. In this profile, the sea floor appears as a surface (marked 'b') of characteristic overlapping hyperbolae. The layers beneath the interval of hyperbolae are different, - apparently smoother. We infer that at the time of deposition of the sediments which make up layer 'b' an event of unknown duration caused the contemporary sea floor to become corrogated. We offer as possibilities: (1) a period of current activity on the sea floor which constructed through the action of winnowing and controlled deposition, a series of large linear dunes (drifts) of regular wavelength; (2) a period of intense erosional solution which dissolved away parts of layer 'b', leaving gullies; (3) a sea-floor avalanche of which layer 'b' is the chaotic debris. The time in the past during which such events took place could be determined by coring. If it could be shown that 'b' is debris from a slump, then its mode of implacement would be inferred to have been more or less instantaneous. However, the regional uniformity in thickness of the layer'b' is very strong evidence against the avalanche hypothesis.

One should note that in this profile the upper series of hyperbolae are actually double. This configuration suggests that a thin layer of material has recently been deposited by a process which left it uniformly draped over the highs and lows. Apparently the episode which created the corrogated surface has come and then gone.

A dramatic example of another sedimentary episode is shown in Figure IV-4C. in this reflection profile the upper series of uniform and parallel layers has been abruptly cut and truncated since their deposition at a time so recent as to have not allowed new sediment to cover the erosional scars. The outcropping in this profile could have been produced during a period of intense scour and erosion from the sole of a submarine slump which has slid further downslope. Either interpretation argues for a recent change in bottom regime, which has not been masked by contemporary deposition.

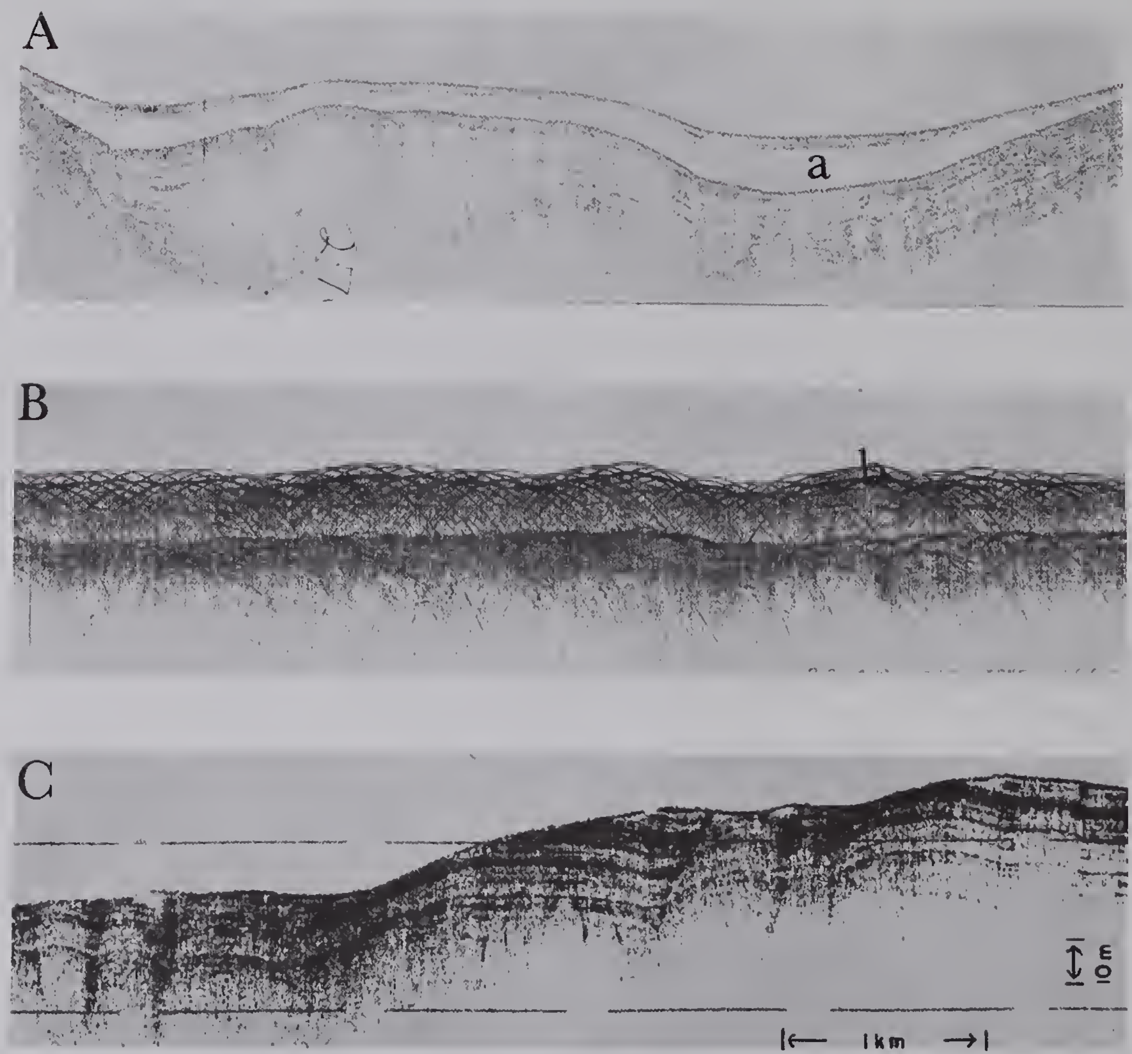


FIGURE IV-4 - Evidence of sedimentary events in reflection profiles

A. A 3.5 KHz reflection profile illustrating differential deposition of a layer of transparent sediment in a broad valley. The thinning of this layer ('a') against the higher relief at the margins suggests that the depositional process was influenced by gravity; i.e. the sediment flowed into place or was drifted into place. Conformable layering at greater depth indicates an abrupt change in depositional process of layer 'a'. RC-12, 1968, Argentine Basin.

B. The hyperbola studded sea floor indicates the presence of regularly spaced linear dunes of probable sedimentary origin, or the chaotic debris of a submarine slump. Thin parallel banding on top of the hyperbola traces,marks a transition to uniform blanket sedimentation (features evenly draped). RC-12, 1968, Argentine Basin.

C. The outcropping of sub-bottom layers indicates a recent episode of intense sea-floor erosion or alternatively marks the source region of a submarine landslide. The outcropping is from a contemporary event, because no recent sedimentation has covered over the exposures. RC-9, 1965, Gulf of Mexico.

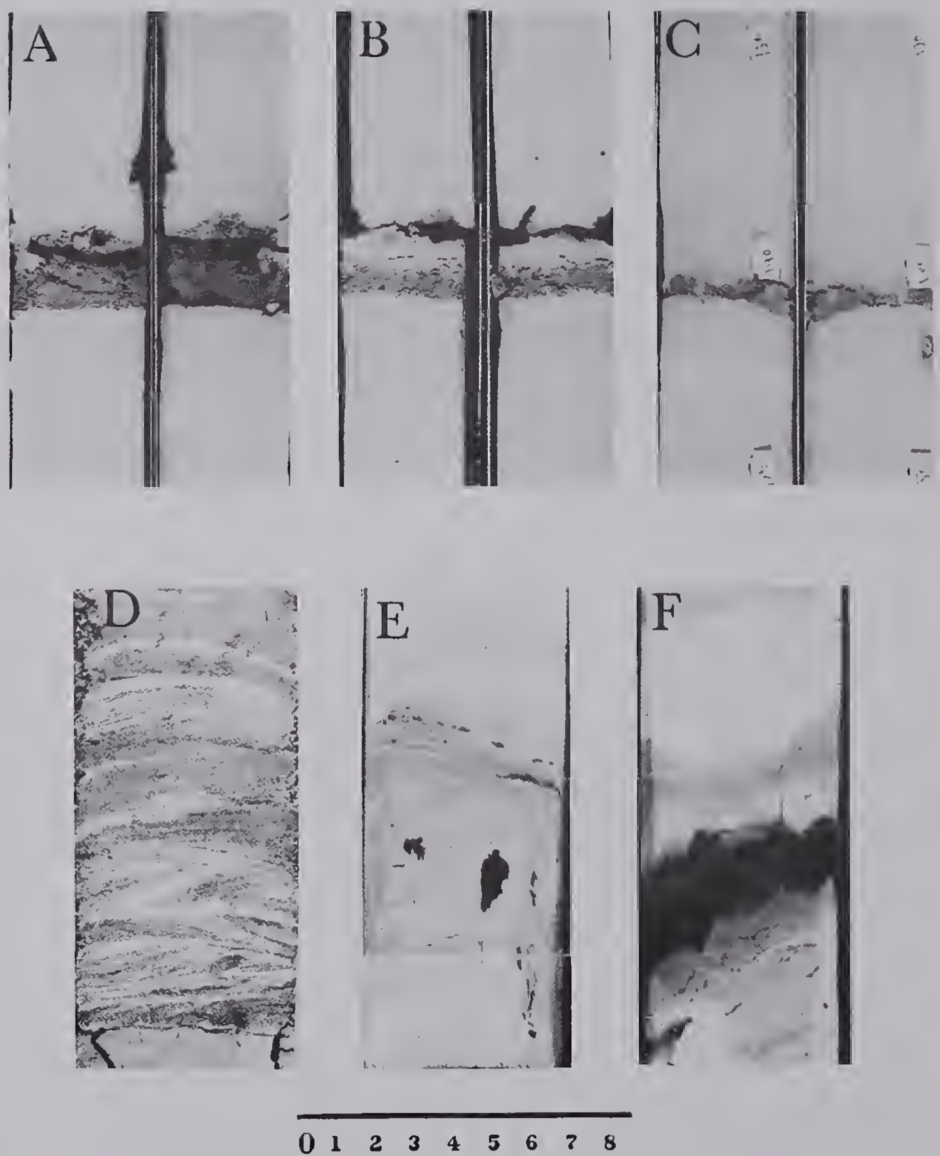


FIGURE IV-5 - Examples of depositional events in cores (Ryan, 1970)

A, B and C. A single layer of volcanic tephra deposited over a distance of several hundred kilometers on the sea bed. Tephra originates as the fine debris of a violent volcanic explosion of a volcano. Note the thinning of the layer and decrease in grain size. Core A is closest to the source, core C is furthest from the source.

D. The typical graded turbidite unit is representative of a single depositional event lasting only a few hours.

E. The micro-breccia is the depositional product of a submarine avalanche.

F. Pene-contemporaneous faulting in soft unconsolidated sediments can result from internal instabilities during compaction or by tectonic deformation. Scale is in centimeters.

4.7-2 Depositional events as revealed in deep-sea cores

Eposidic depositional events can also be inferred from the study of abrupt changes in the lithology or texture of layered sediment recovered in cores. Fig. IV-5A, B, and C show a thin layer (2 centimeters thick) of volcanic tephra in three cores taken several hundred kilometers from each other. This thin horizon of tephra is the product of a huge explosive volcanic eruption, and the entire layer may have been deposited in a period of a few days.

Other episodic depositional events in cores are represented by the typical turbidite unit (Fig. I -5D), a slump structure in the form of a micro-breccia, (Fig. IV-5E), and discontinuities along pene-contemporaneous faults (Fig. Iv-5F).

The periodicity of depositional events can be determined if one can measure the length of time represented in the sequence of layering in reflection profiles or in the column of sediments recovered in a core. In areas of the open ocean, pelagic sediments accumulate at rates varying from a few millimeters per 1000 years to several centimeters in 1000 years. Thus in the reflection profiles of Fig IV-4, a sub-bottom penetration of 20 to 80 meters might cover a span in time of possibly 20 thousand to more than 40 million years. This is certainly a significant unit of time in which to evaluate statistically the re-occurence of sedimentary events for the purpose of determining a periodicity of these events. Such a periodicity is an important consideration to be weighed in predicting further events (the subject of Part V of this report).

4.8 COMMENTS ON WAYS OF INFERRING THE MAGNITUDE OF ANCIENT TURBIDITY CURRENTS

The timing of the sequential breakage of submarine cables downslope one from another has been used to estimate the velocity of submarine flows in historical times. If the sedimentary deposits on the upper surface of the abyssal plains at locations where historical turbidity currents are presumed to have flowed, can be assumed to have been the products of these events, then it might be possible, eventually, to establish relationships between variables of the turbidity current flow such as, current velocity, magnitude, distance from source, and grain size, composition and density of the suspension, with parameters of the deposit such as bedding structures and textures within the turbidite unit. Unfortunately, our comparisons are few, since we have only three historical events with any documentation of current velocity, and no events where the other variables were directly measureable. However, we can point out that on the surface of the Sohm Abyssal Plain there exists a layer of sand and mud of $\approx 10^{11}$ cubic meters in volume which covers an area of 100,000 square kilometers. On the northern part of the plain at a distance of 400 miles from the source, the Grand Banks turbidity current was travelling $\approx$12 knots. The thickness of the resulting turbidite layer in this region exceeds one meter. Within the confines of the Messina Canyon where the turbidity current was also flowing at 12 knots, the thickness of the turbidite layer exceeds three meters.

Despite these observations we do not feel that at the present state of the art, we can realistically infer the velocity of ancient flows from thickness of the turbidite unit. If the areal extent of the flow can be mapped by coring as done in the Tyrrhenian Abyssal Plain (Ryan et al., 1965), Puerto Rico Trench Abyssal Plain (Connolly and Ewing, 1967) or on the Tagus Abyssal Plain (next section of this report), then volumes and thus magnitudes of the sediment fraction of individual past turbidity currents can be calculated. However, unless the density of the turbidity current suspension is known, the magnitude of the actual flow still remains uncertain. Sediment volumes of 10^6 to 10^{11} cubic meters have been calculated for individual turbidite layers on abyssal plains. For an idea of this magnitude, 10^{11} cubic meters of sediment would fill a row of tankers 20 ships wide, running around the equator or would fill Central Park in Manhattan to a depth of 25 km.

Several years ago an attempt was made to infer the actual velocity and density of ancient flows during their passage downslope through the deep-sea channels (Ryan and Heezen, 1964). Recently, this approach has been re-analyzed in a much more thorough and quantitized fashion by Komar (1969).

The method depends on a reasonable assumption that whenever a turbidity current is confined to a channel, its upper surface has a cross-channel slope, owing to the combined effects of the Coriolis and centrifugal forces. This cross-channel surface slope for channel-full flows may cause the difference in the height of the levees that have developed on oposite sides of the channel. An equation can be constructed which essentially balances the Coriolis and centrifugal forces against the pressure force that results from the surface slope. This equation can be used to calculate curves for average velocity versus density, the two variables of the equation and two principal unknowns of turbidity current flow.

Komar has applied the slope equation to measured levee heights on a dozen crossings of the Monterey Deep-Sea Channel along a 100 mile stretch of the channel. In this portion several large bends in the channel provide a good testing ground. Solutions of the slope equation give velocities ranging from 600 to 2000 cm/sec - the same order of magnitude as those of the Grand Banks turbidity currents inferred from the timing of cable breaks. The interested reader is referred to the paper by Komar (1969) because it is one of the first satisfactory attempts to calculate hydrodynamic behaviour of channelized turbidity current flow without resort to empirical channel-flow formulas derived from rivers on land which are not readily applicable to turbidity currents. The shortcomings of the application of the Chezy equation as attempted by Hurley (1964) and Wilde(1965) concerns the evaluation of the drag coefficient, originally derived from empirical data on stream flow and not extendable to the ocean floor environment.

4.9 THE TURBIDITY CURRENT MODEL

An attempt will be made in this section to formulate a model of turbidity current flow. The model is derived from the experience of submarine cable breaks and the analyses of sea floor topography, and is compatible with the composite nature of the eventual depositional facies - the turbidity unit, particularly its lateral distribution and its vertical sequence. The discussion of the model in this report will necessitate many simplifying assumptions and dogmatic statements which would require endless pages of text to qualify and support. We have no doubt that this model will be revised as new data and interpretations become available. Because this report centers on the question of evaluating the hazards of turbidity currents to ocean floor structures, our model contains a conservative "worst possible" approach.

4.9-1 The generation of turbidity currents.

By definition, turbidity currents are self-sustaining gravity-propelled flows of mud and water, and therefore they should flow down any available slope and spread out on a horizontal floor. The flow is apparently initiated when a sizeable mass of loose sediment is thrown into suspension as it cascades downward during sliding or slumping. In fact, the original mass may move first in a slow creep, and only begin to accelerate as it glides over irregular relief and starts to deform and break apart. Apparently in a rather short distance, possibly under an abrupt change to a new thixotropic state, the mass so disintegrates that its material becomes thoroughly suspended, or mixed into a sediment slurry.

Van Andel and Komar (1969) suggest that such a slurry goes through an hydraulic jump at the base of the slope. This jump converts a thin high-speed dense flow, essentially a slump or slide, into a thick low density turbidity current and provides energy for the dispersal of the material. However, we feel that for channelized flows, the dispersal must occur very soon after the initial slumping, well up at the head of the submarine canyon.

For a self-sustaining flow, the dispersed mass of suspended sediment must be of sufficient size and density to overcome all dissipative energy mechanisms. We offer 10^5 cubic meters as a minimum volume for a submarine slump which will generate into a self-sustaining turbidity current capable of travelling down the entire slope to the basin floor. Griggs (1969) has studied the volume of turbidites deposited within the Cascadia Channel during the last 10,000 years, and concluded that the average volume of slumps generating these currents was in excess of 10^7 cubic meters. Many of these young Holocene flows did not fill the channel to the point of overflowing the natural levees. Older Pleistocene flows in the Cascadia Channel (Griggs 1969), Arcadia Channel (Nelson, 1968), Ganges Channel (Ryan, et al., 1967) and Monterey Channel (Komar, 1969) have repeatedly spilled out of the confining levees to strew clastic sediment across the fan, and these flows must have encompassed in excess of 10^8 to 10^9 cubic meters of mud alone.

Additional mechanisms for the dispersal of the avalanche debris into a sediment suspension include the action of breaking internal waves at the upper surface of the flow, internal turbulence due to microtopography on the surface of the sea floor, thermal agitation in the wake of mixing muds once buried at considerable depth beneath the sea floor, and the escape of pore water under confining pressures. Turbidity currents are likely to be highly "cannibalistic" in that they scour up previously-deposited turbidites during their passage down steep slopes and through winding channels. Thus they may grow in volume and mass along the initial part of ther courses.

There is some strong evidence that tsunamis are generated from the abrupt sea floor displacements associated with initial slumping as well as from the movement of the head of the turbidity through canyons and channels.

4.9-2 Channelized flow of turbidity currents

The dynamics of channelized flow of turbidity currents is based on the configuration of the channel cross-section, its downslope gradients, its levee heights, and the distribution of clastic turbidite layer along its floor, on its levees, and across the submarine fan. Velocities of the flow have been discussed in a previous section and range from $\gtrsim$ 45 knots (2240 cm/sec) across the upper fan with gradients of 1:200 to $\approx$10 knots (500 cm/sec) on the distal fan where the gradient is less than 1:1500.

From the studies of the distribution of sediment on deep-sea fans, it can be concluded that, generally, the average flow remains channelized except for a small fraction of the upper sediment cloud which escapes across the levees. The spillage of this part of the flow is greatest in areas of tight meanders. In channels in the northern hemisphere the spillage is predominately over the right hand levee (facing down-slope) as pointed out by Menard (1955), and Hamilton, (1967). The sediment suspension that spills out over the levee, disperses across the fan dropping sediment as the velocity decreases. Menard envisions this deposition to occur from sheet flow. Truchen et al.(1967) and Hamilton (1967), have discovered an undulating surface of regular periodicity and cross-slope orientation on the right hand flank of northwest Pacific submarine fans. The configuration of this surface when examined in high resolution reflection profile resembles large linear sedimentary dunes. These authors have suggested that the undulating relief is a depositional feature from sheet flow during periods of channel spillage, but gullying favored by Hurley (1960), has not been discounted.

The role of sheet flow along the bottom combined with lateral spreading may be the principal factor in the construction and controlled shaping of the submarine fan. Meandering of channels is controlled by the channel dynamics. Regular meander patterns observed along the Indus and Ganges channel systems in the Indian Ocean suggest that the dynamics of channels and rivers may be analogous. One observes active cutting and erosion on the outer bank along bends and dominant deposition on shallower terraces on the inner channel bank (See. Fig. IV-6D).

As many authors have stated, and we concur, the principal process of channel genesis is high speed episodic turbidity currents. That intermittent low velocity tenuous clouds flow through these channels, as favored by Menard (1964), Moore (1966), and Hamilton (1967), we do not deny. However, we believe that the fan is constructed primarily by canyon spillage, and we doubt that the low velocity turbid layers are capable of overridding the confines of the natural levees. If the fan does owe its configuration to processes other than turbidity currents, we doubt that these processes result primarily from gravity-controlled deposition. Instead, we believe the configuration is more likely to be influenced by cross-slope contour following bottom currents, as proposed by Heezen et al., (1966).

4.9-3 Balance between erosion and deposition on the deep sea floor

Channel floors, confined between boundary levees at levels above the base level of the deep sea fan, argue strongly that such channels are dominantly depositional features. In the northern Tufts Plain, for instance, major channels have floors elevated some 20 to 90 m above the general level of the flat areas adjacent to the channels (Hamilton, 1967). Elevation of channel floors has also been observed off California (Buffington, 1952), on the Rhone Cone (by us, unpublished) and on the large Indus and Ganges Cones (Ryan et al., 1967).

Although the levees are primarily depositional bodies, the inner walls of the channels suffer periods of intense erosion and scour during high velocity flows. The erosion is evident in photographs of outcrops of mudstone (Laughton, 1967; Shepard et al.,1969), the recovery of breccias and mud pebbles (Von Rad, 1968), and the truncation of turbidite units (Horn et al., 1970). It is difficult to estimate the extent of erosion, but we would conservatively state that removal of tens of centimeters of sediment per flow is not at all unreasonable

The time of passage for a turbidity current flow past a given point in the channel is a function of the velocity and bulk volume of the flow, and the cross-sectional area of the channel. For channel-full flows at velocities of approximately 20 knots, the time of passage might range from 5 to 30 minutes for a typical cross-section of the Monterey Channel. The range in time corresponds to a flow length of 3 to 30 kilometers.

4.9-4 Arrival of the turbidity current on the abyssal plain and the spreading of the flow across the basin floor

Beyond the end of the channel, the turbidity current can spread laterally and disperse across the basin floor. It seems certain that near the edge of the abyssal plain, many of the channels diverge to form the familiar pattern of a distributary system and that the main part of the flow shifts from one to another distributary as the level of the plain is built up. It is not certain whether a decrease in suspension density is accompanied by an increase in thickness of the flow, or only by its lateral spreading (Hamilton, 1967) as it starts to decelerate and the coarsest particles in suspension begin to drop out. The velocity of the flow at its base may still be sufficient to scour into the sea floor and create flutes, grooves and other sole markings. A certain amount of cannibalism occurs at the proximal margin of the plain, and the lutite stirred up mixes into the overlying suspension.

We estimate that for channel-full flows, velocities on the proximal abyssal plain may reach 10 to 15 knots, and the suspension might be 50 to 100 meters thick with a density in the order of 1.08 to 1.14 gr./cm^3.

We have examined in detail the distribution of turbidite layers across the Tagus Abyssal Plain (off Portugal) to gain some grasp of the dispersal mechanism of turbidity currents once they reach the basin floor. Nine long piston cores on the abyssal plain (see Fig. IV-6) contain a record of over half a dozen turbidity current events, and certain properties of the interbedded pelagic sequences have allowed individual turbidite layers to be correlated from one core to the next. An example of proposed correlations between three cores is shown in Fig. IV-7. From a careful examination of the sedimentary structures and textures of each turbidite unit at the various parts of the abyssal plain cored, the following conclusions can be drawn. Each turbidity current which fills the channel (volume of solids $> 10^8$ cubic meters) spreads itself out over a major portion of the abyssal plain and leaves a sedimentary deposit often covering over 75% or more of the flat basin floor, in an enclosed basin such as this one. The particular facies at one location (proximal or distal) depends (1) on the down-slope gradient from the mouth of the channel to that core site; (2) the absolute depth of the deposition site relative to the deepest portion of the plain; (3) the proximity to the channel source; (4) the volume of the flow; (5) and the size grain distribution of the initial turbidity current suspension.

Along the proximal margin (cores V4-44, 45, and 46) the leading edge of the flows has scoured away from 15 to 20 cm of the previous sedimentary accumulation before deposition on the basal graded unit of the turbidite. Maximum and mean grain size at the base contact is greatest for the proximal cores, particularly those closest to the channel mouth. Sole markings are evident at the sharp basal contact. The lower graded interval is particularly massive, and is followed by a well-sorted interval of parallel laminations. Incomplete intervals, inverse grading, and truncated beds are common. There often occurs an interval of marked convolute lamination, itself truncated and replaced by a very thin interval

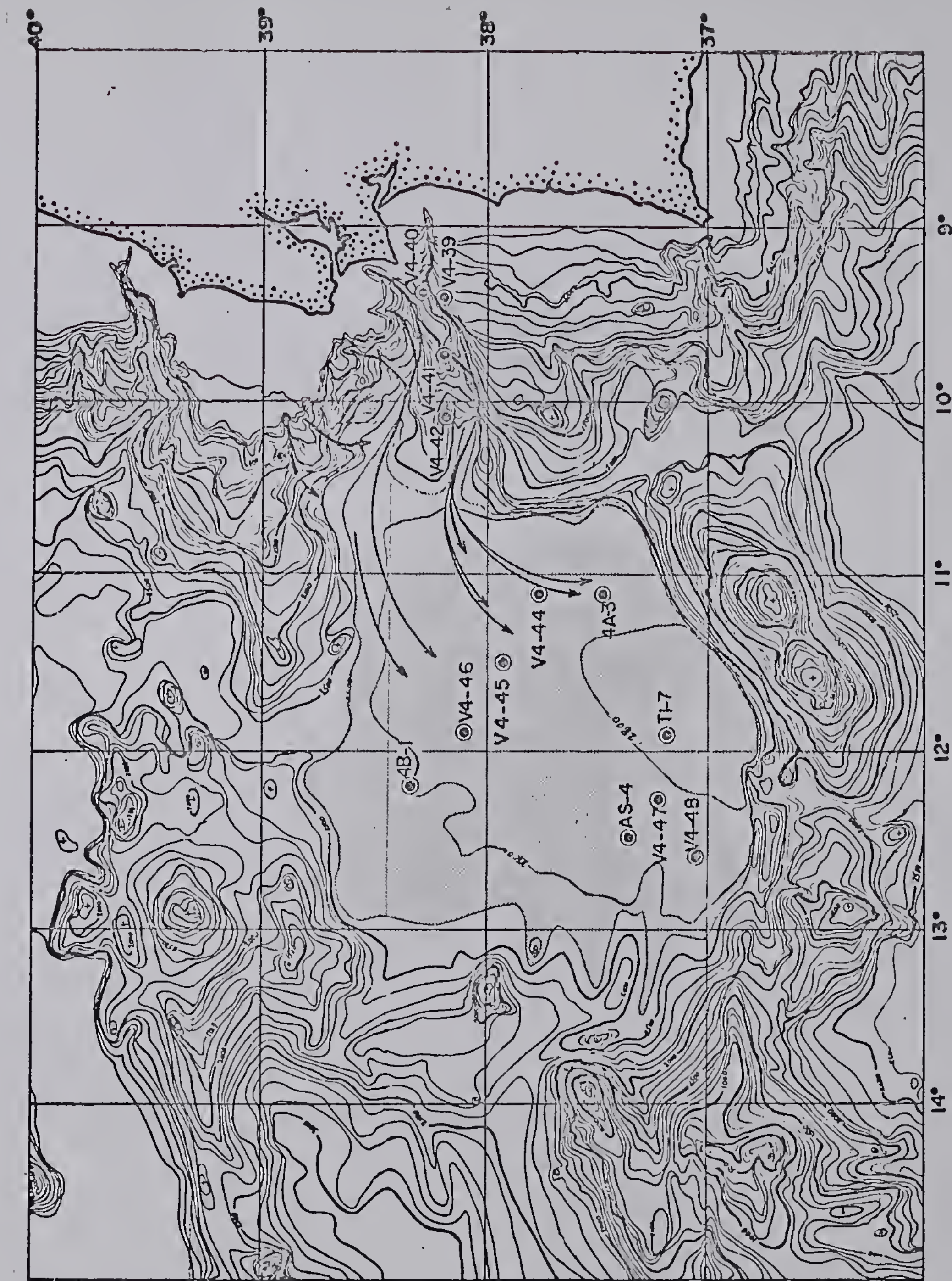


FIGURE IV-6 - The Tagus Abyssal Plain

The shaded area of this map includes the Tagus Fan and Abyssal Plain (adapted from Laughton,1965- preliminary bathymetric map of the eastern North Atlantic). The contours are in fathoms. Arrows indicate the paths of channelized turbidity current flows. The circles indicate the locations of piston cores. The cores from the abyssal plain and from the floors of the canyons contain a sequence of turbidite deposited within the last 100,000 years.

