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Petroleum Geology and Geology of the North Sea and Northeast Atlantic Continental Margin

Proceedings of a conference held at the University of Bergen, December 1973

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Norges geologiske undersøkelse

Geological Survey of Norway

Norges geologiske undersøkelse (Geological Survey of Norway), Leiv Eirikssons vei 39, Trondheim. Telephone: national (075) 20166, international +47 75 20166. Postal address: Box 3006, N-7001 Trondheim, Norway.

Administrative Director: Dr. philos. *Knut S. Heier*

Geological division: Director Dr. philos. *Peter Padget*

Geophysical division: Director *Inge Aalstad*

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EDITOR
Statsgeolog Dr. *David Roberts*, Norges geologiske undersøkelse, P.O.Box 3006, N-7001 Trondheim, Norway.

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Petroleum Geology and Geology of the North Sea and Northeast Atlantic Continental Margin

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December 1973

Edited by
ARTHUR WHITEMAN
DAVID ROBERTS
MARKVARD A. SELLEVOLL

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Preface

The main aim of the Bergen University Conference was to bring together oilmen, university scientists and government officials having a common interest in the petroleum geology and geology of the North Sea Basin and of the North-east Atlantic Continental Margin of Britain and Norway. More than three hundred people from 18 countries attended the conference, and 22 papers were presented. Of these contributions, 18 are reproduced in full in this 'Proceedings' volume, the remaining 4 appearing only in abstract or extended abstract form.

When planning the conference we realized that six years had elapsed since a scientific conference had assembled at the University of Hull, Yorkshire (14th British Inter-University Conference, 1967) to deal with geological and geophysical aspects of exploration in the North Sea and adjacent Shelf Seas and it was thought, considering the great discoveries and advances in knowledge made by oil companies and service companies that, if these companies were not in a position to divulge geological and geophysical data about individual fields, at least they might be willing to contribute to our understanding of regional problems, most of which, because of rigorously applied confidentiality rules, could only be seen dimly through very distorted glasses.

As the papers and the diverse backgrounds of those who attended show (see List of Delegates) we succeeded; perhaps not to the degree we should have liked as far as oil and service company and government contributions were concerned, because a number of the companies and government agencies approached were not willing to present papers. This was either because of their strict application of confidentiality rules or because of commitments to present papers to other conference or elsewhere.

However, those oil companies like British Petroleum, Shell and Phillips and Service companies like Exploration Consultants Limited, who did contribute, advanced our understanding of regional palaegeographic problems by at least an order of magnitude and jointly unveiled a Sub-Upper Cretaceous North Sea rift system, well-known in the industry, which has similar dimensions to the Ethiopian and Gregory Rift Systems of Africa. This first order tectonic feature is now well and truly imprinted on the Geological Map of Europe.

In addition university and government researchers contributed to our understanding of regional problems concerned with the evolution of the Northwest European Continental Margin and provided considerable new and detailed information which has an important part in the piecing together of regional concepts.

Obviously the North Sea geological story is not going to be told in one volume and it is hoped that these proceedings will be viewed as a substantial step towards satisfying our insatiable scientific curiosity concerning the hydrocarbon-rich provinces of the North Sea Basin and the Northwest European Continental Margin.

On behalf of the organizers of the conference I should like to thank the Akademisk Kollegium of the University of Bergen for recognizing the conference officially, so providing us with many facilities: also the Geological Survey of Norway (NGU) who kindly agreed to publish the papers as one volume in their Bulletin series. A special word of thanks must also go to the Students' Union for allowing us to use their excellent conference facilities and last but not least we thank the Bergen City Corporation and the Hordaland County Council for the magnificient hospitality they provided in the form of a banquet and entertainment in Bergen's ancient Haakon's Hall. The splendour of this occasion set against a cold and snowy backdrop of Bergen's encircling arcs will long remain a memory to both overseas and Norwegian participants in the conference.

Arthur Whiteman
Chairman of the Organizing Committee

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Contributors to this Volume

K. Åm, Norges geologiske undersøkelse, Postboks 3006, N-7001 Trondheim, Norway. (Present address: Statens Oljedirektorat, Lagårdsvei 80, 4000 Stavanger, Norway.)

M. Bacon, Institute of Geological Sciences, Marine Geophysics Unit, 13/15 Braefoot Terrace, Edinburgh EH16 6AA, Scotland.

K. Bjerkli, Universitetet i Bergen, Geologisk Institutt, Avd. B, Olav Ryes vei 19, N-5000 Bergen, Norway.

M. H. P. Bott, Durham University, Department of Geological Sciences, Science Laboratories, South Road, Durham DH1 3LE, England.

T. Bugge, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

J. Chesher, Institute of Geological Sciences, Marine Geophysics Unit, 13/15 Braefoot Terrace, Edinburgh EH16 6AA, Scotland.

C. D. Cornelius, Gelsenberg Mineralöl, P.O. Box 30, 43 Essen 1, W. Germany.

A. Dalland, Universitetet i Bergen, Geologisk Institutt, Avd. A, Lars Hillesgt. 19, N-5000 Bergen, Norway.

W. Domzalski, 7, Green Lane, Watford, WD1 4NL Herts, England.

W. W. Dunn, Phillips Petroleum Company, Bartlesville, Oklahoma 74004, U.S.A.

M. V. Edwards, Norsk Polarinstitutt, Rolfstangvn. 12, 1330 Oslo Lufthavn, Norway.

K. Hinz, Bundesanstalt für Bodenforschung, 3 Hannover 23, Postfach 230153, W. Germany.

H. Holtedahl, Universitetet i Bergen, Geologisk Institutt, Avd. B, Olav Ryes vei 19, N-5000 Bergen, Norway.

J. T. Hornabrook, BP Petroleum Development Ltd., P.O. Box 48, 16 Bridge Street, Aberdeen AB9 8HE, Scotland.

D. A. L. Jenkins, British Petroleum Ltd., London, England.

A. S. Laughton, Institute of Oceanographic Sciences, Wormley, Godalming, Surrey, England.

M. Løfaldi, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

G. Maisey, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

D. Naylor, Exploration Consultants Ltd., 31/33 Grosvenor Hill, London, W1X 9HG England.

R. P. Pegrum, Exploration Consultants Ltd., 31/33 Grosvenor Hill, London, W1X 9HG, England.

G. Rees, Exploration Consultants Ltd., 31/33 Grosvenor Hill, London, W1X 9HG, England.

K. Rokengen, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

M. A. Sellevoll, Universitetet i Bergen, Jordskjelvstasjonen, Villav. 9, N-5000 Bergen, Norway.

A. Siedlecka, Norges geologiske undersøkelse, Postboks 3006, N-7001 Trondheim, Norway.

S. Siedlecki, Norges geologiske undersøkelse, Postboks 3006, N-7001 Trondheim, Norway.

F. E. Skaar, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

N. Spjeldnæs, Aarhus Universitet, Geologisk Institutt, Laboratoriet for Palaeontologi og stratigrafi, DK-8000 Aarhus C, Denmark.

E. Sundvor, Universitetet i Bergen, Jordskjevstasjonen, Villav. 9, N-5000 Bergen, Norway.

A. N. Thomas, British Petroleum Ltd., London, England.

B. Thusu, NTNF's Kontinentalsokkelkontor, Hoffsvn. 13, Oslo 2, Norway.

P. J. Walmsley, BP Petroleum Development Ltd., 16, Bridge Street, Aberdeen AB9 8HE, Scotland.

D. R. Whitbread, Cluff Oil Ltd., 13 Park Place, St. James's Street, London, S.W.1, England.

A. J. Whiteman, Universitetet i Bergen, Geologisk Institutt, Avd. A, Lars Hillesgt. 19, N-5000 Bergen, Norway. (Present address: Department of Geology, University of Aberdeen, Aberdeen, Scotland.)

P. Ziegler, Shell International Petroleum Mij., Carel van Bylandtlaan 30, The Hague, Netherlands.

List of Delegates

Aalstad, Inge, Norges geologiske undersøkelse, Trondheim, Norway
Aarseth, Inge, Universitetet i Bergen, Bergen, Norway
Aasheim, Sjur M., Universitetet i Bergen, Bergen, Norway
Aksnes, Gunnar, Universitetet i Bergen, Bergen, Norway
Albriktsen, Per, Universitetet i Bergen, Bergen, Norway
Åm, K., Norges geologiske undersøkelse, Trondheim, Norway
Angelini, G. M., Home Oil of Canada Ltd., London, England
Annoscia, Enrico, Norsk Agip A/S, Oslo, Norway
Anvik, Paul, Universitetet i Bergen, Bergen, Norway
Arcilise, Casper, Phillips Petroleum Company, Oslo, Norway
Askvik, Helge, Universitetet i Bergen, Bergen, Norway
Asphaug, Einar, Universitetet i Bergen, Bergen, Norway
Ayling, M. E., Unionoil Company of Norway, Surrey, England

Baadsgaard, Peter H., Norwegian Gulf Oil Production Co., Oslo, Norway
Baartman, I. C., Danmarks Geologiske Undersøgelse, Copenhagen, Denmark
Bacon, M., Institute of Geological Sciences, Edinburgh, Scotland
Baekeland, Roger R., Petromer Trend Corp., London, England
Bærheim, Reidar, Universitetet i Bergen, Bergen, Norway
Bailer, Gerhard, Deutsche Schachtbau- und Tiefborges., Lingen, Germany
Baker, D. G., Marathon Petroleum U.K. Ltd., London, England
Balling Rasmussen, N. P., Århus Universitet, Aarhus, Denmark
Banister, R. C., Signal Oil & Gas Co., London, England
Barrier, Jean, Esso Exploration Inc., Surrey, England
Beltrandi, M. D., Gulf Oil Corp., London, England
Bendigtsen, Bjørn, Jacob Stolt-Nielsen A/S, Oslo, Norway
Berge, Anker M., Universitetet i Bergen, Bergen, Norway
Bergesen, Knut, Universitetet i Bergen, Bergen, Norway
Bergsager, Egil, Statens Oljedirektorat, Stavanger, Norway
Berthon, M., Elf-Norge, Stavanger, Norway
Bigger, Morton Jr., Unionoil Company of Norway, Oslo, Norway
Biggins, David, Siebens Oil & Gas (UK) Ltd., London, England
Binns, Paul E., Institute of Geological Sciences, Edinburgh, Scotland
Bjerke, Tor, Universitetet i Oslo, Oslo, Norway
Bjerkli, Kristian, Universitetet i Bergen, Bergen, Norway
Bjorøy, Malvin, Universitetet i Bergen, Bergen, Norway
Bjørklund, Kjell R., Universitetet i Bergen, Bergen, Norway
Bjørlykke, Knut, Universitetet i Oslo, Oslo, Norway
Blanton, Edward, Scandinavian Sun Oil A/S, Oslo, Norway
Bobbitt, John E., Scurry-Rainbow (UK) Ltd., London, England
Bockelie, Johan Fredrik, Paleontologisk Museum, Oslo, Norway
Bøe, Reidar, Universitetet i Bergen, Bergen, Norway
Boigk, Heinz, Niedersächsisches Landesamt für Bodenforschung, Hannover, Germany
Bott, M. H. P., Durham University, Durham, England
Bowen, J. M., Shell U.K. Exploration and Prod. Ltd., London, England
Boyle, M. W., Chevron Oil Co. Netherlands, The Hague, Netherlands
Braithwaite, Philip, Sun International Inc., Philadelphia, U.S.A.
Brand, E., Winterschall AG, Kassel, Germany
Brennand, T. P., Shell U.K. Exploration and Prod., London, England
Bruland, Leif, Universitetet i Bergen, Bergen, Norway
Bruton, David L., Paleontologisk Museum, Oslo, Norway
Bugge, Thomas, NTNF's Kontinentalsokkelkontor, Oslo, Norway

Carstens, C. W., Norwegian Oil Consortium A/S & Co., Høvik, Norway
Caston, V. N. D., British Petroleum Co., Middlesex, England
Cesare, F. Di, Agip U.K. Ltd., London, England
Cessford, A. J., Ashland Oil Norge A/S, Houston, U.S.A.
Childs, F. B., Gulf Oil Corp., London, England
Christensen, Ole Bruun, Danmarks Geologiske Undersøgelse, Copenhagen Denmark
Clark, David A. C., Transocean Oil (UK) Ltd., Middlesex, England
Conoco, London, England
Cooper, E. V. H., British Embassy, Oslo, Norway
Cornelius, Carl-Detlef, Gelsenberg Mineralöl, Essen, Germany
Crawford, J. J., North Sea Sun Oil Co., London, England
Culham, W. E., Universitetet i Bergen, Bergen, Norway

Dahle, Anders, NTNF/NORSAR, Kjeller, Norway
Dalland, Arne, Universitetet i Bergen, Bergen, Norway
Dewhirst, N. H., Gaffrey Cline & Associates, Surrey, England
Dinesen, Arne, Danmarks Geologiske Undersøgelse, Copenhagen, Denmark
Discher, R., Elf-Norge, Stavanger, Norway
Dixon, Gerald E., Petroconsultants S.A., Geneva, Switzerland
Dodsworth, Godfrey, Geophysical Service International, Surrey, England
Domzalski, W., Herts, England
Dunn, W. W., Phillips Petroleum Company, Oklahoma, U.S.A.
Durica, Dusan, Czech Geological Office, Prague, Czechoslovakia
Dypvik, Bente Due, Universitetet i Oslo, Oslo, Norway
Dypvik, Henning, Universitetet i Oslo, Oslo, Norway

Edwards, M. B., Norsk Polarinstitutt, Oslo, Norway
Elverhøi, Universitetet i Oslo, Oslo, Norway
Evensen, Christian R., A/S Computas, Oslo, Norway

Færseth, Roald, Universitetet i Bergen, Bergen, Norway
Fails, Thomas, Petromer Trend Corp., London, England
Fairchild, R. E., A. P. Møller, Copenhagen, Danmark
Farestveit, Anders, A/S Geoteam, Stabekk, Norway
Feyling-Hanssen, Rolf W., Aarhus Universitet, Århus, Denmark
Finstad, K. G., Saga Petroleum A/S & Co., Høvik, Norway
Fischer, J., Gelsenberg Mineralöl, Essen, Germany
Fondeur, Madame, Institut Francais du Petrole, Paris, France
Fothergill, Colin, Tricentrol Ltd., London, England
Francis, Georg, Universitetet i Bergen, Bergen, Norway
Francken, C., Ball & Collins (Netherlands) Ltd., Wassenaar, Netherlands
Friis, Arne, J. Stolt-Nielsen, Oslo, Norway

Gammon, John B., A/S Sulfidmalm, Oslo, Norway
Gee, D. G., Sveriges Geologiska Undersökning, Stockholm, Sweden
Gee, E. R., Sussex, England
Gidskehaug, Arne, Universitetet i Bergen, Bergen, Norway
Gittinger, Konrad, Gelsenberg Mineralöl, Essen, Germany
Gjelsvik, Tore, Norsk Polarinstitutt, Oslo Lufthavn, Norway
Gjevik, Bjørn, Universitetet i Oslo, Oslo, Norway
Glasspool, K. R., Union Corp. (UK) Ltd., London, England
Goldsmith, R. W., Robertsen Research Intern. Ltd., Llandudno, N. Wales
Golenko, Nicolas, Norske Fina A/S, Stabekk, Norway
Grammeltvedt, Gudmund, Orkla Grube-Aktiebolag, Løkken Verk, Norway
Grønlie, Gisle, Universitetet i Oslo, Oslø, Norway
Gvein, Øyvind, A/S Sydvaranger, Lysaker, Norway

Haapala, P., Neste OY, Helsinki, Finland
Habarta, Al, Gulf Oil Co. Denmark, Copenhagen, Denmark
Hald, Niels, Mineralogisk Museum, Copenhagen, Denmark
Hanscomb, D. F. A., British Petroleum Co. Ltd., London, England
Hansen, Roar, Norsk Vikingolje A/S, Trondheim, Norway
Hanssen, Bjørn Haug, A/S Kongsberg Våpenfabrikk, Kongsberg, Norway
Halstead, Philip H., Statoil, Stavanger, Norway
Halvorsen, Erik, Universitetet i Bergen, Bergen, Norway
Haugland, Kristen, Universitetet i Bergen, Bergen, Norway
Heim, Johan G., Folldal Verk A/S, Vestby, Norway
Heintz, Natascha, Paleontologisk Museum, Oslo, Norway
Hernes, Ivar, Universitetet i Bergen, Bergen, Norway
Hinz, K., Bundesanstalt für Bodenforschung, Hannover, Germany
Hjelle, Knut, Universitetet i Bergen, Bergen, Norway
Hoch, Ella, Mineralogisk Museum, Copenhagen, Denmark
Høisæter, Trygve, Phillips Petroleum Company, Oslo, Norway
Høy, Tore, Universitetet i Oslo, Oslo, Norway
Hollander, Nils B., A/S Sydvaranger, Lysaker, Norway
Holliday, James A., Esso Exploration Inc., Surrey, England
Holmes, Michael, Berry Wiggins & Co., London, England
Holtedahl, Hans, Universitetet i Bergen, Bergen, Norway
Hopkinson, J. P., Norsk Hydro A/S, Wøyenenga, Norway
Hornabrook, J. T., BP Petroleum Development Ltd., Aberdeen, Scotland
Horpestad, Kåre, Universitetet i Bergen, Bergen, Norway
Howitt, F., British Petroleum Co. Ltd., London, England
Husebye, E. S., NTNF/NORSAR, Kjeller, Norway

Indrevær, Geir, Universitetet i Bergen, Bergen, Norway

Jacqué, Maurice H., Total Marine Norsk, Paris, France
Jaworski, R. K. E., Cities Service International Ltd., London, England
Jenyon, M. K., Seismograph Service Ltd., Kent, England
Johnson, R. J., Burmah Oil (North Sea) Ltd., London, England
Johnstad, Svein Erling, Universitetet i Bergen, Bergen, Norway
Jones, E. L., Mobil Exploration Norway Inc., London, England
Jones, Richard D. E., Invent Incorporated, London, England

Kanestrøm, Reidar, Universitetet i Bergen, Bergen, Norway
Kasim-Al, Faruk, Statens Oljedirektorat, Stavanger, Norway
Kelch, Hans-Jürgen, Deutsche Schachtbau- und Tiefborges., Lingen, Germany
Kilian, E. R., Texaco Production Services, London, England
Kleiven, Arna, Universitetet i Bergen, Bergen, Norway
Kolbjørnsen, Kristian, Geco A/S, Oslo, Norway
Kristjansson, Leo, University of Iceland, Reykjavik, Iceland
Kvale, Anders, Universitetet i Bergen, Bergen, Norway
Kvinnsland, Ole-Jacob, Noroil, Stavanger, Norway

Lang, Hans B., Deutsche Erdölversorgungsges., Düsseldorf, Germany
Larsen, Birger, Danmarks Tekniske Højskole, Lyngby, Denmark
Larsen, Vidar, Universitetet i Bergen, Bergen, Norway
Larssen, Erik J., A/S H. Namtvedt, Godvik, Norway
Laughton, A. S., Institute of Oceanographic Sciences, Surrey, England
Lavik, Håkon, Aftenposten, Oslo, Norway
Lee, H. K., Occidental of Britain, London, England
Lervik, Arne, Statoil, Stavanger, Norway
Lian, H. M., Union Oil Company of California, Los Angeles, U.S.A.
Liboriussen, Jørgen, A. P. Møller, Copenhagen, Denmark
Løkås, Bjørge, D.K.N.V.S. Museet, Trondheim, Norway
Løvfaldli, Magne, NTNF's Kontinentalsokkelkontor, Oslo, Norway
Løvlie, Reidar, Universitetet i Bergen, Bergen, Norway

Lohmann, H. H., Pan Ocean Oil Corp., London, England
Lorentzen, Odd R., Universitetet i Bergen, Bergen, Norway
Loughead, Eric, Hamilton Brothers Oil and Gas Ltd., London, England
Lund, Jens Jørgen, Århus Universitet, Aarhus, Denmark
Lutro, Ole, Universitetet i Bergen, Bergen, Norway
Lysberg, Bjarne, Universitetet i Bergen, Bergen, Norway

Machlaclan, D. I. L., Ranger Oil (UK) Ltd., London, England
Macintyre, Mr., Scottish Univ. Research and Reactor Centre, Scotland
Mæhle, Steinar, Universitetet i Bergen, Bergen, Norway
Maisey, George H., NTNF's Kontinentalsokkelkontor, Oslo, Norway
Mallow, I. R., Phillips Petroleum Company, Oslo, Norway
Mangerud, Jan, Universitetet i Bergen, Bergen, Norway
Marås, Nils, NTNF/NORSAR, Kjeller, Norway
Martyr, J. W., British Steel Corporation, London, England
Maycock, Ian D., Zapex UK Ltd., London, England
McGirk, D. D., Texaco Production Services, London, England
McMichael, L. B., Chevron Petroleum UK Ltd., London, England
McNab, George, Signal Oil & Gas Co. (UK), London, England
Meadows, Mr., Hunting Geology and Geophysics Ltd., Herts, England
Mellingen, Thv., NTNF's Kontinentalsokkelkontor, Oslo, Norway
Miller, T. J., Arco Oil Producing Inc., London, England
Misje, Per, Kåre Misje & Co., Bergen, Norway
Mitra, Rabindranath, Burmah Oil Trading Ltd., Wilts, England
Moehl, W. H., Amoco Norway Oil Company, Oslo, Norway
Morris, D. B., Hunting Geology and Geophysics Ltd., Herts, England
Munck, Fredrik, Universitetet i Bergen, Bergen, Norway
Myers, K. G., Phillips Petroleum Company, Oslo, Norway
Myhre, Lars A., Universitetet i Bergen, Bergen, Norway

Næss, Ole Edv., Universitetet i Bergen, Bergen, Norway
Nagy, Jenø, Universitetet i Oslo, Oslo, Norway
Navrestad, Torstein, Universitetet i Bergen, Bergen, Norway
Naylor, D., Exploration Consultants Ltd., London, England
Naylor, W. V., Esso Exploration Norway Inc., Stavanger, Norway
Neuman, Bjørn E. E., Universitetet i Bergen, Bergen, Norway
Nilsen, Hans R., Universitetet i Bergen, Bergen, Norway
Nilsen, Nils, Universitetet i Bergen, Bergen, Norway
Nordberg, Harald, Norsk Polar Navigasjon A/S, Trondheim, Norway
Norling, Erling, Sveriges Geologiska Undersökning, Stockholm, Sweden
Nysæther, Eigill, Norsk Hydro A/S, Oslo, Norway

Oddy, Mark E., Invent Incorporated, London, England
Ofstad, Kaare, Universitetet i Oslo, Oslo, Norway
Oldroyd, R. W. L., Robertson Research International Ltd., Llandudno, N. Wales
Opheim, Arne, A/S Norcem, Oslo, Norway
Ormaasen, Else, Universitetet i Oslo, Oslo, Norway
Orr, A. K., Marathon Petroleum Co. (UK) Ltd., London, England
Osland, Lidvin M., Tromsø Museum, Tromsø, Norway
Osmaston, Miles F., Surrey, England

Padget, Peter, Norges geologiske undersøkelse, Trondheim, Norway
Pascoe, D. J. C., Superior Overseas Development Co. Ltd., London, England
Pattinson, M. H., Ranger Oil (UK) Ltd., London, England
Pedersen, Asger Ken, Mineralogisk Museum, Copenhagen, Denmark
Pedersen, Lars M., Universitetet i Bergen, Bergen, Norway
Pennington, T. J., Hamilton Brothers Oil and Gas Ltd., London, England
Penttila, W. C., Arco Oil Producing Inc., London, England
Perch-Nielsen, K., Institut for historisk geologi og palæontologi, Copenhagen, Denmark

Pick, M. C., Columbia Gas Development of Canada, Alberta, Canada
Potter, H. C., Apco Oil (UK) Ltd., London, England
Power, P. E., Unionoil Company of Norway, Surrey, England
Prestvik, Tore, Norges Tekniske Høgskole, Trondheim, Norway
Provan, Donald, Stavanger Museum, Stavanger, Norway

Raaen, Knut Chr., Norsk Teknisk Byggekontroll A/S, Høvik, Norway
Raknerud, Lars, Samkjøringen av Norges Kraftverk, Oslo, Norway
Randle, Clive A., Bow Valley Exploration (UK) Ltd., London, England
Rasmussen, Johannes, Færoya Jardfrødisavn, Torshavn, Færøerne
Rendall, Hans, Universitetet i Bergen, Bergen, Norway
Reuter, Fritz W., Exploration Logging Int. Inc., Berkshire, England
Rhys, G. H., Institute of Geological Sciences, Leeds, England
Roaldset, Elen, Universitetet i Oslo, Oslo, Norway
Roberts, David, Norges geologiske undersøkelse, Trondheim, Norway
Roberts, Jeffery, Wood Gundy Ltd., London, England
Roberts-Jones, David Ian, Southern Geophysical Consultants, Bjärred, Sweden
Robie, Robert H., Arco Norway Inc., Oslo, Norway
Robins, Brian, Universitetet i Bergen, Bergen, Norway
Rochet, Jean, B. E. I. C. I. P., Paris, France
Rodwell, J. R., Shell UK Exploration & Production Ltd., London, England
Rønnevik, Hans Chr., Statens Oljedirektorat, Stavanger, Norway
Rong, Olav, Universitetet i Bergen, Bergen, Norway
Rui, Ing., Universitetet i Oslo, Oslo, Norway
Rydberg, Sten, Norsk Hydro A/S, Oslo, Norway
Rygg, Einar, Universitetet i Bergen, Bergen, Norway

Sæbø, Asbjørn, Norsk Polar Navigasjon A/S, Trondheim, Norway
Sæland, Gunnar T., Statens Oljedirektorat, Stavanger, Norway
Saemundsson, Kristjan, National Energy Authority, Reykjavik, Iceland
Samuelson, Lennart, Sveriges Geologiska undersøkning, Gothenburg, Sweden
Sanness, Torstein, Saga Petroleum A/S & Co., Høvik, Norway
Schmeitzner, M., Oljeprospektering A. B., Stockholm, Sweden
Scott, Jim C., Pacific Petroleums Ltd., Alberta, Canada
Sellevoll, Markvard A., Universitetet i Bergen, Bergen, Norway
Seymour, Michael, Sunningdale Oils Ltd., London, England
Sheldon, J. H., NCB (Exploration) Ltd., London, England
Sheridan, D. J. R., Marathon Petroleum Ltd., Dublin, Ireland
Siedlecka, A., Norges geologiske undersøkelse, Trondheim, Norway
Siedlecki, S., Norges geologiske undersøkelse, Trondheim, Norway
Simpson, Dag, Andresens Bank A/S, Oslo, Norway
Sindre, Eyvind, Universitetet i Bergen, Bergen, Norway
Skaldebø, O. A., Universitetet i Bergen, Bergen, Norway
Skarpnes, Oddvar, Universitetet i Bergen, Bergen, Norway
Skoglund, Roland, Sveriges Geologiska Undersøkning, Stockholm, Sweden
Sletten, Kirsten, Universitetet i Bergen, Bergen, Norway
Smith, C. D., Mobil North Sea Ltd., London, England
Smith, Glenn C., MESA (UK) Ltd., London, England
Sørensen, Aimar, Runit, Trondheim, Norway
Sørnes, Anders, Universitetet i Bergen, Bergen, Norway
Solli, Magne, Universitetet i Bergen, Bergen, Norway
Spinnangr, Ånon, Universitetet i Bergen, Bergen, Norway
Spjeldnæs, N., Århus Universitet, Aarhus, Denmark
Steele, Terry, Ivory & Sime, Edinburgh, Scotland
Stenersen, Birger, Kåre Misje & Co., Bergen, Norway
Stevens, George R., Universitetet i Bergen, Bergen, Norway
Stewart, Francis, Norse Petroleum A/S, Bergen, Norway
Stobart, John E., MESA (UK) Ltd., London, England
Stokely, P. W., Sun Oil International Inc., Philadelphia, U.S.A.

Stokstad, Johan O., Oceanic Exploration Company, Oslo, Norway
Stoltenberg, Kr., North Sea Exploration Services, Stavanger, Norway
Storetvedt, Karsten M., Universitetet i Bergen, Bergen, Norway
Sturt, B. A., Universitetet i Bergen, Bergen, Norway
Sund, Tore, Statoil, Stavanger, Norway
Sundvor, Eirik, Universitetet i Bergen, Bergen, Norway
Svela, Per Terje, Det Norske Oljeselskap A/S, Oslo, Norway
Sverdrup, Tor L., A/S Sydvaranger, Lysaker, Norway
Sylvester, Arthur G., Universitetet i Bergen, Bergen, Norway
Syrstad, Erik, Universitetet i Bergen, Bergen, Norway

Tansø, Asbjørn, Amoco Norway Oil Company, Tananger, Norway
Ten Dam, A., Texfel Petroleum Corporation, Paris, France
Thiede, Jørn, Universitetet i Bergen, Bergen, Norway
Thomas, J. B., Amoco (UK) Ltd., London, England
Thomassen, Thomas Chr., Den norske Creditbank, Oslo, Norway
Tonning, Svein A., Universitetet i Bergen, Bergen, Norway
Torske, Tore, Norges geologiske undersøkelse, Trondheim, Norway
Torvanger, Olve, Saga Petroleum A/S & Co., Høvik, Norway
Tryti, Johannes, Universitetet i Bergen, Bergen, Norway
Tveten, Einar, Mineralogisk Museum, Oslo, Norway
Tysseland, Magne, Universitetet i Bergen, Bergen, Norway

Ursin, Bjørn, Runit, NTH, Trondheim, Norway

Vaagen, Øystein, Universitetet i Bergen, Bergen, Norway
Vigran, Jorunn Os, Universitetet i Trondheim, Trondheim, Norway
Villar Del, Jesus Ramirez Copeiro, Empressa Nacional Adaro, Madrid, Spain
Vokes, F. M., Universitetet i Trondheim, Trondheim, Norway

Waagstein, Regin, Danmark Geologiske Undersøgelse, Copenhagen, Denmark
Waering, Erik K., A/S Fred. Olsen, Oslo, Norway
Waites, J. D., Seiscom-Delta, Kent, England
Wallace, Robert, Fairey Surveys, Berks, England
Walmsley, P. J., British Petroleum Co. Ltd., Aberdeen, Scotland
Ward, J. A., Burmah Oil (North Sea) Ltd., London, England
Weder, H., Tyske Konsulat, Bergen, Norway
Westre, Steinar, Universitetet i Bergen, Bergen, Norway
Whitbread, D. R., Cluff Oil Ltd., London, England
Whiteman, Arthur, Universitetet i Bergen, Bergen, Norway
Whittaker, Alun H., Exploration Logging Intern Inc., Berkshire, England
Williams, J. J., Occidental of Britain Inc., London, England
Woollands, M. A., A/S Norske Shell Exploration & Production, Tananger, Norway
Wycherly, W. A., Chevron Oil Netherlands, The Hague, Netherlands
Wyszynski, Z. S., Norges Tekniske Høgskole, Trondheim, Norway

Zdenek, Roth, Central Geological Survey, Prague, Czechoslovakia
Ziegler, P., Shell International Maatschappij, The Hague, Netherlands
Ziegler, W. H., Esso Europe Incorp., Surrey, England
Zimmerle, Winfried, Deutsche Texaco AG, Hamburg, Germany

PART I

North Sea; Petroleum Geology, Geology and Geophysics

The Geological Evolution of the North Sea Area in the Tectonic Framework of North Western Europe

P. A. ZIEGLER

Ziegler, P. A. 1975: The geological evolution of the North Sea area in the tectonic framework of North-Western Europe. *Norges geol. Unders. 316*, 1–27.

The development history of the North Sea area can be subdivided into five stages: 1) Caledonian geosynclinal stage (Cambrian-Silurian) 2) Variscan geosynclinal stage (Devonian–Carboniferous 3) Permo-Triassic intracratonic stage 4) Jurassic–Cretaceous taphrogenic rifting stage 5) Tertiary post-rifting intracratonic stage.

The Jurassic-Cretaceous North Sea central rift system is related to the rifting processes in the Arctic North Atlantic. During the Tertiary this sector of the Atlantic entered the drifting stage. At the same time the North Sea rift system became inactive, leading to regional subsidence. Emplacement of the late Tertiary Rhône-Rhine rift system postdates the central North Sea rift.

P. A. Ziegler, Shell Internationale Petroleum Maatschappij B. V., Carel van Bylandtlaan 30, The Hague, Netherlands

Introduction

From Cambrian to Recent the North Sea area underwent a complex geological evolution during which it formed part of different tectonic provinces and sedimentary basins.

From the viewpoint of basin development we distinguish the following stages in the evolution of the North Sea area:

1. Caledonian geosynclinal stage (Cambrian–Silurian)
2. Variscan geosynclinal stage (Devonian–Carboniferous)
3. Permo-Triassic intracratonic stage
4. Taphrogenic rifting stage (Jurassic–Cretaceous)
5. Post-rifting intracratonic stage (Tertiary)

The superposition of these basins in various combinations in the different areas controls the hydrocarbon potential of the respective sectors of the North Sea.

The geological history of the North Sea area can only be fully understood when viewed against the broad background of the tectonic evolution of NW Europe. The aim of this paper is therefore to retrace with the aid of a sequence of paleogeographic maps in a kaleidoscopic fashion the main development stages of the North Sea within the framework of NW Europe.

Geological information obtained during the recent exploration efforts in the North Sea is integrated with data available from the onshore areas. For the latter the author has drawn heavily on the voluminous compilation of literature dealing with the paleogeographic evolution of Europe. The paleogeographic maps cover a large area and span large time intervals, necessitating much generalisation and simplification. These maps have not been palinspatically corrected; facies provinces are therefore distorted and crowded to various degrees in areas that have been subjected to compression during orogeneic periods. It is hoped that despite these shortcomings the maps convey a ready outline of the main development stages of NW Europe in general and of the North Sea in particular. It should be noted here that a common legend to the paleogeographic maps is presented at the end of the paper, as Fig. 19.

Thanks are due to SIPM, to ESSO Europe and to the Dansk Undergrunds Consortium for releasing this paper for publication. The help and constructive criticism of many of my colleagues are herewith acknowledged.

Caledonian geosynclinal stage

The late Silurian paleogeographic setting (Fig. 1) as summarised by Walter (1972) provides a good starting point for our discussion of the evolution of the North Sea area. This setting precedes the Caledonian orogeny but is eugeosyncline and geanticline. Denmark, Sweden and the Baltic formed part

During this time NW Europe formed part of the Appalachian-Caledonian geosyncline and of the eastward adjacent shelf areas. Ireland, northern England, the northern North Sea and Norway were occupied by the Caledonian eugeosyncline and geanticline. Denmark, Sweden and the Baltic formed part of the Caledonian miogeosynclinal realm. For the Silurian, Walter (1972) postulates emergent areas for southern Norway and parts of the southern North Sea and adjacent Germany. For lack of adequate exposures the paleogeographic framework of Central Europe during this period cannot be fully reconstructed. Although sediments of the Caledonian geosynclinal stage have not yet been reached by the drill in the North Sea their potential significance should not be underestimated since this sequence contains the Cambrian Alum shales, an excellent source rock exposed in southern Sweden (Schlatter 1969).

Variscan geosynclinal stage (Devonian to Carboniferous)

The Caledonian orogeny resulted in the fusion of the North American-Greenland and the North-West European continental masses (Wilson 1966) with the Caledonian fold belt crossing the northern North Sea. In Central Europe the Caledonian orogeny caused an accentuation of the Alemanic-Bohemian geanticline and with this a sharper definition of the Variscan geosyncline.

A speculative connection between the Caledonids of Norway and those of southern Poland along a trend paralleling the Tornquist line (Shurawlew 1965, Fig. 2) has been postulated by von Gaertner (1960) and Znosko (1964) but was later questioned by Franke (1964, 1968). For lack of a sufficiently complete sedimentary record in Denmark this problem cannot be solved at this time. The Tornquist line that marks the boundary between the stable Fenno-Sarmatian platform and the more mobile Western European areas can be recognised as a major structural feature during much of the post-Caledonian history of NW Europe.

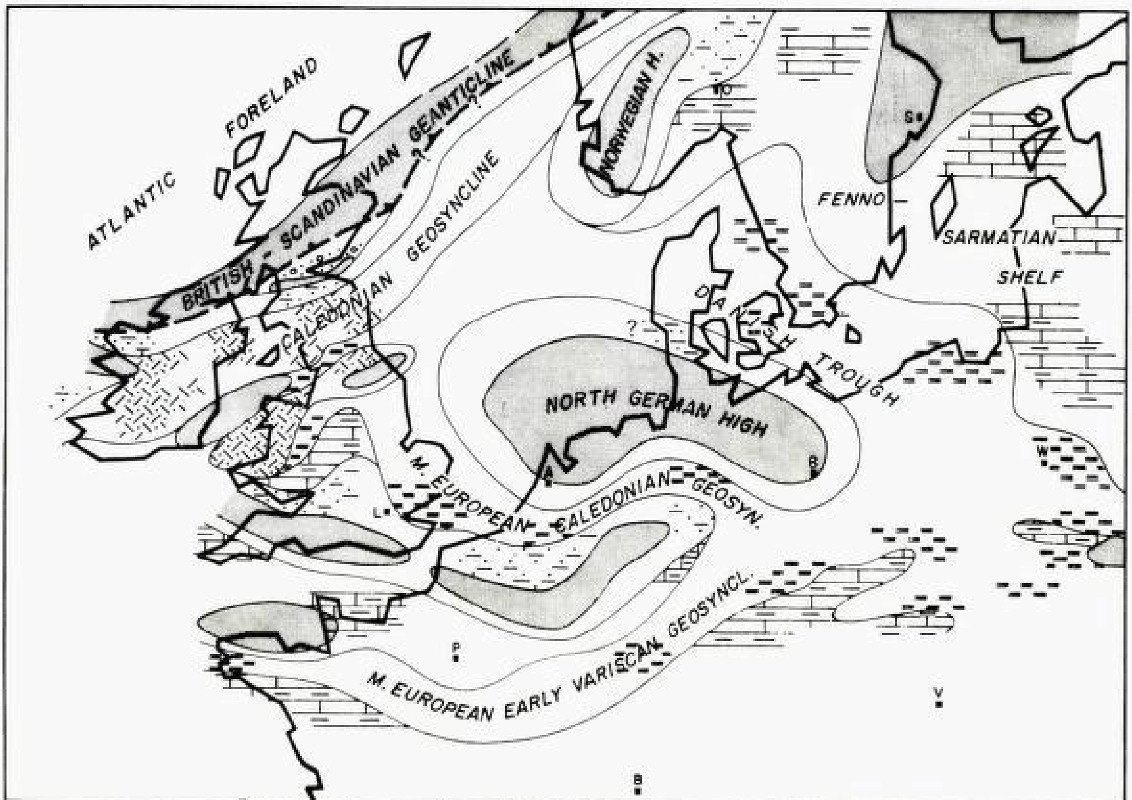


Fig. 1. Paleogeography of NW Europe during the Silurian (Wenlockian) mainly after Walter (1972).

As a result of the Caledonian orogeny the tectonic framework of the North Sea area obtained a new polarity (Fig. 2): the picture was now dominated by the Variscan geosyncline in the south while the Caledonian mountain ranges to the north were rapidly degraded. Late to post-orogenic uplift, associated with a partial collapse of the Caledonian mountain system, resulted in the deposition of the thick, in part lacustrine and bituminous series of the Devonian Old Red sandstone in intramontane basins such as the Orcadian and the Caledonian Cuvette (Bennison & Wright 1972, Allen et al. 1967). A connection between these basins and the scattered outcrops of Old Red sandstone on the Norwegian coast (Nilsen 1973) cannot be ruled out. Old Red equivalent sandstones occur also in the Baltic (Pajchlowa 1970) and have been encountered in a few boreholes in the central North Sea. Marine ingressions from the Variscan geosyncline, which was flanked to the north by a wide, in part reef-bearing, carbonate shelf, reached as far as the central North Sea. Control points are still too sparse to reconstruct details of the Devonian paleogeographic framework of much of NW Europe. However, the London–Brabant massif appears for the first time as a major stable block, a role which it continued to play during much of the subsequent development of the North Sea area.

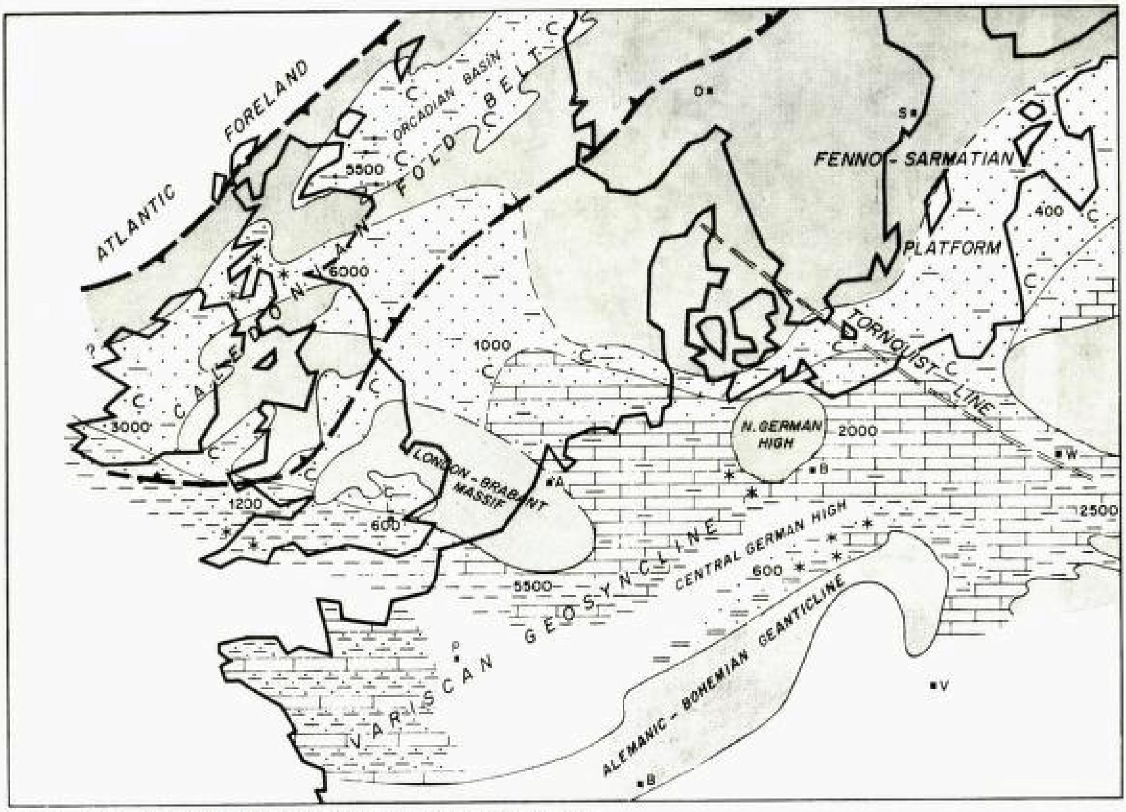


Fig. 2. Paleogeography of NW Europe during the Devonian.

The late Devonian–early Carboniferous Bretonic orogeny resulted in a consolidation of the Alemanic–Bohemian geanticline and in the emergence of the Armorican–central German highs. During the *Lower Carboniferous* (Fig. 3) these highs formed the source of the thick, flysch-like Culm series that were deposited in the Variscan foredeep. During the Visean its distal northern parts were occupied by a wide carbonate shelf which extended from the Irish Waulsortian reef platform to Poland. Marine ingressions reached far to the north into the progressively degraded Caledonian chains. Visean coal-bearing sequences were locally deposited in the central North Sea and northern England. Of special interest is the thick, essentially non-marine Oil Shale sequence in the Scottish Midland valley (MacGregor 1948; Bennison & Wright 1972). Based on the sparse control available, the northern and eastern parts of the North Sea appear to have formed part of a large, positive area during much of the Lower Carboniferous.

At the turn from the Lower to the Upper Carboniferous the Sudetic orogeny led to a further consolidation of the Variscan mountain system with deposition continuing in intramontane successor basins. During the *Upper Carboniferous* deposition of the marine Culm series was restricted to a narrow trough flanking the rising Variscan mountain chains. During the Namurian, marine ingressions

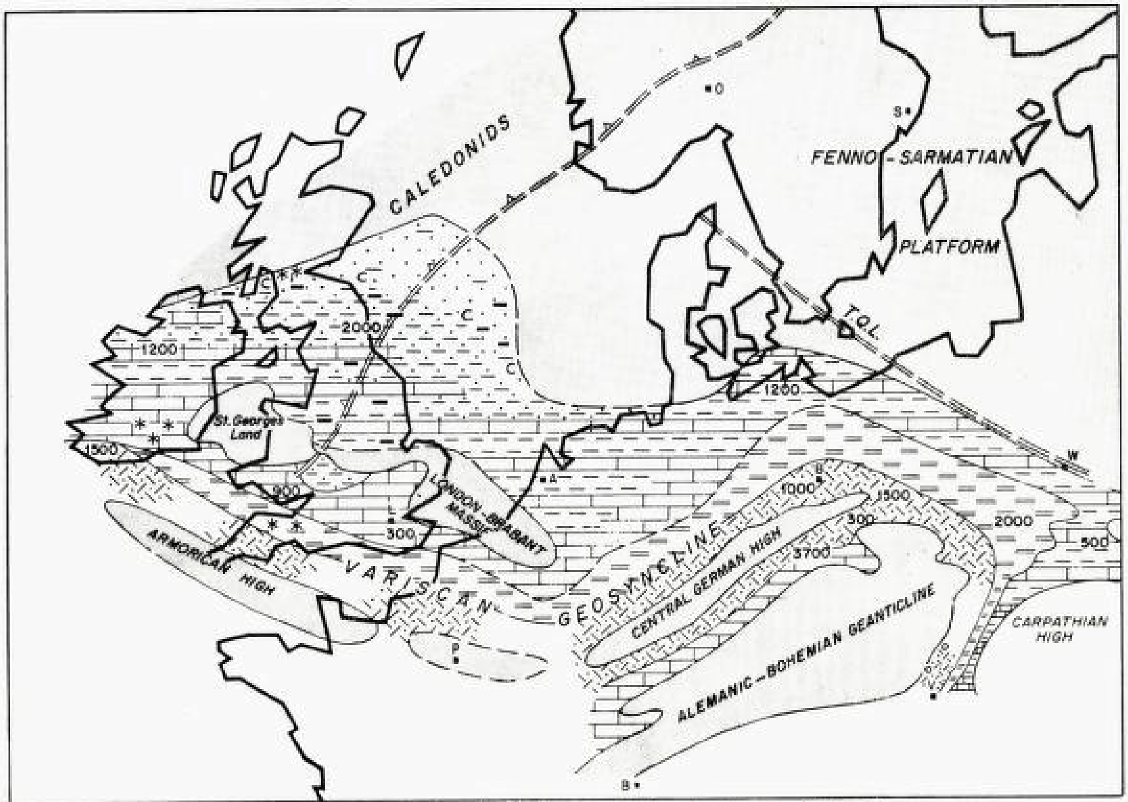


Fig. 3. Paleogeography of NW Europe during the Lower Carboniferous.

originating from the Variscan foredeep inundated much of the Baltic and north German shelf but also reached northern England and Scotland (Yordale and Limestone Coal Group).

Fig. 4 depicts the paleogeographic set-up of NW Europe during the Westphalian. Paralic conditions prevailed in much of the Variscan foredeep and the northward adjacent shelf areas, leading to the deposition of very thick coal-bearing sequences. These are of particular economic interest, especially since they constitute the source rocks for the gas found in the southern North Sea and the adjacent Netherlands and German onshore area (Patijn 1964). These Upper Carboniferous coal-bearing sequences do not appear to extend into the central and north-eastern North Sea.

Permo-Triassic intracratonic stage

The late Carboniferous Asturian and the early Permian Saalian orogenic phases (Variscan orogeny) resulted in décollement folding of the Variscan foredeep and a final consolidation of the Variscan chains with the north European craton. The Permian tectonic framework (Figs. 5 and 6) is characterised by the formation of large post-orogenic intracratonic basins with deposition

Fig. 4. Paleogeography of NW Europe during the Upper Carboniferous (Westphalian).

of red beds and evaporites. During the Triassic a new megatectonic setting came into being as a set of narrow rifts and grabens. This tectonic phase can be related to the collapse and subsidence of the Variscan domain resulting in the establishment of the Tethis and the emplacement of the Arctic–North Atlantic rift system. The Permo–Triassic period is marked by the change from the predominantly compressive Variscan tectonic setting to a period of extension and collapse, the full polarity of which only became evident during the Jurassic and Lower Cretaceous.

During the *Permian* the Variscan fold belt was subjected to post-orogenic uplift leading to its partial collapse and the establishment of continental intramontane basins (Falke 1972) very reminiscent of the post-Caledonian Old Red intramontane basins. This process was accompanied by volcanic extrusions. Similarly, the undeformed Variscan foreland was initially subject to uplift, tilting and erosion, followed by differential subsidence along normal faults, which was accompanied by the widespread effusion of the Lower Permian volcanics. Progressive subsidence led to the establishment of the Upper Permian Rotliegend basins (Fig. 5). The mid-North Sea–Fyn–Grindsted high appeared as the major positive element separating the less well known northern Permian basin from the southern Permian basin.

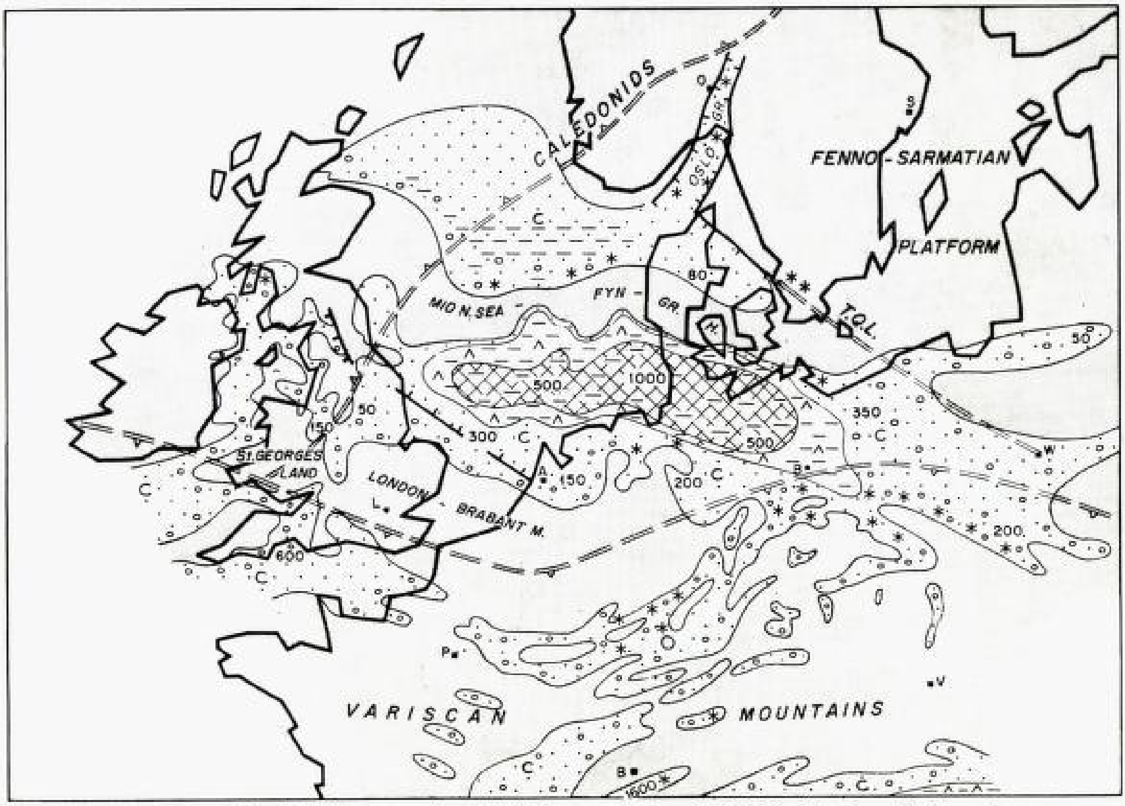
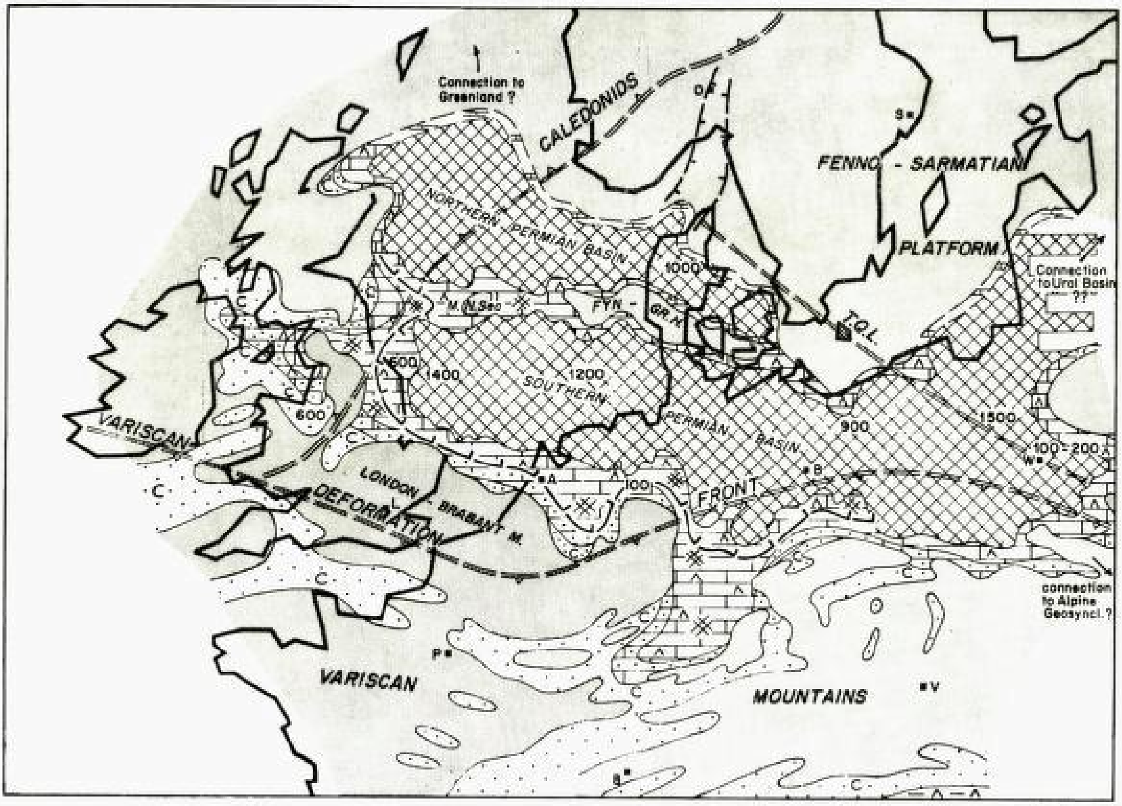


Fig. 5. Paleogeography of NW Europe during the Middle Permian, Rotliegendes.