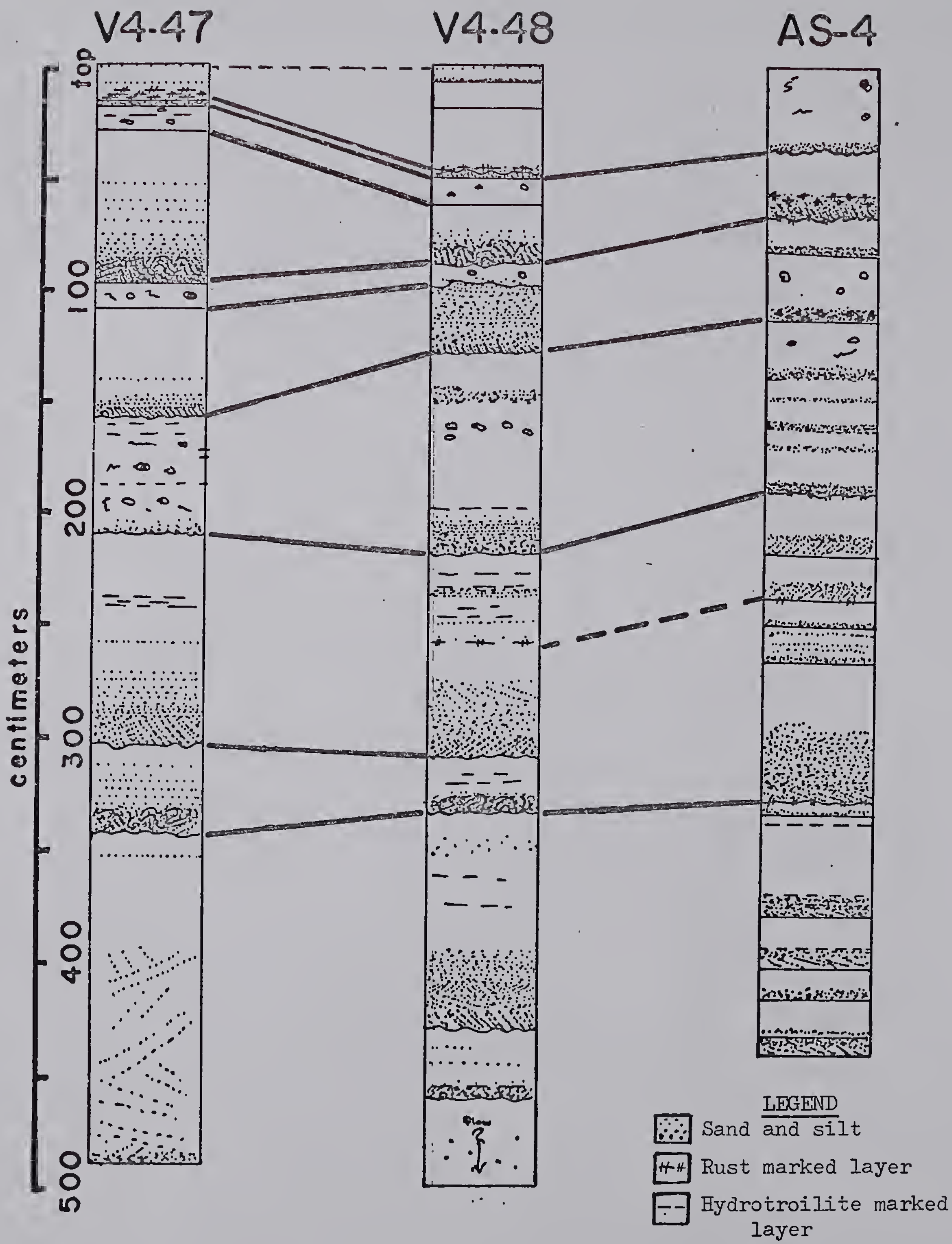


FIGURE IV - 7 - Correlation of individual turbidite layers in cores from the distal part of the Tagus Abyssal Plain.

The graded turbidite units are shown by stippling (note cross-bedding at base of some of the units). Note the sharp irregular (scoured) basal contacts. These three cores are from the distal region of the abyssal plain.

of structureless lutite. The thickness of the entire turbidite unit ranges from 5 to 25 cm. Remarkably, at distal regions of the abyssal such as at core sites Ag-4 and V4-47 and V4-48, Fig. IV-7, the turbidite unit has only a slightly smaller thickness than at sites near the channel mouths. The notable difference in distal turbidites is that 3-5 cm scour at the base is observed, and grading is more complete. On the Tagus Abyssal Plain the average number of turbidites per meter of core is identical, because the lesser total thickness of each turbidite unit is just compensated by the presende of more complete (i.e., thicker) pelagic sequences. At the distal region the grain size decreases, and the sorting improves in the basal unit. The greater thickness is made up by the parallel laminated interval, capped by a relatively thick pelitic interval.

In the deepest part of the abyssal plain, at core site TI-7, the basal graded interval of coarse material is only 1 cm thick. A structureless bed of pelitic sediments, heavily burrowed, makes up the remainder of the turbidite unit. In fact the whole turbidite unit is actually slightly thicker in the distal region than in the proximal region, owing to greater thickness of the pelagic units. No basal winnowing is observed. Burrows in the upper pelitic interval of turbidite units generally do not contain material carried down from overlying pelagic sediments. The numerous burrows indicate a fairly short time interval (tens to hundreds of years) of benthic activity which feasts on rich organic materials and nutrients brought down in the flow. When this material is consumed, the benthic population dwindles because it has no basis of support in the normal pelagic sediments. The burrowing of the pelitic interval down into parts of the graded interval and the absence of material from the graded interval carried up into the overlying pelagic interval is one of the principal arguments in support of very abrupt deposition of both the graded and the pelitic intervals, " as in one casting" (Heezen, et al., 1955; Norin, 1958).

The levelling process of abyssal plain apparently results from the finest grained sediments being able to make their way across the entire basin to reach the lowest depression and pond in it. Some of the fine sediment is cannibalized material of the proximal plain.

The need for a mechanism which can keep the fine-grained mud (4 microns mean diameter) in suspension so that it can flow across the basin has been a serious dilemma for submarine geologists. The thinning of the coarse basal portion of the turbidite and decrease in grain size at the base away from the source is quite satisfactorily explained by assuming that the velocity and thus capacity for suspension also decreases with distance from source. Conversely, the grading with its accompanying upward sequence of structure is explained with a model in which from nose to tail the velocity of the turbidity current gradually slackens. Thus, the deposition at any given point on the sea floor happens from a suspension current flowing above it at a continuously decreasing velocity. However, this implies that the length of the turbidity current increases as it crosses the sea floor, presumably at the expense of its thickness and density. Kuenen(1967) points out that if the nose travels at an average speed of 5 m/sec over a distance of 200 km and a part of the tail at 0.3 m/sec., the nose will take half a day, but the tail will need a week to complete the journey. Hence, at the far end of the plain, deposition will continue for a week.

But, with arguments discussed in Part III, we see that certain structures of the beds, such as convolute laminations and climbing ripples, indicate deposition up to a level well above the rippled horizon, in a matter of hours only. Another problem arises in that Kuenen's tank studies confirm that the fine-grained mud drops out of suspension at velocities less than 10 cm/sec, and thus this mud should not be capable of making its way across the entire plain by simple self-propelled gravity transport in suspension.

One way out of the dilemma is to postulate that the pelitic interval is all composed of redeposited mud as in the analogy of a car driving across the desert. The cloud of dust that follows the car and covers its tracks does not come itself across the desert, but is merely stirred up by the car (Kuenen, 1967). This analogy is, of course, seriously overstated.

The thickness of the pelitic interval on the Tagus Abyssal Plain refutes this proposal because the pelitic interval increases in thickness away from the source at the same time that basal erosion decreases away from the source and becomes entirely nonexistent in the distal plain. One has to accept that the fine-grained sediment is actually transported across the plain to the deepest depression more readily than the high-velocity nose of the original current. Bagnold's (1962), concept of auto-suspension as the driving energetics of the system might remove this dilemma. Physical insight into the auto-suspension mechanism shows that in this type of flow, the current is self-sustaining regardless of the grain-size make up of the suspension, provided that the bottom gradient along the path provides a lowering of the suspension in the direction of flow, more rapidly than the grains can settle out to the bottom by themselves (under Stokes' conditions of settling). The auto-suspension concept merely places a lower limit on the velocity of the turbidity current for any given bottom slope and settling velocity of grains comprising the flow. The authors of this report generally support the auto-suspension concept, but feel it needs further testing in model studies.

We also offer an additional mechanism for the efficient transport of the fine-grained sediment in suspension over long distances. The rapid channelized flow to the basin from shallow regions not only brings in water and sediment at initial high temperature when compared to bottom water temperature, but further compresses the entire sediment suspension, causing an adiabatic warming during descent. A fall of 5,000 m may increase the temperature of the mass in excess of 1°C. Furthermore, the sediment and water may have originally been several degrees warmer at the source region than in the abyss. If, during flow, the turbidity current mass retains its identity, the flow would have a definite heat anomaly which would be compensated by inclusion of water from the masses traversed. Thus as soon as the coarse sands and silt settle from the flow as it crosses the plain, a buoyant force is given to the remaining suspension simply by its thermal anomaly, which is probably compensated by addition of more water from the adjacent mass. A $1^{o}C$ anomaly can balance a suspension density increase of more than 100 mm/l. We offer a suggestion that such tendency toward temporary buoyancy promotes turbulent mixing which assists the lutite to remain fluid and seek out and pond within the lowest depression.

4.9-5 A summary of the turbidity model

In brief summary we review the basin turbidity current model in terms of current velocity, capacity for erosion, eventual sediment thickness, and time period of the total episode

(1) Near the origin, the initial avalanche may be rapidly transformed into a very dense suspension (McCrone 1966) and travel at speeds similar to nuées ardentes (2500 cm/sec). Within the confines of a channel the turbidity current suspension might continue at velocities of 2,000 cm/sec and debouch on the plain at 500 cm/sec. There its velocity will decrease rapidly and it may cross the plain at 100 cm/sec and eventually pond at < 10 cm/sec.

(2) At the site of the avalanche,the depth of erosion may exceed tens of meters over distances of many km. Within the channel erosion probably occurs to a depth of less than one meter. On the proximal abyssal plain, scour and flutes might remove 10 cm of sub-adjacent soft sea floor; on the distal plain, basal scour is virtually non existent. We suggest that where current velocities are below a limit of about 20 cm/sec, controlled deposition will dominate over erosion.

(3) Sediment thickness per flow is highly variable. Flows are thickest on the distal plain, and thinnest in the channels and on the fan. The differential growth of the fan is controlled by the numerous small flows that never reach the plain and by the overriding of the levees during really large flows.

(4) The time duration of the initial slump may be only a few minutes, or fraction of a minute. For instance, during the Grand Banks slump, the cables which crossed the slump area all failed simultaneously even though they covered a region more than 20 km wide. During passage through the canyon the flow may pass a given point for a period of several minutes to more than half an hour. On the proximal plain the flow may last ten hours or more; on the distal plain, many days or weeks.

(5) In many abyssal plain areas there is a water mass at the bottom in which an adiabatic thermal gradient exists beneath a thermal minimum, and the stability of the water column is neutral. This situation facilitates inclusion of water from the mass being traversed and production of vertical turbulence which keeps the fine fraction of sediment in suspension.

Much additional _in situ_ data is needed to answer these very difficult questions on the hydrologic realm of turbidity current flow and deposition. Such data are particularly important for analysis of the flow regime, for the purpose of designing structures for the sea floor, and for predicting the intensity and magnitude of phenomena which these structures might experience.

PART V

PREDICTION AND DETECTION OF TURBIDITY CURRENTS

5.1 PREDICTION OF TURBIDITY CURRENTS

Prediction of turbidity current occurrence at a given site is, in general, very difficult. The problem has many features in common with that of earthquake prediction, in that the record covers too short an interval of time to provide a statistical basis for estimating probabilities, - except perhaps in the most active regions. Yet even this short record provides examples of isolated violent earthquakes, such as that at Charleston in 1886, in regions that otherwise would be rated as exceptionally safe. In the case of turbidity current generation, there is an added degree of uncertainty because a violent shock in a part of the continental slope that has been undisturbed and accumulating sediment for a very long time has a high probability of generating sediment avalanches and turbidity currents. The Grand Banks turbidity current of 1929 is an example.

Since underwater avalanches and turbidity currents are episodic events, the choice of an optimum site for a structure which could be affected by these processes must depend entirely on the evaluation of the likelihood of occurrence of such an episodic event and on any feature of the site that might determine the ability of the structure to resist. There are two approaches which can be taken. First, one can try to choose a site in certain physiographic provinces on the sea floor which are totally inaccessible to avalanches and bottom currents and where we know from oceanographic exploration that there has never been such activity during the whole of geologic time. Such areas would be called geologically "<u>safe areas</u>", and they can be mapped throughout the world's oceans. In fact, they may encompass large areas of the ocean floor. However, geologically safe areas may also exclude vast regions of specific provinces where it is highly desirable to locate certain structures, and thus it may be necessary to choose sites in what may be called areas accessible to avalanches and/or turbidity currents.

Therefore, the second approach is to evaluate sites in "<u>accessible areas</u>" where, geological intuition suggests that for certain predetermined reasons, avalanches or turbidity currents are highly unlikely to occur during the time periods in which the structure is emplanted and in use, even though they may have occurred there in the past. As a corollary to the second approach, it may also be possible to predict sites which are accessible to avalanches and turbidity currents that may possibly be active at some time during the lifetime of a structure, but which geological intuition suggests that the size, intensity, and duration of such events would not significantly affect the usefulness or operation of the sea floor structure. Of course, such an evaluation would be highly dependent upon the engineering characteristics of each individual type of structure.

5.2 PREDICTING PLACES OF OCCURRENCE

In the first approach of selecting geologically safe sites, it is only necessary to consider areas (local or regional) where avalanches and/or turbidity currents could occur and then to try to avoid them entirely. Furthermore, during local pre-site reconnaissance surveys it would be desirable to take long cores in areas selected as safe and try to establish unequivocally a lack of stratigraphic discontinuities for the extent of geologic time recorded in the sedimentary section. For instance, a 20 m core from the abyssal hills regions of the central Pacific may sample a record of continuously-deposited sediments over ten million years. If there were no breaks in the stratigraphic successions of layers which could be attributed to slumps, scour, winnowing, erosion, or episodic deposition, then the likelihood for their occurrence during the next 100 years is extremely small. This conclusion is based on a familiar geologic axiom that "the past is the key to the present".

Regions of the sea floor accessible to avalanches are catalogued in Tables 5-1 and 5-2. The ocean floor is broken down into basic physiographic provinces (after Heezen et al., 1959) and into sub-units of much smaller areal extent. Selected examples of documented rock avalanches, mud flows, gravity gliding, slumping, etc. are cited. In Part VI, a full description of the environmental setting of each physiographic province will be given.

5.3 PREDICTING FREQUENCY OF OCCURRENCE

For avalanches to occur, there must be not only a slope down which the material flows, but also a potential unstable source of sediment. Our knowledge of the extent of deposits of individual turbidity currents (Heezen et al., 1954; Ryan et al., 1965; Conolly and Ewing, 1967; Griggs et al., 1969), indicates that an avalanche of coarse sediment must be of a certain critical size before it can generate a self-sustaining gravity-driven flow. Attempts to artificially trigger self-sustaining turbidity currents on steep slopes (Buffington, 1961) or witness natural occurrences in canyons by free diving or with submersibles have failed because the artificially-induced avalanches have been too small. The inherent critical mass for a self-sustaining flow might well imply that any witness who happened to observe such an event would probably be destroyed by it!

Potential source terrains for avalanches are analagous to reservoirs which must contain sufficient loose sediment such that instability, no matter how it is induced, will result in a gravity-driven, self-sustained flow. We feel that information about such sediment reservoirs is very important in predicting where and when turbidity currents might be generated.

TABLE 5 - 1

AREAS OF THE SEA FLOOR WHERE AVALANCHES OF MAINLY ROCKY MATERIAL HAVE OCCURRED

	ACCESSIBLE AREAS	AREAS OF DOCUMENTED AVALANCHES
CONTINENTAL MARGIN PROVINCE	Walls of submarine canyons particularly those cut into crystalline rocks	Tagus Submarine Canyons, La Jolla Canyon, French Riviera Canyons, Naples Canyon
	Steep continental slopes carpeted with ice rafted debris	Slope south of Grand Banks
	Seaward escarpments of marginal plateaus	Blake Escarpment, Exmouth Escarpment
	Boundary escarpments of submarine banks and coral atolls	Grand Bahama Platform
	The inner wall of seismically active deep sea trenches	Tonga Trench, Calabra and Hellenic Island Arcs, Puerto Rico Trench, Peru-Chile Trench
OCEAN BASIN FLOOR PROVINCE	Walls of channels cut through indurated sediments or basement in abyssal gaps	Theta Gap, Conrad Channel
	Flanks of seamounts and guyots	Grand Meteor Bank
	Aseismic escarpments of oceanic rises and plateaus	90°East Ridge, The Eastern Scarp of Bermuda Rise
MID OCEANIC RIDGE PROVINCE	Steep slopes in rift valley, rift mountains and high fractured plateau	Rift Valley of Mid Ocean Ridge in the North Atlantic, Red sea,Carlsberg Ridge in the Indian Ocean
	Precipitous escarpments of fracture zones	Romanche Fracture Zone, Vema Fracture Zone, Atlantis Fracture Zone, Kane Fracture Zone

TABLE 5 - 2

AREAS OF THE SEA FLOOR WHERE DOWN SLOPE CREEP AND RAPID SLUMPS OF UNCONSOLIDATED SEDIMENTS HAVE OCCURRED

	ACCESSIBLE AREAS	EXAMPLES OF DOCUMENTED MUD SLIDES
CONTINENTAL MARGIN PROVINCE	The continental slope	Continental slope south of the Grand Banks, off Libya, South of Portugal
	Submarine canyons	La Jolla Canyon, Messina Canyon
	Sedimentary aprons such as cones and fans	La Jolla Fan Valley, Rhone Cone, Cascadia Channel, Hudson Apron
	Continental Rise	Continental Rise off Cape Hatteras
	Thick Sedimentary ridge	Blake Bahama Outer Ridge, Central Argentine Basin
OCEAN BASIN FLOOR PROVINCE	Tectonically deformed sedimentary ridges	Mediterranean Ridge, Barbados Ridge
	The proximal abyssal plain boundary near marginal escarpments	Sigsbee Abyssal Plain (near Sigsbee and Campeche Escarpments)
	Flanks of seamounts and atolls	Caryn Peak, Hog Hill
	Abyssal gaps	Theta Gap, Vema Gap, North-west and mid Ocean Canyons
MID OCEANIC RIDGE PROVINCE	Floors and walls of intermon-taine basins	Basins of the Mid Atlantic Ridge
	Floors and walls of fracture zones	Romanche Fracture Zone
	Flanks of rift mountains	Intermontaine basins of the equatorial Mid Atlantic Ridge

For example, today certain rivers, such as the Magdalena in Colombia, South America or the Congo in Africa, debouche directly into heads of submarine canyons. These rivers do not have large deltas on the continental shelf where the products of their discharge settle out. Consequently, the sediment bed load of these rivers and much of the suspended load too, is deposited directly into the heads of the submarine canyons. After sufficient time (which is sometimes only a few years) a large mass of material will accumulate in this natural reservoir. Furthermore, this reservoir is located in a precarious setting within the confines of a steep canyon.

Consequently, any disturbance of the reservoir from earthquakes, mud diapirism, gas discharge or tectonic displacements could initiate a sediment avalance with the potential of developing into a turbidity current. If disturbances occur frequently in relation to the rate of accumulation within the reservoir, avalanches will be small and many of them will not be sufficient to generate self-sustaining flows. The flowage of sand observed by Dill (1964) in the head of Scripps Canyon is one way of depleting a potential resevoir and thus reducing the chances of large scale turbidity currents.

In the case of the Magdalena and Congo Rivers, the annual discharge of sediment from the river is so great that the sediment reservoirs are intermittently emptied - apparently almost annually - by slumps of sufficient magnitude to initiate self-sustaining turbidity currents. There exists a detailed history of turbidity current activity here from the records of cable failures at points where submarine cables cross the canyon axis downslope from the mouth of the river (Heezen, 1956; Heezen, et al., 1964) (see Fig. V-1). It would involve great risk to place structures on the sea floor in this locale.

On the other hand, the Mississippi River and the Nile River now discharge their sediment load into large lacustrine deltas on the inner continental shelf. However, during maximum sea level-lowering of the last glacial advance the ancient delta was at the seaward edge of the continental shelf, similar to the present Congo River such that the rivers discharged their load onto the upper slope. Cores within channels on the Nile Cone and on the abyssal plain on the floor of the Gulf of Mexico contain many turbidite layers deposited during the period of lowered sea level. Conversely, these same cores contain no turbidite layers younger than 12,000 years, when sea level began to rise to its present level and the deltas migrated landward. With confidence, one could predict that today on the Mississippi Cone and Nile Cone there is little potential of turbidity current activity.

There are other suppliers to sediment reservoirs in addition to rivers. Off the coast of California long-shore currents (Dill, 1964; Moore, 1966) move sediment laterally, for considerable distances to canyon heads in which it is trapped. Gravity flowage of sand, creep, and tidal currents are processes which can carry some of this sediment downslope in the canyon heads. Where large accumulations fail abruptly by slumping, either in the canyon head or at some point of concentration further down canyon, a turbidity current may be initiated. Thus long shore shelf processes of sediment transport are important factors in the filling of sediment reservoirs.

A

B

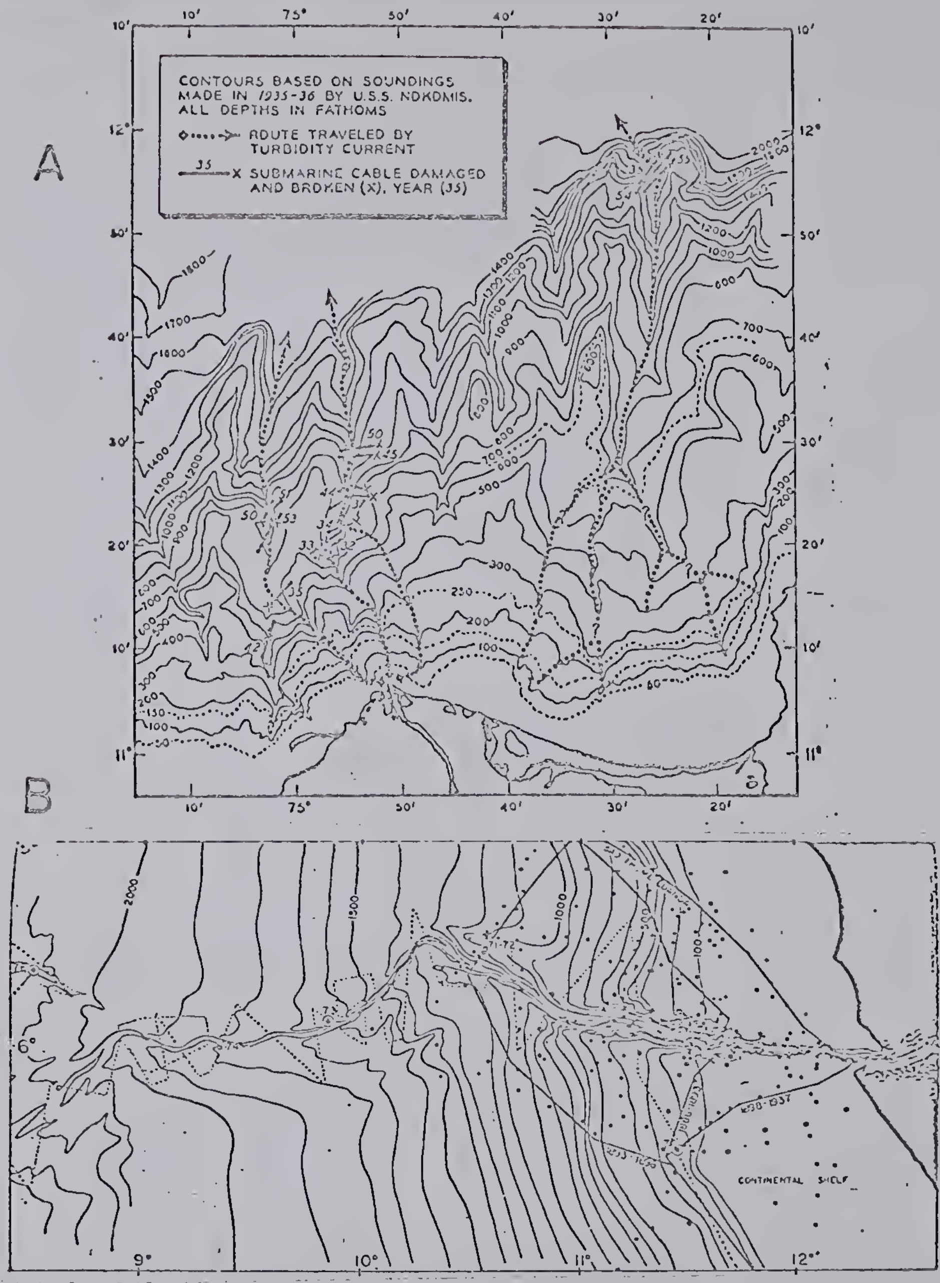


FIGURE V-1 - Example of regions of contemporary turbidity current activity

A. The discharge of the Magdalena River feeds sediments downslope directly into the canyon heads. Cable breaks indicated on the diagram were along a single submarine cable. The failures were caused by turbidity currents resulting from extensive slumps of the Magdalena river bar, (from Heezen, 1963).

B. Similarly the Congo River debouches directly into a major submarine canyon cut across the continental shelf. Submarine cables across this canyon have failed repeatedly. Both the Congo and the Magdalena have a common characteristic in their lack of a modern sub-aerial delta, (from Heezen, Menzies, Schneider, Ewing, and Granelli, 1964)

The study of shelf processes, the exploration of sediment fill in canyon heads, and the examination of intervals of time in the past record of turbidity current deposition in long cores, are important ways of predicting the likelihood of contemporary activity in regions of potential activity (i.e., in non-safe areas). Our feeling is that accumulation in sediment reservoirs is equally as important as the eventual triggering mechanisms. For instance, in places of high seismicity on the inner walls of trenches, in the rift valley of the mid-oceanic ridges, and in fracture zones - the sediment may be shaken loose from steep slopes so often that critical masses will never accumulate. In these regions, small avalanches are probably common, but they may never contain enough loose sediment to initiate self-sustaining turbidity currents. In fact, cores from the floors of fracture zones in offsets of the Mid-Oceanic Ridge contain turbidites deposited at intervals separated by as long as tens of thousands of years of normal pelagic sedimentation. From the record of seismicity, we can conclude that the escarpments and floors of these regions of the fracture zone have experienced many shocks in that time interval. In fact, some of the major fracture zones of the Equatorial Atlantic are filling much more rapidly from their ends with sediment transported by turbidity currents from river deltas such as the Amazon, than from sediment shaken off the fracture zone walls.

5.4 THE DETECTION OF TURBIDITY CURRENTS

Ancient turbidity currents have left their records as graded deposits in cores and, as buried channels by which their paths of flow may be mapped with sub-bottom echo soundings, in some cases supported by additional sampling. Individual turbidity currents may carry sediments of unique mineralogy or those containing certain tracers (e.g., rare earth metals, unique fossils or some identifiable lithologic property) which can help in mapping the entire flow from source to final resting place. Preliminary attempts have been made by Heezen et al.(1954); Ryan et al.(1965); Schneider and Heezen (1966); and Conolly and Ewing (1967), to map past flows.

The detection of contemporary or future turbidity currents is a problem which has received very little attention. We suggest some possibilities.

As described in Section IV, the classical method of detecting turbidity current activity has been from the study of sequential, down-slope failure of submarine cables. The problem that has arisen in confirming this activity is that there are only very limited ways of proving that a graded deposit of sediment on the surface of a submarine fan or on an abyssal plain in the region of the cable failure, was actually deposited on the day of cable failure and not some 10, 100 or even 1000 years previously.

In the case of the 1908 Messina earthquake and turbidity current (Ryan and Heezen, 1965) the broken ends of the submarine cables were recovered and contained, wedged into them, samples of a pumice-rich sand. This material was of identical mineralogic make-up to a surface layer of graded sand within the canyon, across which the cable broke. Other gravel layers below the surface within cores from these regions contained sands of a different mineralogic composition.

In the case of an earthquake and subsequent cable failure in 1967 in the Solomons Sea, the cable repair ship recovered broken pieces of cable with sand and twigs imbedded into them and the twigs contained green leaves (Bruce C. Heezen, personal communication).

Following the 1964 great Alaskan earthquake, one of our vessels detected a thick cloud of suspended sediments above the floor of the Aleutian Trench by light scattering measurements. Such clouds might be the tenuous suspension of a turbidity current triggered by the quake, from which the suspended sediment had not completely settled. It is not yet known how long fine grained sediment stays in suspension above an abyssal plain, but we would speculate that at least for several days, if not weeks, a detectable layer of increased light scattering would be present.

Another way of monitoring a recent turbidity current event would be by studying the rate of repopulation of the sea floor by bottom dwelling benthic fauna. The upper pelitic interval of the turbidite unit is often extensively burrowed by organisms which feed on organic matter carried into the basin in suspension (Heezen et al., 1955, 1957; Griggs et al., 1969).

If the normal intensity of burrowing activity could be established, then the extent of burrowing or the extent of surface activity as revealed by tracks, trails, and mounds would be an index of the amount of time lapsed since deposition of the youngest turbidite layer in the basin. We would expect that a flow which covers an area of 4000 km^2 might so destroy the benthic community that it would take many years to repopulate this area. Study of the benthic life (Griggs et al., 1969) might permit a resolution of a few tens of years in determining the history of previous events.

A better resolution can be obtained by measuring the detailed thermal gradient in the sea floor in regions where a recent event is suspected of having either eroded the bottom or left a deposit. A thermal measurement made on the Sirte Abyssal Plain in the Mediterranean Sea in 1965 by Lamont-Doherty scientists revealed a marked temperature inversion within a 4 m thick graded sand layer on the surface of the plain. The measured thermal gradient indicated that this layer of sediment was a few tenths of a degree warmer than either the bottom water of the basin or the mud below it. Model studies (Ryan, 1970) explain this apparent anomaly by suggesting that this layer of sand was deposited from a turbidity current approximately 4-6 months prior to the measurement of the thermal gradient. The initial high temperature of the graded layer is conceivably caused by the adiabatic heating of the turbidity current suspension as it compresses during its descent through the submarine canyon to the abyssal plain. Regardless of the mechanism responsible, thermal disturbances are introduced in the areas of erosion and of deposition. These are preserved according to the specific heat and conductivity of marine sediment. They afford a shorter and admittedly crude time scale for estimating the history of depositional events. We feel that this technique might hold some promise for confirming recent rapid deposition from contemporary turbidity currents.

Another method would be to listen with passive sea-floor seismic or acoustical arrays for events of long duration (many hours) whose source of acoustic energy moves across the sea floor for tens or hundreds of miles. More direct methods would employ the use of anchored bottom current meters (Nowroozi et al., 1968), sea-floor tidal gauges (Isaacs et al., 1966), or even simpler indicators that would be affected by strong flow in channels.

5.5 DEVELOPMENT OF PREDICTION AND MONITORING TECHNIQUES

The prediction of turbidity currents must be based on two types of measurements, (1) those which can map potential reservoirs of critical masses of sediments, and (2) those which can evaluate potential triggering mechanisms. For mapping the reservoirs, one can employ new high resolution seismic reflection techniques (Hersey, 1963; Moore, 1966) with low frequency sparker or air gun sources (Ewing and Tirey, 1961) or higher frequency (3.5 or 12 khz) echo-sounders. Besides measuring the thickness and thus volume of sediments in reservoirs, potential instability is revealed in reflection profiles by signs of differential settling during composition and by indications of active movement as inferred by displacements along faults (slump scars) as shown in Fig. V-2. Such displacements are probably caused by upper flowage of entrained interstitial fluids, and particularly in deltaic environments by volumetric expansion caused by organically derived gases (Monroe, 1969). The vast majority of submarine slumps occur in sediment which have been rapidly deposited. The burial of large quantities of organic material such as kelp and eel grass in the heads of submarine canyons upon subsequent decomposition leads to the formation of gases which saturate the interstitial waters and which can even be seen to bubble up from the bottom in shallow marine environments (Dill, 1964).

The growth of mud and shale diapirs is usually directly associated with high pressure gas. Diapirs suspected of being of this type are found sometimes in deep water in areas of high rates of sedimentation such as on the Magdalena and Mississippi Cones, but most commonly they occur within the river delta (Shepard et al., 1968).

Potential sediment reservoirs can be forecast by mapping the rate of lateral transport of sediment across the shelf and estimating the time of filling of the canyon heads. Similarly, the amount of discharge of the bed load and suspended load from major rivers can be monitored and used to calculate the rate of accumulation of sediment in deltas. It would be worthwhile to undertake programs to measure actual rates of accumulation by implanting stakes into the bottom and revisiting these sites annually, or by placing sensitive pressure transducers on the bottom, which can telemeter the weight of the overburden as the sediment buries them.

The second approach involves the recognition of potential triggering mechanisms. Earthquakes are known to have initiated marine slumps that have generated into turbidity currents (Heezen and Ewing, 1952). Maps of seismicity that show the locations of earthquake epicenters are becoming available for all areas of the world,

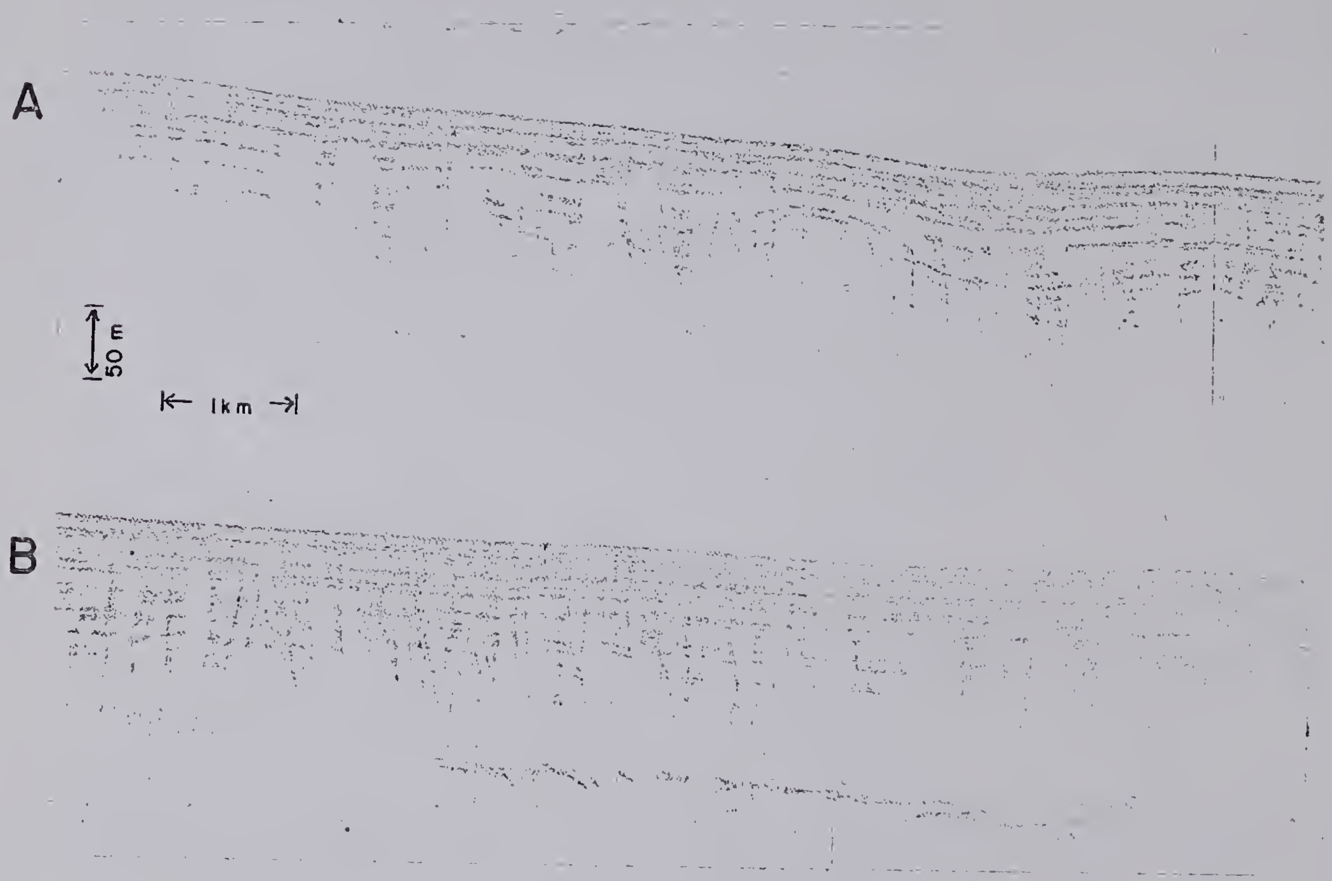


FIGURE V-2 - <u>Contemporary displacements of sedimentary strata</u>

A & B. Examples of layered sediments of the continental rise in the Argentine Basin. These records are 3.5 KHz profiles taken on R/V Conrad, in December 1968. The offsets of the horizontal layering are interpreted as indications of post-depositional displacement. Some of these offsets reach to the sea floor and indicate that this activity is contemporary. The movement of this sediment is aided by the release of entrained pore water, initially trapped in rapidly deposited high porosity sediments. Such potentially unstable sediment bodies may be sources for future turbidity currents.

and their completeness and reliability are becoming better as each year passes. Ocean bottom seismographs can be used to study local seismicity, and in particular, arrays of these instruments can detect the actual motions of the sea bed (Francis, 1968). We believe that certain types of earthquakes accompany displacements on the sea floor where one crustal layer overthrusts another. Such displacement may be more effective in initiating large slumps, than those of purely dilational collapse or horizontal strike slip.

In other words, a shove against a sediment pile is an effective way of creating the original horizontal acceleration to set in motion the eventual avalanche. Dill (1964) finds that near earthquakes in the heads of canyons generally produce no apparent surface displacements on faults, but merely shake the sediment, causing it to settle and compact, releasing some of the entrained gas. These earthquakes apparently do not result in down-slope flow of the canyon sediments.

Studies of the first motions of earthquakes, the depth of the sources of seismic energy, and the magnitude of energy released, coupled with knowledge of the precise location of the previously active faults on the sea bed would aid in predicting whether future seismic activity might trigger submarine landslides. It might be particularly useful to carry out advance geodetic surveying in regions of known seismic activity along sea bed faults (such as within fracture zones or on the inner wall of deep-sea trenches) and obtain first hand experience about the magnitude of earth movements and the susceptibility of the sea bed to slumping.

In this regard it might also be advantageous to use vehicles which can be located at various places of interest on the ocean floor to monitor this environment before structures are designed for these locations. The vehicles could be equipped with tiltmeters, seismometers (long and short period), current meters, light-scattering meters, gas analyzers, and sediment traps) and could be retrievable after months or years of service. Picking a site on the sea-bed might in many ways be analogous to choosing an optimum site on land, and despite all the empirical laws of civil engineering, the first hand experience of individuals who have lived on such a site can be most valuable, particularly when assessing episodic phenomena. It would be desirable to start gaining experience about the sea bed in advance of actually placing structures in these environments. Furthermore, we might scan data from seismic arrays in the sea bed which have been emplaced already for many years and look for intermittent unexplained acoustic events which might be attributed to rock falls or sustained turbulence near the ocean floor.

5.6 RECOMMENDATIONS FOR IMMEDIATE USE

In order to learn more details of turbidity current flow, it would be very helpful to witness first-hand an actual event, monitor its course with tracers, measure its velocity with meters, and map its deposit by coring. Certain low-velocity flows have been detected in reservoirs and lakes (Forel, 1885; Grover and Howard, 1938). In fact, lakes such as Lake Leman in Switzerland offer great potential for a further understanding of the hydrologic realm of turbidity currents. Sedimentation in lakes has been studied from cores and seismic reflection profiles (Houbolt and Jonker, 1968; Serruya et al., 1967). Off of the mouth of the Rhone River a sublacustrine channel with natural levees leads down to a depth of about 200 meters, where the levees disappear. The channel continues to a depth of 280 meters where it debouches onto a sublacustrine fan which extends further to a central abyssal plain with a maximum depth of 310 meters. Within this miniature environment there are direct analogues to the deep ocean. We would encourage a full scale assault on the sedimentary processes of such environments as a first step procedure to understanding dynamical sedimentation in channels, fans, and on abyssal plains. Such a program should include provisions to monitor in situ current velocities and densities within the channel and across the fan and plain over long periods of time. It is known that turbidity currents occur rather frequently in Lake Leman and thus the chance of getting data quickly is expedited by choosing lakes instead of starting in the deep sea.

Another program deserving immediate attention would be to artificially trigger a large avalanche within a submarine canyon, deep-sea channel, submarine fan, and abyssal plain sedimentation system using explosives, and then monitor the downslope flow of the sediment suspension with an array of instruments planted on the sea bed. The first witnessing of such an event would offer a valuable test to our inferences as discussed in Part III of this report.

Furthermore, the artificial triggering of avalalanches could be used as it is at ski resorts to free slopes of sediment before the implantation of structures. In the deep-sea environment, the sediment accumulation is so slow that this procedure would only have to be done once. In the process of creating artificial avalanches, valuable empirical data would become available on the susceptibility of the sediment carpet to failure.

PART VI

DISTRIBUTION OF TURBIDITY CURRENT DEPOSITS IN THE OCEAN

6.1 TYPE PROVINCES IN WHICH TURBIDITY CURRENT DEPOSITION MAY OCCUR

Depositional processes and particular physiographic provinces are closely related. To illustrate this a few studies of depositional environments will be examined in detail. In particular, we shall discuss the deep-sea distributary system which includes the submarine canyon on the continental slope, the deep-sea channel on the continental rise, the submarine fan, and the abyssal plain. Since turbidity currents are gravity-propelled flows, their downslope travel carries sediment from the continent to the deep-sea floor. The energy characteristics of individual turbidity currents depend not only on the magnitude of the initial slump, but also, even more importantly, on the available downslope routes for the sediment transport.

Clastic sediments derived from the continent may have a considerable residence time on the continental margin before their ultimate deposition on the ocean floor. The continental margin includes those provinces associated with the transition from continent to ocean floor. Many margins, particularly in the Atlantic and Indian Oceans, are made up of a continental shelf, continental slope, and continental rise. Generalized physiographic province maps of the Atlantic, Pacific and Indian Oceans, and of the Mediterranean Sea are shown in Figs. VI-2,3,4,and 5. Fig. VI-1 is an index for these maps and identifies the various provinces shown.

6.1-1 Continental shelf province

The continental shelf is the platform extending off the continent to a depth of about 200 m, where there is often a marked change in slope, known as the shelf break, beyond which lies the steeper continental slope. Though by no means uniform, the average slope of the continental shelf is only 0°.07' and is slightly steeper along the inshore half of the shelf. Flat plains are rare. Slight irregularities in relief are common and are the result of many different processes, many of the processes being types well represented in continental topography.

For the purpose of this study, we will concern ourselves with three categories of irregularities, - erosional channels, troughs, and linear ridges of regular wavelength

Beyond the mouths of a few major rivers, small erosional channels occur, extending seaward.

The Hudson Channel (not canyon) on the east coast of the United States is an example of such a channel. It most likely represents a drowned extension of the Hudson River, which during periods of lowered sea-level was subaerial. Such drowned extensions of rivers are important indicators of ancient transport routes of detritus across the shelf. Their terminations often mark the sites of major sediment reservoirs which, in turn, are potential sites for initiation of large scale slumps. On many shelves, the drowned valleys are discontinuous, parts of them having been buried by more recent longshore sediment dispersal. Burial attests to the fact that they are non-functional relics.

Troughs, sometimes hundreds of meters deep, occur most commonly at high latitudes (i.e. Labrador, Alaska, Norway, Antarctica). Such troughs are believed to have been gouged into the shelf by the flow of glaciers, and are significant because their courses also mark the transport routes of large volumes of sedimentary material carried in the ice. The outer terminals of these troughs may be likened to terminal moraines, and most certainly mark the location of massive seidmentary reservoirs. In fact, the Grand Banks turbidity current originiated in a series of slumps at a site just seaward of the glaciated trough known as the Laurentian Channel.

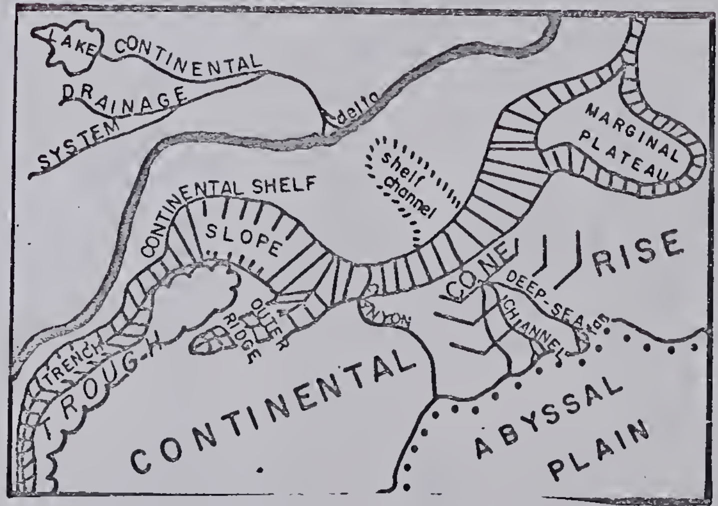
Therefore, even though such channels offer no contemporary hazard to underwater structures situated along their courses (since they are presently inactive as transport routes), the associated terminal regions near the shelf break are the locations of potential downslope flows by catastrophic failure. There are many mechanisms that can induce slumping. In the environments just discussed, natural erosional phenomena associated with the glacial epochs may provide examples of local over-steepening of the upper continental slope by excessive sediment concentrations in very restricted zones. Since slumping is particularly common at the heads of submarine canyons near the mouths of major rivers, such as the Magdalena, slumping would be likely to occur in the vicinity of these relic river mouths.

The third type of irregularity on the continental shelf significant to this study includes the whole class of linear ridges of regular wavelength. These are associated with the migration of coarse-grained sediments by longshore currents in traction carpets.

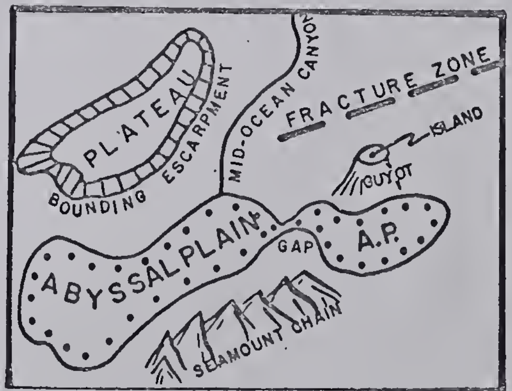
Sand waves with amplitudes of several meters have been noted in the region of the Grand Banks, in the English Channel and North Sea, and on many other shallow shelves. The presence of such sedimentary structures attests to periods of active longshore sediment transport which may effectively trap and thus concentrate unconsolidated materials

GENERAL PROVINCE MAPS OF THE ATLANTIC & PACIFIC OCEANS & MEDITERRANEAN SEA: EXPLANATION - MAPS 1,2,3

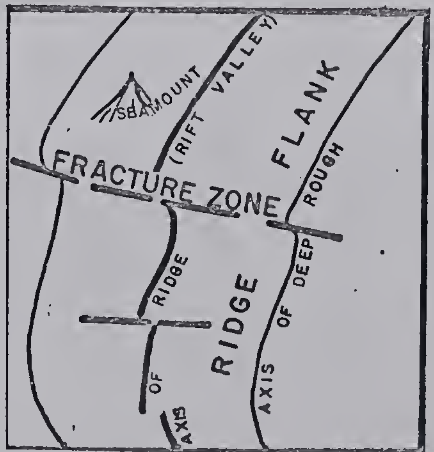
CONTINENTAL MARGIN PROVINCE



OCEAN BASIN FLOOR PROVINCE



MID-OCEANIC RIDGE PROVINCE

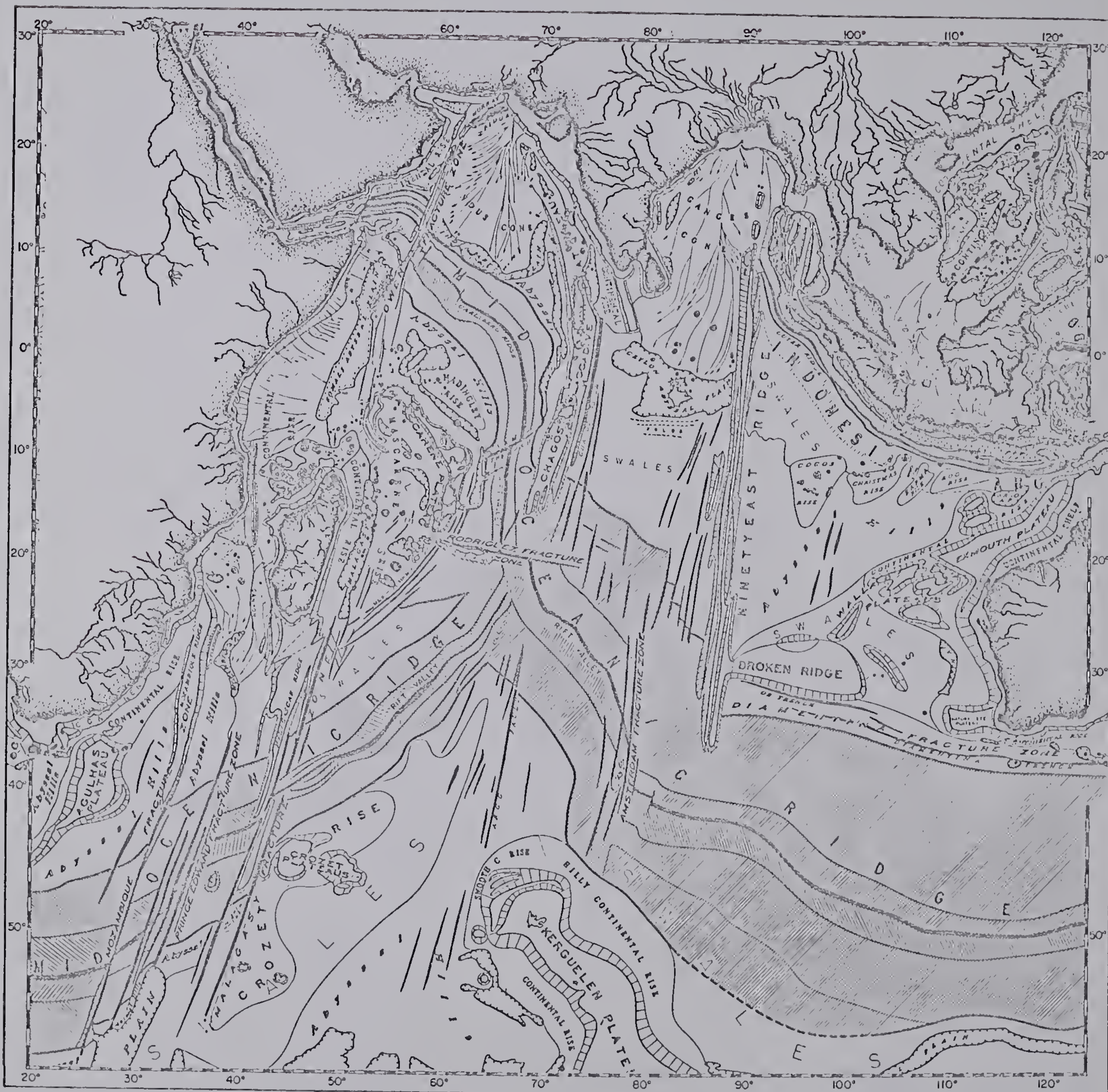


GENERALIZED PROVINCE MAP OF THE ATLANTIC OCEAN. Prepared for this report from a tracing of the map "Atlantic Ocean Floor", 1968; by B.C.Heezen and M.Tharp; as drawn by H.C.Berann for the National Geographic Magazine, vol.133.