Variscan Germano-type deformations are postulated by Franke (1967, 1968) for the Polish–Danish furrow, with the Tornquist line forming the boundary between the stable Fenno–Sarmatian platform and the rapidly subsiding intracratonic Permian basins that also encroached on the Variscan fold belt. In the southern Permian basin Rotliegend dune sands (Glennie 1972) are the primary gas reservoir in the southern North Sea and the adjacent Netherlands and German onshore areas. These sands grade northward into Sebka shales and evaporites. The latter gave rise during the Mesozoic to salt diapirism in the German Bight and the adjacent onshore areas.

The configuration of the northern Permian basin is less well known, and for want of sufficient well data no reliable facies pattern can as yet be drawn up. A significant element of the northern Permian basin is the volcanic Oslo graben (Wurm 1973).

Continued subsidence of the arid Rotliegend basins, possibly below sea-level, resulted finally in the catastrophic ingression of the Zechstein seas, the origin of which is still open to speculation (Fig. 6). No concrete evidence has been obtained to date for a connection between the northern Permian basin through the northern North Sea to the marine Permian series of Greenland (Maync 1961) and Spitsbergen (Harland 1961), nor has the possible connection via



Limit Zechstein 2 – Salt

Fig. 6. Paleogeography of NW Europe during the Upper Permian, Zechstein.

the Moravian Gate to the marine Permian of the eastern Alpine geosyncline been established. The least likely link to Permian seas is via the Russian platform to the Ural foredeep and in a roundabout way to Spitsbergen. However, the clearly marine character of the Z–1 and Z–2 carbonates (Füchtbauer 1962) as well as the large amount of evaporites contained in both the southern and northern Permian basins leaves little doubt that these basins had at least a narrow connection to the open seas. The deposition of thin shelf carbonate and sulphate sequences was restricted to the margins of the Zechstein basins. These were offset by thick, basin-filling halite sequences that reached thicknesses in excess of 1000 m.

Zechstein carbonates form a significant gas and oil reservoir in onshore areas but play a subordinate role in North Sea hydrocarbon prospects. Diapirism of the Zechstein salts, both in the southern and northern Permian basins, strongly influenced post-Triassic sedimentation.

The *Triassic* period meant for much of the North Sea area a return to a continental depositional regime. The Permian structural pattern still dominated the paleogeography of northern Europe. However, the emplacement of new graben systems resulted in significant modifications (Fig. 7). These grabens probably formed in conjunction with early movements along the Arctic North Atlantic rifting zone (Hallam 1971).

It is likely that the Central Graben system of the North Sea was established during the Triassic; however, at this time it had not yet developed into a dominant structural feature. Similar rapidly subsiding Triassic grabens in the North Sea area are the Horn and Glückstadt grabens in the Danish offshore and northern Germany, respectively, and the Danish–Polish furrow which is bounded to the east by the Tornquist line. Similar grabens developed in the Celtic Sea and the Western Approaches and along the Atlantic seaboard of Scotland and Ireland. Normal to the northern margin of the Brabant–Rhenian massif the Emsland and Weser depressions formed during the Bunter. Their flanking highs were actively uplifted and subjected to erosion, as illustrated by the Hardegsen unconformity (Wolburg 1961, 1962). Progressive downwarping and widening of the northern and southern Permo–Triassic basin resulted in the gradual burial of the mid-North Sea–Fyn–Grindsted High. Further degradation of the Variscan mountains and progressive subsidence of the intramontane basins brought about a link-up between the north European basin and the Tethian basin (Boselli & Hsü 1973). During the Röt time marine incursions reached the southern North Sea via the Moravian Gate (Polish furrow).

The Muschelkalk transgression entered into the North European basin through the Hessian Depression as well as the Moravian Gate (Tokarski 1965). During the Triassic no marine connections existed between the North Sea and the marine areas of north-eastern Greenland (Birkelund 1973) and Spitsbergen (Parker 1967; Defrentin–Lefranc 1969; Harland 1973; Cod & Smith 1973).

In these parts of the North European basin, where thick Zechstein salt deposits coincided with Triassic depôt centres, halokinetic movements were triggered during late Triassic times. These movements continued to influence at least on a local scale the depositional pattern of the Jurassic, Cretaceous and Tertiary.

In the future Alpine geosynclinal domain the erosion and collapse of the Variscan mountain system led to the deposition of the thick, graben-bound Permian Verrucano that is frequently associated with volcanic extrusion (Trümpy 1965, 1972). Permian marine series are known from the southern Alpine facies realm only (Rau & Tongiorgi 1972). During the Triassic these graben systems were enlarged, leading to a general marine transgression and the deposition of thick, carbonate sequences in the Eastern and Southern Alpine facies realm (Trümpy 1971). From this incipient Alpine geosyncline (Tethis basin), marine transgressions reached northward into the Alpine foreland.

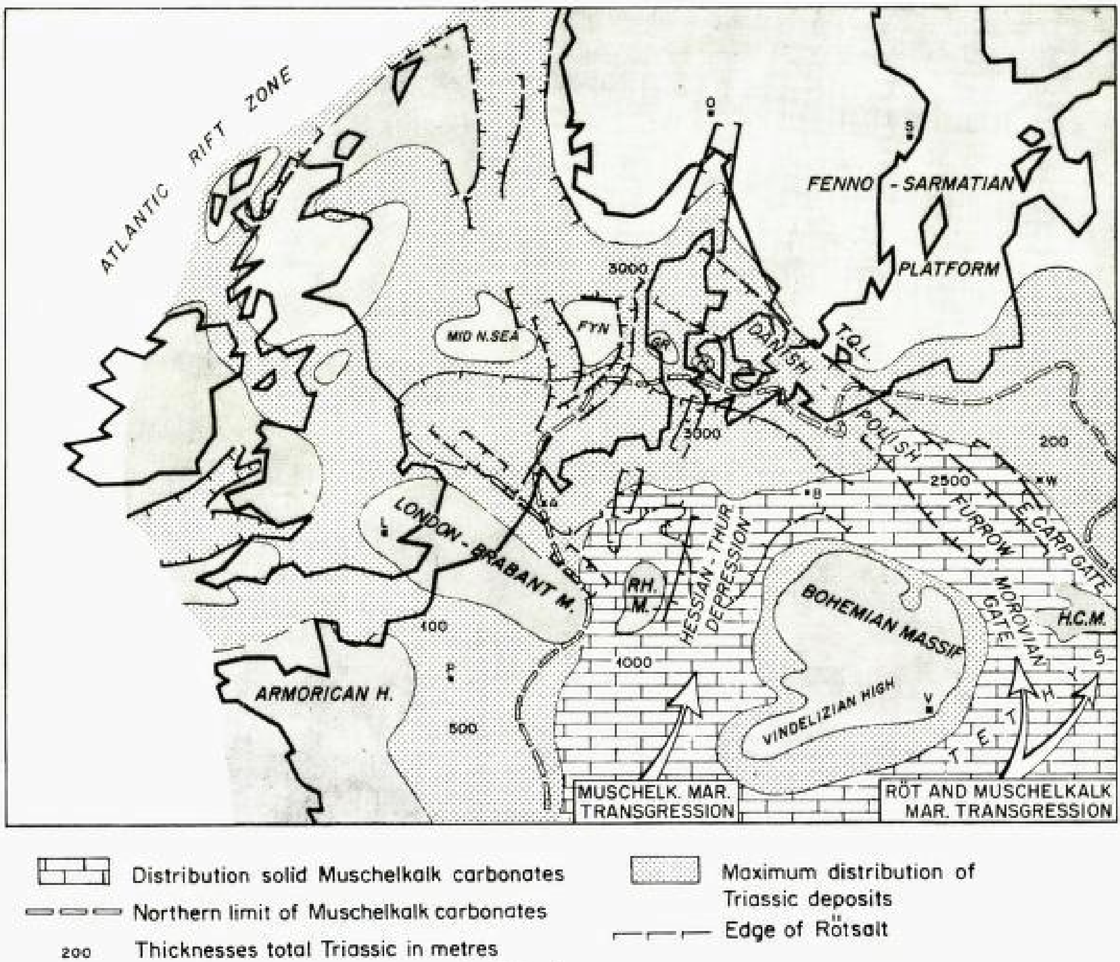
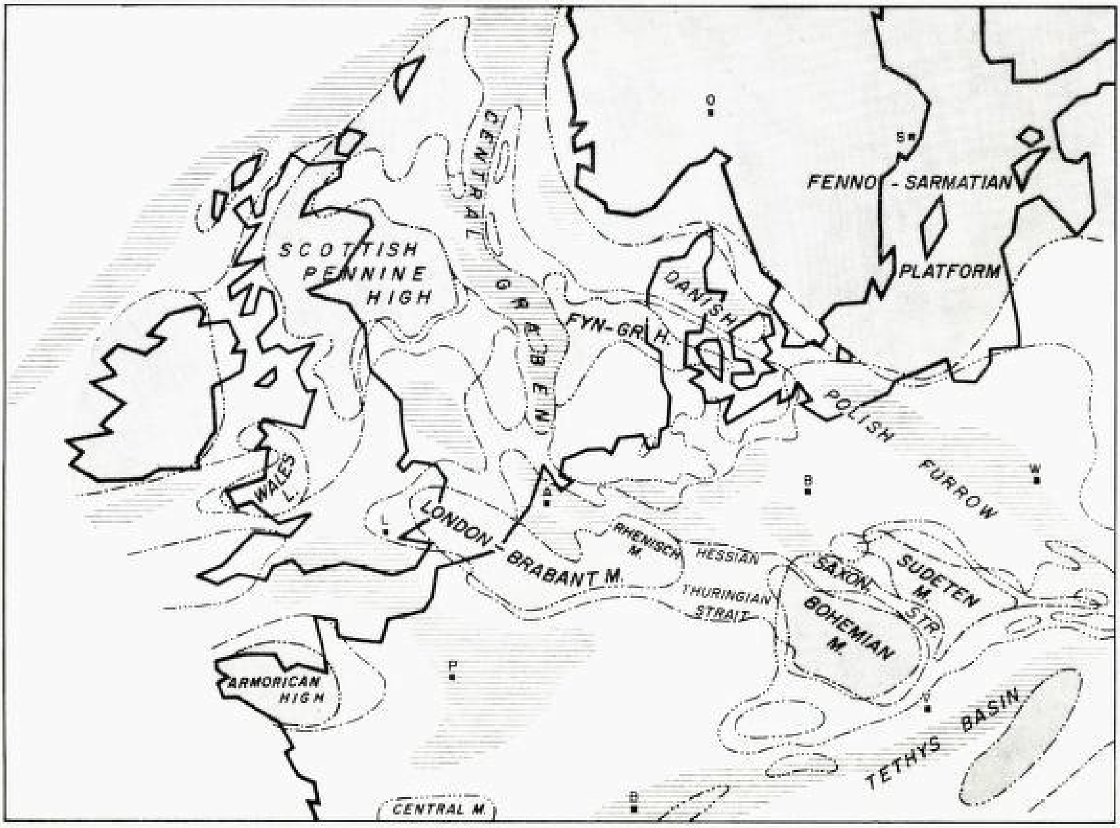


Fig. 7. Paleogeography of NW Europe during the Triassic.

Taphrogenic rifting stage (Jurassic-Cretaceous)

The late Triassic topography of NW Europe was characterised by an extremely low relief (Trümpy 1971). The Early Kimmerian (Rhaetian) movements marked the transition from the Triassic depositional framework to the Jurassic sedimentary pattern (Rusitzka 1967, 1968). With the Liassic transgression, marine conditions returned to large parts of NW Europe. Marine ingressions into the North Sea area originated chiefly from the newly established North Atlantic Seaway (Hallam 1971) through the northern North Sea, but also from the Tethis via the Paris basin and southern England, via the Hessian–Thuringian depression and in the Toarcian also through the Moravian Gate (Köbel 1968).

Fig. 8 presents an outline of the Jurassic framework of NW Europe, indicating areas of maximum total Jurassic subsidence. A comparison of the approximate depositional edges of the Liassic, Dogger and Malm conveys



APPROXIMATE EDGES OF JURASSIC BASINS IN NW-EUROPE

Liassic
Dogger
Malm
Jurassic depot centres

Fig. 8. Approximate edges of Jurassic basins in NW Europe.

an impression of the paleogeographic changes that occurred in the general North Sea area during the Jurassic. Main features of the Jurassic paleogeography are:

a. Opening up of the Arctic North Atlantic seaway during the Pliensbachian and its continued widening during the Middle and Upper Jurassic (Hallam 1971); the North Atlantic remained, however, in a pre-drifting stage.
b. The establishment of the North Sea Central Graben system as the dominant rift system. The Horn and Glückstadt grabens became largely inactive. The Danish–Polish furrow continued, however, to subside rapidly.
c. Flanking the London–Brabant–Rhenish–Ardennes massif as well as the Bohemian massif, marginal troughs (so-called 'Randtröge', Vogt 1962) developed. During the Dogger and Malm the narrow Silesian strait transected the Bohemian massif in a NW–SE direction.

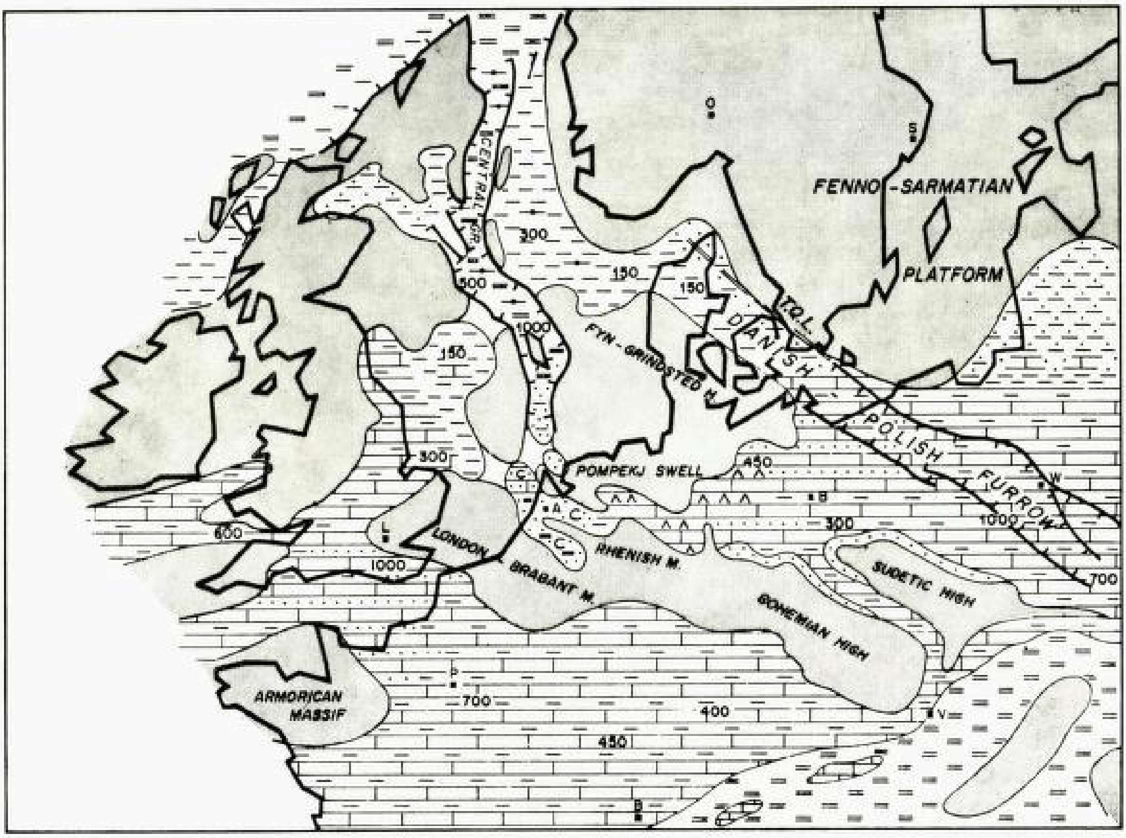
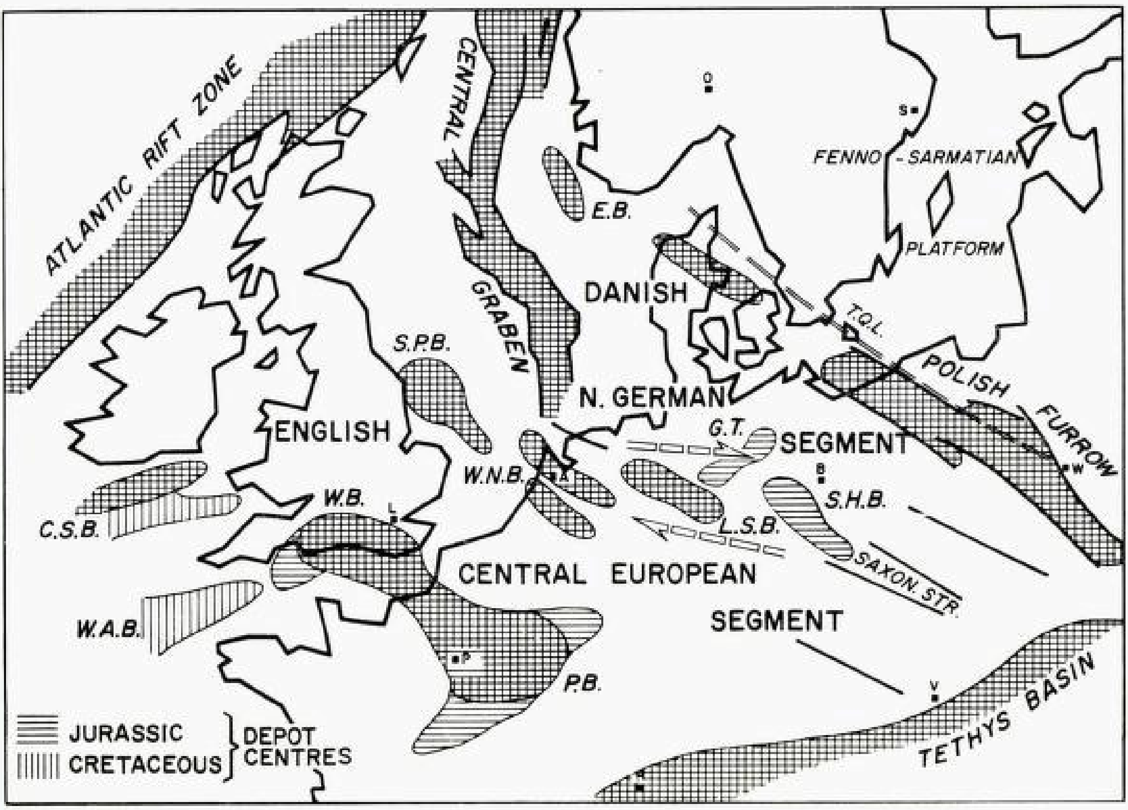


Fig. 9. Paleogeography of NW Europe during the Upper Jurassic.

d. The Alpine geosyncline reached a 'Mediterranean paraoceanic' pre-orogenic stage during the Jurassic with eugeosynclinal (oceanic) troughs separated by more stable platform areas (Trümpy 1971). Subsidence of large areas along flexures and normal faults is documented from the northern margin of the geosyncline and is indicative of an extensional stress setting (Trümpy 1960; Köbel 1968).

The overall impression gained is the one of NW Europe being subjected as a whole to extensional stresses. This led to its partial fragmentation whereby the North Sea central rift and the Polish furrow represent the major fracture zones in the Alpine foreland (Fig. 9). The North Sea Central Graben system should be considered as an offshoot of the Arctic North Atlantic rift system. The Variscan massifs appear to have prevented a southward extension of the North Sea graben system. Instead a series of en échelon tensional depressions developed mainly along their northern margin. Emplacement of these marginal troughs probably followed pre-existing fault patterns and was possibly in response to deep-seated transform movements between the Danish–North German segment (North) and the Central European–English segment (South). The Tornquist line formed the boundary between the stable Fenno–Sarmatian



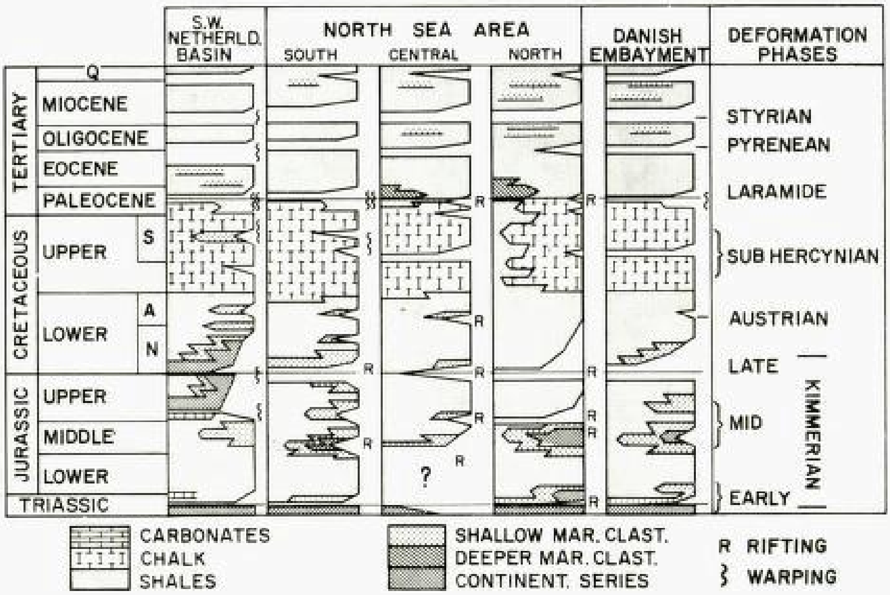
JURASSIC–CRETACEOUS TECTONIC FRAME WORK OF NW–EUROPE

C.S.B.	Celtic Sea Basin	*S.P.B.*	Sole Pit Basin	*G.T.*	Gifhorn Trough
W.A.B.	Western approaches Basin	*W.N.B.*	West Netherlands Basin	*S.H.B.*	Sub-Hercynian Basin
W.B.	Wead Basin	*L.S.B.*	Lower Saxony Basin	*E.B.*	Egersund Basin
		P.B.	Paris Basin		

Fig. 10. Jurassic–Cretaceous tectonic framework of NW Europe.

platform to the east and the metastable blocks to the west, the southern boundary of which was formed by the Alpine geosyncline.

The Jurassic record of the North Sea is not yet well enough known to develop a comprehensive story of the evolution of its Central Graben system at this stage. Halokinesis as well as several stages of downfaulting and differential subsidence of the fragmented graben floor, coupled with uplifting of the rift margins, make the deciphering of events rather difficult. In the northern North Sea there is clear evidence of continuous differential subsidence of the graben floor since the Triassic and well into the Jurassic. A major rifting phase locally corresponding to an angular unconformity within the Dogger is documented from the central North Sea. In the southern and northern North Sea this phase is expressed by regressive–transgressive clastic cycles often in concordance with the underlying marine Liassic sequences.



STRATIGRAPHIC DIAGRAM NORTH SEA AREA

Fig. 11. Stratigraphic diagram of the North Sea area.

A second phase of rifting, this time better documented in both the southern and northern North Sea, occurred during the transition from the Middle to the Upper Jurassic. In the North Sea Central Graben the Upper Jurassic is often represented by organic deeper-water shales which indicate that the raft system had developed into a submarine trough (Fig. 10). At the Jurassic–Cretaceous boundary a further major rifting phase occurred throughout the Central Graben (late Kimmerian phase). All of these three main phases are recognisable through much of the North Sea area either as unconformities, disconformities or regressive–transgressive cycles (Fig. 11). Only in the deepest part of the Central Graben, a veritable taphrogeosyncline (term used in the sense of Trümpy 1960), do more or less continuous Upper Jurassic–Lower Cretaceous sequences occur. During the successive periods of downfaulting of the graben floor the graben margins were uplifted and subjected to erosion, thus conforming to the rift model drawn up by Illes (1970). Erosion on the highs flanking the Central Graben cut down, e.g. in the Central North Sea, as deep as the Zechstein and locally even into the Devonian.

Drilling in the Central Graben has as yet yielded only sparse evidence of rift-volcanism during the Jurassic and Lower Cretaceous. In the marginal troughs flanking the Variscan massifs the above-described rifting phases can roughly be recognised as either regressive–transgressive cycles or as unconformities associated with minor warping of the basin floors followed by rapid subsidence (Fig. 12).

Warping and temporary uplifting of the basin floors may have been due to

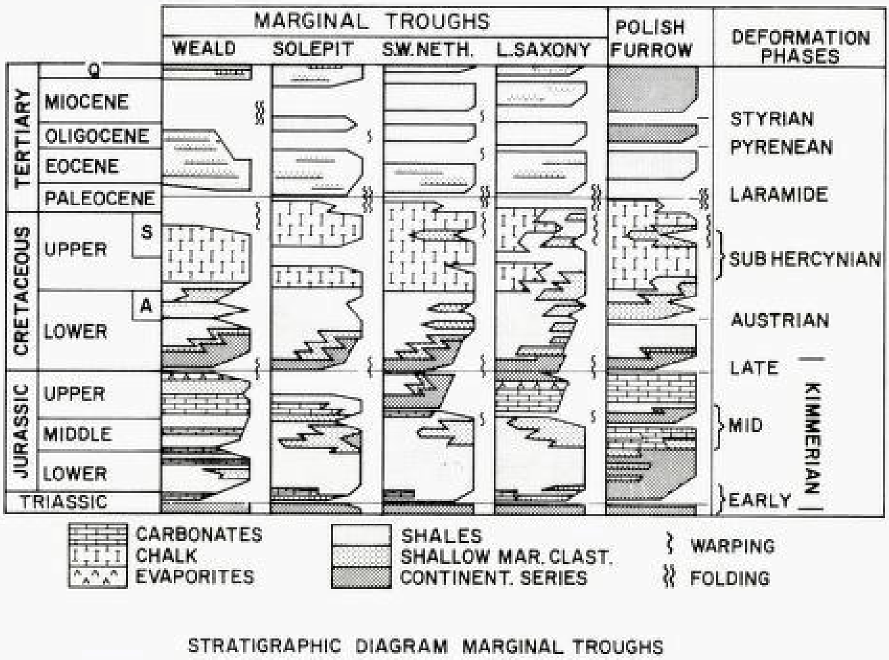


Fig. 12. Stratigraphic diagram of the marginal troughs.

slight jarring of these basins in response to the postulated deep-seated transform movements between the North German–Danish segment and the Central European–English segment (Fig. 9).