FIGURE VI-2

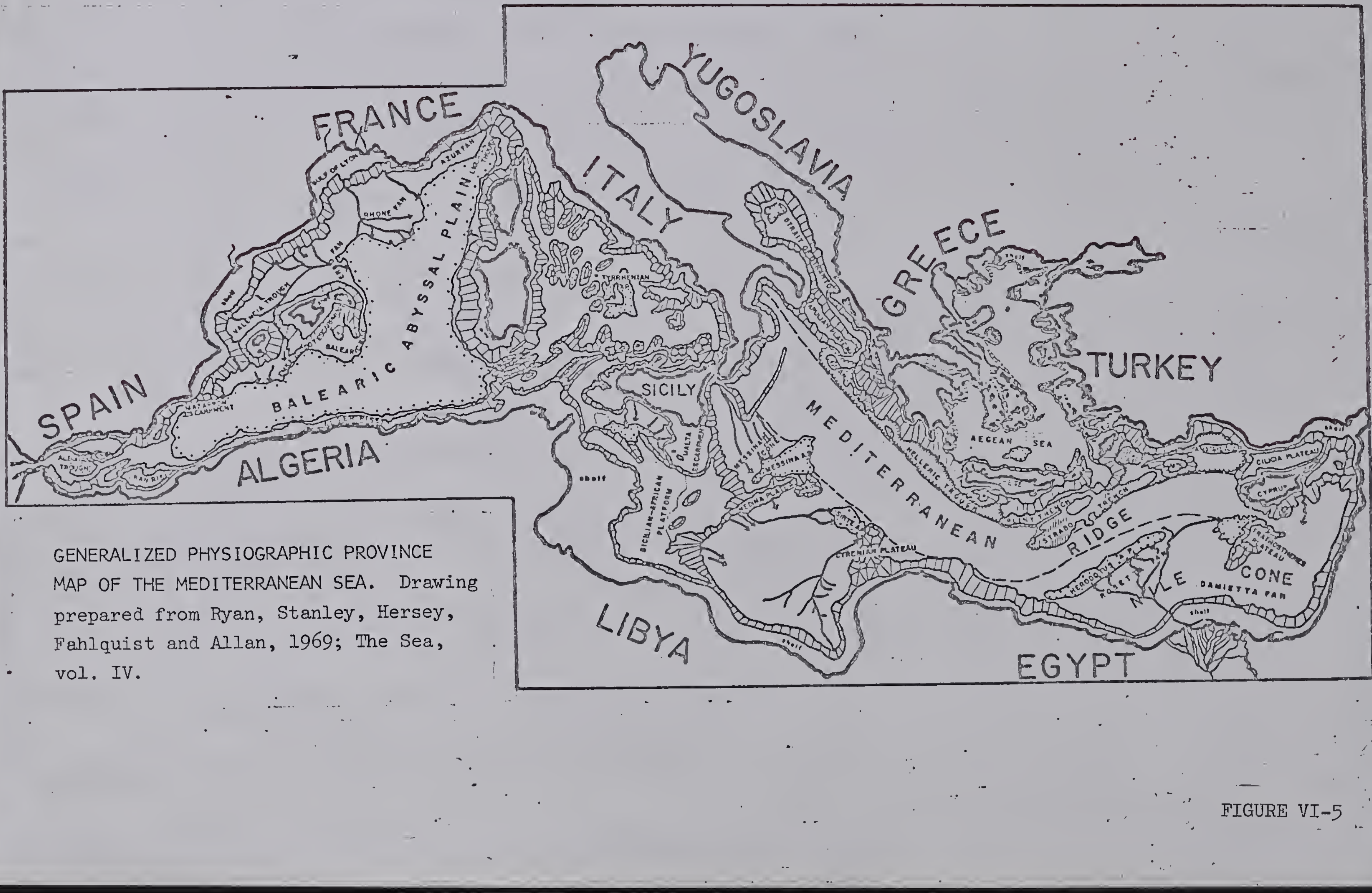
GENERALIZED PROVINCE MAP OF THE PACIFIC OCEAN. Prepared for this report from a tracing of the map "Pacific Ocean Floor", 1969; by B. C. Heezen and M. Tharp; as drawn by H.C. Berann for the National Geographic Magazine, vol. 136.

FIGURE VI-3



PHYSIOGRAPHIC PROVINCE MAP OF THE INDIAN OCEAN. (slightly modified for this report) This map accompanies the "Physiographic Diagram of the Indian Ocean" 1965, by B.C.Heezen and M.Tharp; Geological Society of America.

FIGURE VI-4



GENERALIZED PHYSIOGRAPHIC PROVINCE MAP OF THE MEDITERRANEAN SEA. Drawing prepared from Ryan, Stanley, Hersey, Fahlquist and Allan, 1969; The Sea, vol. IV.

FIGURE VI-5

in reservoirs. For example, longshore transport on the continental shelf of western Africa contributes material to a sediment trap within the Congo Canyon that has cut completely across the shelf. Much of the contemporary activity of turbidity currents within the Congo Canyon is believed to be linked with this active sediment source, which is most likely even a greater contributor of sediment than is the Congo River estuary*. Longshore transport also carries sediment into canyon heads which may cut back only a mile or two into the shelf edge.

In conclusion, the relative flatness of the continental shelf offers no place for contemporary turbidity current activity. However, certain dynamic processes of sediment distribution, both relic and presently active, have resulted in the development of local sediment reservoirs seaward of the shelf break, which may empty themselves in a catastrophic manner, as they have evidently done many times in the past.

6.1-2 Continental slope province

The main features of the continental slope are the submarine canyons which dissect it deeply in many areas. For a discussion of submarine canyons, the reader is referred to Shepard (1963), Shepard and Dill (1966) and Guilcher (1963). The continental slope has an average inclination of about 4.5° for the first 2000 m of descent.

On the basis of topographic gradient alone, it is apparent that the continental slope is favorable for slumping. However, when one estimates the volume of sediment which has been swept to the deep ocean floor by downslope processes, it vastly exceeds the volume of the denuded areas incised into the slope. Such observations indicate that if the continental slope is the source of the slumping which initiates the turbidity currents that carry this sediment to the ocean basin floor, most of the entrained sediment within the individual turbidity currents is not original slope material. Rather, the sediment has had a residence time on the slope, and the turbidity current process is a naturally-occurring means of removal of the transient materials.

The significance of the continental slope lies therefore in its role as a temporary reservoir and as a funnel by which flowing sediment is channeled into canyons. The slope receives sediments in its reservoirs from steady-state processes of sediment dispersal and releases the sediment from those reservoirs in catastrophic flows. From the viewpoint of placing structures on the seabed, the continental slope is probably the most hazardous area; since the reservoir zones as well as the transport routes in the canyons must be avoided.

*The fact that the head of the Congo Canyon extends as an embayment into the coastline attests that the river is also an important source of sediment.

6.1-3 Continental rise province

The continental rise lies at the base of the continental slope. The gradients on the rise average 1:300, but vary from 1:50 to 1:1000. The surface of the rise is generally smooth, though in some areas it exhibits a corrugated pattern of linear continental rise hills resembling sedimentary drifts.

In general usage, "continental rise" denotes the thick sedimentary apron at the base of the continental slope. In this apron we can distinguish two different structures which may have different origins.

First is the abyssal cone, which is a wedge-shaped body of sediment built out from the slope at its base. The cone often results from a single distributary. The basic configuration of an abyssal cone, if compared to an alluvial fan, suggests that it has been constructed from downslope dispersal of sedimentary materials radiating seaward from a single source of supply. Abyssal cones are found seaward from many major river deltas (i.e. Mississippi, Congo, Magdalena, Rhone, Nile, Ganges, Indus, Amazon), as well as (less commonly) at the mouths of canyons which have no relation to present-day river courses. Ewing, Ericson and Heezen (1958) have made a distinction between the massive composite apron which they call a cone, and smaller features which they call deep-sea fans. The fans radiate outward from individual channels on the cone, and one cone may contain several fans. The Nile Cone, for instance, has two fans, the Damietta and Rosetta, each of which has been built by a separate distributary system of channels. The abyssal cones and associated fans are implicitly connected with the turbidity current process, and it is generally agreed that these wedges of sediment originate from gravity-transported sediment in suspension. Differences of opinion center around whether these features are primarily constructional products of high-speed catastrophic flows or the result of low-speed, low-density, steady-state, downslope dispersal in nepheloid layers.

Other structures which are distinguishable from abyssal cones in their basic morphology are prism shaped continental rises, such as those off the east coast of the United States and the west coast of Africa (Somali). Seismic reflection proviles reveal that such prisms of sediment are not simply a series of coalescing cones. Instead the continental rise appears as a broad smooth-surfaced wedge built against the slope and over the flat-lying ocean floor sediments in the manner of a linear asymmetrical sedimentary drift (Ewing and Ewing, 1964). Parts of the rise of the eastern margin of the United States are detached from the continental slope and occur as outer ridges (i.e. the Blake-Bahama, Hatteras, and Caicos Outer Ridges). The exact mechanism for the production of the outer ridges is not clearly understood. Heezen et al., (1966) have suggested that these bodies are more likely the constructional products of contour-following, sediment-laden, geostrophic currents associated with strong thermohaline circulation, rather than the result of downslope sediment dispersal solely under the influence of gravity. Heezen (1968) states that "recent investigations indicate

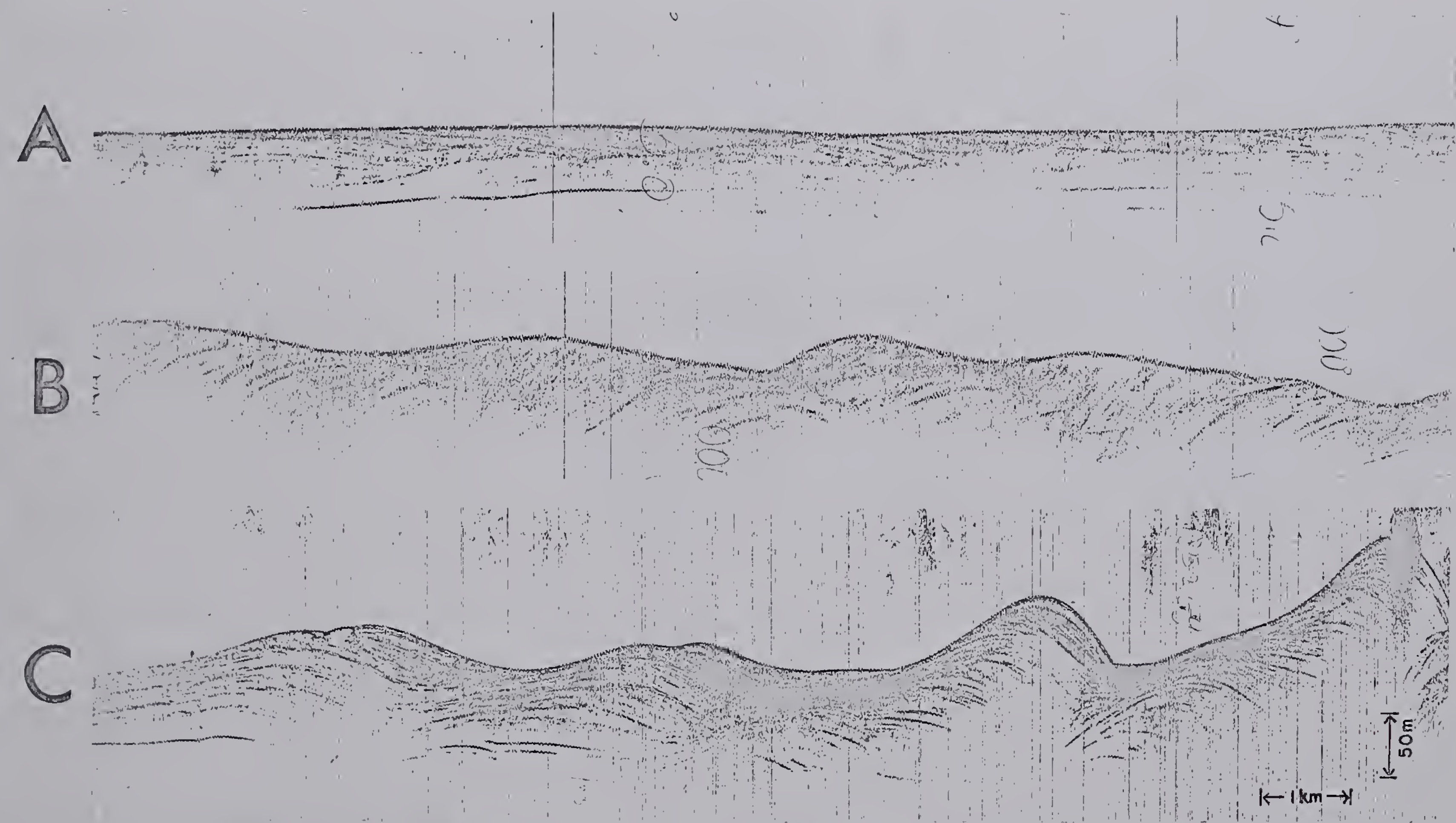


FIGURE VI-6 - Migratory sediment drifts

Examples of migratory sediment drifts typical of the Lower Continental Rise Hills province, but also found on the flanks of detached outer ridges. (ROBERT D. CONRAD-12)

A. Extremely long wave-length hills demonstrating macro-bedding.
B. Steeply dipping depositional bed forms.
C. Migrating lutite waves on the Argentine Outer Ridge.

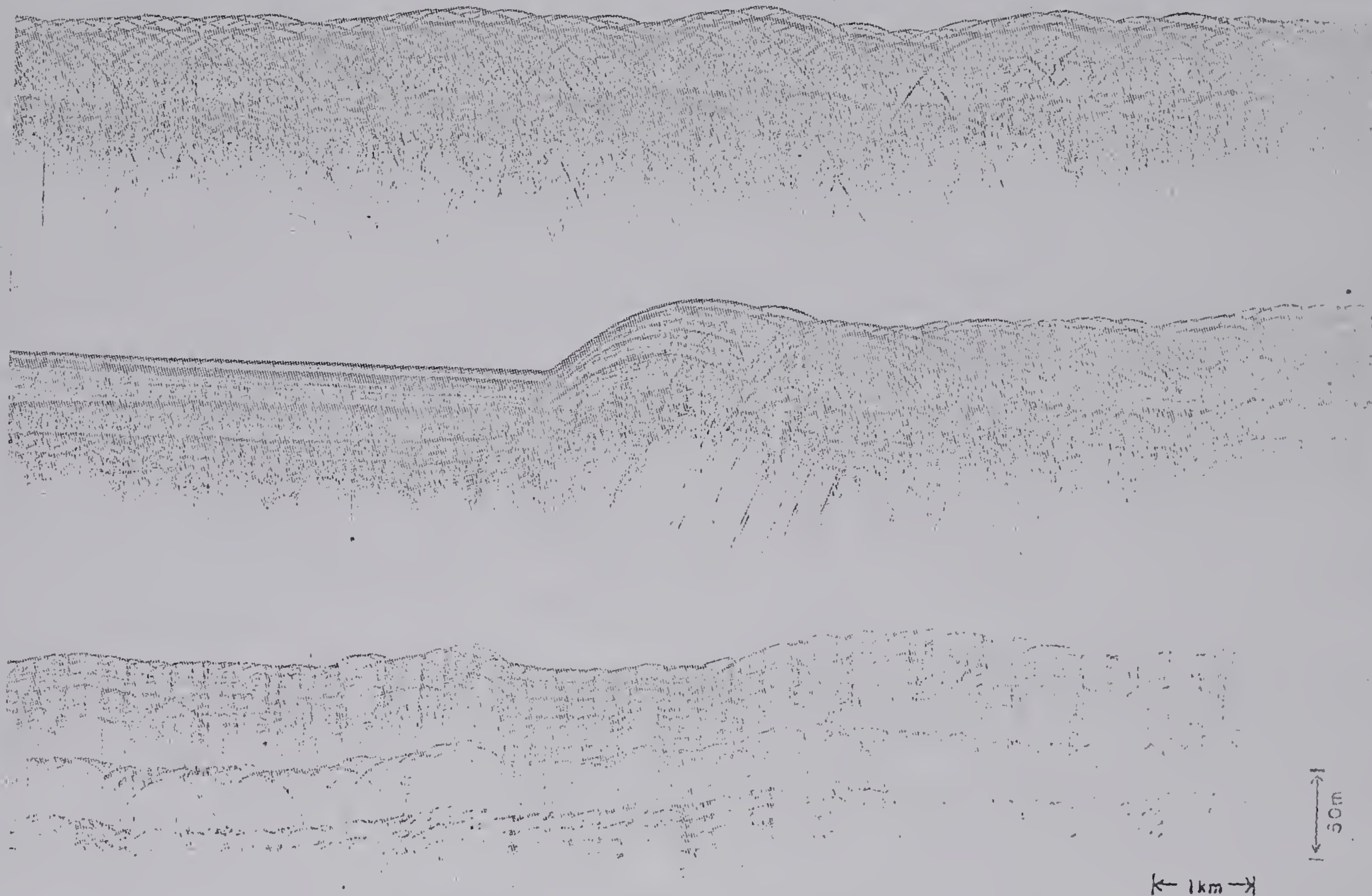


FIGURE VI-7 Examples of corrugated surfaces on and beneath the seabed.

These surfaces are created through processes of winnowing and the concurrent [illegible] waveforms such as dunes and ripples. These examples are from the [illegible] (ROBERT D. CONRAD-12, 1968) that contour-following bottom currents involved in the deep thermohaline circulation of the world ocean are perhaps the most important factor in the transportation and deposition of sediments on the continental rise and that the currents probably shape the continental rise and associated sedimentary drifts, the outer ridges."

Two observations are important to the main discussion of this paper concerning turbidity currents and their effect on sea floor structures. First, the sediment which makes up the rise and outer ridges is originally derived from the continents and thus must travel with some downslope component. Second, the deep thermohaline circulation contains, in suspension, a diffuse cloud of fine-grained material,--the nepheloid layer (Ewing and Thorndike, 1965; Eittreim et al., 1969). Some of this suspended sediment may be scavenged directly from the sea floor as the result of interface dynamics (Bagnold, 1963) and vertical eddy diffusion, but since the rise is constructional, the net transport must be from the continent to the rise.

The question arises whether the nepheloid layer observed today in vigorously circulating boundary currents is maintained exclusively as the result of lifting sediment from the sea floor beneath the current and temporarily maintaining it in suspension by vertical eddy diffusion, or if the presence and configuration of the nepheloid layer attests to some steady-state net input of sediment into the deep boundary current. The most likely source for input into the boundary current would be from the lateral overflow from downslope channelized turbidity currents. In other words, the rise may be constructed and "shaped" by sediment-loaded contour currents, but the input of this material into the deep sea is most likely to be primarily by turbidity currents. The unanswered question remains: "What is the true significance of the observed nepheloid layer vis-a-vis the role of turbidity currents?" Particularly puzzling are light scattering observations in the Caribbean near the base of the Magdalena Cone which show no nepheloid layers, although there have been recent slumps and turbidity currents on this cone inferred. Similar measurements revealed a lack of a bottom nepheloid layer on the Messina and Sirte Abyssal Plains in the eastern Mediterranean despite the occurrence of recent and contemporary turbidity current activity in this area (Ryan and Heezen, 1965). Perhaps true turbidity currents create suspensions of very limited vertical thickness, as suggested in previous sections of this report, and the maintenance of thick nepheloid layers (up to 2000 m in thickness on the continental rise off Hatteras) result from a type of vertical eddy diffusion, supported by geostrophic flow and unsupported by intermittent short-lived turbidity current flow.

From the point of view of effect on sea-floor structures, contour currents are capable of winnowing, rippling, and re-working loose surface sediment, possibly to depths of more than 0.5 m. Post-glacial rates of accumulation on the rise reach 50 - 100 cm/1000 yr in some areas. Thus small scale drifts could conceivably form around objects placed on the sea floor and erosion may even undermine the supports of such objects. Figure VI-6 shows large lutite waves in the Argentine Basin where the subsurface layering indicates a progressive migration of these waves during deposition. These very large scale features demonstrate convincingly the ability of the present shape of the sea floor interface to control future deposition, - a process which would be applicable to protruding man-made objects as well.

Figure VI-7 illustrates smaller scale corrugations at the sea floor interface and also within sub-bottom layers, which appear in reflection profiles as regularly-spaced overlapping hyperbolae. The cause of the reflection hyperbolae is believed to be regularly-spaced ripple troughs of about 50 m amplitude which focus the acoustic energy from wide-beam echosounding transducers. The appearance of corrugated surfaces at discrete levels in the sub-surface, and their absence at other levels indicates that the depositional process responsible for these features is intermittent. If the ripples are the products of boundary currents, then the deep circulation has apparently varied in intensity in the past. Recent unpublished studies at Lamont-Doherty have demonstrated that in some areas of the North Atlantic, periods of intense thermohaline circulation at the sea floor are associated with changes in the earth's climate caused by the growth of glaciers and the drop in sea level.

The 12kHz reflection profiles in Fig. VI-8 show examples of differential erosion and deposition. In profile A corrugations are seen only within the deep axis of a trough, indicating that current activity was confined to this narrow zone. Profile B shows a wedge of superficial transparent sediment whose distribution is clearly related to the pre-existing sub-surface morphology. The result of such differential erosion and deposition would be to create scour channels around foundations of sea-floor structures and to smother other parts under local drifts. Optical apertures and sensitive pressure transducers would be severely effected, and hydrodynamic strumming might interfere with acoustic experiments. However, under some circumstances, rapid burial of deep-sea cables under sea floor drifts could be advantageous.

Other examples of controlled deep-sea erosion and deposition are shown in Fig VI-9. Profile A shows the development of regularly spaced sedimentary waves identified as "lower-continental rise hills" (Heezen et al., 1959) on the margin of a deep sea fan. It is an interesting fact, although we have no clear dynamic explanation for it, that in all cases investigated at Lamont-Doherty, the hills or the fans appear always to migrate obliquely uphill. Profile B shows a straight line crossing of a single erosional channel on the Rhone Fan. At the center of this profile, one can see where the ship passed directly over a tight bend in the channel, such that the reflection from the outside wall at the bend was maintained. Such erosional channels are believed to be carved into the cone by intermittent turbidity current action. Profile C on the Mississippi Cone shows a large section of the sea floor from which as much as 10 m of superficial loose material has been removed by very recent erosion. The extent of the erosion appears to be more severe to the right on the upper part of the cone, and may be controlled by the particular contour level of the geostrophic flow in this region of the Gulf of Mexico.

In summary, the continental rise, whether an apron of sediment deposited from steady-state bottom currents, or a wedge of sediment in an abyssal cone, is a most dynamic province. The placement of structures ought to take into consideration the whole spectrum of hydrodynamic processes related to sediment transport and sea floor erosion.

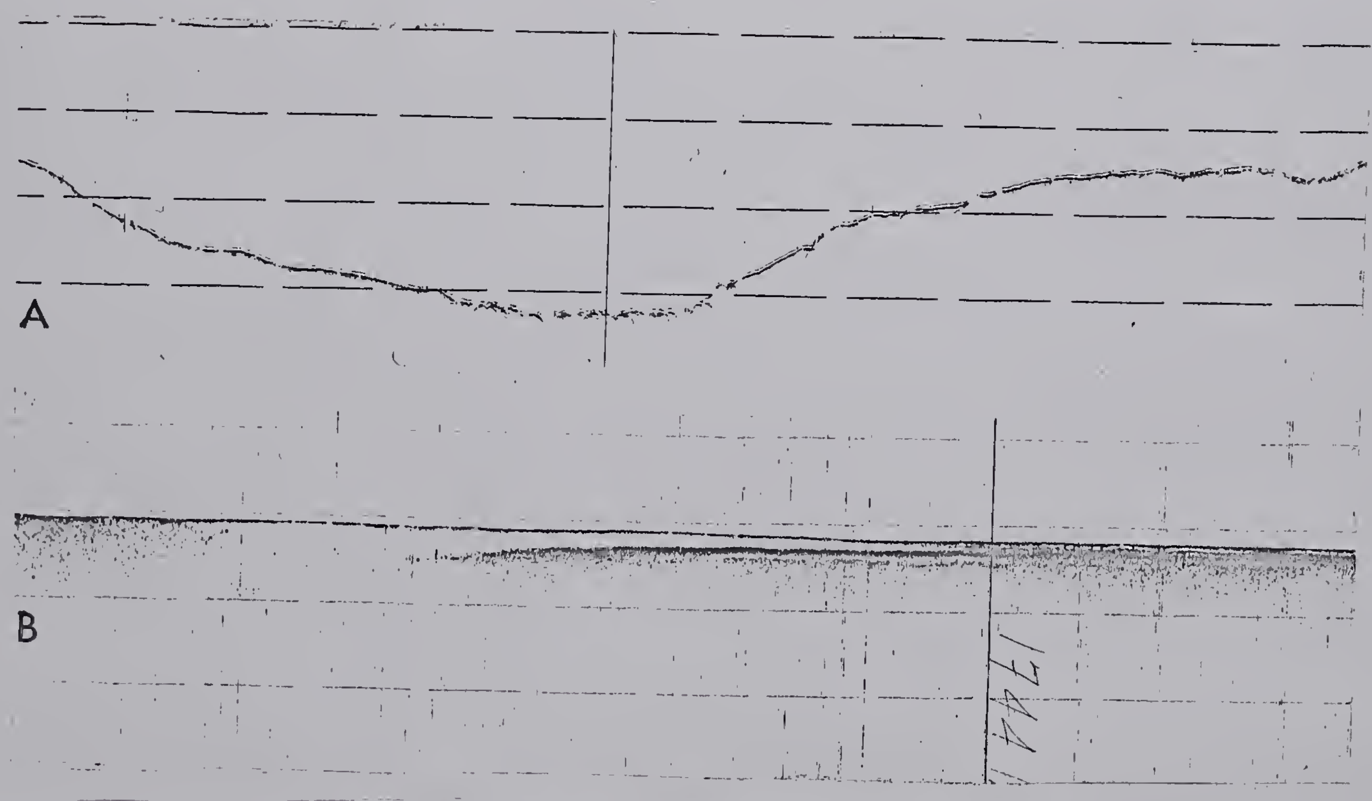


FIGURE VI-8 <u>Examples of differential erosion and differential deposition</u>

In profile A the occurrence of a corrugated sea floor surface only along the axis of a small channel indicates that the concomitant winnowing and erosion by bottom flow is confined to the deep axis of this valley. (ROBERT D. CONRAD-9, 1965, Gulf of Aden)

In profile B a lens of transparent sediment exhibits a marked preference in its deposition within a depression in a fashion which smooths the regional relief. (ROBERT D. CONRAD-9, 1965, Lower Mississippi Cone, Gulf of Mexico)

FIGURE VI-9 - Example of contemporary preferred deposition and controlled erosion. A. The upslope migration of broad low dunes is noted in many cases on the right-hand (looking downstream) flank of deep-sea fans. B. This profile shows a down axis cross-section of a single tightly meandering deep-sea channel on the Rhone Fan. The configuration of the meanders appear to be in a steady-state equilibrium with the average dynamics of the Rhone Fan turbidity currents. C. Outcropping of loosely consolidated marine muds on the Mississippi Cone. The level of cutting has created a new seafloor surface which shows little respect for previous bedding plains. The most marked erosion is where the slopes are greatest.(ROBERT D. CONRAD-9, 1965)

6.1-4 Abyssal plain province

Seaward of the slope and rise lie the broad flat plains which mark the final sites of gravity-controlled sediment deposition. Abyssal plains were first observed by Maurice Ewing on an expedition of the Woods Hole Oceanographic Institution's Atlantis in the summer of 1947. This was the first expedition on which the conventional fathometer had been adjusted so that it produced good records down to truly oceanic depths, permitting accurate continuous recording of the ocean floor contours. Ewing reported that "about midnight on July 27 we entered a great plain at 2,900 fathoms, and this we crossed for the next two and a half days... It is very level..." (Ewing, M., 1948). A second Atlantis expedition re-examined this plain and found it to be "almost unbelievably flat and level..." (Ewing and Sisson, 1949), and took sediment cores that proved the existence of graded beds with coarse sands at the base.* We now know that in the Atlantic, Indian and northeast Pacific Oceans, abyssal plains are as flat as 1:10,000. Abyssal plains adjoin all continental rises and can usually be distinguished from the rises either by a distinct change in bottom gradient or by a marked change in acoustic property, related to sub-bottom layering and reflectivity.

Every line of evidence suggests that abyssal plains are produced by turbidity currents spreading out on the sea floor. The plains contain coarse sand and shallow water fossils that are clearly derived from the continents and shallow shelves. All the known abyssal plains occur only where there are no topographic barriers between the plain and the continent (Heezen and Menard, 1963). Thus abyssal plains are rare in the Pacific, where typically the continents are bordered by island arcs and trenches (see Fig. VI-3). Although much of the deposition by modern turbidity currents has been in great fans at the mouths of submarine canyons, an equal volume of sediment resides on the ocean basin floor in the form of the vast flat-lying plains which have buried pre-existing basement relief (Ewing and Ewing, 1964). Abyssal plains are also found in marginal trenches, in marginal basins, and even in some lakes and reservoirs. When two adjoining abyssal plains, lying at different depths, are interconnected, the passage which joins them is termed an abyssal gap (Heezen and Laughton, 1963).

As mentioned earlier, the flow dynamics of an individual turbidity current from the continent to the distal abyssal plain vary considerably along its path and are dependent upon the local sea floor gradients, as well as the magnitude and composition of the initial turbulent suspension. A spectrum of different phenomena will accompany the flow and a variety of different sedimentary deposits will be found. To illustrate the contrasts and complexities, we will discuss in the following section the entire depositional environment from source region to distal abyssal plain for two type areas. One, in an open basin configuration in the Pacific Ocean, typifying the abyssal cone province; and the other in a small enclosed basin in the Mediterranean Sea, typifying the filling of topographic depressions by sediment ponding.

*Underway seismic reflection measurements, made each hour, yielded water depth to ±0.5 fm.

6.2 PHYSIOGRAPHY AND SEDIMENTARY PROCESSES OF THE LA JOLLA SUBMARINE CANYON, FAN VALLEY, SUBMARINE FAN, AND SAN DIEGO TROUGH

The depositional environments of the La Jolla Canyon, fan valley, fan and San Diego Trough are well known from closely-spaced sounding lines, observations from submersibles, numerous cores, and continuous reflection profiles (Shepard, et al., 1969). This region has been investigated more thoroughly than any other comparable submarine environment, and may be considered as a type area for many of the depositional processes discussed in this report.

Since this region off the western coast of California is typical of many of the continental margins of the Pacific and Indian Oceans, and of limited regions of the Atlantic, a brief review is in order.

The bathymetry of the La Jolla submarine canyon, fan valley, and submarine fan is discussed in detail by Shepard and Buffington (1968). Figure VI-10 shows selected topographic maps of the canyon, with details of the fan valley and its terminus on the San Diego Trough. The contours of the submarine relief in the canyon and the locations of samples examined (von Rad, 1968; Rees et al., 1968; Shepard et al., 1969) are shown in Fig. VI-10 A. The La Jolla Canyon and its concordant tributary, Scripps Canyon, continue seaward beyond their 150 fm (274 m) juncture to a depth of about 280 fm (512 m) cutting through rock formations. There the gradient abruptly decreases, such that the canyon becomes a deep-sea channel. (Shepard uses the term "fan-valley rather than "deep-sea channel", and we will retain that usage in discussing the part of the La Jolla distributary system on the La Jolla fan). Seaward of the canyon, the fan-vally cuts semi-consolidated deposits, which are also found in the San Diego Trough. The trough is an oblong, flat-floored basin that is completely filled with sediments to its sill, and has a uniform, unidirectional slope of about 1:1000 to the south. This gradient indicates a steady-state sedimentary equilibrium controlling deposition of fine-grained silty and sandy sediments, since part of the recent sediment supply which enters the trough from the continent by way of the canyon-fan-valley system bypasses the trough and spills over to the southwest to fill the adjacent San Clemente Basin (Hand and Emery, 1964), while part is retained and deposited on the fan proper. The San Diego Trough is what is called an "open-ended"basin.

Along its lower course, the fan valley has a meandering inner channel, shown in Fig. VI-10 B (Shepard and Einsels, 1962; Shepard and Buffington, 1968). The valley rim consists of marginal embankments, or "natural levees." These levees are virtually continuous along the north side of the fan valley, but intermittent on the south side. There are small terraces on the inner side of winding turns within the channel. On the outside of some of the bends, observers from submersibles have noted extremely steep cliffs and vertical walls with outcrops of erodable semi-consolidated sediment (Buffington, 1964; Moore, 1965). Isolated cobbles and boulders (up to 1 m in diameter), as well as rubble blocks of semi-consolidated clay, apparently broken off the precipitous walls, have been found in the fan valley (Shepard et al., 1969).

During a dive at 450 fm (823 m) an extraordinary number of small cracks and narrow ridges were seen in the channel fill (Shepard and Dill, 1966). Some of the cracks and narrow ridges appeared to extend diagonally up-valley, and to resemble tension crevasses on the sides of glaciers. These cracks might indicate slow downslope creep of parts of the channel fill, which would be surprising on a regional slope of less than one degree.

Recent investigations by seismic reflection profiling and by extensive coring (Moore, 1966; von Rad, 1968; Shepard et al., 1969) have shown that the youthful Holocene fan deposits are relatively thin, with older, semi-consolidated sediments exposed in only a few places within the channel axis. Generally, on the valley floor there is a surface layer of flocculated mud, easily set into suspension by eddies from the motion of submersibles. Many deep-sea holothurians and brittle stars have been observed here. Both of these observations suggest that at present the valley floor is not an area of high current velocities. An upright tin can with an attached siliceous sponge (photographed by Dill) indicates many years of tranquillity.

From 560-600 fm (1010-1080 m) the morphology of the fan valley with its several smaller distributaries becomes rather indistinct, having relative channel depths of only 1-3 fm (2-5 m). Here the predominantly sand-filled channel merges with the relatively flat floor of the San Diego Trough (as shown in Fig. VI-10 C). At this merger, large drifts with wavelengths of 15-30 m and heights up to 1 m have been found, together with erratic blocks of semi-consolidated clay, driftwood, and parts of shallow water plants such as kelp and eelgrass.

The San Diego Trough contains at least 1000 m of rather flat-lying post orogenic sediments (Moore, 1966). Descriptions of the distribution of sediments and structures within the sediments were given by von Rad (1968), Table VI-1. The near surface materials in the San Diego Trough consist mostly of poorly-sorted clayey silt, with some well-sorted, mostly fine to very fine sands. The sands are distributed in an irregular pattern. On the outer peripheral part of the La Jolla Fan and the proximal San Diego Trough, the sands form very thin, finely-laminated and cross-bedded layers (the thickness ratio of sand:mud being 0:0.1). Thicker layers of fine to medium sand, many of them graded, are found only in wedge-like concentrations along the La Jolla Canyon and fan-valley axis. These fan-valley sediments have a greater sand:mud thickness ratio (0.25:4.0).

Along the narrow, meandering canyon and fan-valley, many of the layered sand beds contain a variety of sedimentary structures (Figs. VI-11 and VI-12). Many of these structures are similar to those frequently observed in flysch beds. They include graded bedding, parallel lamination, small-scale ripple lamination, convolute lamination, and basal sole

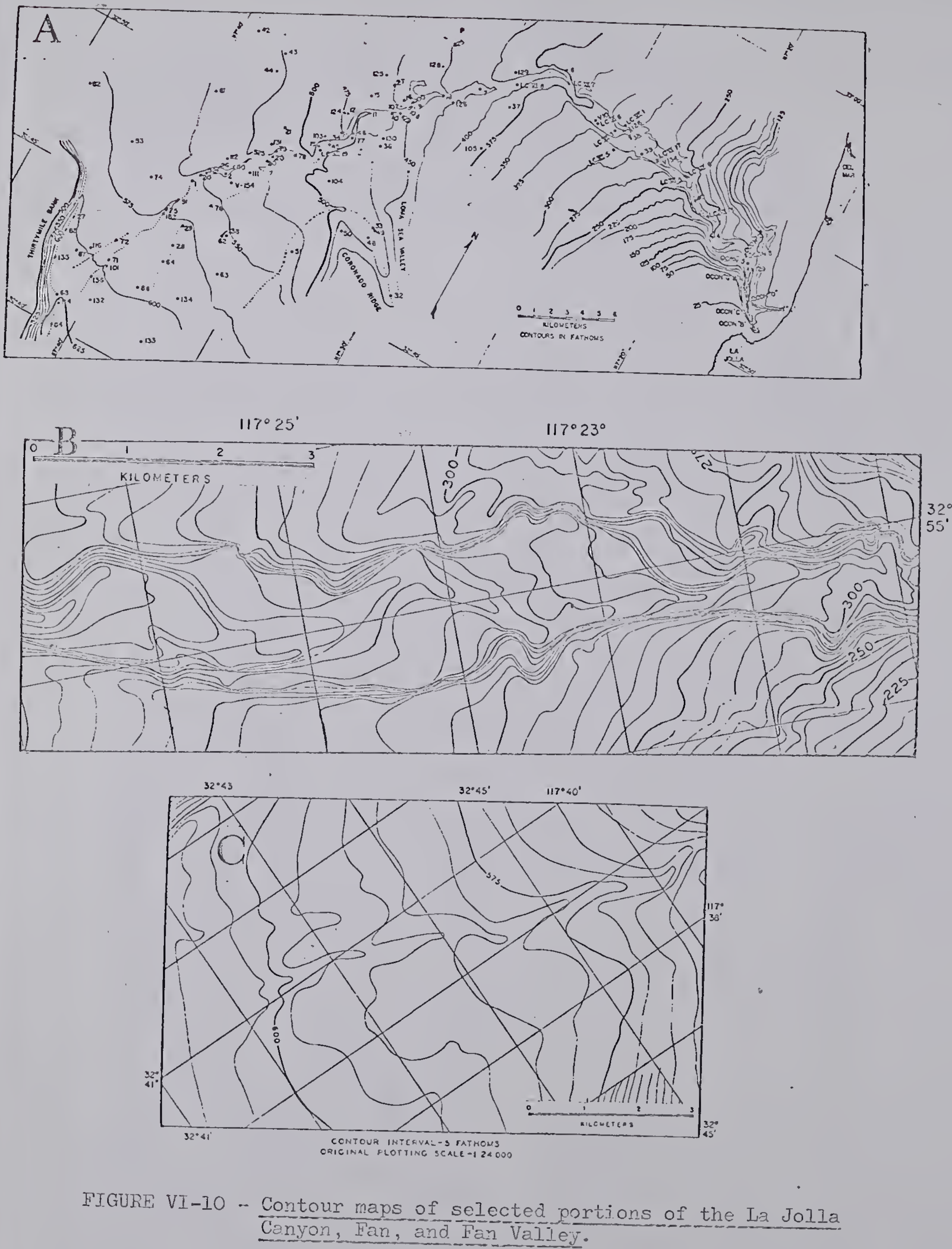


FIGURE VI-10 - Contour maps of selected portions of the La Jolla Canyon, Fan, and Fan Valley.

A. The entire distributary system from Scripps Canyon to the San Diego Trough.

B. Gentle meanders within the La Jolla Fan Valley.

C. The mouth of the Fan Valley where it debouches into the San Diego Trough.

TABLE VI-? —*Comparison between the Facies of the Bavarian Flysch and San Diego Trough** ** †

Characteristics	Bavarian Flysch (GF = Gault Formation)	San Diego Trough (SDT) and La Jolla Fan (LJF (LJC = La Jolla Canyon, FV = La Jolla Fan-Valley)
I. Tectofacies and Basin Morphology		
1. General bathymetry and basin shape	Oblong, narrow, marine deep-water trough (± restricted) elongated parallel to the tectonic strike (*11*)	
2. Dimensions: length	**>500 km (?)	about 100 km
width	*Several tens of km (>50 km??)	about 20 km
area	>10,000 km²	about 2000 km²
3. Depth of basin	**deeper than neritic (>200 m)	about 500–1300 m
4. Basin relief, axial slope, nature of borders (sill depth)	GF**: *flat* sea floor (bed-by-bed correlations!) sloping to the E (?). Trough bordered by (submarine) geanticlinal ridges or island arcs (?) to the N and S	LJF+FV: gradient 1–2%. Meandering channels with lateral terraces and levees. SDT: flat-floored, unidirectional gradient (0.5%) to the S. SDT bordered by the sediment supplying continent to the E, open to the ocean (W), but locked from oceanic influences by submarine ridge (bank).
5. Tectonic setting and basin history. (See also fig. 13.)	*Synorgenic* trough, formed during and after the first phases of orogeny of the Northern Calcareous Alps (Middle Cretac.) **Filling* of trough from distant source areas, possibly independent of tectonic events (orogeny).	SDT: tectonically unstable fault-trough (graben ?) within Continental Borderland (*2*), downfolded and blockfaulted during Miocene diastrophism (*2, 7*). Some recent earthquake activity. Pleistocene and Holocene basin fill slightly affected by recent folding and faulting (*7*). SDT rapidly filled during Pleistocene. At present filled to overflowing (*2, 7*). Relation to structural history of Peninsular Ranges and Ventura and Los Angeles Basins. See fig. 13!
II. Lithofacies		
1. Location and nature of source areas	**A few point sources (?submarine canyons and fans) only. Distant crystalline source area (near distal end of trough ?), partly covered by carbonate rocks (GF).	One major (LJC) and a few minor lateral point sources for the sands (for example Coronado Canyon). Nearby, high-relief (>1000 m) crystalline source area (plutonic rocks)
2. Sediment supply (availability)	Abundant and readily available supply of coarse- and fine-grained, continent-derived clastic sediments	
3. Sediment dispersal and reworking	**Mainly longitudinal (in axial part), for GF proven by bed-by-bed correlation (*3*).	** Lateral supply along FV+LJF, longitudinal dispersal along axis of SDT (*15*)
4. Mechanisms of sand transport, (paleo-) current system	*Periodic marine bottom currents with pulsating velocities? (GF: *3, 12*), possibly also episodic turbidity currents (?)	*Tide-related, periodic bottom currents (v_{mean} 5–10 cm/sec, $v_{max.}$ >25 cm/sec) (*15*).* At times swift, partly erosive currents of unknown origin (inferred from incidental evidence, see text)
5. Direction of sediment transporting currents	**Trough-parallel, ± constant in space and time for one formation. Several reversals in Cretaceous.	*In center of SDT longitudinal (periodicity in time not known)
6. Total thickness of sediment fill	Total (Bavarian Flysch): 1000–1500 m (compacted)	Central SDT: 420 to >1100 m (7) LJF: in general about 50–420 m
7. Rate of deposition	*GF: relatively slow (2.5 cm/1000 yrs: *3*) Reiselsberger Sandstein: 2–max. 7 cm/1000 yrs	**Rapid (−160 cm/1000 yrs) during Pleistocene times (*7*); very slow recent rate (ca. 8 cm/1000 yrs for uncompacted sediments, or about 4.5 cm/1000 yrs for compacted solids) on outermost LJF (*15*).
8. Rhythmicity of sedimentation	Monotonous rhythmic cycles of (graded) psammite-pelite series. Extraneous deep-sea sands (graywackes), interbedded with (hemipelagic?) shales	Extraneous deep-sea sands alternating with allochthonous (not hemipelagic!)* turbidite muds (*7*)

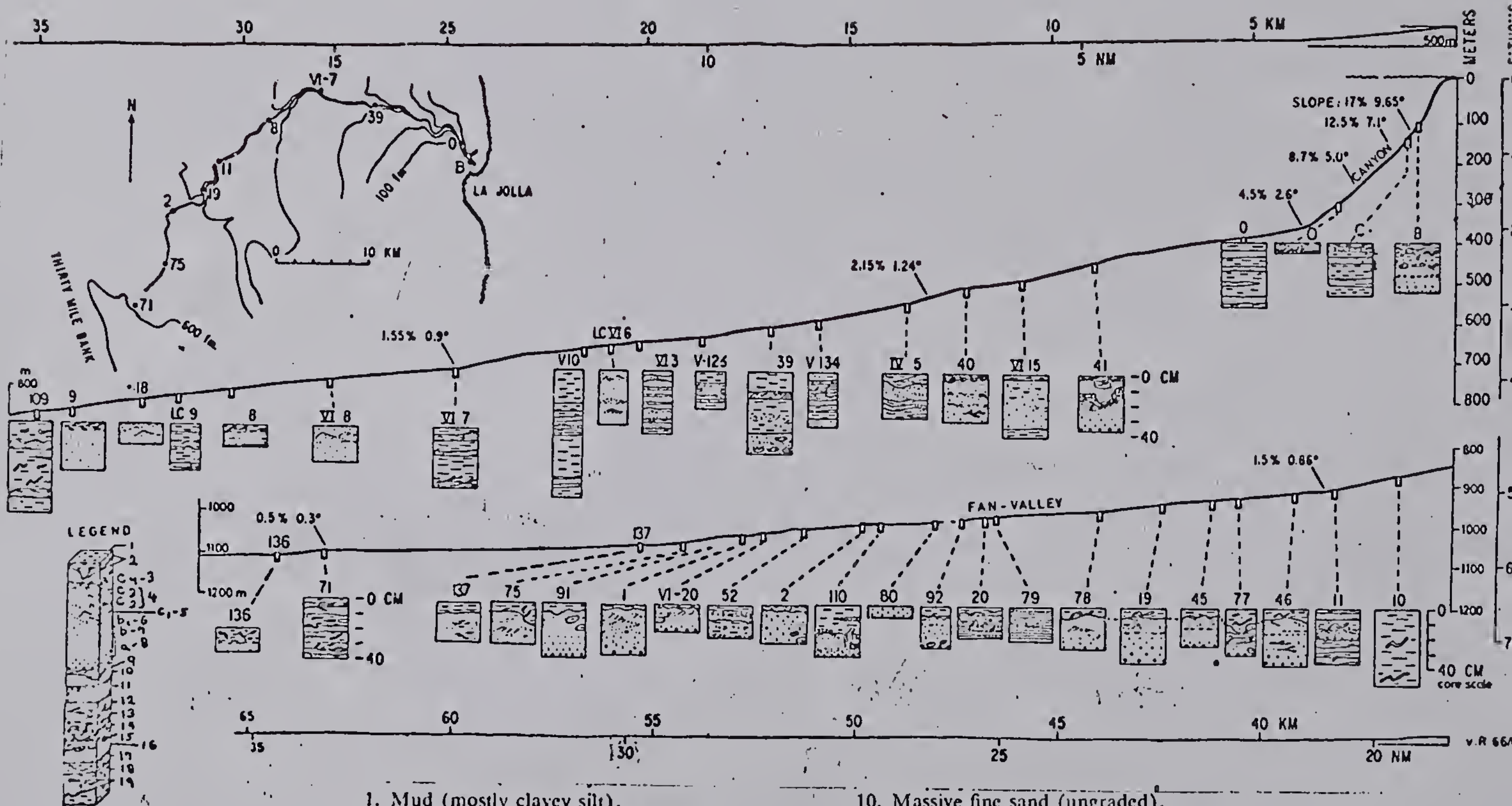
(from Von Rad, 1968)

TABLE VI-2† (Continued)

Characteristics	Bavarian Flysch (GF = Gault Formation)	San Diego Trough (SDT) and La Jolla Fan (LJF) (LJC = La Jolla Canyon, FV = La Jolla Fan-Valley)
9. Psammitic pelite (sand:mud) ratio	GF (3): average 0.5 for beds A-AL (fig. 7): 1.3	FV: 0.25-4, LJF+SDT: <0.25 (uncompacted surface sediments!)
10. Thickness of sand(-stone) beds	GF (3): average: 110 cm (5-250 cm)	LJC+upper FV: up to 25>cm LJF+SDT : 2-10 cm
11. Consistency of individual beds and volume of sediment/bed (arrangement of lithic fill: 11)	**GF (3): individual graywacke beds correlated over >110 km Up to 4×10^9m³** per bed (3)	**Discontinuous, elongate, lense-shaped bodies (shoestring sands) along FV; max. extent 4 km* (see text); max. volume *0.24×10^6m³/sand layer; "ponded" sediments in SDT (7)
12. Character of boundaries of sand(stone) beds	Lower b.: always distinct Upper b.: generally less distinct	lower b.: always sharply defined (erosional) Upper b.: partly gradational, partly distinct
13. Horizontal grading	GF (3): slight, but continuous downcurrent decrease in grain size and bed thickness	Inconsistent along FV SDT: ?
14. Texture: average grain size and sorting of psammites	GF: graywackes poorly sorted very fine to fine (med.) sandstones, sometimes bimodal (one mode: clay matrix). Sorting slightly improving towards bed tops	Sand layers: well sorted, very fine to fine (medium) sands (inherited from winnowed littoral sands). Some silt layers poorly sorted (mixed by bioturbation). Standard deviation slightly decreasing towards the top in most cases
15. Composition a) light mineral associations	GF (3): compositionally mature quartz-graywackes, very high quartz: feldspar ratios. Other formations more immature	Immature (sub)arkoses, quartz: feldspar-ratio very low (1.3): disintegration of intermediate plutonic source rocks
b. heavy mineral associations	GF: stable zircon-tourmaline association (grain size dependent: "granular variation")	Immature amphibole-epidote ass. no granular variation observed
	Mineral associations ± evenly distributed and uniform over large parts of the depositional basin	
16. Sedimentary structures a) internal structures	Complete or incomplete sequence of flysch-type graded sand beds: basal massive graded division (a), succeeded by the divisions of parallel (b), small-scale current-ripple cross-lamination (c) and convolute lamination (1, 15). GF: graded beds more prevalent than ungraded, laminated beds ("laminites")	LJF: micro-scale slump structures, large intraform. mud pebbles, erosional channels, layers of plant debris (kelp, wood etc). Parallel and cross-lamination much more common than grading (15) SDT: predominantly "laminites" (?)
b) sole markings	Abundant scour marks (flute casts), drag marks, load casts, and so forth.	Irregular sand-filled scour marks, scour and fill structures, load casts
III. Biofacies 1. Macrofauna	Macrofossils extremely rare (exotic shells) in psammites	Shell fragments in coarse sands, but rare in sands
2. Microfauna	Autochthonous benthonic faunas in pelites ("flysch-type" arenaceous foraminifera), allochthonous shallow-water microfauna in graywackes (9)	In deep-water sands: allochthonous benthonic foraminifera, displaced by currents (tractive or suspension) from the neritic to upper bathyal biofacies zones (10, 8). Downslope sedimentation shown by the crossing of faunal lines
3. "Ichnofacies" (13) (trace fossils)	In psammites: surface-feeding trails ("Weidespuren") and feeding burrows ("Fressbauten") much more common than "repose trace fossils" ("Ruhespuren") (13)	? (not yet studied in detail)

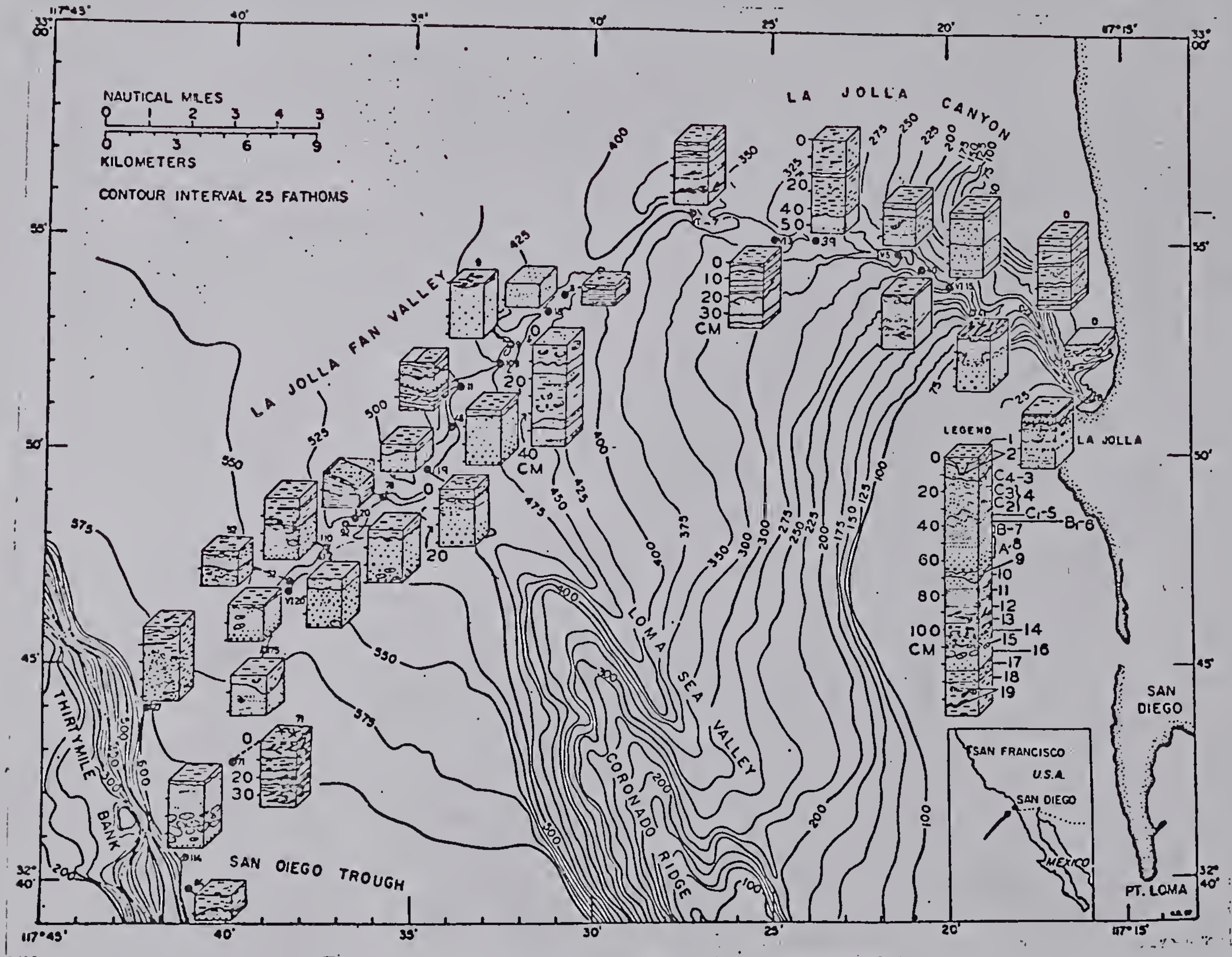
* inferred from alleged (incidental) evidence; alternative explanation possible.
** inferred from suggestive evidence.
† for key to references, see list on footnote to table 1.