Phases similar to those recognised in the North Sea Central Graben are also evident in the Danish–Polish furrow where the Dogger phase corresponds to a marine transgression, the Callovian–Oxfordian phase to an unconformity, and the Late Kimmerian phase to a basin-wide unconformity (Fig. 12). The above-postulated regional tectonic model is in good agreement with the occurrence of similar, largely extensional phases in the Alpine geosyncline that resulted in an accentuation of positive platforms and negative areas. Of particular interest are the early mid-Jurassic phase (Trümpy 1960), which is recognised, e.g., in the Briançonnais as an erosional phase associated with the shedding of breccias in the Sub–Briançonnais, and the Oxfordian phase that is documented by, e.g., the Brèche du Télégraphe in the Sub–Briançonnais (see also Gwinner 1971; Fig. 13). In the Alpine geosyncline the stress pattern changed during the Lower Cretaceous from essentially extensional to a compression (Trümpy 1965). The occurrence of Neocomian flysch in the Penninic facies realm of the eastern Alps is thought to be linked to early compressive movements resulting in the rising of the Penninic Geanticline (Gwinner 1971, Wunderlich 1967).

The extrusion of ophiolites in the Swiss Alps is dated as Lower Cretaceous. These extrusives are considered by some authors as being contemporaneous with early folding phases (Trümpy 1958, 1965). The gradual change of the

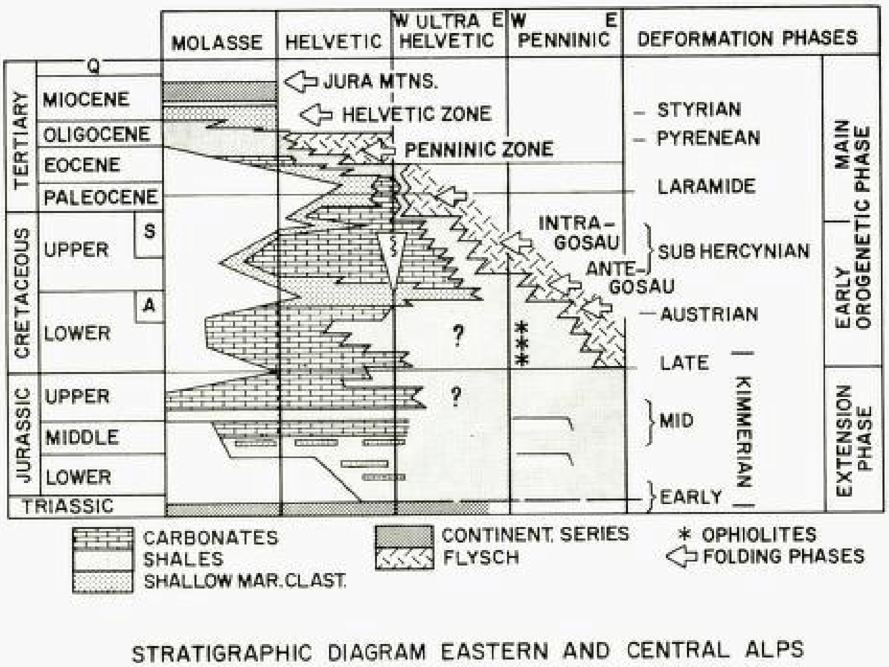


Fig. 13. Stratigraphic diagram of the eastern central Alps.

stress pattern during the Lower Cretaceous, however, remained restricted to the Alpine Geosyncline. In the North Sea area and in central Europe the Jurassic setting that was accentuated by the Late Kimmerian tectonic phase persisted through much of the Cretaceous period (Fig. 14). In the North Sea Central Graben only minor rifting movements can be recognised during the Lower and Upper Cretaceous; the Graben itself, however, continued to subside rapidly.

The last phase of rifting accompanied by uplifting of the graben margins and rapid subsidence of the Graben itself occurred during the Paleocene (Laramide phase; Fig. 11). In the North Sea the gradual abatement of rifting movements during the Cretaceous was accompanied by widespread transgressions resulting in the inundation of the uplifted flanks of the central rift during the Upper Cretaceous. The Upper Cretaceous development of the North Sea area is thus transitional between the Jurassic–Lower Cretaceous taphrogenic stage and the Tertiary post-rifting stage.

During the Late Kimmerian phase the Variscan massifs were once more consolidated into a continuous barrier reaching from southern England to Poland (Fig. 14). In the marginal troughs this phase is recognised as an unconformity accompanied by minor warping, followed by rapid subsidence of the respective basins and by the deposition of thick clastics. During the Lower Cretaceous a gradual transgression over the northward adjacent highs can be observed. The effects of the Albian–Aptian Austrian phase that are readily recognisable in the marginal troughs are only vaguely reflected in the

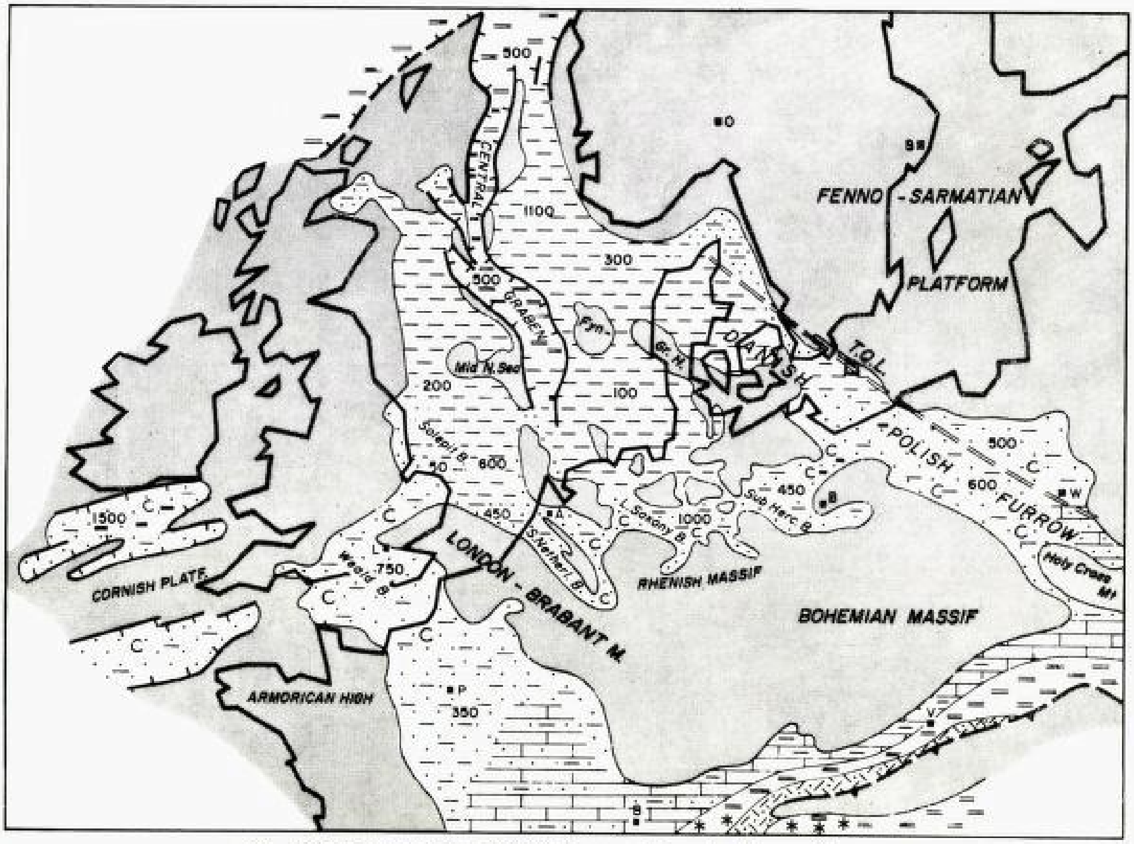


Fig. 14. Paleogeography of NW Europe during the Lower Cretaceous.

North Sea Central Graben, but are clearly evident in the Polish furrow Figs. 11, 12). As in the North Sea area the Upper Cretaceous transgression is very widespread in central Europe (Fig. 15). First inversion movements (term used in the sense of Voigt 1962) occurred in the marginal troughs as well as in the Polish Lowland during the Senonian (Subhercynian phase), leading to the uplifting of the axial zones of the basins and the development of secondary depôt centres flanking the uplifted basin centres (Voigt 1962; Heybroek 1964; Pozaryski 1960). Large-scale inversion involving also the southern part of the North Sea Central Graben took place during the Maastrichtian to Upper Paleocene, resulting in uplifting of the basin fill either along steep reverse or normal faults and/or by regional warping above the erosional level. These inversion movements are time-correlative with the last phase of downfaulting of the central and northern parts of the North Sea Central Graben and are thus not compatible with the above-developed tectonic model. However, in the Alpine geosynclinal realm compressive movements became dominant during the Upper Cretaceous with flysch deposition spreading in the Maastrichtian from the Penninic into the Ultrahelvetic facies domain (Fig. 13). It is hypothesized that the forces causing these Late Cretaceous–Early Tertiary Alpine

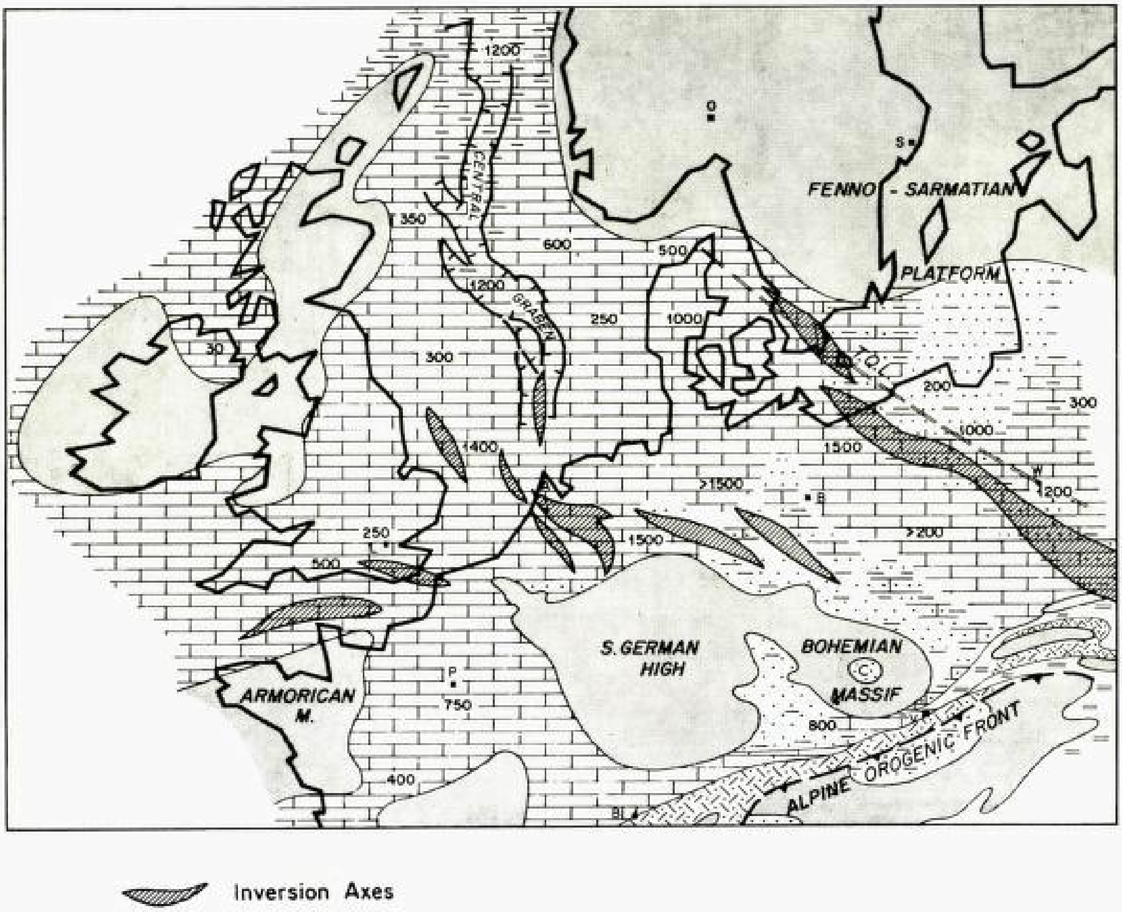


Fig. 15. Paleogeography of NW Europe during the Late Cretaceous.

orogenic movements had reached the dimensions to exert a compressive stress on the Alpine foreland causing the inversion of the marginal troughs that acted like shear-pins in the otherwise rigid platforms. However, once inverted these basin became largely inactive.

Only minor inversion movements are recognised in the marginal troughs flanking the Variscan Massifs during the post-Laramide, main Alpine orogenic phases with the exceptions of, e.g., the Weald Basin (Gallois & Edmunds 1965), which was mainly deformed during the Miocene and in which only mild warping occurred during the Laramide phase (Fig. 12).

An explanation for the contemporaneity of rifting movements in the North Sea and possibly in the North Atlantic and compressive, orogenic movements in the Alpine geosyncline has to be sought on an even larger scale than is considered in the present paper.

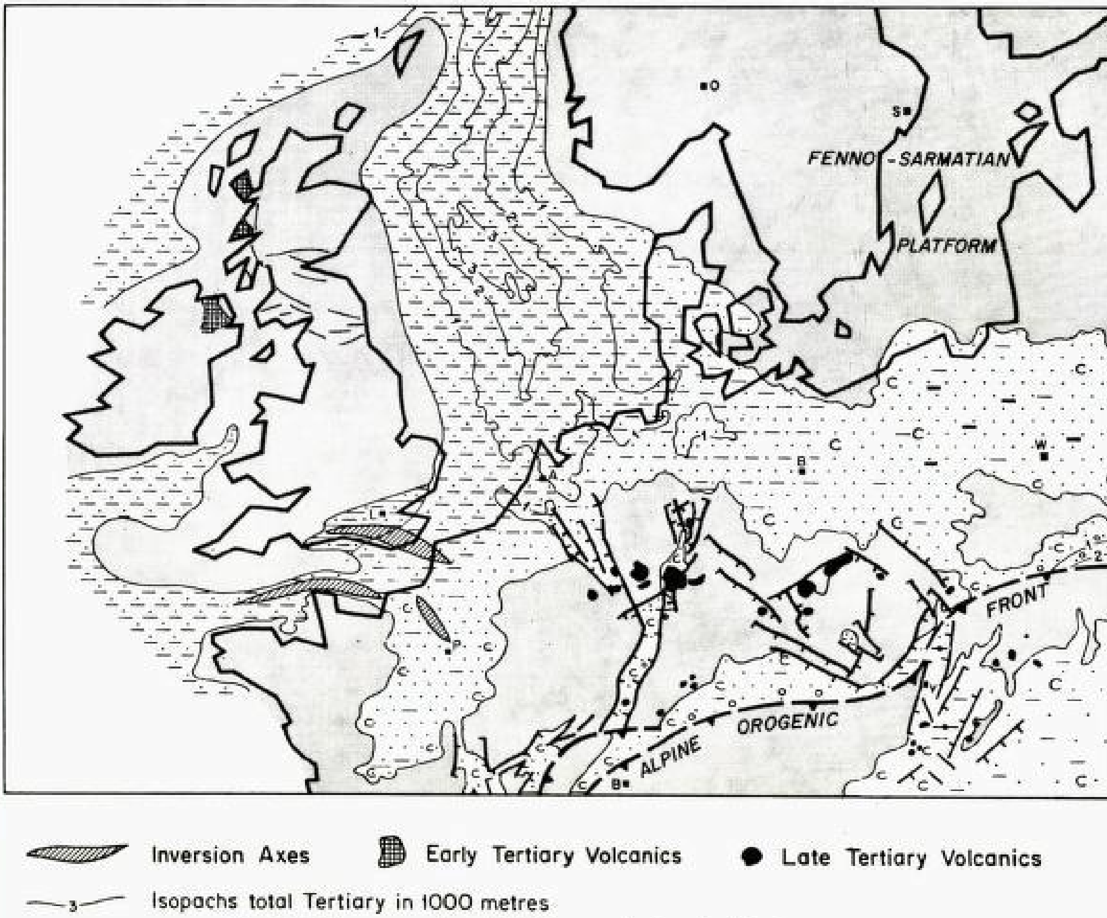


Fig. 16. Paleogeography of NW Europe during the Tertiary.

Post-rifting intracratonic stage (Tertiary)

The taphrogenic stage of the North Sea, that lasted possibly since the Triassic but definitely since the Jurassic, came to a close in the Late Paleocene. Final parting of the European and the North American–Greenland plate may have been effectuated during the Late Paleocene or Eocene (Pitman & Talwani 1972), thus initiating the drifting stage in the Arctic North Atlantic. Late Paleocene to Eocene flood basalts extruding during the final rifting and early drifting phase are known from Scotland (Rayner 1967; Richey 1961; Mitchell & Reen 1973) as well from Greenland (Brooks 1973). With the Arctic North Atlantic entering the drifting stage during the Late Eocene (Pitman & Talwani 1972), extensional stresses apparently ceased to influence NW Europe.

The North Sea area was dominated by regional subsidence resulting in a symmetrical saucer-shaped intracratonic basin, the axis of which coincides with the now inactive central rift valley. Maximum Tertiary thicknesses of up to 3.5 km (Dunn et al. 1973; Heybroek et al. 1967) occur in the central parts of the North Sea.

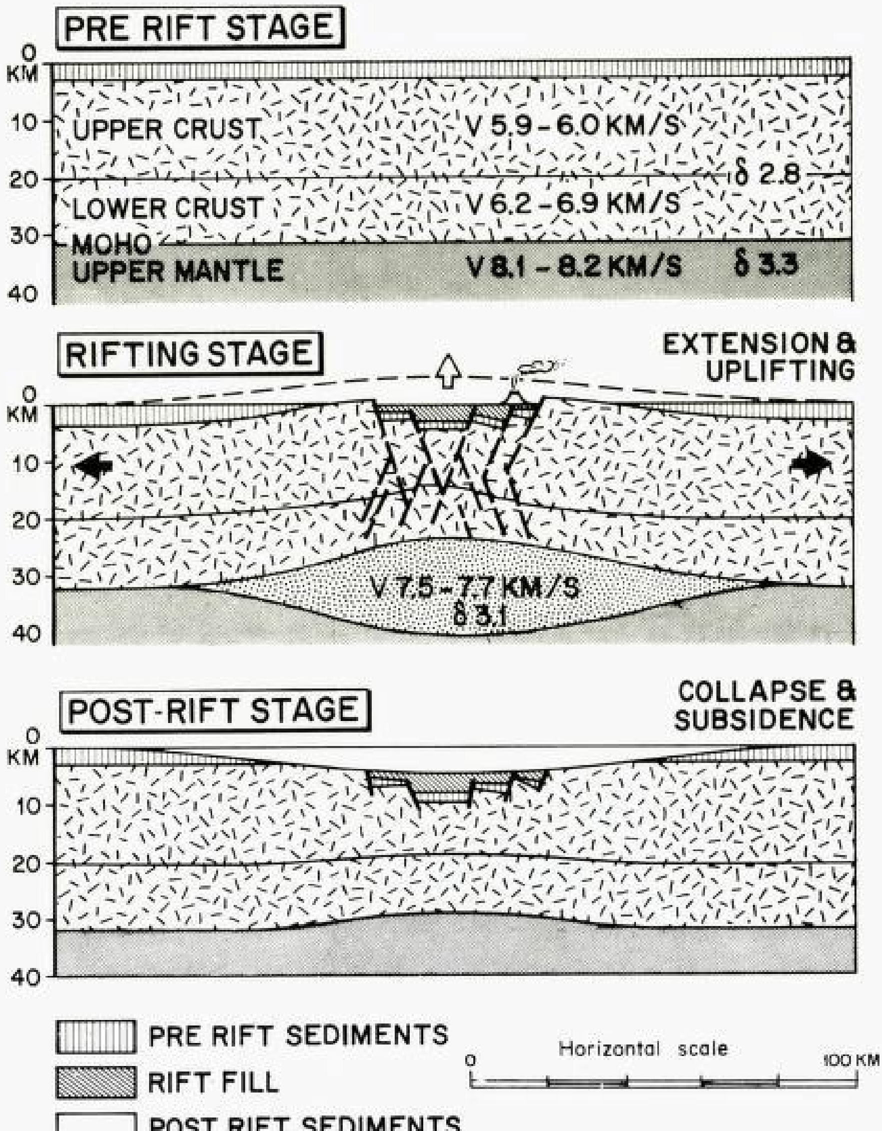


Fig. 17. Conceptual model of rift development.

ATTEMPT AT A GEODYNAMIC INTERPRETATION

The Tertiary subsidence pattern of the North Sea basin can be explained by the conceptual rift model as summarised in Fig .17.

The *'Pre-rifting' Stage* (Fig. 17a) represents a standard continental crust in isostatic equilibrium. The 'Rifting Stage' model (Fig. 17b) was fashioned according to the refraction data obtained by Ansorge et al. (1970) and Ahorner et al. (1972) for the Rhine Graben area and incorporating the more theoretical considerations of Illies (1970) and Laubscher (1970).

The most significant element in the 'rifting stage' model is the rift cushion

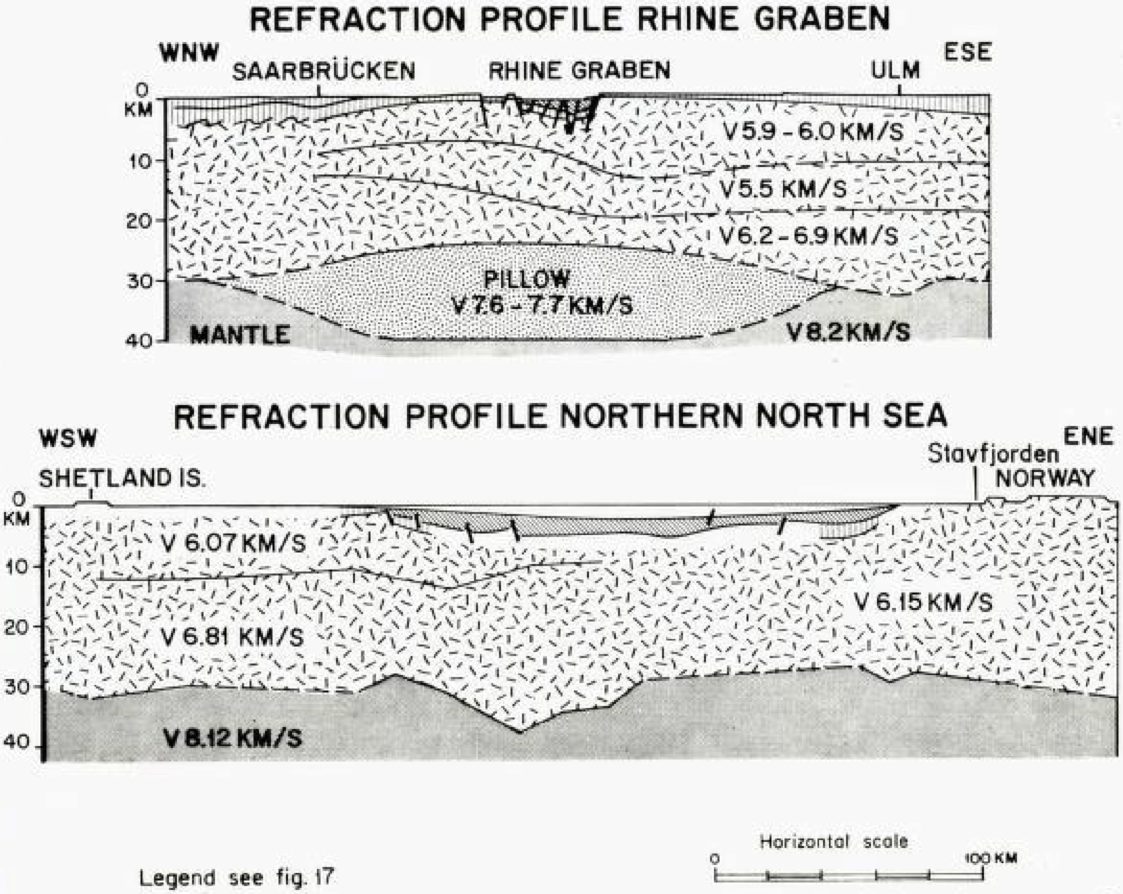


Fig. 18. Comparison of a refraction profile across the Rhine Graben and one across the northern North Sea.

that has a p-velocity of 7.5–7.9 km/sec and a density of 3.1 (Meissner et al. 1970). Similar rift cushions have been observed in, e.g., the Baikal graben (Artemiev & Artyushkow 1971) and the Red Sea rift (Drake & Girdler 1964). These cushions centre under the central rift valley and pinch out laterally under the uplifted flanks of the rift.

In Fig. 17b an unbroken rift dome (stippled line) is reconstructed. Artemiev & Artyushkow (1971) estimate the crustal extension resulting from uplifting of a correspondent 200 km wide and 3–4 km high rift dome to amount to some 200 m only. Laubscher (1970) estimates the extension across the Rhine Graben rift, which has similar dimensions, to be in the order of 5 km. This indicates that the uplifting of a rift area is probably only a subsidiary cause of the Central Graben development, but that the primary cause of the rifting is regional extension (see also Osmaston 1971).

It is hypothesized that intracontinental rifts such as the North Sea Graben or the Rhine Graben are initiated as the result of regional extensional stresses and that progressive rifting results in 'necking' of the crust, causing decompression of the mantle and the formation of a rift cushion through fractional distillation from the mantle. This, however, is only possible if a temporary decoupling between the crust and the mantle is assumed (Artemiev & Artyushkow 1971). Osmaston (1971) postulates as an alternate a model with coupling between the crust and the mantle and fracturing extending into the mantle.

Emplacement of the low density rift cushion causes uplifting of the rift zone and, with this, secondary subsidence of the central rift valley. This is furthermore thought to be linked to the initiation of the rift volcanisms. This stage was reached in the North Sea rift in the Middle to Upper Jurassic and lasted through much of the Lower Cretaceous.