(from Von Rad, 1968)



1. Mud (mostly clayey silt).
2. Slump structure and microfaulting.
3. (c_4) Convolute laminae.
4. (c_3) Wavy laminae.
   (c_2) Deformed cross laminae.
5. (c_1) Current-ripple cross laminae.
6. (b_1) Inclined laminae.
7. (b) Parallel laminae.
8. (a) Massive graded bedding.
9. Erosive channel filled with clayey silt.
10. Massive fine sand (ungraded).
11. Semiconsolidated Pleistocene(?) clay.
12. Layer of plant debris (kelp, surf grass, wood, *etc.*)
13. Layers of coarse sand, gravel, and pebbles.
14. Layers with gravel and shells.
15. Mud balls (rounded fragments of semiconsolidated clay).
16. Fine to medium sand.
17. Sandy silt.
18. Clayey silt.
19. Bioturbation by burrowing organisms (worms, mollusks, echinoids, *et*

FIGURE VI-11 The sedimentary structures seen in box cores taken along the La Jolla Canyon and Fan Valley. The cores were taken within the canyon and fan valley. (Shepard, Dil and Von Rad, 1969)



1. Mud (mostly clayey silt).
2. Slump structure and microfaulting.
3. (c_4) Convolute laminae.
4. (c_3) Wavy laminae.
   (c_2) Deformed cross laminae.
5. (c_1) Current-ripple cross laminae.
6. (b_1) Inclined laminae.
7. (b) Parallel laminae.
8. (a) Massive graded bedding.
9. Erosive channel filled with clayey silt.
10. Massive fine sand (ungraded).
11. Semiconsolidated Pleistocene(?) clay.
12. Layer of plant debris (kelp, surf grass, wood, *etc.*)
13. Layers of coarse sand, gravel, and pebbles.
14. Layers with gravel and shells.
15. Mud balls (rounded fragments of semiconsolidated clay).
16. Fine to medium sand.
17. Sandy silt.
18. Clayey silt.
19. Bioturbation by burrowing organisms (worms, mollusks, echinoids,

FIGURE VI-12 Block diagrams of box cores from environments outside of the La Jolla Canyon and fan valley. For interpretation of symbols see Figure VI-11. (Shepard, Dill, and Von Rad, 1969) markings. Photographs of some of these structures are shown in Fig. III-5. Cores from the outermost part of the fan valley have distinct layers of plant debris.

The sand layers in the fan-valley province occur as finger-like linear sand lenses on and outside the levees and as shoe-string sands along the winding course of the valley floor (von Rad, 1968). Continuity of individual layers is lacking. A few layers can be traced with certainty, but for only a few km. A single bed of graded sand found in five closely-spaced cores had a total volume of $.24 \times 10^5$ m^3, which compares well with the amount of sediment flushed out annually from the head of Scripps Canyon (1.7×10^5 m^3) as calculated by Chamberlain (1960) and discussed by Dill (1964), and with the volume of sand moved annually by littoral longshore drift along the coast of Southern California from north to south, which Ingle (1966) has estimated to be about 1.2×10^5 m^3.

Small-scale slump structures, micro-faulted cross-lamination, and semi-consolidated clayey pebbles (called "mudballs") were found in some cores from the fan-valley (Fig. III-8B). Micro-channels 5-15 cm wide, cut into parallel-laminated sand, and later filled with homogeneous clayey silt, indicate the activity of strong, erosive bottom currents that did not subsequently deposit sandy material. Cores from the outermost part of the fan-valley contained distinct layers of plant debris.

In only two places along the axis of the fan-valley has the recent sediment cover been found to be missing or anomalously thin. Over much of the fan and trough a thin layer of clayey silt lies on the sea floor. The fine-grained sediment cover seem to indicate that the sedimentary environment in this province has been more tranquil recently than in the past. Hence, it is possible that seafloor structures could be placed safely on this part of the outer fan-valley and within the San Diego Trough, and the structure would remain essentially uneffected for many years, even though this is in the path of hypothetical turbidity currents.

In the head of the Scripps Canyon, where the slopes are much steeper, the environment appears to be considerably more dynamic.

6.3 THE HEAD OF THE SCRIPPS CANYON

The sedimentary material that fills the head of Scripps Canyon is somewhat different in composition and structure from that found on the broad, sand-covered continental shelf into which it is cut. Canyon sediments are composed of interbedded layers of organic debris and highly micaceous sands and silts; whereas the shelf sediments are relatively clean, fine-grained quartz sands. The physical composition of the different types of material entering the canyon head, which eventually form the individual layers of the fill, is dependent on (1) bottom currents, (2) length of time exposed prior to burial, (3) time of burial, and (4) position of the steep seaward face of the nearshore wedge of sand formed by material along the beach by littoral currents. Nearshore sands, once accumulated, are moved by traction and saltation towards the canyon head, primarily by swell-induced bottom currents.

The resulting canyon fill is one of an interbedded deposit with lenses of fine sand, and organic material forming a cohesive matted sedimentary deposit along the entire length of the canyon (Fig. VI-13).

In the canyon, broken debris from brown algae and surf grass accumulate, together with clastic sediments. When compacted, these materials form a springy interwoven mat containing silts and sands. In general, the organic material becomes more abundant and the layers thicker in a seaward direction.

This mixture develops a sedimentary structure of considerable internal strength. Under increasing overburden thickness, the inter-layered organic mat stretches and deforms so as to distribute the stress throughout the entire sediment body of the fill. Very heavy objects, experimentally loaded on this fill settle very slowly into the sediment, showing that stresses are redistributed slowly by internal distortion without abrupt collapse or shear failure. The composite nature of canyon fill allows it to act as a resistant sedimentary unit as long as the organic material is fresh.

After a time, this material begins to decay and lose its strength, converting slowly into an unstable mass which cannot sustain the high angle of repose developed at the time of its depostion. Therefore this initial deposition of a perishable substance contributes in large measure to a condition of future instability which will lead to slumping and eventual sudden removal of much of the fill of the canyon. Once the canyon has emptied, the cycle begins again.

Contributing to the gradual loss of stability of the sediments are the bacterial breakdown of the organic portion of the fill and loss of internal strength when the sands become mobilized by the ensuing gravity creep. As the abundant organic material intermixed with sand breaks down, it is slowly converted into a slick, highly reduced, black, gelatinous mass. During decomposition, there is continual emission of gas bubbles from the sediment fill.

Removal of sediment from the canyon head by some process of slumping is almost continual. This conclusion is based on SCUBA divers' in situ measurements of downcanyon gravity creep relative to wall rock. Three basic types of flowage apparently take place, sometimes in combination; (1) slow gravity creep analagous to that of glaciers, (2) relatively rapid but localized small-scale slumps and slides, and (3) sand flows.

The following observations concerning the dynamic removal of sediment from the head of Scripps Canyon and the relationship of this removal to turbidity currents and subsequent downslope activity in the

La Jolla Canyon have been made by Dill (1964). Moving sediment is eroding the rock walls of Scripps Canyon down to depths of at least 165 ft. (52 m). Truncated burrows of marine animals, lack of subaerial weathering products, polished rock surfaces, gouge marks, hanging valleys, overhanging cliffs, and the inclination of the rock bottoms of hanging tributaries, are all evidence of contemporary submarine erosion, and these are observed only in areas of moving sediment. Striations, drag marks, and gouge scars on the rock walls of eroded areas have the same angle of inclination as the critical angle of repose of the sediment in the channels where they are found, approximately 30°

A rich marine population develops in exposed rock areas not subject to erosion by moving sediment. The relatively narrow transition zone between the rock surfaces that are covered with and eroded by animals and plants and the undercutting surfaces that are eroded by moving sediment forms a slope on the canyon wall that is about equal to the critical angle of repose of the sediment fill.

Studies of the physical properties of the sediments, during the period January 1958 to January 1964, indicate that turbidity currents were probably not active agents in the removal of silt to sand-sized sediments from the head of Scripps Canyon. Earthquakes apparently have had little to do with failures that caused the removal of sediment from the head of Scripps Canyon. Instead, the generation of gas and the breakdown of the plant material that initially gave strength to the matted sedimentary fill cause the instability that results in gravity creep of the canyon sediments.

Gravity creep of the matted sedimentary fill is capable of moving downslope large objects such as automobile bodies, cement blocks, boulders, lobster traps, and tires. The downslope tipping of marker stakes placed in the sedimentary fill of the canyon head shows that there is differential movement within the deposit. Movements of marker stakes relative to reference points on the canyon walls show that approximately half of the movement observed in the fill is by gravity creep and internal distortion; the other half by sliding of the entire fill of the canyon along the rock floor of the channel.

The principal cause of instability leading to progressive slumping is a relatively rapid overloading (one day to two weeks) of marginally stable canyon sediments when nearshore sands are moved rapidly into the canyon head by storm-generated orbital motion of bottom water. Slumping and sliding occur along planes of weakness. The main mass of the fill retains its strength during a slump, as is illustrated by the fact that it broke a 2 -inch steel cable streched across the head of Sumner Branch.

Thus the head of Scripps Canyon provides an interesting example of contemporary dynamic filling and subsequent flushing of a local "reservoir"; whose flushing, however, is insufficient over the time scale observed, to trigger large scale evacuation of the canyon proper.

Table VI-2 shows the properties of sediments from Scripps Canyon (Dill 1964). We are witnessing a more or less steady-state filling of the upper most part of the canyon (from 200-1000 m in depth) by intermittent fillings and flushing in the steep canyon head. The canyon head processes are related to nearshore processes on the shelf, involving longshore drift of beach sands and the production of organic fill. The lower La Jolla fan-valley is today the site of periodic reworking by deep tidal currents and the gradual downslope diffusion of fine-grained sediment in tenuous suspension. Apparently the canyon is gradually filling to critical size where, somewhere along its length, one mighty slump could attain sufficiently high velocity to transform the material into a turbidity current flow. Such a major slump has apparently not occurred in the past several thousand years.

6.4 THE DEPOSITIONAL ENVIORNMENT OF AN ENCLOSED BASIN - THE TYRRHENIAN ABYSSAL PLAIN

In the Mediterranean Sea there are several enclosed basins whose deepest depressions are floored by accumulations of flat-lying sediment. The configurations of these sediment layers have been studied in cores, seismic reflection profiles, observations of submersibles, and by deep-sea photography. Although some properties of these sediments are comparable to those found in the La Jolla canyon, fan, fan-valley, and San Diego trough, there are other characteristics so markedly different as to merit description.

The principal reason for emphasizing the similarities and contrasts is that many investigators have concluded that sand and other coarse-grained sediments formed on fans and on flat abyssal plains in troughs have been transported by turbidity currents through canyons and their deep-sea channel continuations. If this is so, why would there be major differences in the texture and configuration of the depositional products?

For example, in the center of the Tyrrhenian basin there is a small abyssal plain (about 6000 km^2, - approximately the same surface area as the La Jolla Fan - San Diego Trough - San Clemente Basin complex). Several submarine canyons extend out to the abyssal plain from Sardinia, Sicily and Italy (Fig. VI-13 A). The basin topography is rugged except in local depressions where levelling and smoothing have been effected by sediment fill. The central abyssal plain is one such place. It is divided into two basins connected by a narrow abyssal gap (Fig. VI-4B). The depth of the eastern plain is 1858-1877 fm (3565 - 3603 m), slightly more shallow than the western plain. Seismic reflection profiles have shown the unconsolidated sediment beneath the centers of the two plains to be more than 1 km thick.

A prominent canyon, originating off Naples, debouches onto the northeastern corner of the eastern plain. Photographs of the floor of the canyon and observations from submersibles (Piccard and Dietz, 1961) show precipitous cliffs with outcrops of erodable semi-consolidated sediments, as well as terranes containing loose cobbles and mud balls, similar to those in the La Jolla fan valley. Where the Naples Canyon reaches the abyssal plain, photographs from the TRIESTE showed large sedimentary waves with crests 3 - 4 m apart.

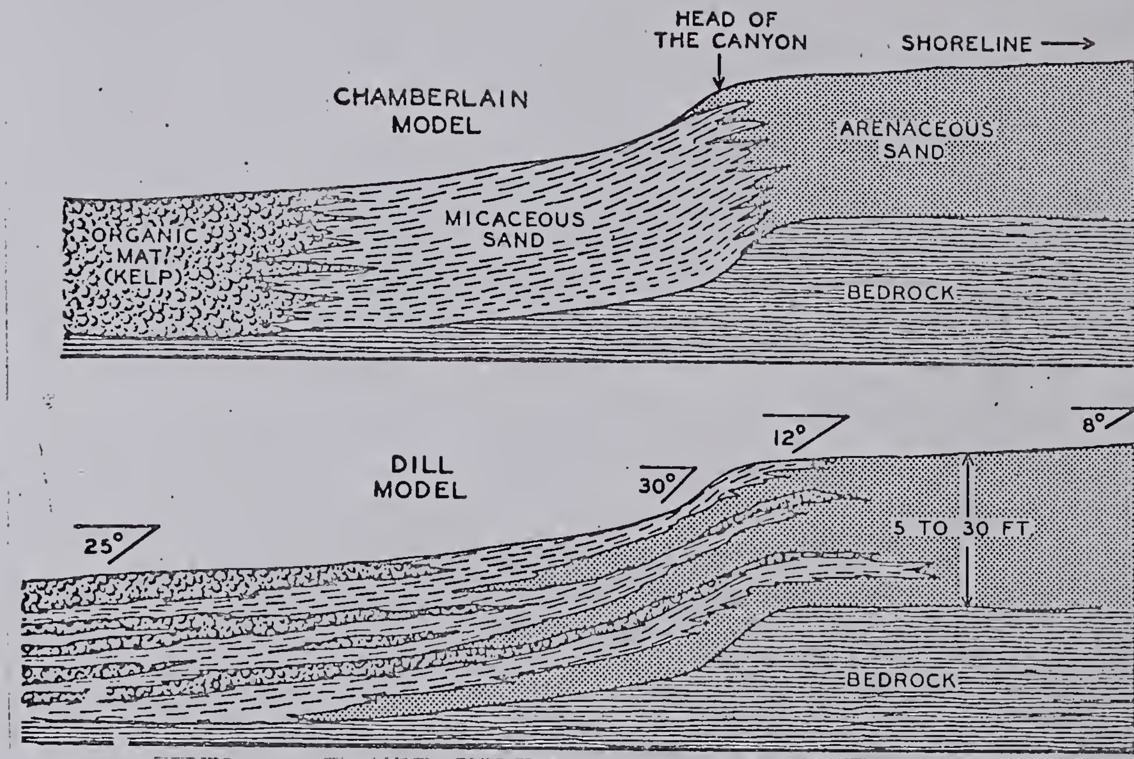


FIGURE VI-13 Sediment Models

A generalized schematic diagram by Dill of the stratigraphy of the sedimentary fill in the head of Scripps Canyon compared with the model proposed by Chamberlain (1960). The nature of the interbedded layers of sand and plant detritus throughout the entire length of the canyon head is based on observed structure exposed in slump scars (Dill, 1964, p. 93).

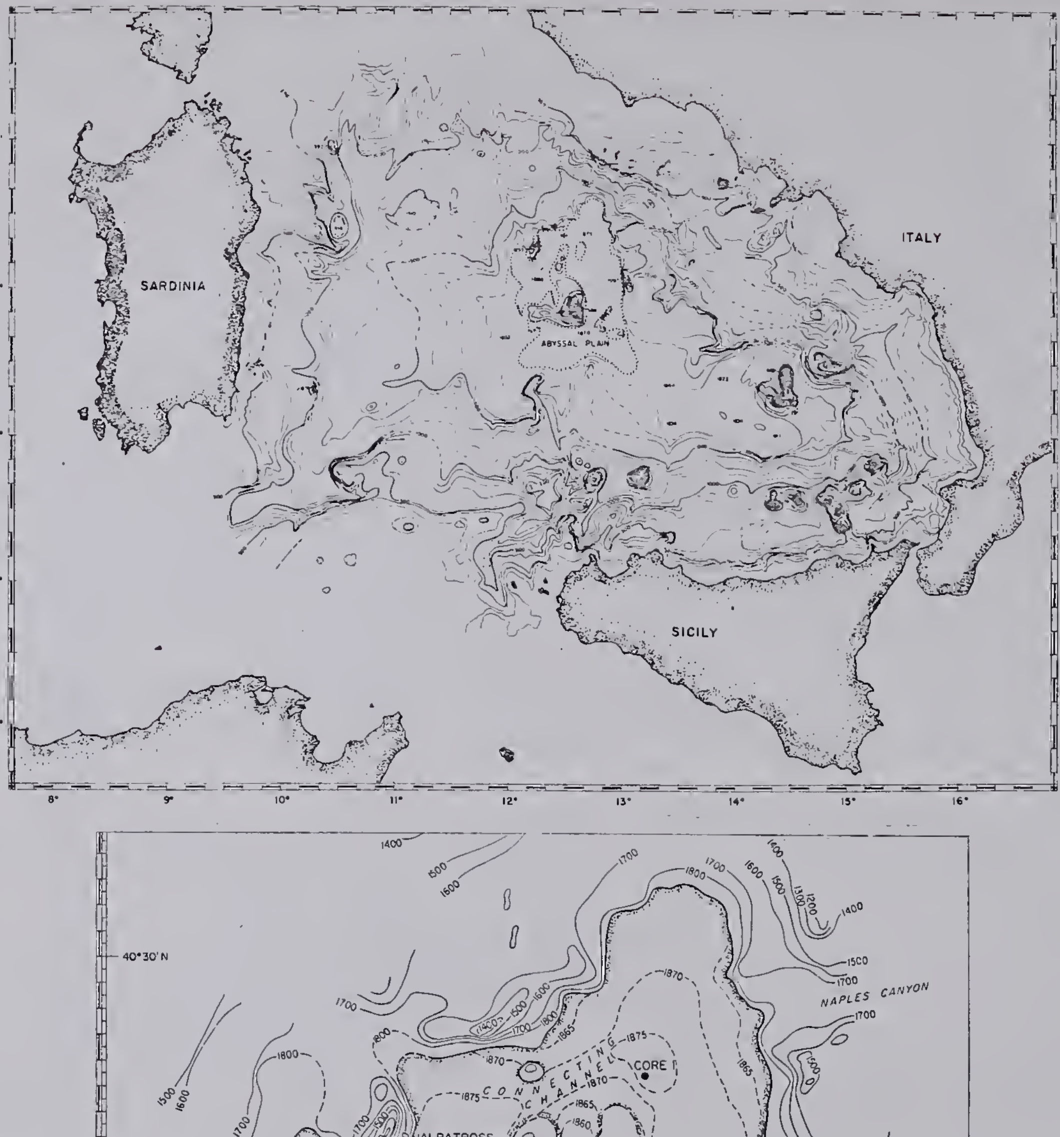


FIGURE VI-14 - Tyrrhenian Basin

Contour maps of the Tyrrhenian Basin in the Mediterranean Sea. Lower map shows an enlargement of the central abyssal plain. (Workum and Hersey, 1965)

TABLE VI-2

Comparison of observed and measured properties of sediments from the head of Scripps Submarine Canyon with those needed to develop a state of "spontaneous liquefaction" (from Casagrande, 1938; Terzaghi and Peck, 1948, p. 100-105; and Terzaghi, 1956, p. 13).

Requirements for spontaneous liquefaction	Characteristics of Scripps Canyon sediments in areas of intermittent mass movement
1. Temporary excess hydrostatic pore pressure.	Temporary excess pore pressure is possible. However, the permeability of the sediment is low enough so that there is some entrapment of gas within the sediment. The high gas content would have a tendency to cushion any sudden shocks and prevent an extremely rapid buildup of excess hydrostatic pressures except in local areas.
2. Sediment must have a metastable structure that will lose most of its internal strength when subjected to a suddenly applied shear stress.	The sand areas in the shallow heads of the canyon have void ratios that are almost equal to or slightly greater than the critical void ratio. Therefore, the change in volume at collapse would be very small and not develop large excess pore pressures if the sediments are subjected to a suddenly applied shear stress. The earthquake of June, 1963, did not cause any substantial collapse and verified at that time the lack of a metastable structure that could lead to "spontaneous liquefaction." Large heavy objects (rocks, hammers, and lead weights) placed on the surface of canyon sediments remain supported even when subjected to violent shock or strong blows. However, on slopes greater than 30°, the micaceous sediments will flow or slump when overloaded and carry the objects along with them until a stable slope is reestablished.
3. No creep movement prior to failure (movement relieves metastable structure)	Reference stakes placed in the canyon sediments show that they are slowly moving downslope and that this movement is accelerated just prior to a major slump.
4. Sediment must have low cohesion and be in the silt to sand size.	Undisturbed *in situ* vane shear measurements and laboratory vane shear, unconfined shear, and triaxial shear tests on typical canyon sediments indicate that they have a measurable cohesion. The inorganic portion of canyon sediments range in size from silt to fine sand.
5. Homogeneous in plane of failure.	The mixed origin and interbedded nature of the sediment filling all parts of the canyon preclude it from being classified as homogeneous in any plane of failure (Figures 8, 9, and 27).
6. Slope cannot exceed angle of internal friction nor angle of repose for cohesionless grains (otherwise it has cohesion).	The surface slopes of sediments accumulating in the canyon head often exceed 35° and have been observed on occasion to rest with an angle of repose of 54°. The cohesion necessary to maintain such steep slopes is derived from the sea grass and algae incorporated within the sediment mat.
7. Need triggering mechanism to start increase in pore pressure.	Triggering mechanisms are available in the form of rapidly deposited sediment brought into the heads of the canyon by rip currents during storms and periods of large swell. Earthquakes are a possible but not a common triggering mechanism (the strongest earthquake shock in 10 years had little effect).
8. Sediment must be 100% *water* saturated.	The high gas content and active bubbling observed just prior to slumps precludes the possibility of a state of 100% water saturation.
9. Spontaneous liquefaction should develop when subject to rapid buildup of pore pressure.	A series of tests on canyon sediments that had built up over a period of 6 months showed that they do not fail when subjected to rapid increases in pore water pressure. Neither jet boring nor high pressure air applied slowly or rapidly caused "spontaneous liquefaction." An increase in bubbling during the air jet tests indicated that the entire sediment fill of the canyon was affected over an area of at least 800 square feet.

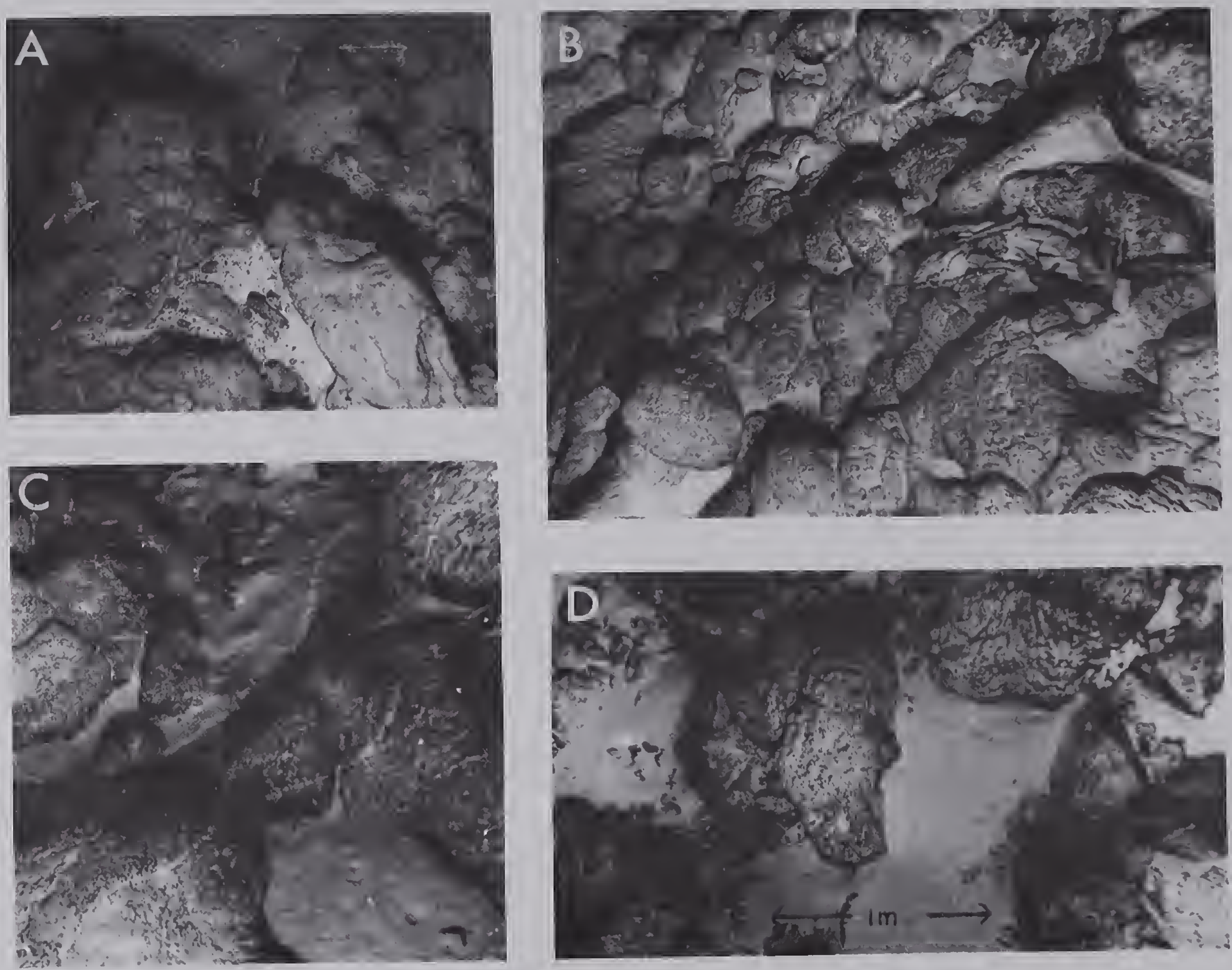


FIGURE VI-15 - Pillow lavas

Pillow lavas of the sea floor crust in the Mid-Oceanic Ridge Province. This floor might offer sites where foundations can be secured directly to the basement rocks. A. V-18-115-3; B. V-18-112-14; C. V-18-112-24; D. V-16-70-2.

Cores from the abyssal plain contain individual beds of coarse-grained clastics interbedded in lutites: nearly two dozen beds per ten meters of the sedimentary sequences. Ryan et al. (1965) demonstrated that individual beds within this sequence could be traced laterally as continuous horizons and could be correlated with identical sequences in other cores taken on both the eastern and western parts of the plain. Individual beds could be correlated on the basis of unique mineralogy, interbedded volcanic tephra layers of differenct indices of refraction, and certain indentifiable lithologies, as well as by tracing the continuity of shallow sub-bottom layering of the abyssal plain between the core sites.

Virtually all the coarse-grained layers are graded, most with complete sequences starting with the basal massive graded unit, and a thick upper pelitic section. No convolute laminations have been observed in the Tyrrhenian cores. As the base of a few of the graded beds, grooves resembling scour marks are evident. Climbing ripple lamination and cross-bedding structures are very common. The coarse fraction consists of fine sand and medium to coarse silts. Sorting is best at the extreme base of the graded intervals. The middle intervals are poorly sorted, and sorting improves towards the top. Most of the beds are bimodal, with a clay matrix as one of the dominant modes.

In addition to the vertical grading, notable and consistent changes in thickness of certain of the graded beds occur in the cores with an increase in distance from the source. A decrease in thickness of the basal graded interval is accompanied by a decrease in median grain size in samples taken just above the base contact. Curves representing this decrease in grain size with increasing distance from source, both for the Naples Canyon and the Sardinia Canyon are remarkably similar in shape for each of the different correlatable layers. In many instances the decrease in thickness of the graded interval is matched by a decrease in the thickness of the parallel-laminated interval, and by a better development of the cross-bedded interval. The pelitic interval increases with distance from source. However, in general, the entire bed shows an overall decrease in thickness with distance from the source region.