The '*Post-rifting*' *model* (Fig. 17c) is largely based on the present-day configuration of the northern North Sea as depicted by the regional cross-section from the Shetland Islands to the Norwegian shore (Fig. 18). This section combines the deep refraction data obtained by Sørnes (1971) and reflection seismic results for the shallower sedimentary layers. In the line of the section unfortunately no information is available on the position of the central rift. valley floor. The 'post-rifting' stage is characterised by the absence of the rift cushion under the former rift zone and by the presence of thick post-rifting sediments, deposited in a symmetrical saucer-shaped basin, the deepest parts of which coincide with the former rift valley.

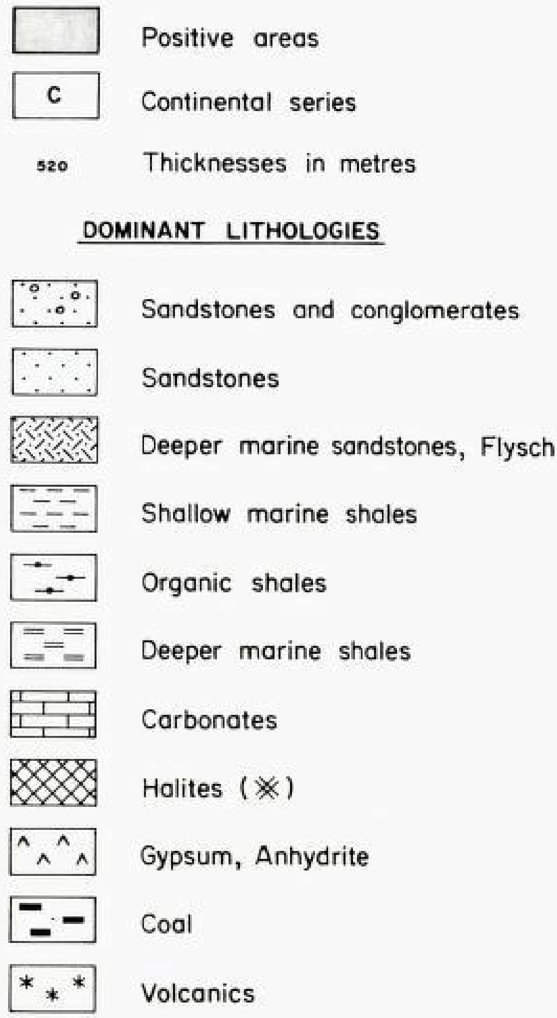
The subsidence of the rift zone is explained by the gradual resorption of the low density rift cushion by the mantle. This is in keeping with Osmaston (1971), who considers by the formation of a rift cushion a 'reversible thermal effect upon the uppermost mantle material'. It is further speculated that the rift cushions can only exist as such during the active rifting stage during which rifting movements result in progressive decompression of the mantle.

Once rifting movements cease the resorption of the rift cushion proceeds, resulting in a gradual ascending of the Moho and a corresponding subsidence of the rift dome. In view of the 'necking' of the crust and erosion on the rift margins, a large saucer-shaped basin develops, the width of which corresponds roughly to the width of the original rift dome. However, insufficient refraction data are available from the North Sea to date to fully verify the above hypothesis.

During the Tertiary, central Europe was dominated by the Alpine orogeny, the main phases of which are summarised by Fig. 13. In the North Sea regional disconformities correlate roughly to these main orogenic phases that are reflected in the marginal troughs by minor inversion movements. A notable exception is the Weald basin that underwent major inversion during the Miocene.

During the Eocene the Rhône–Rhine Graben system was emplaced in the Alpine foreland (Illies 1970). Its further development during the Oligocene and Upper Tertiary is largely concomitant with the Alpine folding phases. Other graben systems developed during the Tertiary in the Panonic basin with the Vienna basin practically crossing the Alpine orogenic front. Uplifting and

Fig. 19. Legend to the paleogeographic maps.



fracturing of the Variscan massifs led to widespread volcanism during the Mio–Pleiocene (Knetsch 1963). The northernmost branch of the Rhône–Rhine Graben system is the Rur Graben that reaches the southern Netherlands (Heybroek 1974). There is no evidence of a Late Tertiary reactivation of the North Sea central rift system. In view of the age disparity between the Rhône–Rhine Graben system, which is in an active rifting stage, and the North Sea central rift system, which is in a post-rifting stage, the two should not be considered as part of one megafracture system dissecting western Europe. However tempting such a speculation may be, their apparent continuity is largely fortuitous.

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Geothermal Aspects of Hydrocarbon Exploration in the North Sea Area

CARL-DETLEF CORNELIUS

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A statistical analysis of the distribution of hydrocarbon phases indicates that in a time-temperature diagram these phases are restricted to hyperbola confined areas. Thus, the 'liquid-window-concept' has been improved by inferring a time correction. Based on a survey of recent temperature conditions and the reconstruction of paleo-temperatures, it has been applied to the North Sea region. Furthermore, the survey of present geothermal gradients has made it evident that the region north of the Variscan orogen is a type region of an inverted geothermal realm. This can be explained by the abundance of source beds and other sediments of low grade lithification in the subsiding areas. The maximum temperatures have been preferentially determined by the vitrinite (coal) reflectivity method; because it allows measurements without physical–chemical alterations of the organic matter and because the reflectibility depends on the gradational changes of aromatization, which is one of the significant processes during the formation of crude oil. A Table for the determination of time-corrected temperatures from vitrinite reflectivities has been added to this paper.

The relationships between recent and paleo-temperatures have been systematized and some paleo-thermal events have been interpreted:

a. The temperature high above the Bramsche Massif which formed during the Austrian orogeny.
b. The Mid-Jurassic event which led to the formation of the gas deposits in the Rotliegend of the British part of the southern North Sea.
c. The subrecent re-coalification processes responsible for the persistent formation of gas deposits in the 'Buntsandstein' beds north of the Rotliegend reservoirs.
d. The importance of the early Tertiary heat dome along the Lofoten high axis has been taken into consideration.

Many, or maybe all geothermal events affecting the North Sea region since the Cambrian, indicate a time-space relationship to the alternating phases of sea-floor-spreading along mid-oceanic ridges; respectively, compressional orogenic phases caused by plate collisions. Positive temperature anomalies formed preferentially at discontinuities. Some of them are related to the irregular course of the North Sea graben system.

C.-D. Cornelius, Gelsenberg Mineralöl, P.O.Box 43, Essen 1, W. Germany

1. Introduction

The purpose of this paper is to report the results of investigations of changing temperature conditions in the North Sea region. They refer to worldwide statistical data and special experiences gained from the genetically closely related areas of Northwestern Germany as well as the Netherlands. Subsequent to an introductory discussion about the importance of the temperature parameter in the hydrocarbon genesis model, special results of present temperature measurements in the North Sea region will be presented. Finally, the genesis of hydrocarbon deposits in the North Sea region will be elucidated,

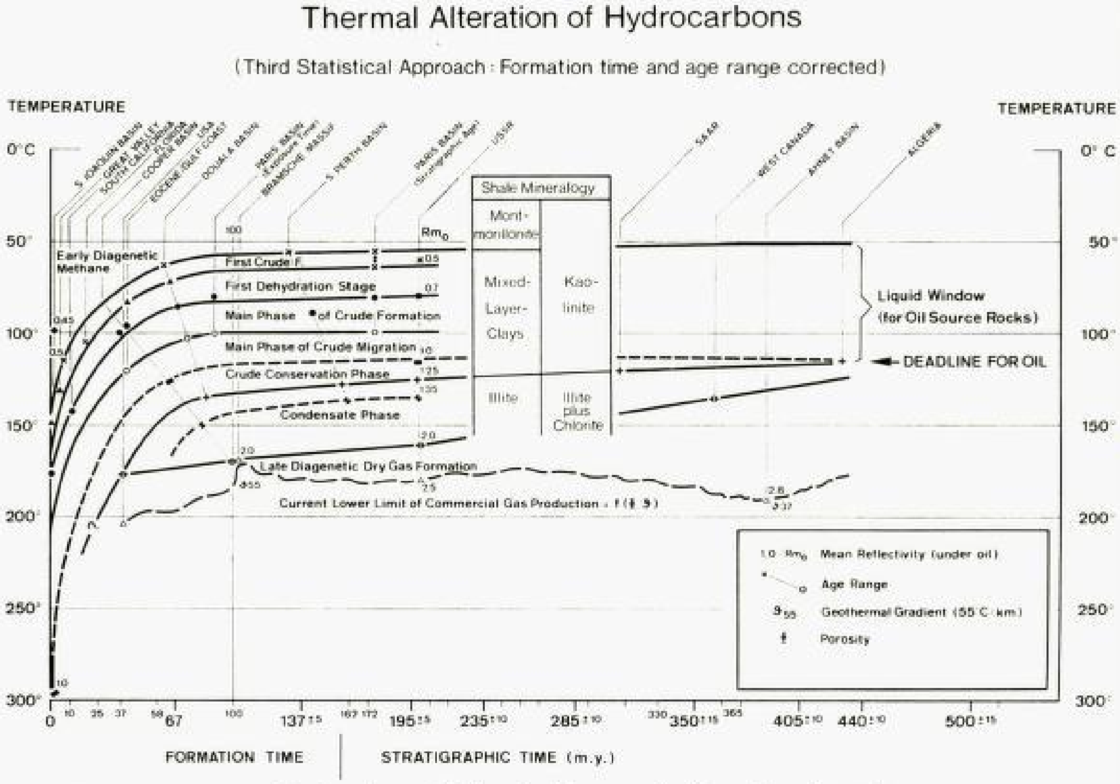

Fig. 1. Thermal alteration of hydrocarbons (third statistical approach: formation time and age range corrected).

The values of hydrocarbon alteration are taken or derived from Alpern (1969), Bartenstein et al. (1971), Castano (1973a), Correia (1969), Demaison (1973), Evans & Staplin (1970), Johns & Shimoyama (1972), Landes (1967), Leythaeuser & Welte (1969), Louis & Tissot (1967), Phillipi (1965), Price (1973), Reel & Griffin (1971), Robert (1971), Seibold (1973), Ting (1973), Wassojewitsch et al. (1969).

The 'shale mineralogy window' is based on discussions in papers by Burst (1969), Dunoyer de Segonzac (1964 & 1970), Frey & Niggli (1971), Johns & Shimoyama (1972), Kubler (1968), Leplat (1973), Ludwig & Hasse (1973), Weaver (1960), Weber (1972).

based on the determination and the evalution of maximum temperatures.

I am indebted to Gelsenberg AG for consenting to the publishing of the results of my investigations; to the staff of the exploration division for technical support and especially to Dipl. Geophysicist J. Deist, for fruitful discussions; furthermore, to Dr. U. Franz, Technical University, Munich, for the translation of the manuscript.

2. The temperature parameter in the hydrocarbon formation model (Figs. 1 & 2)

Subsequent to Sokolow's (1948) publication an exponentially growing number of writers have been explaining the crude oil formation by transmutation of kerogen in the low temperature realm (Figs. 1 & 2). Kerogen is a highly

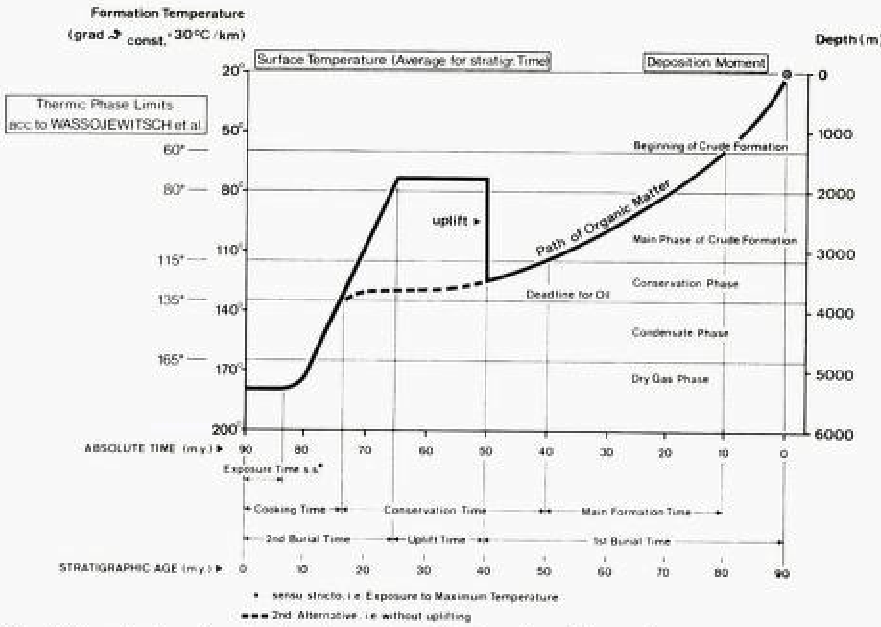

Fig. 2. Terminology for the time intervals of hydrocarbon diagenesis.

polymeric organic matter, disseminated in the sediment and insoluble by organic solvents. The maceral groups essential for the hydrocarbon formation are liptinite and to some extent, vitrinite. Heating of kerogen in an unoxidizing, hydrogen-bearing environment results in the dissociation of hydrocarbons from these macerals providing their grade of aromatization does not exceed a certain limit (Welte 1973, also 1965 and 1972, Louis & Tissot 1967, McIver 1967, Califet et al. 1969, Tissot 1969, 1971, Espitalié et al. 1973).

Pusey's (1973) 'Liquid–Window–Concept' is known as the hypothesis of the limitation of crude oil deposits to a distinctive temperature range. However, much more subtle differentiations were proposed by Landes (1967) and Wassojewitsch et al. (1969), who distinguished a number of phases confined by isotherms.

Landes (1967)		Wassojewitsch et al. (1969)	
		Early diagenetic methane	< 60°C
Oil and gas	< 93,3°C	Initial phase of crude formation	< 80°C
Light oil and gas	< 121,1°C	Main phase of crude formation	< 115°C
Mostly gas	< 149°C	Final phase of crude formation	< 135°C
Oil phase-out zone	< 177°C	Late diagenetic methane	<(180°C)

We prefer the Russian model and modify it by adding the following horizons and stages:

a. Johns & Shimoyama (1972) followed a proposal of Burst (1969) and the perception of Perry & Hower (1972), saying that the dehydration process during the illitization is set forth stepwise in two phases. Drawing conclusions from studies in the U.S. Gulf Coast Eocene, they defined the second dehydration phase as the main phase of oil migration, proceeding in a temperature range between 120°C and 130°C. Seibold (1973) claims that migration in the Douala-Basin of Cameroun begins already at temperatures around 105°C. However, careful consideration must be given to the fact that the solubility of the various hydrocarbons in water as well as their affinity to form colloidal soaps is different and, furthermore, is changing with salinity and pressure (see Chapman 1972, Berry 1973, Cordell 1973, Hobson 1973, Price 1973).

b. Robert (1971, 1973) appears to interpret the 'final phase of oil formation' as a conservation phase. We incorporate this idea, particularly as several authors believe in an upper temperature limit for crude oil formation of approximately 115°C. However, Meinhold (1972) assumes that crude oil formation processes are only extenuated around this temperature (see also Fig. 2 for the importance of the conservation phase).

c. The conservation phase is bounded by the 'deadline for oil' (M. Teichmueller 1971). The breakage of straight chain hydrocarbons will commence no later than at this isotherm (Abelson 1963).

d. The question of the 'deadline for gas' has been modified into the question for the 'current lower limit of commercial gas production' by reservoir engineers (Landes 1967). This limit depends on the porosity which is a function of the geothermal gradient (Maxwell 1964). However, we could prove that hydrocarbon impregnation will prevent a porosity decrease in quartzitic sandstones (Pernow 1969). This observation is of common interest especially since it supports M. Teichmueller's (1971) general statement that the velocity of coalification processes exceeds that of inorganic diagenetic processes.

It remains a remarkable fact that large gas deposits become scarce above temperatures of 150°–165°C (M. Teichmueller 1971, Bartenstein et al. 1971); although Meinhold (1972) places the main gas formation phase into a temperature range of ± 170°C and Geodekjan (1972) speculates about gas formation even at temperatures between 300° and 400°C. One reason for the formation of giant gas fields is the rapid dissociation of gas from fat coal and ess coal (Fig. 7); another reason is the transition of liquid hydrocarbons into gaseous ones. Conversely, the destruction of gas deposits is caused by an increase of jointing and shaly cleavage (Fourmarier 1970) yielding a loss of the sealing characteristics of overlying shales, as well as by an increase of the reservoir pressure, both resulting mainly from temperature increases ($p \cdot V/T$ = constant; for additional information about problems of diffusion see Rudakow 1965, Kroepelin 1967, Nesterow & Uschatinskij 1972).

Thus, the evolution of organic matter is well known, and it can be ordered into a sequence of phases of formation and transformation; crude oil genesis has been brought into a systematic scheme of specific diagenetic and metamorphic steps: between early and high diagenesis (according to M. Teichmueller 1971 and Robert 1973); the catagenesis, principally the meso-catagenesis

(Wassojewitsch et al. 1969); the anchi-metamorphism (preceding the metamorphism sensu strictu according to Harrassowitz, see also Correns 1949); the 'initial' metamorphism of Baker & Claypool 1970), the 'eometamorphism' (Landes 1967), 'the low-grade' metamorphism (Dunoyer de Segonzag 1970), the 'very-low-stage' metamorphism (Kisch 1973, in reference to Winkler 1970). The terms of diagenesis are applied in this paper because Winkler speaks of metamorphism only above temperatures of about 200°C. (Further information about time-dependent mineral temperatures may be obtained from the remarks about the clay mineral diagenesis, as well as from the apatite fission track ages of Wagner & Reimer 1973.)

The temperature intervals, however, are not conclusive. The Anglo–American literature of the sixties was dominated by Phillipi's (1965, 1969) doctrine that crude oil formation begins at a minimum temperature of 150°C, and by the Russian findings that crude oil is still stable at temperatures between 170°C and 180°C (Hedberg 1964). On the other hand, in the Eastern Hemisphere, the concept was developed (since Louis & Tissot 1967) that crude oil formation is limited to a range between approximately 60°C and 135°C (Wassojewitsch et al. 1969). Albrecht & Ourisson (1969) shifted the base line of the 'liquid window' in the Douala-Basin even down to 105°C. Pusey (1973) expanded his 'window' concept with a statistically broader range, between 65° and 150°C in a rather pragmatic attempt to summarize the different results of various authors from various regions. In that way he made it more consistent, yet unfortunately he restricted the economic value of the whole concept.

My alternative model of the distribution of hydrocarbon phases (Fig. 1) might be considered more useful because it is based on empirical data gained from detailed investigations of approximately twenty sedimentary basins. Ten of these have the rank of case histories with two or more values from each profile, even though, in two cases, (U.S.A. and U.S.S.R.) a conclusive time-correction could not be accomplished. In two other cases (Cooper Basin and S. Perth Basin, Australia) reliable coalification values were available. Incorporation of additional vitrinite reflectivity values yielded a sufficient number of appropriate data to present a composite model, which is based on the combined evaluation of hydrocarbon as well as coal maturity values. This combination appears to be tolerable because the degradation grade of kerogen (Welte 1973), the maturation grade of crude oil (Califet et al. 1969) and the reflectivity of coal (see M. Teichmueller 1971, for details about the index of refraction) all depend on the grade of aromatization (Oelert 1972, see also Leythaeuser & Welte 1969).

The curves in Fig. 1 are isolines of maturity and diagenesis grades. We are dealing with iso-reflections, a term that has been introduced by coal petrographers (Bartenstein et al. 1971); it is synonymous with the term iso-apostilbs (Paproth & Wolf 1973). Within certain limitations, discussed later, iso-voles or iso-volatiles, i.e. isolines of volatile constituents of coal, can also be utilized (Kuyl & Patijn 1961, Patijn 1961, Boigk et al. 1971, M. & R. Teichmueller

1966, 1971a, b). In addition, the maturity grades of kerogen and crude oil can be characterized by 'iso-maturities' or 'iso-crack-values'. A superimposed term would be 'iso-aromatization grade'. The term 'geochronotherms' was introduced by Karzew in 1968 (cit. Meinhold 1972) because their loci are functions of the parameters temperature and time.

In an initial attempt to accomplish a statistical analysis, we referred to the stratigraphic position, i.e. to the geologic age of source rocks. From this it could be learned that only the source rocks from young continuously downwarping basins indicate causal relations. Thereupon the geologic ages of up to 150 m.y. old samples were corrected by the formation times. Finally, hydrocarbon distribution profiles (published from various basins) were time-corrected. (The dextrally inclined lines of the diagram indicate the age range.) These corrections yielded evenly bent hyberbolic curves which indicate the validity of an exponential function. Based on geological considerations it could be demonstrated that the Arrhenius equation (1) is valid for the formation and the maturation of hydrocarbons:

$$\frac{dC}{dt} = - C_i \cdot A \cdot e^{-\frac{E}{RT}} \qquad (1)$$

C = kerogen concentration at the process time t
C_i = initial kerogen concentration
$-$ = negative sign; becauce $C < C_i$ (degradation)
A = Arrhenius– or impact– or frequency factor (number of molecular groups reacting with each other per time unit)
e = base of the natural logarithm
E = activation energy
R = universal gas constant
T = absolute temperature (°K)

The discussion of this equation (1), based on laboratory findings and theoretical assumptions, is dealt with by Huck & Karweil (1955), Karweil (1956), Abelson (1963), Deroo et al. (1969), Hanbaba & Juentgen (1969), Karweil (1969), Phillipi (1969), Tissot (1969), Lopatin (1971), Tissot (1971), Tissot & Pelet (1971), Eglinton (1972), Tissot et al. (1972), Welte (1972), Johns & Shimoyama (1972), Bostik (1973). A detailed interpretation of the equation (1) will be presenterd on an other occasion. The reader's attention is drawn to the apparently straight line boundary between wet and dry gas. The time relationship of the clay mineral diagenesis (Fig. 1) and the water solubility of hydrocarbons is not sufficiently clarified yet. However, it can be recognized that time is gradually substituted for temperature in an exponentially decreasing mode during the first 60 m.y. of kerogen degradation. Thus, the introduction of the time factor allows a relativation of the various 'liquid-window' concepts. (Strange to say, Wassojewitsch et al. 1969 substituted time only for the geothermal gradient.) Moreover, the processes abstracted in the Arrhenius equation are irreversible and therefore their products are fossilizable.

From this point of view, the quantification of the time factor must be considered the most important task of a geologist trying to solve problems. These questions have been explained within the framework of the Russian model (Fig. 2). The crude oil formation period makes up only approximately 1/3 of the geologic history of certain source rocks. According to the Arrhenius equation degradation in a source rock during this time interval is 27%. This change of concentration can be called the recovery factor of the source rock. In the simplified model the previously formed oil accumulation remains unchanged even through an uplift phase.

Only at temperatures above 135°C is there sufficient energy supplied to initiate the cracking process of crude oil, especially of the straight chain hydrocarbons. This point marks the boundary between formation time and cooking time, the latter term here only being used for the gasification period of crude oil formation. The coal geologists consider the exposure time under maximum temperature conditions to be the most crucial factor for the mode of hydrocarbon formation. Experience has shown that the liquid phase will not be totally transformed into the gas phase if the exposure time interval of the cooking period is comparatively short. In this case the formation of deposits with a high gas/oil ratio is favoured. After all, the recovery factor increases during the cooking stage because solid kerogen in the source rock discharges additional gas during this phase.

The time intervals during which kerogen is degraded ought to be distinguished from the exposure times: in most cases the geologist can only make an estimation of maximum values from the subsidence curve. The stagnation and uplift phases ('hiati') creating a reduced energy level are mostly beyond recognition; however, they should be substracted from the periods of maximum subsidence, using 'best guess' assumptions.

The practical usefulness of the above-described geochronothermal concept lies in the fact that the temperature ranges of hydrocarbon phases can be recognized in any sedimentary basin or sub-basin, provided that the subsidence history is known. Thus, determined isotherms can be incorporated in the 'liquid-window' diagrams, where depth values relate to geothermal gradients (see, e.g., Pusey 1973). These diagrams allow a prognosis about lacking, sufficient or over-maturation, as well as crude oil qualities and gas contents; thus providing the basis for well planned drilling programmes. For instance, any kind of drilling activity will be irrelevant unless crucial temperatures for the formation of oil or gas have been reached in source rocks above the basement. The application of heavy rigs for ultra-deep wells appears to be useful only in rapidly foundered areas or in areas with a comparatively low geothermal gradient.

3. The current temperature distribution in the North Sea region

3.1 TEMPERATURE DETERMINATIONS (Fig. 3)

The geothermic specialists from various geological surveys, mining companies and geophysics departments of universities have made detailed investigations

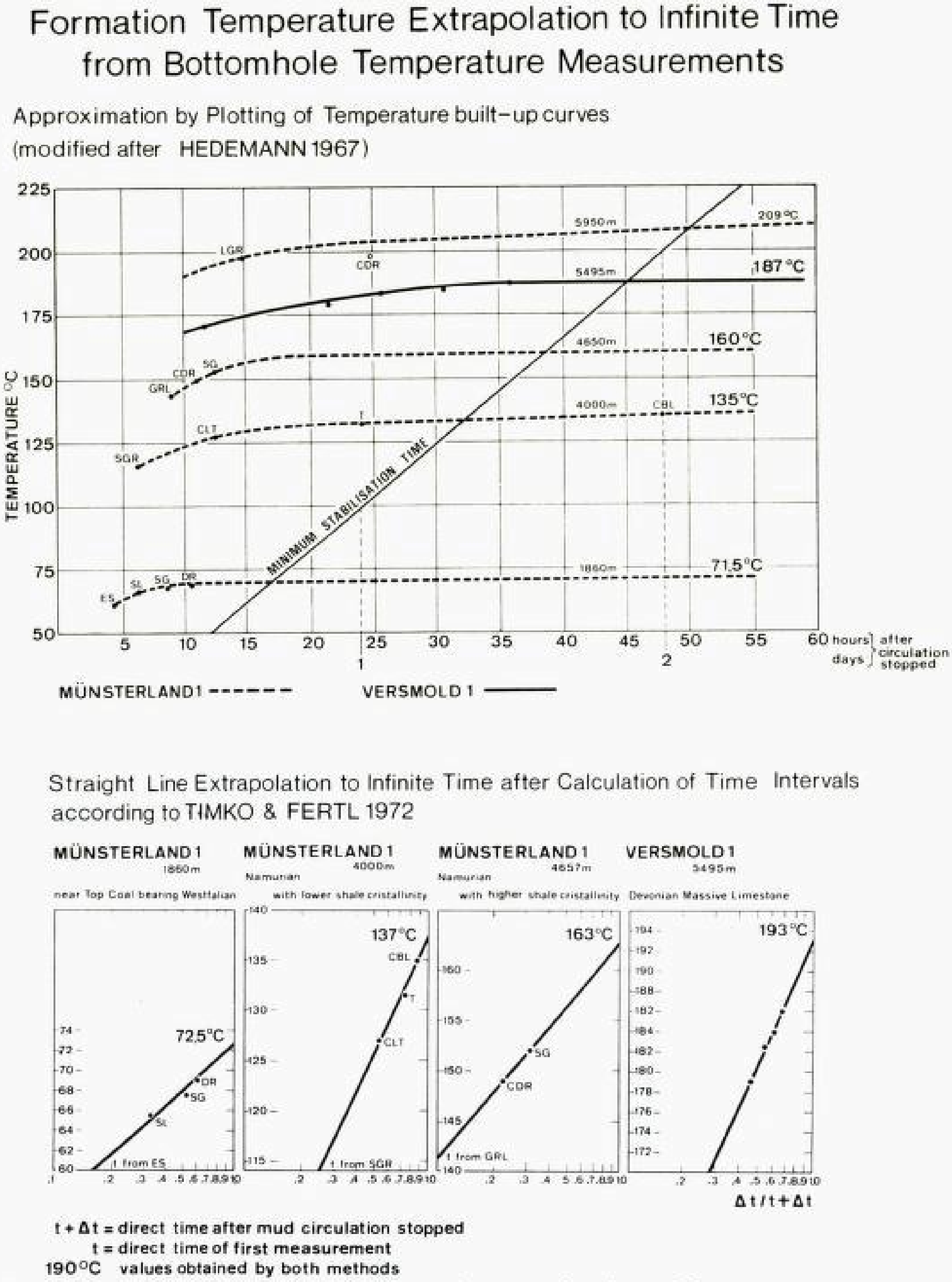

Fig. 3. Formation temperature extrapolation to infinite time from bottomhole temperature measurements.

to determine the true formation temperatures of adjacent onshore areas (e.g. Braaf & Maas 1952, Creutzburg 1964, Heine 1962, Kappelmeyer 1961, Parasnis 1971, Quiring 1936, Swanberg et al. 1972.) On the basis of their findings they handle the petroleum geologists' temperature data from deep wells with some scepticism; the justification for this might be discussed in the following. Their objections refer to the grade of accuracy and the comparability of such measurements as well as to the temperature compensation times and the depths of reference.

The results of preliminary investigations by Gelsenkirchener Bergwerks A.G. dealing with this group of questions have been summarized by Heine (1962). According to this, West Germany's Coal Mining Association presented in 1956 'guidelines for the determination of true formation temperatures' for the purpose of standardization. Gelsenberg has been using and comparing temperature-measuring devices constructed by Huegel (1942), Schlumberger et al. (1957), and the Coal Mining Ass. and Deilmann Corp. These are mercury, electric and Negative Temperature Coefficient (NTC) resistivity thermometers with a reading accuracy of $\pm$ 0.1°C. The results differed from each other by values of up to 1.6°C in general less than 1°C. Hence, the usefulness of the mapping of isotherm intervals of $<$ 2°C must be critically examined.

Recently, the Boliden Aktiebolag of Sweden produced a Hewlett-Packard quartz thermometer with a tolerance of 1/1000°C for measurements in their shallow exploration holes (Parasnis 1971). Unfortunately, such precision thermometers are scarcely ever used.

Their application is useful if all other parameters determining the true formation temperatures are under control. However, the geothermist should proceed pragmatically, i.e. he should also include less accurate data — as well as data from drilling companies with a lesser interest in geothermal problems — in his investigations providing that the density of data will be sufficient for a good geothermal evaluation of his area of study.

The results of the application of refined measuring techniques by geothermists show clearly that exact temperatures can be obtained only several years after the termination of drilling (see for instance Kappelmeyer 1961, Meincke 1966, Demaison 1973). Lachenbruch & Brewer (1959) made the statement that temperature data with an accuracy of $\pm$ 0.1°C could be obtained after 6 years, and that with an accuracy of $\pm$ 0.01°C only after as much as 50 years .However, this concept holds true only for 'dry holes' and it would be irrelevant for petroleum exploration unless these authors had proposed a method of extrapolation.

(1) Formation interval testing has shown that the temperature of the formation liquid will adjust to that of the formation within a comparatively short period, provided that the influx of uncontaminated formation water is considerably great in relation to the thickness of the reservoir interval. However, attention must be paid to the cooling effect of adiabatically expanding gas from reservoirs (Kunz & Tixier 1955) or source rocks (Heine 1962). Conversely, mud and mud filtrate disturb the natural temperature balance if the formation

is permeable. For this reason temperature logs for the location of top cement or lost circulation reveal only relative data (Hedemann 1968). Nevertheless, they give some information about changes in the thermal gradient (see chapter 3.2).

(2) Temperature data with a tolerance of $\pm$ 1°C are considered to be precise enough to be used for the models of crude oil formation and the temperature distribution in the North Sea region as described in this article. Furthermore, the exploration geologist is obliged to develop ideas about the formation of hydrocarbon deposits immediately after drilling the first exploration well(s) in a sedimentary basin, since he cannot wait several years for the stabilization of natural formation temperatures. Where he was successful, FIT's (Formation Interval Tests) or in an annual cycle shut-in pressure measurements will provide him with reservoir rock temperatures of an accuracy tolerance of $\pm$ 0.1°C at best. These will change within the years only for technical reasons: influx of higher temperature bottom water or of cooler injection water.

(3) The mud with an average temperature of about 30°C — on its way down — heats the bedrock in the upper part, and cools it in the lower part of the hole. After the circulation stops the assimilation process of the mud temperature to that of the formation in a given depth will operate according to the following equation (Lachenbruch & Brewer 1959):

$$T_m = k \cdot e^{\frac{t}{t-s}} + T \quad (2)$$

T_m = temperature measured after the time t
k = inclination coefficient determinable by plotting
T_m over $\frac{t}{t-s}$ on semi-logarithmic paper
t = time elapsed after reaching a definite depth respective after mud circulation stopped
s = drilling time = total drilling time ./. t
T = true formation temperature, graphically extrapolated for $t = \infty$

Equation (2) is clearly designed for measurements taken right after the termination of a well; hence, it is not applicable to the multitude of abandoned wells. Confining himself empirically to control measurements with an accuracy tolerance of $\pm$ 1°C in observation wells after 3 days as well as several weeks, Heine (1962) found that a temperature balance had been achieved within only 3 days.

Hedemann (1963) drew graphs of the temperature build-up curves for particular logging series in the Muensterland 1 well by plotting the mercury–maximum–thermometer values (T_m) of respective sondes and runs over the times (t) elapsed after stopping the mud circulation (Fig. 3). He had to determine the true formation temperatures graphically since he used arithmetic scales for both the temperature and the time axis. Fig. 3 shows that the estimation of the asymptote is more subjective as the temperature equalization time increases. Later, Hedemann (1967, 1968) thought that from the grade of conformity of the equalization curves conclusions could be drawn as to their degree of probability in uniform facies areas. Moreover, the examples selected from the results of Carboniferous well tests in Northwest Germany demonstrate that temperature measurement data tend to assimilate to true formation temperatures around 100°C as quickly as within one day, and to approx. 200°C within about two days. From that the conclusion can be drawn that Heine could have measured true formation temperatures after less than 3 days in wells of the Ruhr District, penetrating rocks of a similar facies.

Hedemann's graphic solution is mathematically unsatisfactory. Hence, his data as well as all other bottom hole temperatures (BHT) have been revised according to Timko & Fertl's (1972) method, in which, besides T_m, only the measurement time intervals have been utilized (Fig. 3). This method has the advantage of providing a mainly graphic solution including a minimum of calculation. The values gained by applying this method on impermeable bedrocks are 1.5–3.0% higher than the graphically determined ones (Fig. 3).

This solution requires three test values. In case only two values can be obtained — as at the final depth of Muensterland 1 — an exact determination of the true formation temperature cannot be made (Fig. 3).

The slopes of the linear functions plotted on semi-logarithmic paper indicate the heat conductivity of the bedrock at the test locations: directly if it is impermeable, and indirectly if it is infiltrated by mud. Gentle slopes mean low heat conductivity: an example is that of the coal-bearing Westphalian in the upper parts of the Muensterland 1 section. At a depth of 4000 m in the coal-barren Namurian (Fig. 3) the conductivity is higher. It decreases parallel to an increase in crystallinity of the clay texture and the increasing interstitial water content probably associated with it. The steepest curve of Fig. 3 is a characteristic of a limestone of high heat conductivity.

Steep curves are also encountered in zones of lost circulation where the temperature balance process is slow. In such cases even abrupt bends can be seen on semi-logarithmic paper. They can be explained by the sudden infiltration of more strongly warmed mud which is brought about by the swab-effect of the heaving of the sonde.

A fourth objection of the critics of borehole temperature tests refers to the reference depth. In most cases, the sonde does not quite reach the bottom and, in general, the thermometer is placed a few metres above the base of the sonde; details are provided by the well-surveying companies.

3.2 TEMPERATURE PROFILES (Fig. 4)

For the construction of a well temperature profile the surface temperatures are used in addition to the true formation temperatures. The geothermists put much effort into the determination of the so-called neutral depth, referred to as the depth where daily and annual temperature variations are no longer influential. For our investigations, secular variations, i.e. variations in a scale of geologic periods and changes of the reference level by erosion or subsidence

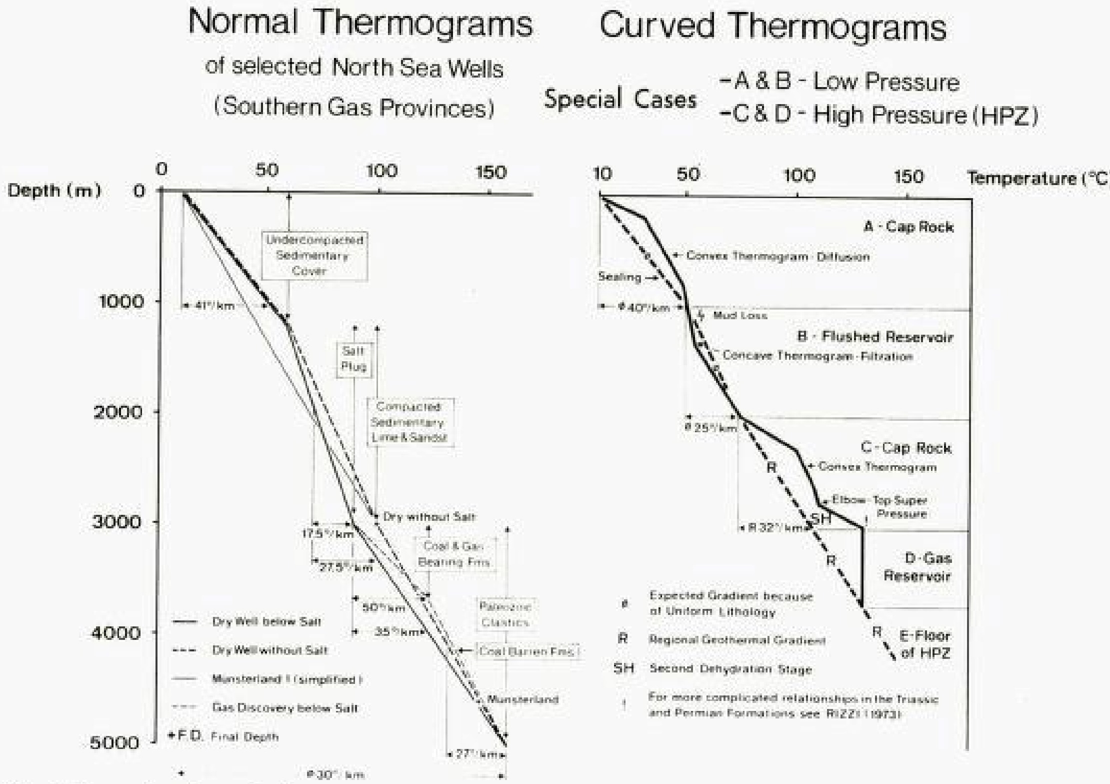

Fig. 4. Normal and curved thermograms.

(Fig. 10), are of particular interest. Relative to the accuracy tolerance of well temperature surveys, surface temperature changes of $> 1.0\,°C$ are of some importance. Taking into account that in the North Sea area nearly all utilized formation temperatures were measured at depths of approx. 3000 m, it becomes clear why the determination of the neutral depth (varying between 5 m and 25 m) is mostly neglected.

In the Ruhr District average surface temperatures (T_s) are determined according to the following equation taken from Heine (1962):

$$T_s = 10.76\,°C - 0.006 \cdot h \quad (3)$$

where h is elevation above msl in (m), so that T_s would be 10.0°C at an elevation of 126.6 m. Accordingly, average Ruhr District soil temperatures T_s vary between 10.0° and 10.4°C. Defant (1961) published sea-floor temperatures in the North Sea of 9°–10°C in water depths of up to 25 m, decreasing to 6°C at a depth of about 100 m. On the basis of these data and following the example of Braaf & Maas (1952), as well as of Harper (1971), a standard surface temperature of 10°C is used for the design of temperature profiles in all close-to-coast lowland or shallow water wells of the North Sea region. In all other areas data from climatological atlases or from previous workers (e.g. Quiring 1936, Creutzburg 1964) have been used.

Temperature profiles — also called thermograms by hydrologists (Bredehoff & Papandopulos 1965, Sejdidow & Gawrilow 1973) — are graphic presentations of temperature changes versus depth. From these diagrams the geothermal gradients and the interval gradients can be determined for any depth. According to the known equation (4) these gradients depend largely on the heat conductivity of the various rock units:

$$q = -\lambda \cdot \vartheta \quad (4)$$

q = heat flow ($cal \cdot cm^{-2} \cdot s^{-1}$), world average* of $1.5 \cdot 10^{-6}$
λ = heat conductivity ($kcal \cdot cm^{-1} \cdot s^{-1} \cdot {}^{\circ}K^{-1}$)
ϑ = geothermal gradient (°C/km), world average* of 25°C/km

Heat conductivity data for rocks of the area of study have been published by Benthaus (1959), Kappelmeyer (1961), Creutzburg (1964 and 1965) and Hueckel & Kappelmeyer (1966). From the 166 mineral heat conductivity values determined by Horai (1971), theoretical heat conductivities can be calculated by a factor analysis of all minerals contributing to a rock's composition. However, such calculations always yield higher values than the effective values of porous rocks dealt with by petroleum geologists. The latter have been investigated especially by Huang (1971), subsequent to work of Meinhold (1968) and Lewis & Rose (1970). Accordingly, the heat conductivity is inversely proportional to the geothermal gradient which itself is proportional to the bulk porosity as well as being inversely proportional to the grade of compaction, but proportional to the amount of oil or gas infiltration of the sediment. In low pressure gas reservoirs the geothermal gradient attains its maximum. Complications in permeable rocks are due to the fact that the heat flow is brought about not only by diffusion but, to a larger extent, also by convection (Kappelmeyer 1968, Meinhold 1968).

On the basis of this brief introduction to the principles of heat flow, the interval gradients encountered in the North Sea region will now be discussed (Fig. 4). For the most part, the thermograms reveal a characteristic subdivision into three or four portions. In the southern North Sea and adjacent areas high gradients are found predominantly above salt plugs of high conductivity. In the central and northern parts of the North Sea region they are caused by undercompaction of the Tertiary sediments. Low gradients are encountered in the middle parts of sediment sequences characterized by a considerably high percentage of carbonates or, especially in the southern North Sea, of evaporites. The gas source rock parts of the Carboniferous are responsible for higher gradients. Finally, the sterile 'basement' — speaking in terms of petroleum geology — reveals lower gradients again.

In the southern North Sea the geothermal gradient intervals are especially well marked in gas-bearing sections. Here, the production conditions are optional if in the Carboniferous layers the amount of gas-rich coal with a low heat conductivity is high, and if the Rotliegend reservoirs are covered by salt

* according to Lee & Uyeda (1965)

of high heat conductivity. The distribution of these reservoirs, however, must be surveyed by means other than geothermal methods.

The evaluation of the above information also enables the petroleum geologist to interpret the more complicated unpublished thermograms from oil-bearing or dry wells of the North Sea region. In this article only the curved thermogram of Fig. 4 may be discussed since they describe cases of pressure-bound convection. Case C — determination of over-pressure zones — is familiar to any petroleum engineer from the publications of Lewis & Rose (1970) and Timko & Fertl (1972). The reader's attention may be drawn to Rizzi's (1973) findings which have not yet been integrated into the Anglo–American literature. He found that not only the undercompaction of young clastic sediments but also formation water locked in clastic intercalations of evaporites could produce over-hydrostatic pressures. This holds true especially for the 'Buntsandstein' of the southern North Sea region.

Cases A and B rely on the experiences of the hydrogeologists (Sejdidow & Gawrilow 1973) and of the present author. In cases of vertically upward infiltration, i.e. especially through accelerated diffusion, and in cases of lateral migration in inclined beds as well as in cases of overlying permafrost soil, the thermograms are conversely bent. The perfect sealing capacity of a lithologically homogeneous impermeable bed is indicated when the thermogram is a straight line. Concave thermograms indicate downward infiltration, i.e. under hydrostatic pressure in a reservoir rock below dehydrated clays. Thus, lost circulation and also occasionally increased salinities of the underlying beds can be predicted. Of course such predictions can be made only if the wells are logged comparatively often, as, for instance, the first exploration well in a sediment basin, or if changes of the thermal gradient can be determined from continuous measurements of the mud temperatures. This is now a standard procedure applied by the mud engineers for the detection of over-pressure zones in undercompacted Tertiary sediments of the North Sea region.

To some extent surface soil temperatures reflect the subsurface geothermal conditions because of upward heat diffusion. This rule is applied for the geothermal prospection in marginal areas of North Sea (Paul 1935, Poley & Steveninck 1970 and 1971, Geertsma 1971). Fig. 5 shows the oldest ever published soil temperature profile above a salt plug of the Germanic basin; determined from temperature measurements from two-metre deep holes (Paul 1935). In the lower part of the figure, a profile is presented which was drawn later relying on results from deep salt and oil wells and from potassium mines (Bentz 1949, supplemented by oil field data according to Richter-Bernburg & Schott 1959).

Furthermore, the salt plug rocks have been classified after their heat conductivities, based on Creutzburg's (1964) measurements (supplemented by the estimated value for salt clay). On the surface above the plug, increased soil temperatures can be measured in zones of increased heat conductivity unless they are sealed by clay. Above subsurface pockets of clay or faults, soil temperature decreases of 2–3 °C are common. Paul had matched the soil temperature minima with the flanks of the salt plug. This holds true for a

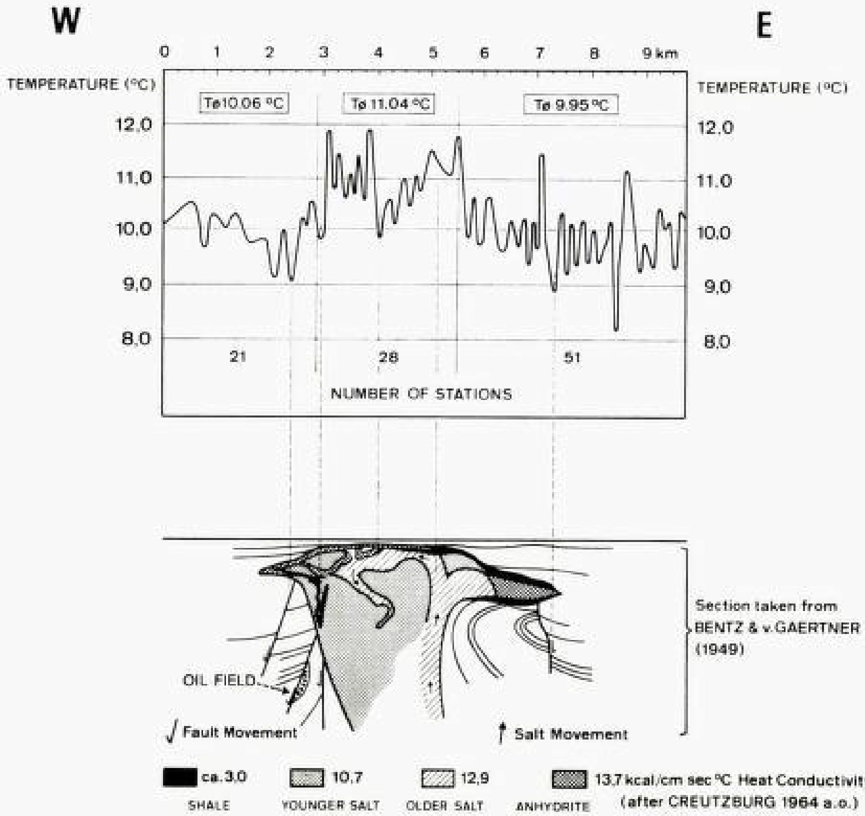

Fig. 5. Soil temperature survey over a salt plug near Hanover, Germany, by M. Paul (1935).

depth of approximately 500 metres because the salt overhangs are hard to separate from the surrounding sediment, as illustrated in Fig. 5. Accordingly, for a temperature analysis Selig & Wallick's (1966) cylindrical salt plug models can be applied even to salt mushrooms.

Near Groningen and in the Peel-Horst area (in the south-easterly continuation of the Oostzaan High, see Fig. 6), Poley & Steveninck (1970) have located salt domes and faults at depths of more than 1000 metres. As in the case of the Hannover salt plug they were indicated by temperature increases of 1–2°C, although these were unnoticeable from surface features. Originally they measured the temperature at a depth of 2 m. Later, however, they realized that anomalies can be mapped even from measurements at depths as shallow as 5 cm.

Recent remote sensing techniques allow the reconnaissance of temperature anomalies of this order from satellites. Even on simple black and white satellite photographs of N.W. Germany — published by Heuseler 1973 — some of the numerous salt plugs in this region, as well as the Northern rime of the Lower Saxonian Basin, seem to show up. Thermal infra-red pictures, however, have not yet been taken.

Geertsma (1971) explains the temperature anomalies above faults by changes of the heat conductivity along the fault planes: a decrease caused

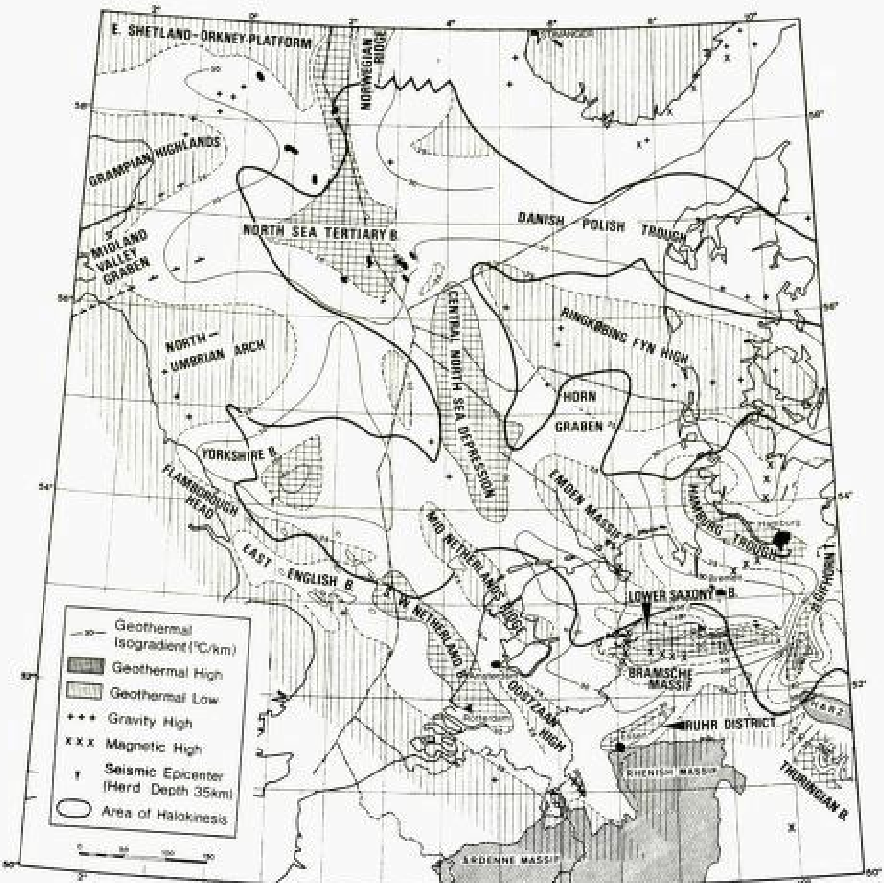

Fig. 6. Geothermal gradients, North Sea area.

by rupturing of highly lithified rocks; an increase by compaction of less lithified rocks due to shear stresses building up along the fault. An increase should also result from mineralization along the fault. These interpretations, however, are restricted to heat diffusion. Due to prevailing 'open' tensional faults in the area of study, convection is common and increases of temperatures are proportional to increases of permeabilities; in the Ruhr District and South Limburg we are even dealing with hot brines ascending along faults.

3.3 GEOTHERMAL ANOMALIES OF THE NORTH SEA REGION (Fig. 6)

Temperature surveys of the North Sea region have been published only in a few cases. From adjacent areas geothermal isogradient maps of the Saar District have been presented by Hückel & Kappelmeyer (1966), and of the Thuringian Basin by Meincke (1966). The first specific geothermal gradient map of the North Sea region was presented by Harper (1971). It is based on mean gradients, which were determined from bottom hole temperatures (BHT) building up over a maximum time interval of 48 hours. This map covers the North Sea up to 59° N, and the Dutch onshore lowland. A maximum accuracy was obtained by plotting 2°C/km intervals (see above 3.1).

The author's map (Fig. 6) is also based on mean gradients. Offshore, they have been gained exclusively from BHT's, mostly from wells with depths of around 3000 m. True formation temperatures could be extrapolated from BHT's only in rare cases when operating companies took sufficient interest in geothermal problems. For the author's map, 2°C/km intervals were also utilized; however, the map was simplified by plotting only 5°C/km intervals (Fig. 6) which does not, in the author's opinion, alter its essential value.

Hardly any difference can be observed between the author's and Harper's map concerning Norwegian and British waters. This is good support for his assumption that gradients can be determined from unextrapolated temperatures with an accuracy tolerance of less than 10%.

In many wells, according to explanations given in chapter 3.1, standstill periods of 48 hours were even sufficient for the regeneration of true formation temperatures. Furthermore, for these parts of the North Sea the author had available only a limited amount of supplementary data. Of these data the Ekofisk field reservoir temperature of 129.4°C from an average reservoir depth of 3.170 m has been of special interest because the calculated gradient of 39°C/km as well as the published gradient of 36.8°C/km (OGJ) could be effectively incorporated into the map.

Apparently, in the Dutch, German and Danish waters as well as in the adjacent continental areas, the author had many more data available than Harper. In addition, the author was able to incorporate preferentially precise measurements from various oil fields and mines in this area. The majority of these data have been taken from the publications of Quiring (1936), Braaf & Maas (1952), Fabian (1955, 1963, Benthaus (1959), Patijn (1961, 1964), Heine (1962), Hecht et al. (1962), Stheeman (1963), Hedemann (1963, 1967, 1968), Creutzberg (1964), Lee & Uyeda (1965), Simmons & Horai (1968), Kimpe (1973) and Tunn (1973).