The average sand-mud thickness ratio for Tyrrhenian Abyssal Plain cores is 0.4 and the average thickness of the graded beds is about 25 cm. Correlations between cores suggest that each of the layers blankets a major portion of the abyssal plain. The large lateral dimensions of these layers indicate that turbidity currents descend the canyon along a gradient so steep that they dissipate very little energy prior to arriving on the abyssal plain.

Thus, each successive flow behaves the same hydrodynamically; for there is no equilibrium slope on a major fan or cone to provide an environment in which some of the suspended sediment can sort itself out and settle to the bottom. Instead, each flow dissipates only after arriving at the level of the plain and after dispersing its flow over a large lateral area.

In the case of the Tyrrhenian Abyssal Plain, the particular hydrodynamic realm at any site is dependent more on the local topographic setting and total transport history than on flow magnitude: whereas in the case of the La Jolla Fan and the San Diego Trough, the magnitude of the flow is important in controlling the particular forms of sedimentation at any location. On the floor of the Tyrrhenian Basin, the abyssal plain provides an equilibrium gradient for all flows that arrive at the plain, since they all have abundant energy remaining. On the La Jolla Fan, the present topographic gradient down the canyon, fan-valley, and across the San Diego Trough is apparently in equilibrium for only one flow size, and in non-equilibrium for all others.

6.5 TURBIDITY CURRENT ACTIVITY IN THE MID-OCEANIC RIDGE PROVINCE

The mid-oceanic ridge is the broadest and largest mountain range on earth. It extends as a more or less continuous feature through the Arctic, Atlantic, Indian, Antarctic, South Pacific and East Pacific Oceans, for a total distance of over 60,000 km. The ridge is a wide, fractured swell with an elevation of 1 - 3 km above the adjacent ocean basin floor. The lateral boundaries of the mid-oceanic ridge are defined by the first abrupt scarp, or abrupt gradient change, between the abyssal hills of the ocean basin floor and the mid-oceanic ridge (Heezen and Ewing, 1963).

Seismic reflection profiles made across the Mid-Atlantic Ridge (Ewing and Talwani, 1964), the Mid Indian Ridge (Ewing et al., 1969), and the East Pacific Ridge (Ewing et al., 1968) reveal that the sediment veneer is very thin. In the carbonate zone at depths of $<$ 2500 fm, the regional thickness is $<$ 100 m. Nevertheless, in many instances the local configuration of the sediment accumulation seems to be controlled by the steepness of the individual hills. Where hills are low and of long wavelength, carbonate sediments are often draped over the volcanic basement surface. In the central regions, including the rift valley, slopes and peaks are sediment-free, exposing surfaces of pillowed volcanic lava flows (see Fig. VI-15). Along many traverses in regions of high local relief, there are crossings of intermontaine basins where sediment has been ponded in flat-bottomed valleys. The flat bottom strongly suggests local transport of sediment from highs in a sufficiently fluid condition to form level surfaces (Heezen and Fox, 1966). Such transport could be achieved by locally-generated intermittent turbidity currents or by the wafting of sediment from the mountain flanks by deep sea currents. Bottom photographs taken in the crest provinces have shown ripple marks and scour marks in loose sediment drifted across lava pillow (Heezen et al., 1959)

It is not entirely clear what role turbidity currents play in the accumulation of sediment in the intermontaine basins. Van Andel and Kumar (1969) investigated layered sediments in small valleys on the western flank of the Mid-Atlantic Ridge at 22°N and found that the volume of sediment on the floor of each valley is related to the area of its catchment basin. They concluded that an average of 15 m of sediment has been transferred from each catchment area to its valley floor. Furthermore, in this region of the Mid-Atlantic Ridge the removal of sediment from hills and its redeposition in valleys are the work of sliding and turbidity currents. In one such valley (North Pond) sediment cores revealed a high frequency of recent flows.

Kagami and Ewing (in preparation) examined massive graded layers in cores from two small basins at 40°N near the base of the eastern flank of the Mid-Atlantic Ridge, and also concluded that this material was emplaced by turbidity currents of local origin. The most recent flow occurred about 700,000 years ago, and past flows apparently occurred at intervals of several hundred thousand years.

The slides and turbidity currents would effect only the downslope areas of the individual valleys, and the likelihood of contemporary activity would have to be based on examination of the individual basins. Since the flat floors will receive most of each major flow originating in the catchment basins (some graded beds intermontaine basins exceed 5 m in thickness), isolated areas on flanks or on peaks of hills might offer safe sites for structures which could be placed on or attached to exposed basement rock.

Fracture zones are extremely long, narrow strips of the ocean floor marked by anomalously deep troughs and often bounded by assymetrical ridges. The great fracture zones of the eastern Pacific were discovered early in the 1950's (Menard and Dietz, 1952; Menard, 1955), and were shown to offset large segments of the mid-oceanic ridge and ocean basin floor laterally as well as vertically (Mason and Raff, 1961; Menard, 1964, p. 64). Deep clefts in the mid-oceanic ridge have also been discovered and mapped in the Atlantic Ocean (Heezen and Tharp, 1961; Heezen _et al_., 1964), and Indian Ocean (Matthews, 1963; Heezen and Nafe, 1964; Heezen and Tharp, 1965).

Fracture zones are important features to be considered when planning sites for sea floor structures, not only because of the steep and rugged nature of the bounding escarpments, but because certain parts of these strips are loci of contemporary episodic events.

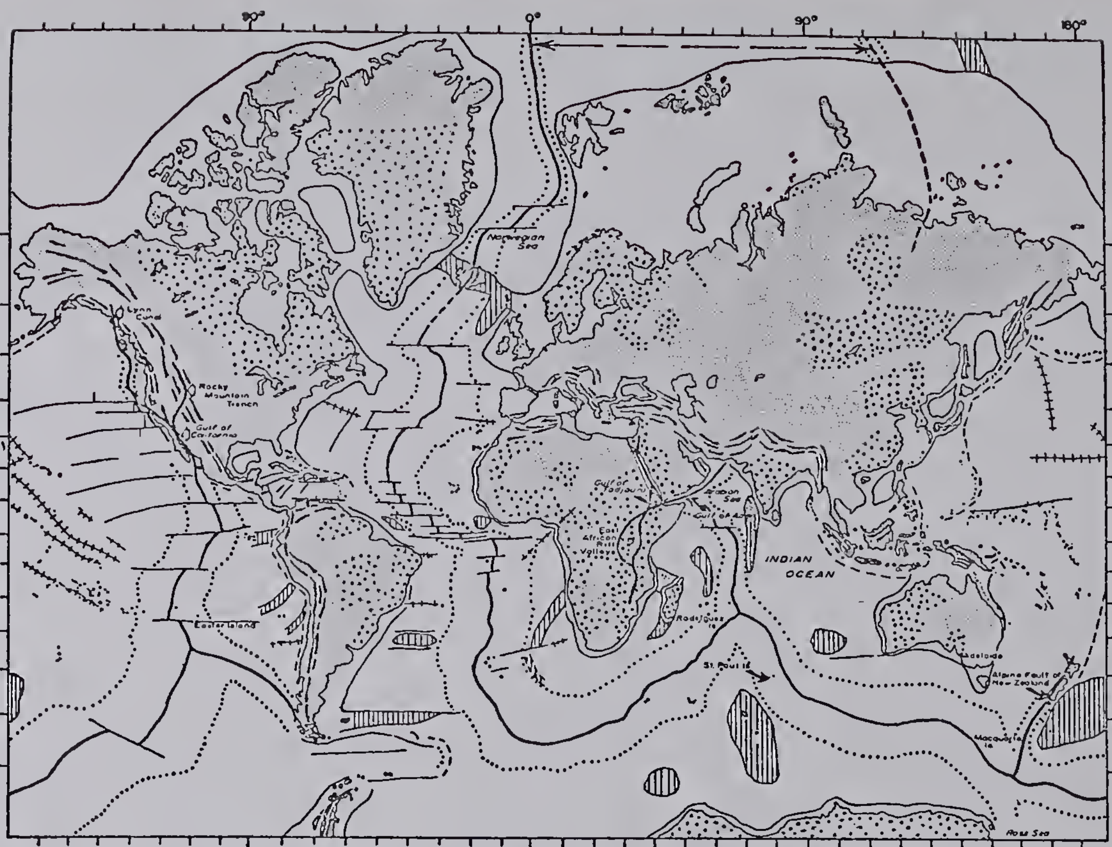
For instance, a narrow belt of seismic activity is associated with the crest of the mid-oceanic ridge. The ridge, which encircles the globe in a worldwide system, is not continuous in detail, but is interrupted and offset by the fracture zones. Nearly all the earthquakes on the ridge occur either directly along the median crest or along those parts of the fracture zones that are between crests of the ridge (Sykes, 1967). Only a very few earthquakes occur on other parts of the fracture zones, and the ocean basins are even less active seismically.

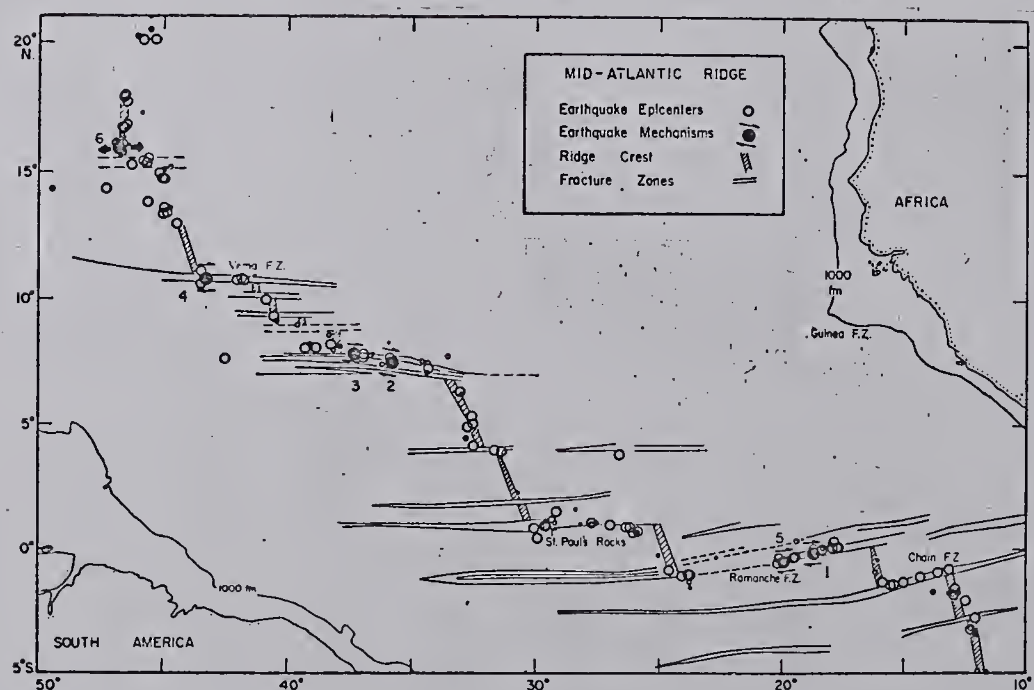
The occurrence of earthquakes predominantly in the offset portion of the fracture zone cannot be explained by active offset movements of the whole ridge crest by transcurrent faulting (see Fig. VI-16). Wilson (1965) has proposed a class of shear faults which he calls transform faults. In the transform hypothesis where the expansion of previously offset ridge crests is produced by axial accretaion (Heezen, 1960), the sense of motion of the growing sea floor is opposed only along that portion of the fracture zone which offsets the ridge crests. The transform hypothesis, which is generally accepted today as a requirement of an expanding ocean floor, implies that the great active tectonic shears on the ocean floor occur exclusively in the offset region of the fracture zones. Shearing is clearly demonstrated in metamorphosed rocks dredged from the escarpments which have undergone mylonization, brecciation, and hydrothermal heating. Structures placed within the offset region of fracture zones would be subject to numerous earthquake shocks, tilting of the sea bed, and rock falls. On some of the oceanic ridges, shearing movement along transform faults is in excess of 10 cm/yr.

The other end of the fracture zone, the extension on the ocean basin floor, is a locus for events of another kind. Particularly in the Atlantic Ocean, the deep floors of the fracture zone extensions (often $>$ 5000 m) are the sites of recent sediment accumulation from turbidity currents originating on the continental margin. For example, a fracture zone at 24°N now recieves sediment from flows travelling across the Hatteras and Nares Abyssal Plains which recently reached a sill at the distal margin of the Nares Basin. A disequilibrium sea floor gradient across portions of the Nares Abyssal Plain now allows large masses of the fine-grained portion of contemporary turbidity current flows to be funnelled into the narrow, deep fracture zone (Fig. VI-17). Similarly, a fracture zone at 8°N receives sediment which bypasses the Damarrera Abyssal Plain off the Amazon Cone.

These unusual drainage configurations imply that a major turbidity current occurring along the southeast coast of the United States might drain not only onto the Hatteras Abyssal Plain, but also arrive several days later in the 24°N fracture zone. Similarly, any flows off the Amazon Cone may eventually pond in the 8°N trough. Thus, placing an ocean floor structure in the extension of such fracture zones would increase areally over 10,000-fold the possibility of sea floor catastrophe.

We would strongly recommend that careful consideration be given to avoiding sites in portions of fracture zones that offset the crests of the mid-oceanic ridge or which lie in the drainage basin system of major abyssal plains or sedimentary cones.





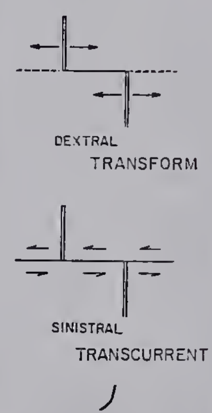


FIGURE VI-16 - <u>World-wide Mid-Oceanic Rift System</u>

A. The world-wide rift system of the Mid-Oceanic Ridge. Upper diagram shows the median axis of the tectonically active Mid-Oceanic Ridge and its offset on fracture zones. Heezen, 1963.

B. Earthquakes are confined to the rift valley, rift mountains, and offset portions of the fracture zones. Sykes, 1967.

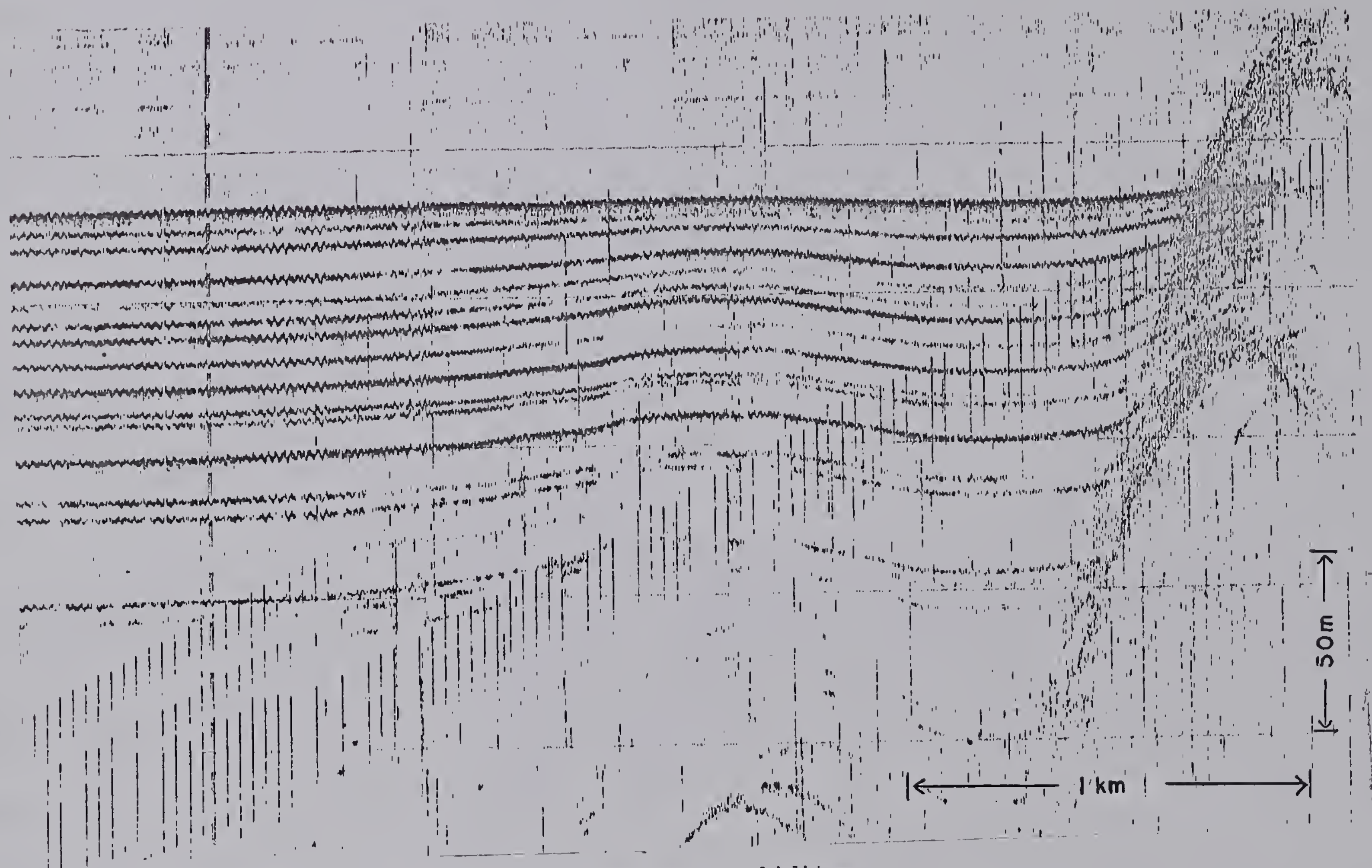


FIGURE VI-17 - Mid-Atlantic Ridge basin turbidites

Horizontally ponded turbidites in an intermontaine basin of the Mid-Atlantic Ridge, demonstrating significant turbidity current activity in this region in the past. The materials which comprise these turbidites are of local origin within the catchment basins of the particular isolated depression. R/V Kane-9, 1968. 24° N Fracture Zone.

6.6 TURBIDITY CURRENT ACTIVITY ON ARCHIPELAGIC APRONS

Menard (1956) noted that many of the large Pacific Ocean seamounts and islands have broad smooth aprons which rim their bases. The aprons appear to have been formed by accumulation. It is not known with certainty to what extent the apron is the result of extensive, very fluid lava flows, and to what extent it is a blanket of sediment eroded from the seamount and/or island. It is believed that the aprons are probably interlayered structures, with lava flows, ash, tuffs, and also debris carried down by turbidity currents.

As pointed out by Menard (1964) many of the Pacific Islands lie close together, forming clusters or ridges. Lava flows and ash falls collect between these peaks and construct linear groups, such as the Hawaiian, Marquesas, Samoan, and Society Islands. With the development of a chain or cluster of islands (or even large isolated islands) a significant secondary structural change is often noted, and the change strongly effects the subsequent sediment distribution pattern. This change is the development of an encircling moat and arch around the island, apparently resulting from a re-arrangement of magma below the upper crustal surface, in response to the massive downward pressures of the island piles themselves. The moat and related encircling arch, once formed, restrict the final resting place of the sediments as they move down from the islands.

In the case of the Hawaiian islands, their core of seamounts lies in a linear chain upon a piece of sea floor crust of Mesozoic age (Hayes and Pitman, in press). Geological investigations on the exposed peaks of the Hawaiian Islands indicate that the volcanic ridge was constructed in a sequence from west to east, with summits of seamounts emerging in the late Cretaceous. The moat at the eastern end did not appear until after the end of the Eocene (Schreiber 1969). There is as much as 1 km of sediment in the moat northeast of the ridge. If the deposition rate within the moat were constant, it would average about 1 cm/650 years. However, cores indicate extensive turbidite activity, with periods of low deposition alternating with sudden turbidite flows.

The Hawaiian Islands are, as are most of the Pacific Islands, loci of considerable seismic and tsunami activity, so that sediment in an unstable condition will be triggered into downslope motion at fairly frequent intervals. For example, in a core taken in the northeast portion of the moat north of Maui (Schreiber, 1968), more than 30 separate turbidites are seen per 10 m of deposition. All the turbidites appear to be latest Pleistocene or younger and contain reworked upper Miocene fauna. The beds are fairly thin (10-60 cm), but on many of them the contact surface indicates scour (sediment removal by the leading edge of the flow). This high frequency of turbidite activity, together with the presence of reworked older sediments, indicate the probability of a post-Miocene slope change at some point on the upper part of the ridge, - allowing vast quantities of reworked sediments to move downslope. Other cores taken in the Kauai and Maui deeps and around the eastern end of the moat region contain similar turbidites and slump breccias.

At other points around the Hawaiian Arch, sedimentation rates have varied with time and from site to site (Fig. VI-18). The seamount hills northeast of Oahu (in the moat) are complex, probably a combination of slumps, faulting, and extensive channeling. On the south side of the ridge, the moat is discontinuous, owing to the large number of seamounts; and therefore the sedimentation rate is highly variable. These clusters of seamounts block and channel the ridge sediments and have, in addition, developed their own "aprons" of sediment, ponded lavas and ashes. Between these areas the moat is filling with sediments as it is to the north and the east.

The material moving onto the slope and into the moat on the south side of the ridge is somewhat different from that on the north and east sides of the ridge. On the south side, the particles are predominantly highly altered ash, scoria, and palagonites, containing a somewhat higher precentage of $CaCO_3$ (organic) fragments than elsewhere. The sediments often appear to be welded or intergrown due to extensive alteration (in situ) to harmotome and zeolites, and they are semi-consolidated over large areas of exposure. Part of the reason for this is that surface water and winds move much of the volcanic ash directly out from the ridge as it is formed. A second reason for this difference in sediment composition is the nature of the bottom water. Water moves in from the west and southwest ("The Mid-Pacific Mountains", Hamilton, 1956) and sweeps around the ridge (and somewhat over it between Oahu and Kauai) at a fairly rapid rate, carrying many fine fragments off to the northeast (Fig. VI-18). Since the water is also somewhat warmer than usual, it allows $CaCO_3$ bearing (shallow water) sediment to remain essentially unaffected. Part of this $CaCO_3$ then appears to enter the local pore waters, causing rapid chemical alteration of the ash into a semi-consolidated mass.

An area such as the Hawaiian Ridge is not simple to describe, especially since its entire mode of origin is not yet fully understood. However, since the Pacific islands seem to be of approximately the same age, and of roughly similar origins, it can serve as a model of its class, with its encircling sediment basins and archipeligic aprons.

6.7 THE GENERAL DISTRIBUTION OF HAZARDOUS AND SAFE AREAS

Transport routes for turbidity currents are predictable, based on knowedge of the local physiographic setting and the recorded history of sedimentary deposits as revealed in cores. The continental shelves are essentially free of contemporary high-velocity flows except for a very few places where canyons have cut their way across the entire shelf. The continental slope acts as the main source terrane for the initiation of turbidity currents, and the numerous incised canyons mark the individual transport routes where turbidity current activity is concentrated.

However, many safe areas do exist on the slope. Probably the optimum sites in this province occur in intercanyon areas where seismic profiling and deep sea coring show the surface layer of loose sediment to be the thinnest, or absent. Small basement peaks are only occasionally found on the slope, and these might afford excellent sites in an

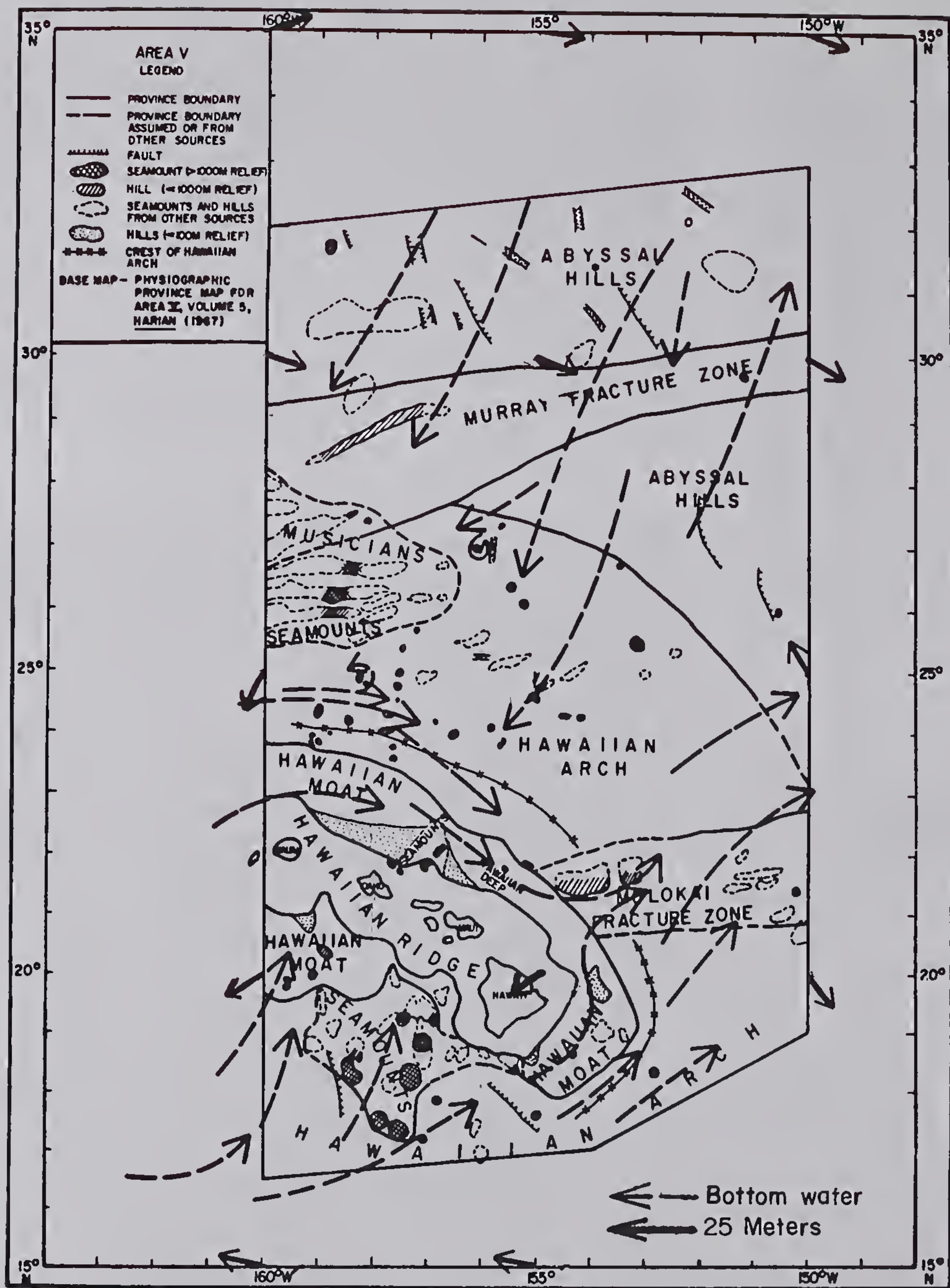


FIGURE VI-18 - Hawaiian chain province map

Approximate province map of eastern end of the Hawaiian chain, showing water current directions as developed from 20 camera stations (Schreiber, 1968 and Gurikova 1966).