Lee & Uyeda's (1965) data from the British Isles and Swanberg et al.'s (1972) data from Norway have not been incorporated into the map because they are outside the North Sea Basin, and because their number is insufficient for a special survey of these inhomogeneous continental areas. The British values are around 32.3°C/km, the Norwegian ones around 15°C/km. The latter gradients vary in a range between 9.45° and 22.39°C/km (Swanberg, personal communication).

In the southern North Sea and the easterly adjacent continental areas of our map, the density of data is sufficient; whereas in the central part of the North Sea — south-west of the Montrose–Ekofisk line — it can be considered fairly sufficient. From the Norwegian part of the North Sea, north of the 58th latitude, only scattered data are available.

In addition to the isolines of geothermal gradients, regional geological informations, e.g., basement highs and depocentres, have been entered in the map including information from structural analyses of Ahorner et al. (1972), Bartenstein et al. (1971), Boigk (1968), Gunn (1973), Kent & Walmsley (1970), Lindström (1971), Ramberg (1971), Sorgenfrei (1969) and Thiadens (1963). Comparisons with other articles in this volumes indicate that the British nomenclature for structural units in some cases has been changed; however, this does not require a revision of the essential structural information.

The isolines of geothermal gradients delineate the regional geological units. Usually thermal highs coincide with subsiding areas, whereas thermal lows can be encountered above basement highs; this implies that elevations of the basement are inversely proportional to the geothermal gradients. For this reason the North Sea region can be considered as an inverted geothermal realm, contrary to normal geothermal realms like, e.g., the Rocky Mountains (Klemme 1972) or the Sirte Basin. South of the northern margin of the Variscan orogen is a normal geothermal realm (Fig. 6, detailed information concerning the Thuringian Basin).

Second order structural elements such as anticlines and synclines yield normal temperature distributions within the basins of an inverted temperature realm. This is even more evident in the Ruhr District, which has been more thoroughly investigated than offshore areas in the North Sea region. In the Ruhr District, for instance, deeper shafts can be sunk in synclines because they have lower gradients than the anticlines. The major reason for high gradients in anticlines must be selective heat flow, parallel to the bedding planes. In addition, convection along faults, which is indicated by hot brines, might be taken into account. Such faults coincide with anticlines as in South Limburg and in the Peel Horst. Also, the Ekofisk anticline appears to be distinguished from its surroundings — with a temperature gradient of approximately 35°C/km — by having a considerably higher gradient of approximately 39°C/km. However, this observation can not be generalized as long as reliable temperature values are available only from reservoirs and structural highs of the North Sea. As previously mentioned in chapter 3.2, heat flow parallel to bedding planes, as well as the special situation over salt domes with high heat conductivities and heat-damming below insulating undercompacted clays, may be of some importance.

Thus, details of the geothermal situation can be explained; but what about a synthesis? According to Ahorner et al. (1972) the focus of the 1931 Doggerbank earthquake was at a depth of 35 km (Fig. 6), indicating the crustal thickness in the whole North Sea region. Relying on Neuhaus's (1968) assumption that temperatures at the Moho are around 900°C, a mean gradient of approx. 25°C/km can be extrapolated for the North Sea crust; this corresponds to gradients over basement highs in our map.

Heat conductivities in such consolidated basement blocks outside the coal-bearing Variscan orogen can be considered uniform. In the latter area, however, and in areas of more recent subsidence with a considerable content of heat insulating Jurassic and Paleocene source rocks as well as undercompacted clays (especially Eocene tephra), thermal gradients are higher. Contrary to the Upper Rhine Graben (Haenel 1970) and other rift systems of the earth, crustal thinning does not seem to be responsible for the increase of gradients; especially since mean gradients of 35°C/km in subsiding areas of the North Sea are low in comparison with those of 60°C/km in genuine rifts. So the explanation through heat-damming remains feasible, as earlier suggested by Harper (1971).

The situation in the southern and eastern parts of Harper's map was so different to our's that a revision of his interpretation appeared to be necessary:

a) Harper offers no structural or stratigraphic explanation for the thermal anomaly east of Yorkshire. Yet reference should be made to the earthquake centre at its southern margin, as well as to the coal content of the northern shallow syncline, which subdivides geothermically the Central North Sea High ('Northumbrian Arch').
b) According to Harper the Permian evaporite basin is characterized by a uniform E–W striking negative thermal anomaly. In Fig. 6 this trend can no longer be recognized since it appears to be dismembered by a series of NW–SE striking anomalies. Furthermore, halokinesis covers a much larger area within abundant positive heat anomalies.
c) The central North Sea depression can also be traced geothermically into the area of the Dutch oil discovery F 18–1 from where it continues to Lower Saxony through NE Holland.
d) The central North Sea depression is separated from the SW Netherland Basin (which possibly continues into the Rhine Graben through the 'Niederrheinische Bucht' and the Rhenish Massif) by the Mid-Netherlands Ridge.

More structural details, which are not actually discussed here, can be gathered from the map. High gradients of up to 55 °C/km in the Lower Saxony Basin and the Gifhorn Trough cannot be explained by heat-damming alone. In the case of the Lower Saxonian tectogene (Boigk 1968) a supply of fossil heat by upwarping of the basinal infill from a higher to a lower temperature level could be taken into consideration (Fig. 13). In Fig. 6 several gravimetric and magnetic highs have been entered on the map which, as in the case of the Oslo Graben (Ramberg 1971), indicate crustal thinning and the formation of heat domes during the geologic past. The relationship between recent and previous temperature fields of the North Sea region will be discussed in chapter 5.

4. Determination of maximum temperatures (Figs. 7-9)

There are many methods and even more literature on this particular subject. In 1971 the Commission on Petrography of Organic Matter in Sediments and Application to Geology was founded by coal petrologists of the ICCP. Initially the purpose of this Commission was the measurement of coalification ranks encompassing analyses of hydrocarbons (particularly n-paraffins and CP-indices, cit., e.g., Welte 1965, Leythaeuser & Welte 1969, Leplat 1973, who studied samples from the discussed area, among others) and of their melting or boiling points (Jacob 1967, Silverman 1971). For calibration there are utilized: shale mineralogy (Fig. 1), mineral temperatures, electron-spin-

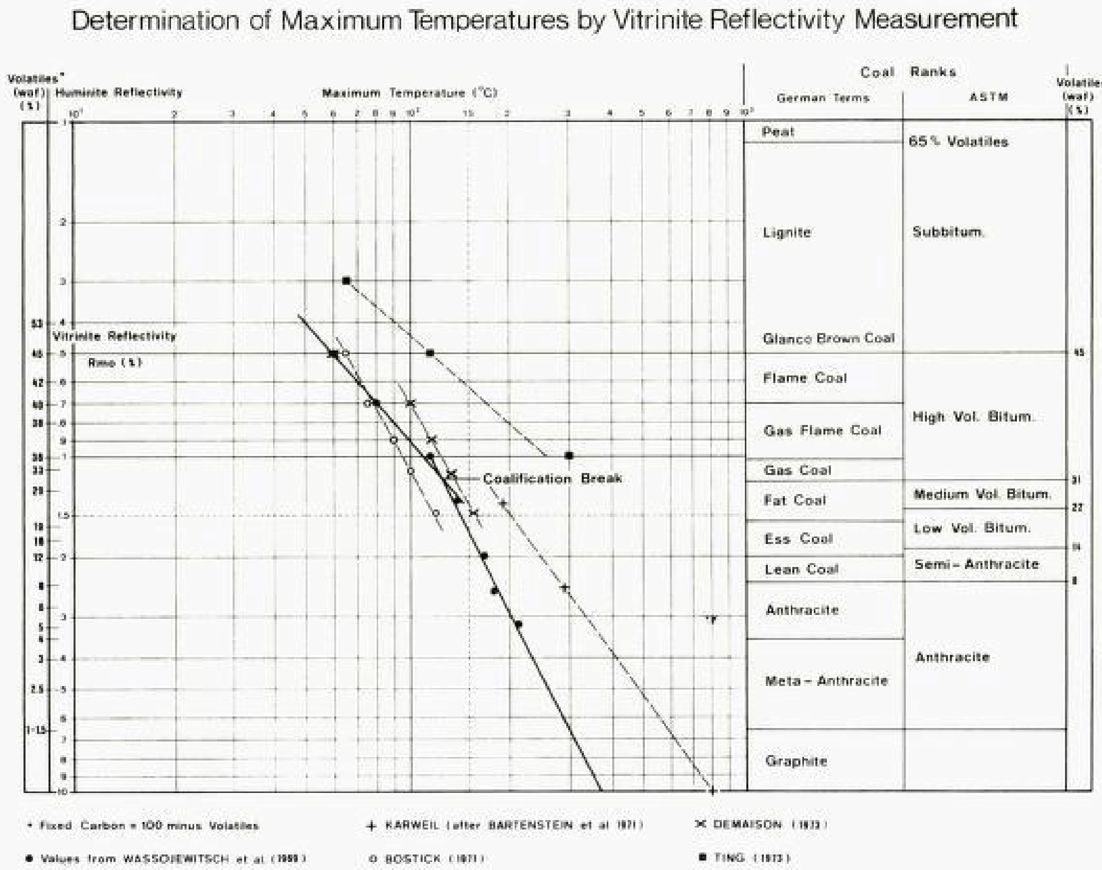

Fig. 7. Determination of maximum temperatures by vitrinite reflectivity measurement.

resonance (ESR, Pusey 1973), pyrolysis (Wehner & Welte 1969, Leplat 1973, Somers et al. 1973) and fission-track-ages (Wagner & Reimer 1973). According to our experience, the usefulness of the methods is limited to those which are based on a laboratory treatment by which the physicochemical state of the sample is modified and in which equation (1) as well as the geologic time factor are not taken into account (cit., e.g., Hanbaba & Juentgen 1969). Consequently, pyrolysis yields maximum temperatures, which are always too high, and ESR temperatures compare to those determined from coalification ranks only for samples from young and continuously down-warped basins.

As early as 1873, Hilt, a mining engineer from Aachen, published his coalification rule according to which the volatile percentage of pit coal decreases with increasing depth: near Aachen by 1%/100 m. Today, the coal petrographers prefer the optical measurement of the mean reflectivity under oil of polished vitrinite-bearing samples. The reflectivity values are correlated to the volatile percentage of the coal as determined in the laboratory. Compared with other methods the vitrinite reflectivity measurements have the advantage of not being bulk analyses for all coal macerals and all organic constituents

In addition, they allow a differentiation of the various coalification ranks of vitrinite. The youngest population can be interpreted as autochthonous only if the constituents are not exclusively represented by reworked matter.

From the bibliography of the above-mentioned Commission the following references are recommended for a basic study of the matter and/or for a study of the particular relevancy to the North Sea: Alpern (1969, 1970), Alpern & M. Teichmueller (1971), Alpern et al. (1972), Ammosov (1967), Bartenstein et al. (1971), Boigk et al. (1971), Bostik (1973), Castano (1973b), Caye & Ragot (1972), Hedemann & R. Teichmueller (1966), Jacob et al. (1970), Jones et al. (1972), Karweil (1973), Koch (1970), Kuyl & Patijn (1961), Paproth & Wolf (1973), Patteisky & M. Teichmueller (1960, 1962), Robert (1971, 1973), M. Teichmueller (1971), M. Teichmueller & R. Teichmueller (1971a) and Wolf (1969, 1972).

The diagram of Fig. 1 shows that the vitrinite rank runs parallel to the maturation of liquid hyrdocarbons at least for mean reflectivity values of up to about 1.35% ('iso-aromatization grade'). If the order parameters are known, the time-factor can be derived from Fig. 1. Such a procedure permits the time-calibration of Fig. 7, which is used for the determination of maximum temperatures from mean vitrinite reflectivities. Data from published approaches to a calibration of the coalification-rank thermometer have been plotted on double-log paper for the purpose of presenting a graphic analysis. It appears that the calibration values of Wassojewitch et al. (1969) are located on two straight lines with differing slopes. The point of intersection is situated at about 31% volatiles, i.e. approx .1.2% R_m. Apparently, this point is equivalent to the coalification break of Stach (1953) and Patteisky & M. Teichmueller (1962) which has been defined as the border-line between weak and strong coalification as well as between gas-rich and gas-poor coal. Differing slopes indicate differing reaction-velocities. The 'deadline for oil' seems to coincide with this coalification break.

Since the calibration values published by other authors do not show the coalification break, two additional approaches will be discussed (Figs. 8 & 9).

(1) Density values for coal of varying coalification ranks have already been compiled by Fülöp (1967). They were supplemented with data published by Ruhrkohle (1969) and with the precise density value of clean graphite (Correns 1949), and plotted in such a manner as to indicate the relationship between subsidence and alteration. In Fig. 8 the coalification break is defined by the blurred density minimum. A distinct differentiation, however, can be found only beyond the anthracite stage. Thus, from the diagram it can be concluded that the Formation Density Log can be used, if at all, for direct coalification logging below the liquid window only.

(2) A more successful approach to the problem was made by Stahl (1973, see also Boigk et al. 1971) in the Laboratory for Isotope Studies of the Federal Geological Survey of Germany in Hanover. It is based on the

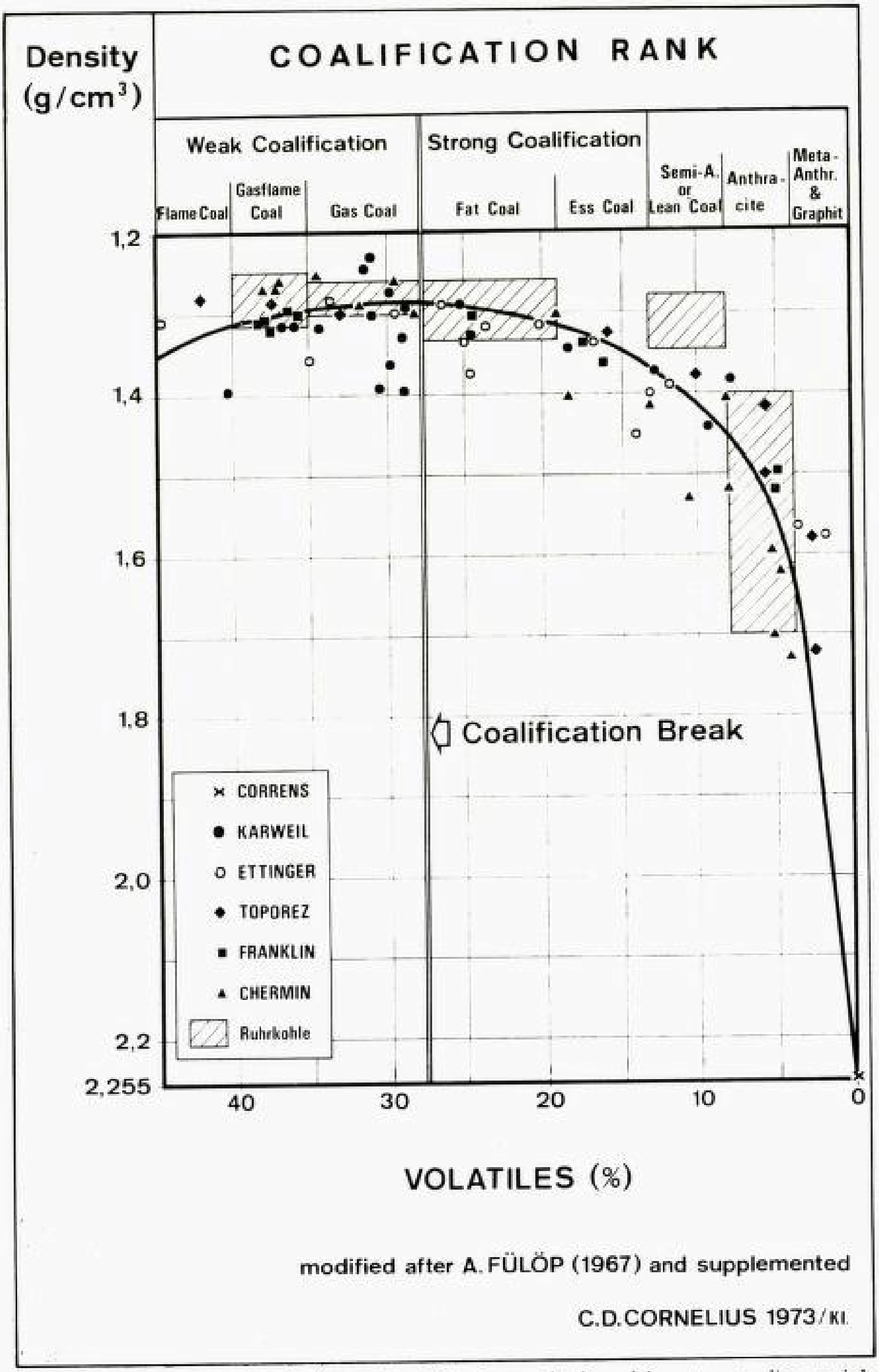

Fig. 8. Coal density versus volatiles and coalification rank, from laboratory studies, mainly after Fülöp (1967).

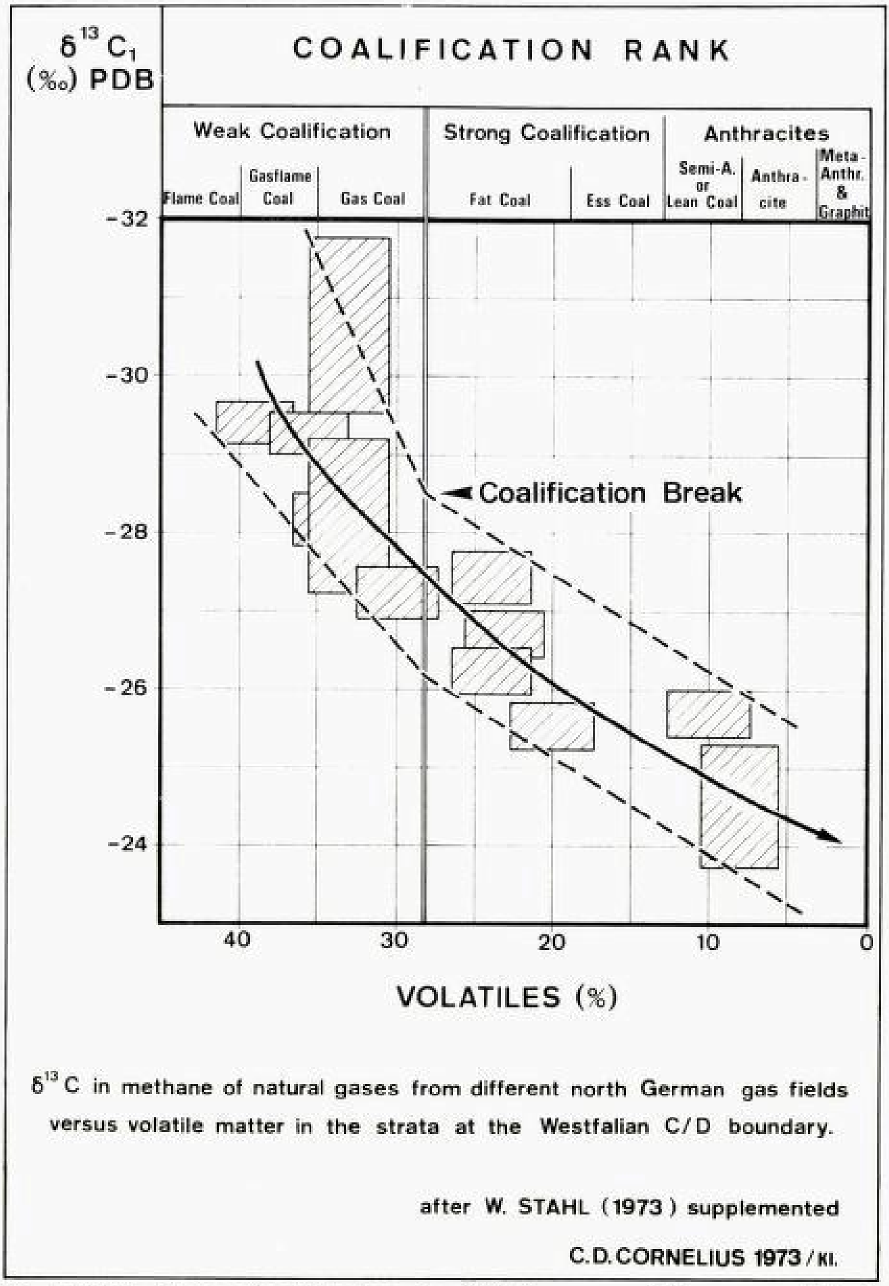

Fig. 9. Carbon isotope ratios of methane from NW German gas fields versus volatiles in coal seams near Westphalian C/D boundary, mainly after Stahl (1973).

correlation of volatiles percentages with carbon-isotope ratios of methane (Fig. 9). The δ^{13}C-ratio increases proportionally to the coalification rank, i.e. the gas becomes the heavier, isotopically, the more the coal is altered. This trend slows down in the range of intensified coalification (Fig. 9) thus confirming the results of Fig. 7. Although Stahl assumes a continuous change of the isotope ratio, a possible break is indicated in Fig. 9, again coinciding with Stach's (1953) coalification break. (Silverman's (1971) correlation of carbon-isotope ratios with the boiling-temperatures of the various constituents of crude oil proves other discontinuities but could not be evaluated for our purposes because of abolition of the time-factor.)

Accordingly, the validity of the rule of the coalification break and, at the same time, the Russian calibration appears to be approved. The author would like to make the following additional remarks:

a) Bostik's (1973) maximum-temperatures, assumed for North West German coals with evidently longer maturation times than those of the Russian samples, appear to be based on reaction rates which are too low (because they are based on Karweil's frequency factor for reactions in a solid state). However, he already had doubts about the reliability of maximum temperatures derived from vitrinite reflectivities of less than 0.6%. Demaison (1973) made the same mistake for shorter coalification times.

b) The shortest coalification times were found by Ting (1973). His samples from the Upper Miocene of Texas represent a maturation stage from before the coalification break and indicate still greater reaction velocities than the Russian samples, possibly because of better permeabilities. Ting's uppermost value was obtained in the laboratory.

c) Karweil's maximum temperatures from the contact-metamorphic aureole of the acidic laccolith of Uchte within the Lower–Saxony tectogene (see Bartenstein et al. 1971) have been calculated with shorter reaction-times after the coalification break than those on which Wassojewitsch's calibration values appear to be based. Faster reactions seem to be possible over batholiths because the related tensional faults allow thermal convection (chapter 3.2). A similar case was already suggested by Teichmueller & Teichmueller (1966) for the laccolith of Erkelenz at the SE corner of the Oostzaan High (Fig. 6), where exposure times of 5 m.y. (and more) were calculated. In the case of Uchte no time factors were published. If, however, the temperature values of 190° and 290°C from Fig. 7 are checked by transferring the respective R_m-values of 1.4% and 2.48% into Fig. 1, time-values of 45–40 m.y. will be obtained which are irrelevant to batholiths. From an earliest possible intrusion age of 100 m.y. — during the Austrian orogenic phase — and from the fact that the laccolith of Uchte was already cooled at the early Upper Campanian, 75 m.y. ago, the conclusion can be drawn that the thermal event was effective for less than 25 m.y. If, finally, the calculation is based on a temperature of about 800°C for anatexis and consequently a straight line parallel to the line integrating the Russian values is drawn in Fig. 7, maximum temperatures of 300°C and 400°C, respectively, result from the above-mentioned R_m-percentages. Hence, there is good proof that the heat event associated with the intrusion of this laccolith was short-lived.

Once more, these case histories from the broader North Sea region prove the significance of the determination of the time factor. If it cannot be obtained from the reconstruction of the burial history, other methods of maximum temperature determination are to be applied. Although the experiences with such methods are limited so far, the following observations may be summarized.

I. The vitrinite-thermometer of Fig. 7 will be supplemented by humanite–reflectivity measurements and by fluorescencse spectra of sporinites for weak coalification ranks (Ottenjann et al. 1973).

II. Methods of colorimetric and transparency investigations of sporinites have been introduced by Correia and by Staplin, both in 1969. These are optical methods like the vitrinite reflectivity method. Although their data are less accurate, they can be evaluated by computer processing. To the author's knowledge, time-corrections can be made only after calibration by vitrinite reflectivities. Apparently, scientists are oblivious to the fact that this method was already proposed by Kirchheimer as early as in 1934. He calibrated the palynomorphs' conservation stages with experimental temperature values which were too high, again because of a neglecting of the time factor.

III. Mineral temperatures determined from authigenic minerals are very useful as, for instance, the temperatures of more than 300°C (Bartenstein et al. 1971, Stadler & R. Teichmueller 1971) indicated by pyrophyllite from Oxfordian sediments above the Bramsche Massif (Figs. 6, 12 & 13); especially if they can be time-corrected according to equation (1). The zeolite-zones in the sense of Winkler's (1967) mineral facies concept can be used as a thermometer (Kisch 1969, 1973) only if they are calibrated by vitrinite reflectivities (Castano 1973a) or after a reconstruction of the burial history of Kisch's samples, which has yet to be undertaken.

IV. Temperatures from hydrothermal mineral veins are thought to be very reliable for the study of crude oil formation (Florowskaja & Pikowskij 1973). However, in our opinion they are of minor importance because they are limited to 2-dimensional anomalies. Certainly, the veins indicate thermal convection, but generally they produce only narrow zones of thermal contact metamorphism. This could be proved from ESR-measurements along basaltic dykes of the Lower Rhine valley carried out by Fauth from the Ruhrkohle Research Center in Essen. Since they affect the adjacent rock only over short distances, they produce atypical mineral temperatures which, referring to the concept of Fig. 2, appear to be too high. An exception may be made for mercury because of its mobility which, indeed, is quite similar to that of the hydrocarbon gases, as suggested by Dikenstejn et al. (1973, see also Tunn 1973) from an investigation of the Rotliegend gas reservoirs of the southern North Sea. This may also be conclusive for inert gases like nitrogen and helium (Müller et al. 1973). On the other hand, the mineralization of the ore veins within the Lower Saxony gas field Rehden (Fabian et al. 1957, for location see Fig. 12) most probably started already at temperatures above 170°C, which have been calculated from the adjacent fat coal seams.

V. For these reasons, the organic inclusions of hydrothermal minerals as described by Mueller (1964, 1973) from Derbyshire/England and elsewhere, are considered to be overtempered compared to their environment and — contrary to his suggestions — their temperatures are not conclusive for the formation temperatures of any oil accumulation; not even for the smallest British oil fields.

VI. Statistical comparisons of maximum-temperatures determined from various methods are, so far, insignificant because the methods and their respective data are incompatible unless they were calibrated by vitrinite reflectivities. Comparative studies of mineral facies, illite crystallinity and coalification ranks remain promising for the future (Frey & Niggli 1971).