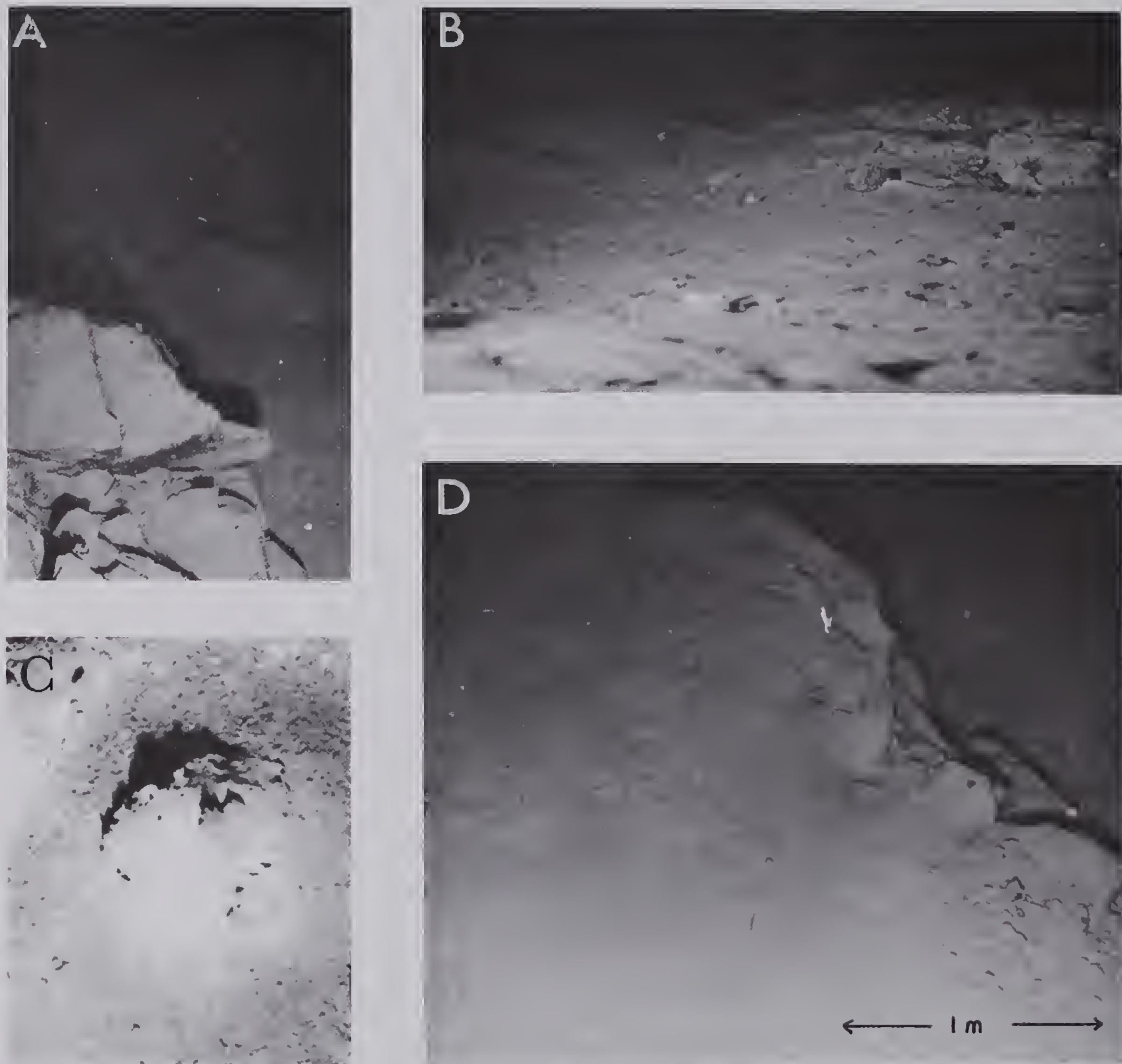


FIGURE VI-19 - Outcrops of lithified sedimentary rock on the ocean floor

Such lithified rocks (limestones, cherts, and mudstones) might provide optimum sites at which to anchor structures to the seabed.

A and D: R/V Eastward 9-66, Blake Escarpment
B: Cable Enterprise, No. 5, 90°E Ridge, Indian Ocean
C: R/V Chain-61, Mediterranean Ridge

essentially sediment-free environment. We would strongly recommend that sea floor cables be laid perpendicular to the strike of slopes so as to take the most direct route to the deep sea floor and at the same time avoid crossing canyons.

Similarly, on the continental rise or abyssal cone, any structures should be placed in inter-channel areas, preferably on small elevated peaks or protruding seamounts. The crests of lower continental rise hills, though areas of active deposition from contour currents, are still likely to be above the level of turbidity current flow, and thus should be free of the extreme turbulence associated with these intermittent high-velocity events.

Sites on the levees or within deep sea channels should generally be avoided, as these are regions where the flows are expected to be concentrated and the forces most severe. Conversely, the distal abyssal plain is to be preferred to the proximal abyssal plain, even if the net sedimentation there is slightly greater from any single flow. Many small flows may never travel across the entire abyssal plain and thus leave this site undisturbed. Within most abyssal plain provinces there occur individual protruding peaks which may rise several tens of many hundreds of fathoms above the level of the plain. Sites on these peaks are advantageous, as they are easy to locate and are completely free from gravity-propelled transport.

In the mid-oceanic ridge, crestal regions of high seismicity and volcanic activity should be avoided; and elevated mountainous regions free of flat-lying accumulations preferred. Bottom photographs in the North Atlantic show many locations on elevated peaks where volcanic lava flows cover the seabed (Fig. VI-15). Structures fastened directly to this rock foundation would offer the maximum security. Near the crestal regions the pillow lavas are generally fresh and unweathered without significant manganese crusts. However, as one approaches the flanks, basement rock may in some places become loose and friable as the result of weathering and metamorphism. Samples of the rock material would show whether or not the foundation is suitable for structures.

6.8 SPECIAL CONSIDERATIONS

In conclusion, we would emphasize that the sea floor environment presents many unknowns because it is a poorly-explored region of our planet. For designing structures and selecting sites on the seabed, it is necessary to draw on selective insights from the "state of the art" knowledge of this environment.

Two basic provinces have special considerations from the point of view of catastrophic phenomena. The first province is the general distributary province of submarine canyons, cones, fans, fan-valleys, and abyssal plains. In this province, sediments are dispersed as the result of the passage of turbidity currents. Therefore a basic approach in the selection of sites is to avoid transport routes within the distributary system. Sea floor coring can provide an excellent indication of these routes.

This province apparently has an amplifying effect. Even a small slump occurring in an uphill direction, with transport access to a selected region, can produce serious effects in the distributary province. Thus one must predict the likelihood of slumps and turbidity currents in the entire "upstream" source region.

The second basic province is that of isolated catchment basins such as the intermontaine basins of the mid-oceanic ridge, deep sea trenches, or fracture zones. In these regions the areas of potential accumulation are limited in size compared to the area of the total catchment basin. Therefore it is generally easy to predict the sites of accumulation and to avoid them. However, it is much more difficult to predict the potential source terranes of the future slumps and /or turbidity currents. In the catchment basin the problem is not so much being in the path of turbidity currents, as being proximal to the source of future slope failure.

We suggest that in the province of the distributary system, structures be located at sites where the local seabed is either isolated from, or elevated above transport routes of gravity-propelled flows. In catchment basin provinces, we suggest that sites be located where engineering studies show the slope to be stable, where sediment is thin or lacking, or where artificial means can be used to shake down and remove any suspected loose material.

As for anchoring structures, we suggest the use of cables attached to bed rock or subsurface lithified layers (Fig. VI-19), to take advantage of the inherent buoyance of the structures. The sea floor muds in the deep basins have a temperature very near freezing. We suggest feasibility studies to find out if structures could be supported on stilts which consist of thin rods cooled by refrigeration and which, in turn, would create a type of artificial permafrost beneath the foundation.

It is recommended that marine geologists work with the engineers so that they can contribute to discussion of special environmental features of the area in which any subsea structures are to be emplaced.

PART VII

SUMMARY AND GENERAL RECOMMENDATIONS

7.1 Summary of currently available information concerning the occurrences, causes, magnitudes and velocities of turbidity currents.

The downslope flow of gravity-propelled avalanches and turbidity currents is hazardous to sea floor structures within their path. Although there are large regions in the world ocean where these processes are non-operative, the safe emplacement of structures within such areas as continental margins, abyssal plains, fracture zones, and the intermontane basins of the Mid-Oceanic Ridge would require careful evaluation of the distribution, frequency of occurrence, causes, magnitudes and flow dynamics of these catastrophic events.

The aspects of the hydrologic flow regime of turbidity currents most pertinent to the emplacement of structures on the ocean floor include: mechanisms of generation, downslope velocities, turbulence of the flows, magnitudes, constitution and densities of the sediment suspensions, and the time duration of the turbidity current event. The mode of turbidity current flow suggests that the parameters of velocity, density, and thickness would vary along the path of the flow, depending on such variables as topographic slope, proximity to source, channel width (for confined flows), and general basin geometry.

The breaking and disruption of submarine telegraph cables, when not caused by a tectonic disturbance at the site of the break, provide direct indications of powerful sedimentary transport and episodic deposition. Table VII-1 briefly summarizes published evidence of turbidity current-related cable breaks and indicates the appropriate literature citations.

As this report has shown, however, the parameters we need can, in many instances, be evaluated only in light of inferences drawn from the details of sedimentary deposits. The reliability of the inferences depends to a great extent on the breadth of our knowledge of *in situ* depositional processes in the ocean, the dynamics of which are visualized by deduction from geophysical observations of the ocean floor and from direct study of sediment samples obtained through coring and drilling. Consequently, further understanding of turbidity currents will come only hand in hand with increased general understanding of all dynamical processes in the ocean. A crash program aimed at exclusive study of catastrophic events in the modern ocean will have the same shortcomings as would inferences about similar phenomena made by geologists from narrow studies of flysch deposits on land.

7.2 Immeduate use recommendations

Implementation of the following "immediate use" recommendations would add pertinent information to our state-of-the-art knowledge of dynamical processes in the ocean. The order in which the recommendations are listed indicates our view of relative priorities.

TABLE 7.1-1 Evidence of Turbidity Currents

	DATE	APPARENT CAUSE	INFERRED APPROXIMATE VELOCITY	REMARKS	SOURCE
Grand Banks and Laurentian Cone (North Atlantic)	1929	Earth tremor	12-45 knots	Cable breaks; estimated volume of transported sediments $10^{11} m^3$	Heezen & Ewing, 1952
Orleansville (Mediterranean)	1953	Earth tremor	10-40 knots	Cable breaks	Heezen & Ewing, 1955
Kadavu Passage (Fiji Islands)	1953	Earth tremor	20 knots	Twisted and abraded steel hawser; telephone cable broken, transported and buried	Heezen, 1959 Heezen & Wellman, 1962
Messina Straits (Mediterranean)	1908-1909	Earth tremor	12 knots	Cable breaks	Ryan & Heezen 1965
Tyrrhenian Abyssal Plain (Naples Canyon, Mediterranean)	-	unknown	No estimate	Extensive large sandwaves of 3-4 m wavelength, indicative of high velocities. Sediments contain many extensive, correlatable turbidite layers.	Ryan _et al._, 1965
Congo Cone (South Atlantic)	1897-1937	Sediment instability	No estimate	Apparent seasonal effects, many repeated events (est. 50/100 yr.)	Heezen _et al._, 1964
Magdalena River (Caribbean)	1935-6 1945	Sediment instability	No estimate	Estimated volume of one slump in the study, $3x10^8 m^3$. Green grass recovered in a cove 440 km from the river mouth.	Heezen, 1956

TABLE 7.1-1 continued

	DATE	APPARENT CAUSE	INFERRED APPROXIMATE VELOCITY	REMARKS	SOURCE
Alaska (North Pacific)	1908-1943	Sediment instability	No estimate	Cable breaks	Heezen, 1959 Heezen and Johnson, 1966 1969
Cascadia Channel (North Pacific)	-	Earth tremor	No estimate	Estimated volumes of slumps on the order of $5x10^3m^3$ up to 10^7m^3	Griggs, 1969
Hawaii (Pacific)	-	unknown	No estimate	At least 30 separate events in the past 15,000 years	Schreiber 1968
Monterey Deep-Sea Fan (Pacific)	-	unknown	No estimate	$10^8 - 10^9m^3$	Komar, 1969
Gulf of Corinth (Mediterranean)	1904-1957	Sediment instability	No estimate	Probably seasonal. Cable breaks, with entangled logs, branches and weeds	Heezen _et al_ 1966
New Britain Trench (Pacific Ocean)	1966 1968	Earthquake near unstable deltaic deposit	15-28 knots	Breaks occurred where cable crossed deep-sea channels	Krause _et al_ 1970

(1) Monitors should be installed on the sea bed. The first instrument packages can be very simple and inexpensive. When a land speculator seeks property, he can explore it, examine it, and survey the surrounding regions, but the people who have settled the property previously are his best source of advice concerning the frequency of earthquakes, hurricanes, floods, blights, or the prospect of a new highway development, -- all of which can affect the value of this property.

Monitors can serve the same function for those who seek to build on the ocean floor. In addition to recording the day-to-day environment, monitors can be witnesses to episodic phenomena. Furthermore, the time domain of potentially hazardous episodic phenomena is, in many instances, smaller than that which can be read from time spectral analysis of layering in deep sea cores. Consequently monitors may be the only guides to correct interpretation of recent events when making analyses of the frequency spectrum of episodic dynamical sedimentation.

Parameters to be measured by monitors might include: rate of sediment accumulation, rate of sediment reworking, tilting of the structure, earthquake activity, acoustic background signals, sea-floor currents, opacity and light scattering in bottom waters, absolute temperature and temperature variations, activity of benthic populations, and corrosion and degradation of the monitoring package itself.

(2) The rate of discharge of sediment into the ocean should be tabulated in a major data center. Such an index could include measurements of bed load, the suspended load discharge of all major rivers, longshore current drift, rates of outbuilding of deltas, rates of filling of intermediate reservoirs, and rates of organic productivity. The above information is a vital prerequisite for predicting future turbidity current activity.

(3) "Safe areas" in the ocean, where structures could be emplaced without any danger from catastrophic events should be mapped in detail. The distribution of safe areas can be assessed from the examination of deep-sea cores in major repositories, in combination with detailed topographic mapping, and continuous seismic reflection profiling. The research could be conducted either in-house with contracted access to data libraries, or could be contracted out entirely to major research institutions or groups specializing in marine geology or geophysics.

7.3 Long Range Research Projects

Since there undoubtedly will be a need at some time in the future to consider the emplacement of ocean floor structures in areas accessible to turbidity currents, more knowledge needs to be gained on the hydrologic character of sediment flows. For this purpose, planning should consider an actual full-scale experiment in the artificial triggering and *in situ* monitoring of a catastrophic event.

It is beyond the scope of the present report to explore the detail of such an experiment. However, the data return to engineers, physicists, and those concerned with waste disposal, as well as to marine scientists, is of such potential value that the planning and execution of the experiment should allow for the broadest possible cooperation between the scientific and engineering communities.

Other long range projects might include (1) the assessment of techniques for anchoring structures to bedrock outcrops, and (2) the tethering of bouyant structures to cables anchored to subbottom lithified interbeds, to bedrock or to "frozen-in" pilings. With bouyant anchoring, seabed structures could be released at the first sign of an upslope avalanche or turbidity current. Thus, they could float upwards to safety, be recovered and subsequently re-emplaced.

Under long range planning, we would also suggest surveys at trial sites for seabed structures, combined with a major analysis of relevant survey and data interpretation techniques. Since a great amount of new knowledge would be gained from such comprehensive, detailed surveys of limited areas of the ocean floor, we recommend that considerations of immediate scientific and engineering return be allowed to outweigh those of geographical priority. The first sites under consideration, if they involve expendable structures, might best be located in environments where our knowledge is the most limited. The results of both preliminary pre-emplacement surveys and the post-emplacement monitoring could thus provide an early evaluation of the magnitude of dangers from submarine avalanches and turbidity currents.

The dynamics of submarine sedimentation are varied and complex, and not yet fully comprehended. This report is but a first step towards the evaluation of potential hazards from episodic sedimentary events. Many critical questions are left unanswered. There is a need for new ideas and for more research on the deep ocean floor --- the least explored region of our planet.

SUPPLEMENT TO THE BIBLIOGRAPHY

Bouma, A. H., 1962. Sedimentology of some flysch deposits. A graphic approach to facies interpretation. Elsevier, Amsterdam, 168 pp.

Buffington, E. C., 1964. Structural control and precision bathymetry of La Jolla submarine canyon, California. Marine Geology 1 (1) p. 44-58.

de Smitt, V. P., 1932. Earthquakes in the North Atlantic as related to submarine cables (abstract): Amer. Geophys. Union Trans., p. 103-109; unpublished Western Union Company Report, 1932; revised 1936.

Chamberlain, T. K., 1960. Mechanics of mass sediment transport in Scripps Canyon. Ph.D. thesis, Univ. of California, Scripps Inst. Ocean., La Jolla, Calif. 200 pp.

Dzulynski, S., and E. K. Walton, 1965. Sedimentary features of flysch and graywackes. Elsevier, Amsterdam. 274 pp.

Ewing, J., M. Ewing, T. Aitken and W. J. Ludwig, 1968. North Pacific sediment layers measured by seismic profiling. In Knopoff, L., C. L. Drake, and P. S. Hart (Editors), "The crust and upper mantle of the Pacific Area," Amer. Geophys. Union Monograph Ser., no. 12, p. 147-173.

Ewing, M., 1948. Exploring the Mid-Atlantic Ridge. Nat. Geographic Mag., v. XCIV, no. 3, p. 275-294.

Ewing, M., and R. A. Davis (1968) Lebensspuren photographed on the ocean floor; J. B. Hersey (Editor); in "Deep-Sea Photography," Johns Hopkins Press, Baltimore, p. 259-294.

Ewing, M., J. I. Ewing, and M. Talwani, 1964. Sediment distribution in the oceans: the Mid-Atlantic Ridge. Geol. Soc. Amer. Bull., 75, 1, p. 17-36.

Ewing, M., H. Kagami, and J. Ewing , in preparation. Mirrol Trough: sediment distribution in the northeast Atlantic abyssal hill area.

Ewing, M., and R. F. Sisson, 1949. New discoveries on the Mid-Atlantic Ridge. Nat. Geographic Mag., v. XCVI, no. 5, p. 611-640.

Forel, F. A., 1885. Les ravins sous-lacustres des fleuves glaciaires. Compt. Rend., 101, p. 725-728.

Griggs, G. B., 1969. Cascadia channel: the anatomy of a deep-sea channel. Ph.D. thesis, Oregon State Univ., Corvallis.

Guilcher, A., 1963.. Continental shelf and slope. In Hill, M. N. (Editor) "The Sea," v. 3, Wiley, New York, p. 281-311.

Gurikova, Z. F., 1966. Computations of the surface and deep ocean currents in the North Pacific Ocean in summer. Okeanologiia, Akad. Nauk, SSSR, 6 (4), p. 615-631.

Hand, B. M., and K. O. Emery, 1964. Turbidites and topography of the north end of San Diego Trough, California. J. Petrology, v. 72, p. 526-542.

Hayes, D. E., and W. C. Pitman III, in press. Magnetic lineations in the North Pacific. Geol. Soc. Amer. Mem. 126.

Heezen, B. C., 1960. The rift in the ocean floor. Scientific American, 203 (4), p. 98-110.

Heezen, B. C., 1968. The Atlantic continental margin. Jour. Univ. Missouri at Rolla, 1, p. 5-25.

Heezen, B. C., and P. J. Fox, 1966. Mid-oceanic ridge, In Fairbridge, R. W. (Editor) "Encyclopedia of Oceanography," Reinhold, New York, p. 506-517.

Heezen, B. C., and A. Laughton, 1963. Abyssal plains. In Hill, M. N., (Editor), "The Sea, v. 3," Wiley, New York, p. 312-364.

Heezen, B. C., and H. W. Menard, 1963. Topography of the deep-sea floor. In Hill, M. N. (Editor) " The Sea, v. 3," Wiley, New York, p. 233-280.

Heezen, B. C., and J. E. Nafe, 1964. Vema trench: western Indian Ocean. Deep-Sea Res., 11, p. 79-84.

Heezen, B. C., and M. Tharp, 1961. Physiographic diagram of the Indian Ocean, the Red Sea, the South China Sea, the Sulu Sea, and the Celebes Sea. Geol. Soc. Amer.

Heezen, B. C., and M. Tharp, 1965. Tectonic fab ric of the Indian Ocean. Phil. Trans. A, v. 258, p. 90-106.

Heezen, B. C., M. Tharp, and M. Ewing, 1959. The floors of the oceans. 1: the North Atlantic. Geol. Soc. Amer Spec. Pap. 65, 122 p.

Houbolt, J. J. A. C., and J. B. M. Jonkers, 1968. Recent sediments in the eastern part of the Lake of Geneva (Lac Léman). Geol. en Mijnbouw, v. 47, no. 2, pp. 131-148.

Hubbert, M. K., and Rubey, W. W., 1959. Role of fluid pressure in mechanics of overthrust faulting. Geol. Soc. Amer. Bull., v. 70, p. 115-166.

Hurley, P. J., 1964. Analysis of flow in Cascadia deep-sea channel. In Miller, R. L. (Editor) "Papers in marine geology," Shepard commemorative volume. Macmillan, New York, p. 117-132.

Hurley, P. J., 1964. Analysis of flow in Cascadia deep-sea channel In Miller, R. L. (Editor) "Papers in marine geology," Shepard commemorative volume, Macmillan, N.Y., p. 117-132.

Jones, O. T., 1954. The characteristics of some Lower Paleozoic Marine sediments. Proc. Roy. Soc. (London) Ser. A., v. 222, p. 327-332.

Komar, P. D., 1969. The channelized flow of turbidity currents with application to the Monterey deep-sea fan channel. J. Geophys. Res., v. 74, p. 4544-4558.

Kullenberg, B., 1954. Remarks on the Grand Banks turbidity current. Deep-Sea Res., 1, p. 203-210.

Laughton, A. S., 1967. Underwater photography of the Carlsberg Ridge. In, Hersey, J. B. (Editor) "Deep-sea Photography," Johns Hopkins Press, Baltimore, p. 191-206.

Mason, R. G., and A. D. Raff, 1961 Magnetic survey off the west coast of North America, 32°N. latitude to 42°N. latitude. Geol. Soc. Amer. Bull., v. 72, p. 1259-1266.

Matthews, D. H., 1963. A major fault scarp under the Arabian Sea displacing the Carlsberg Ridge near Socotra. Nature, v. 198, p. 950-952.

McCrone, A. A., 1966. Thixotrophy and the mobilization of clays in slump and turbidity flows. J. Sed. Petrol., v. 36, p. 270.

Menard, H. W., and R. S. Dietz, 1952. Mendocino submarine escarpment. J. Geol., v. 60, p. 266-278.

Milne, J., 1897. Suboceanic changes. Geog. Jour., v. 10, p. 129-146; 259-289.

Moore, D. G., 1965. Erosional channel wall in La Jolla sea-fan valley seen from bathyscaph *Trieste II*, Geol. Soc. Amer. Bull., v. 76, p. 385-392.

Moore, D. G., 1966. Structure, litho-orogenic units and post-orogenic basin fill by reflection profiling, California continental borderland. Ph.D. thesis, Univ. of Groningen, 151 pp.

Nelson, C. M., 1968. Marine geology of Astoria deep-sea fan. Ph.D. thesis. Oregon State Univ., Corvallis.

Piccard, J., and R. S. Dietz, 1961. Seven miles down.. G. P. Putnam and Son, 349 pp.

Rees, A. I., U. von Rad, and F. P. Shepard, 1968. Magnetic fabric of sediments from the La Jolla submarine canyon and fan, California. Marine Geology, v. 6, p. 145-168.

Ryan, W. B. F., 1970. The floor of the Mediterranean. Ph.D. thesis, Columbia Univ., New York.

Ryan, W. B. F., 1970. The floor of the Mediterranean. Ph.D. thesis, Columbia Univ., New York.

Ryan, W. B. F., B. C. Heezen, and D. J. Hider, 1967. A meandering deep-sea channel. Amer. Geophys. Union Trans., v. 48, p. 142.

Ryan, W. B. F., D. J. Stanley, J. B. Hersey, D. A. Fahlquist, and T. D. Allen, in press. Tectonics and geology of the Mediterranean Sea, In Maxwell, A. (Editor), "The Sea, v. 4," Wiley, New York.

Schreiber, B. C., 1969. New evidence concerning the age of the Hawaiian Ridge. Geol. Soc. Amer. Bull., v. 80, p. 2601-2604.

Serruya, C., O. Leenhardt, and A. Lombard, 1966. Études géophysiques dans de Lac Léman, interpretation geologique. Arch. Sc. Geneve, v. 19, p. 179-196.

Shepard, F. P., 1954. High velocity turbidity currents, a discussion. Proc. Roy. Soc. (London) Ser. A, v. 222, p. 323-326.

Shepard, F. P., 1963. Submarine canyons. In Hill, M. N. (Editor) "The Sea, v. 3," Wiley, New York, p. 480-506.

Shepard, F. P., and E. C. Buffington, 1968. La Jolla submarine fan-valley. Marine Geology, v. 6, p. 107-143.

Shepard, F. P., and R. F. Dill. 1966. Submarine canyons and other sea valleys. Chicago, Rand McNally, 381 p.

Tolstoy, I., and M. Ewing, 1959. North Atlantic hydrography and the Mid-Atlantic Ridge. Geol. Soc. Amer. Bull., v. 60, p. 1527-1540.

Truchan, M., C. L. Windisch, G. R. Hamilton, 1967. Detailed bathymetric and seismic reflection survey of a portion of the Delgada deep-sea fan and its channel (Abstr.) Amer. Geophys. U. Trans., v. 48, p 126.

Wilde, P., 1965. Recent sediments of the Monterey deep-sea fan. Ph.D. thesis, Harvard Univ., Cambridge, Mass.

Wilson, J. T., 1965. A new class of faults and their bearing on continental drift. Nature, v. 207, p. 343-347.

Wüst, G., 1936. Schichtung und zirkulation des Atlantischen ozeans das Bodenwasser unde die Stratosphäre. Wiss. Erg. Deutsch. Atlant. Exp. "Meteor" 1925-1927, v. 6, p. 1-288.

ERRATA

Page 15. Reference 191 should be Houtz, R. E., and H. W. Wellman, 1962.

Page 45, paragraph 1, line 10. The Stoneley reference should be 1956, not 1957.

Page 74, paragraph 2, line 3. The reference should be Ryan and Heezen (1965).

Page 76, paragraph 4, line 10. The reference should be Truchan et al. (1967).

Page 97, paragraph 3, line 5. 0° .07' should read 0° 07'.

page 129, paragraph 3, line 2. The first reference should be Ewing, Ewing, and Talwani, 1964.

Page 129, paragraph 3, line 15. The reference should be Fox and Heezen, 1965.

Page 130, paragraph 2, line 1. The reference should be Ewing, M., H. Kagami, and J. Ewing (in preparation).