5. The geothermal history of the North Sea region (Figs. 10-13)

After the determination of a sufficient number of maximum temperatures from a well, a paleo-thermogram (Fig. 10) can be drafted following the procedure explained in Fig. 3. In some cases, uniform paleo-interval temperatures can be observed which show no gradient. Such interval temperatures may be explained by homogeneous lithologies and their uniform heat conductivity characteristics (equation 4). Jones et al. (1972) discovered in onshore and offshore Northumberland and Durham that Carboniferous coal measures show no thermal paleo-gradient where they are overlain by heat-insulating shales, but distinct gradients where they are overlain by heat-conducting sandstones. Differing observations were published by Patteisky & M. Teichmueller (1962). In summary, the interdependency of vitrinite reflectivity and lithology is not completely understood.

The paleo-thermograms of Fig. 10 differ from each other and from the actuo-thermogram by their slopes. Six cases are possible; all have been confirmed by studies in the broader North Sea region and elsewhere:

(1) Case P — present temperatures = maximum temperatures: in continuously down-warped basins, e.g., U.S. Gulf Coast and main parts of the Central North Sea basin and Graben.

(2–5) Cases L, M and B — maximum temperatures > present temperatures: uplifting after reaching the maximum paleo-gradient(s) at the margins and elevated areas of the North Sea Graben.

(2) Case L — paleo-gradient < actuo-gradient, temperature difference to be recognized in the upper parts of sections only; in lower parts cit. (1), e.g., Oklahoma (Pusey 1973).

(3) Paleo-gradient = actuo-gradient: appears to be possible along basin flanks, possibly realized in S. Texas (Pusey 1973).

(4) Case M — paleo-gradient > actuo-gradient: the most common case outside the central graben zone (examples to be discussed below).

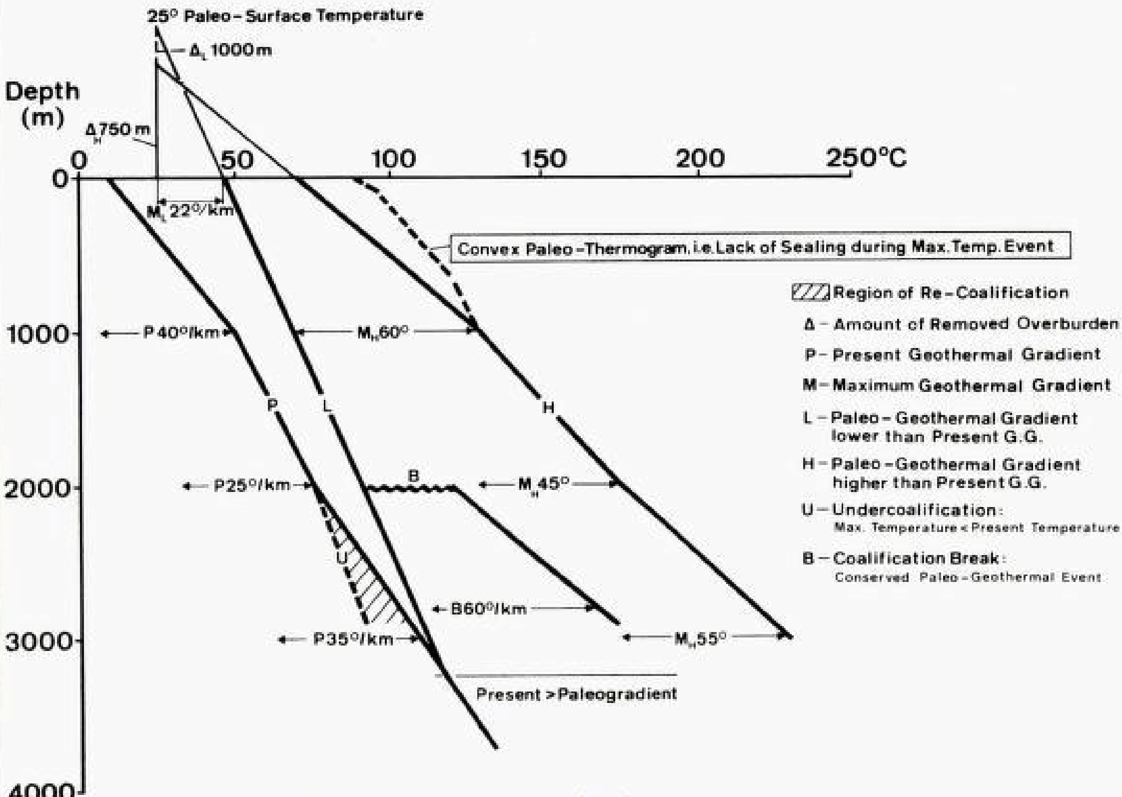

Fig. 10. Comparison of present with paleo-temperatures.

(5) Case B — earlier paleo-gradient > later paleo-gradient: combination of case L in the upper part of the sections and of case M in the deeper parts, i.e. a paleo-thermal event encountered in the well; examples cit. (4).

(6) Case U — maximum temperature < actuo-temperature, i.e. the present temperature prevailed for a shorter effective time than the maximum temperature, so that the final maturation rank was not yet reached: 're-coalification' at Groningen (Patijn 1964), called 'main-coalification' by Hedemann (1967); another example is the current formation of gas in Bunter reservoirs east of Yorkshire (Fig. 11).

Other compilators may, of course, consider (3) as a subcase of (2); or (5) as a subcase of (4), and (6) as a subcase of (1). For cases (2–5) the paleo-heat flow can be calculated according to equation (4). If good approximations to both the heat conductivity of the eroded section and the paleo-surface temperature at the time of a paleo-thermal event can be achieved, the amount of removed overburden can be graphically determined from Fig. 10. Consequently the time of the paleo-thermal event can be determined from the paleo-thermogram.

Some representative paleo-thermal events of the North Sea region will be

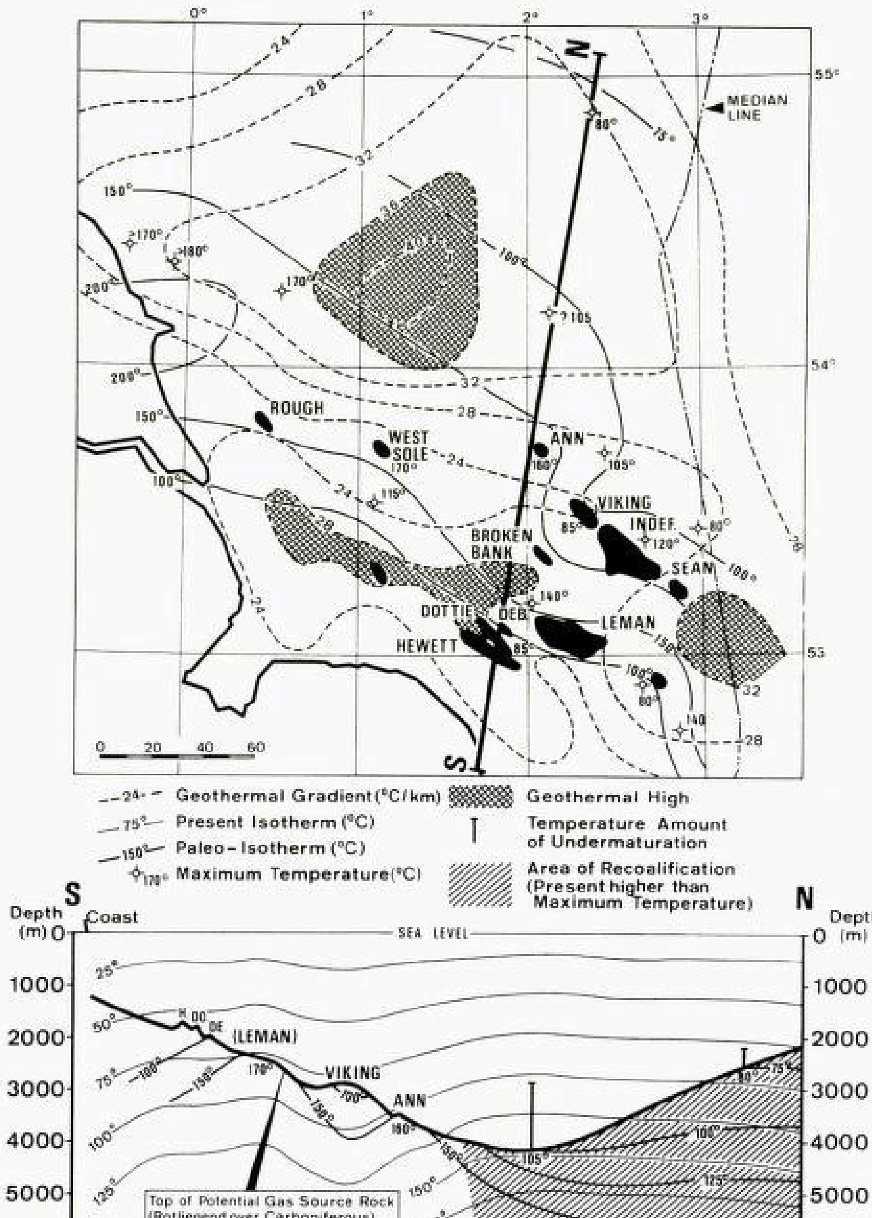

Fig. 11. Present and paleo-temperatures, East Anglian gas province.

discussed in chronological order to give an outline of the thermal history of this region.

a) The Palaeozoic history of the North Sea region was interpreted by McKerrow & Ziegler (1972). More details on its evolution during the early Palaeozoic era have been provided by Dewey (1971), Gunn (1973) and Hallam (1973), especially for the Midland Valley graben. The paleo-gradients for this era can be determined by mineral temperatures only. Following the hypothesis that the Grambian orogeny was associated with subduction of the Proto–Atlantic crust under the eastern margin of the Canadian Shield, a paleo-thermal gradient of about 50°C/km could be presumed from comparison with the recent Andean thermal realm. For the Early Caledonian (Wurm 1973) or Taconian orogeny — with axial strikes of around 100° in the Stavanger area — which might be the result of collision of Gondwanaland with the Baltic Shield, the author assumes a gradient of more than 30°C/km based on the shallowness of folding. The main Caledonian orogen, which is thought to be the result of a subsequent collision of the Baltic and the Canadian Shields, displays mineral temperature gradients between 17.5° and 25°C/km, i.e. of an intermediate range in comparison with the 'hot' Hercynian mountains and the 'cool' Alps (Zwart 1967).

b) The Caledonian molasse, i.e. the Old Red sediments of the Orcadian basin, requires an investigation by the experts in organic metamorphism. However, the number of maximum temperature data is not yet sufficient to prove an estimated paleo-gradient of about 35°C/km.

c) The status of investigations in the Hercynian area is much better than in the Orcadian. Based on the studies of Wolf (1969, 1972) and Paproth & Wolf (1973) as well as Weber (1972), even pre-orogenic gradients of about 60°C/km could be estimated for parts of the Rhenish geosyncline. During the orogeny, heat waves migrated through the geosyncline starting from the Bohemian Massif and directed towards the Ruhr District (Krebs & Wachendorf 1973), which produced heat domes characterized by gradients up to 250°C/km (Zwart 1967, see also Hall 1973). But gradients of 150°C/km could be established even for the synclines calculated from coalification ranks published by Paproth & Wolf 1973).

d) Even better data are available from the Hercynian molasse, the Upper Carboniferous coal measures of the mining districts. Kuyl & Patijn (1961) determined a paleo-gradient of 50°C/km from South Limburg. The same value was derived from Thiadens' (1963) coalification ranks for the Oostzaan High (Fig. 6). North-east of this, Corle 1 revealed a paleo-gradient of less than 40°C/km, whereas Zeddam 1 yielded only 22°C/km. The high paleo-gradients are restricted to NW–SE striking highs which most probably formed not earlier than the Mesozoic. Patijn's (1964) re-coalification in the area around Groningen becomes clear by the assumption that the low gradients are the original ones. R. Teichmueller (1973) estimated the paleo-gradient of the Ruhr District to be as high as 70°C/km. This gradient must have been developed during subsidence providing that the coal measures show almost uniform coalification ranks through anticlines and synclines. For instance, coal seam 'Sonnenschein' is characterized by the level of the coalification break (Fig. 7) almost throughout (Patteisky & M. Teichmueller 1962).

e) Pb–Zn ore mineralization of Asturian age, i.e. post-molassian open faults in the Ruhr District (Pilger 1956) indicates gradients of around 100°C/km. According to the definition given under chapter 4. IV such mineralization is brought about by two-dimensional convection which has no effect on the coalification rank of intersected coal beds.

f) During the Mesozoic era, the Eurasian Plate drifted over the northern hemisphere as demonstrated by Dietz & Holden (1970) and other workers. Following Turcotte & Oxburgh's (1973) membrane-theory, we may assume that the formation of the Viking Graben began during the Lower Jurassic at a paleo-latitude of 45°N. The graben extended progressively southward, but discontinuously, without acquiring a thermal identity — as defined in Fig. 6. (The Upper Rhine Valley Graben opened at the same latitude during the Early Tertiary.)

g) In the southern North Sea, the central graben is bounded by NW–SE striking structures (Fig. 6). From these, the East Anglian gas province is geothermically well studied (Fig. 11). In addition to the present temperatures the maximum temperatures have been calculated based on Robert's (1971) coalification ranks. The Rotliegend gas reservoirs are restricted to a zone of maximum temperatures from 100°C up to 170°C at the top of Carboniferous source rocks. The maturation studies are not sufficient for precise calculation of paleo-gradients but from a temperature profile a gradient around 65°C/km may be conjectured. This gradient should have been reached when the Carboniferous of the Leman field was covered by approximately 1850 m of sediment.

Referring to Kent & Walmsey's (1970) information on Leman and other gas fields, the paleo-thermal event was tentatively dated as early Middle Jurassic. Should this assumption be correct, the positive heat anomaly below the East Anglian Basin, i.e. the Sole Pit trough, would have built up simultaneously with the opening of the Bay of Biscay and the possible onset of sea-floor spreading in the southern North Atlantic.

Situated to the south-east is the Oostzaan High (Fig. 6) and, at its south-eastern corner, the above-mentioned laccolith of Erkelenz near Aachen. This was dated late-Hercynian by Teichmueller & Teichmueller (1971b), but was considered to be of Permian or Cretaceous age in earlier papers (Patteisky & M. Teichmueller 1962). If one supposes that the author's as well as Burek's (1973) recent concept of heat waves originating not only from sites of orogenic activity but also from sites of sea-floor spreading is correct, post-Hercynian dating of the Erkelenz laccolith finds new support in as much as it might have a close genetic connection to the opening of the North Atlantic Ocean. The north-eastward overhang of the laccolith seems to indicate centrifical movements in connection with the Biscay spreading.

h) The laccolith of Bramsche (Fig. 13) also reveals a northward overhang. The strike of the main axis of this pluton is parallel to the axial trend of the Austrian orogeny (Fig. 13). The best age estimate for the intrusion is

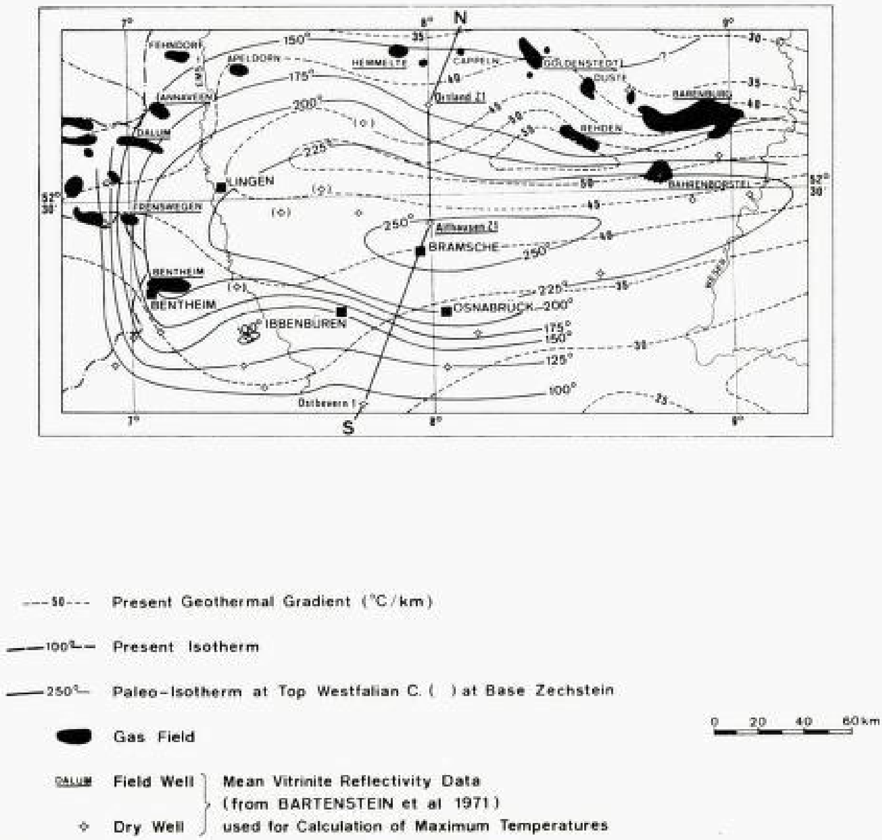

Fig. 12. Bramsche massif, map of present and paleo-isotherms.

Middle Cretaceous. Fig. 13 includes the actual gradients of Fig. 6 with maximum temperature data deducted from vitrinite iso-reflections given by Bartenstein et al. (1971). The related isotherms were fitted into a cross-section prepared by the same authors (Fig. 13). Both diagrams demonstrate that the paleo-temperature high, with gradients up to 65°C/km, was located on top of the Bramsche Massif, whereas the actuo-temperature high with gradients up to 55°C/km is situated some 55 kms further north within the Lower Saxony Basin, just south of the gas province. The situation could be interpreted by northward heat migration which lasted for 100 m.y., but related thermograms do not allow such a conclusion. Because of this a coalification break of type B (Fig. 10) at the base of the Upper Campanian substage refers to the already mentioned upwarping of the Lower Saxony Tectogene (Boigk 1968) which, therefore, should preferably be called an aulacogen — during the sub-Hercynian Phase. The possibly contemporaneous intrusion of the Uchte laccolith has been discussed in chapter 4 c.

i) During the final stage of the Laramide Phase, i.e. during the late Paleocene, the well-studied spreading of the North Atlantic shifted into the Reykjanes Ridge axis. Contemporaneously the Lofoten Islands charnockites were

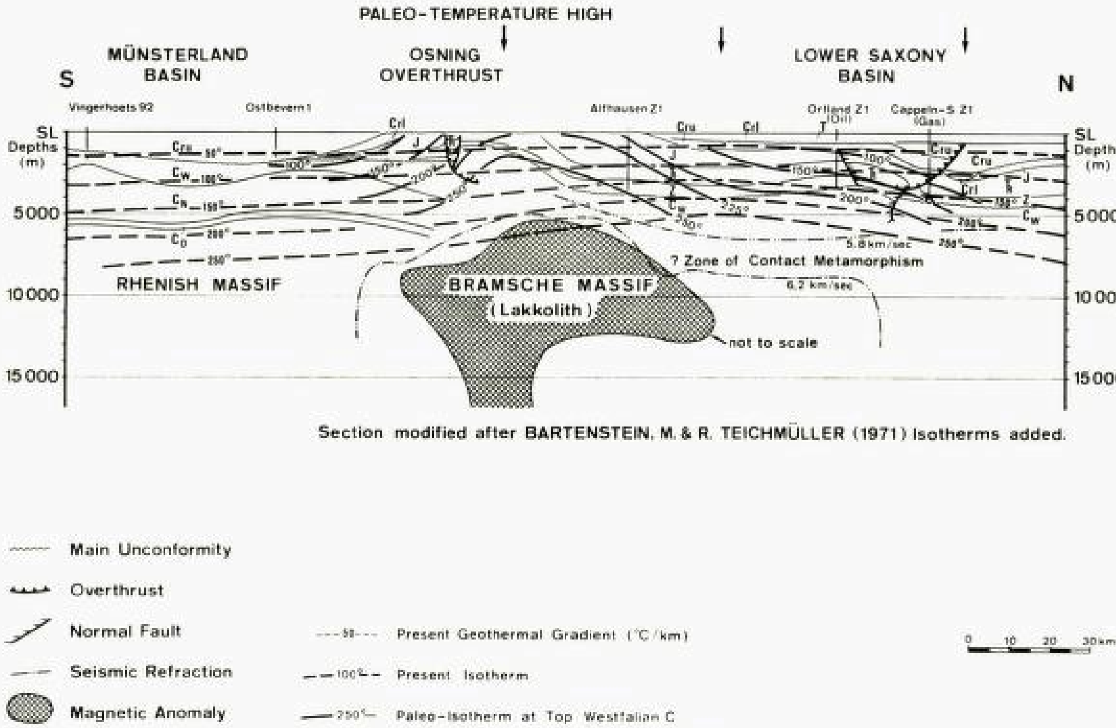

Fig. 13. Section through the Bramsche massif; present and paleo-temperatures.

emplaced along faults parallel to the ridge axis, thus subdividing the northern part of the Nordland Basin. The assumption that a paleo-gradient of 50 °C/km originated from that event was based on coal-petrographic studies from Andøya, although ESR-temperatures as well as hydrothermal quartz temperatures would indicate higher gradients. The ESR-measurements, however, have probably been disturbed by the resinite-compounds of the Upper Jurassic coal and the veins are to be interpreted as untypical, 2-dimensional, local heat anomalies (see chapter 4, IV).

In summary, the main results of our investigations are that the examined paleo-thermal events indicate a time-space relationship to the compressional phases of the various orogenies or to the spreading-phases along mid-oceanic ridges. Some overturns which have been determined are directed about north and, therefore, are explained as a result of tangential forces produced by the a.m. tectonic events. The paleo-thermal events cannot only be studied optionally within the coal-bearing belt of the Carboniferous but they also appear to be restricted to that zone. One reason for this restriction may be a thinning of the northern margin of the Hercynian crust characterized by high paleo-thermal gradients which provided for repeated thermal upwelling. It seems natural that the North Sea Graben finds its southern end in this zone. Many additional detailed investigations will be necessary before the thermal history of the North Sea region can be completely understood.

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North Sea Basinal Area, Europe — an Important Oil and Gas Province*

W. W. DUNN

Dunn, W. W. 1975: North Sea basinal area, Europe – an important oil and gas province. *Norges geol. Unders. 316*, 69–97.

The North Sea covers the offshore part of a major sedimentary basin which extends from Norway, Scotland, and Denmark across northern Germany and the Netherlands into eastern England. Information gained from exploration efforts over the last 10 years shows that the North Sea covers several smaller sedimentary and structural basins of different geologic ages, but for descriptive purposes these can be divided into southern and northern areas. The rocks range in age from Paleozoic to Tertiary and consist of sandstone, shale, carbonates and evaporites. The most important reservoir rocks are the Lower Permian sandstones of the Rotliegendes Formation, the Upper Permian dolomites of the Zechstein Formation, the Triassic sandstone of the Bunter Formation, the Jurassic sandstones, the Maestrichtian–Danian chalk, and the Paleocene and Eocene sandstones. Significant shows of hydrocarbons have been found in 10 formations. The main source rocks are Carboniferous coal measures, Mesozoic shale and carbonates, and Tertiary shale and carbonates. The significant traps are folds and fault blocks associated with salt movement and basement faulting.

Exploration activity received its initial impetus in 1959 from the discovery of a major gas field, Schlochteren, onshore in northern Netherlands. In the early 1960s the passing of legislation favorable for the acquisition of exploration acreage offshore added further stimulus to the exploration pace. The majority of this activity was concentrated initially in the southern area, and resulted in the discovery of the first offshore commercial gas field at West Sole in 1965. This discovery was followed rapidly by other gas discoveries in the United Kingdom and the Netherlands culminating in the Leman Bank field, a major gas field by world standards. Interest and activity lagged, however, in the northern area despite reported small oil and gas discoveries in Denmark, and the discovery in 1968 of the Cod gas-condensate field in Norway. In late 1969, oil production was established at the Ekofisk field in Norway. With this discovery and subsequent confirmation as a major field, exploratory interest has shifted to the north.

W. W. Dunn, Philips Petroleum Company, Bartlesville, Oklahoma 74004, U.S.A.

I realize that many of you are thoroughly aware of the events leading up to the discovery of oil in the North Sea. However, I am sure that there are others present today who are not familiar with the story. Therefore, I plan first to sketch briefly the historical background, and to high-light the noteworthy events. I then plan to outline the regional geology. Next, we will examine in some detail the geological and geophysical information on the Ekofisk area in Norway, and finally, I will give you a status report on the offshore production installations at the Ekofisk Field.

* This account is an edited version of the speech given by Mr. W. W. Dunn at the Bergen Conference. It is based on a speech presented by Mr. Dunn on 18th April 1972 to the Annual Meeting of the American Association of Petroleum Geologists held in Denver, Colorado, and later published in modified form in the *Oil and Gas Journal*, January 1973 (Dunn, Eha & Heikkila – 'North Sea is a tough theater for the oil-hungry industry to explore.')

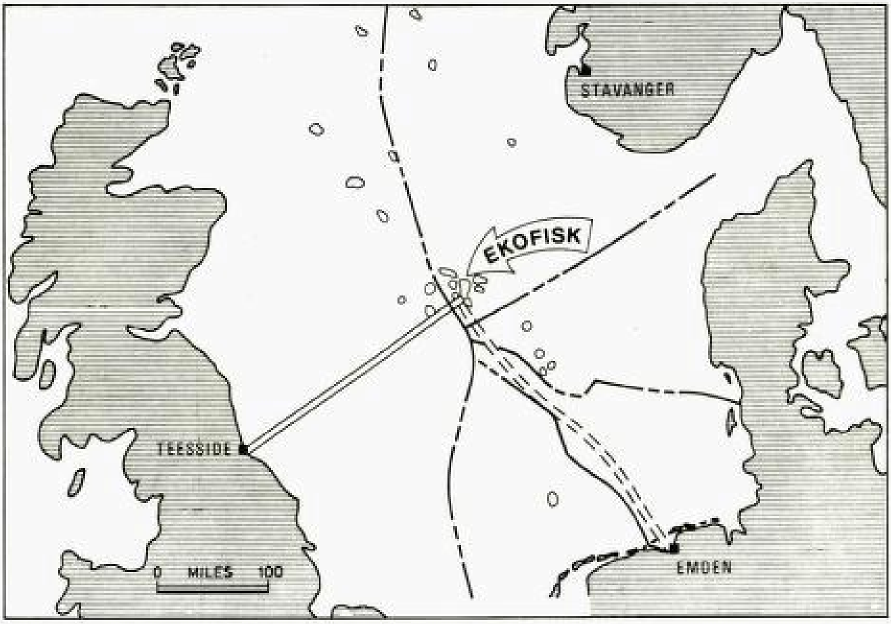

Fig. 1. Location of the Ekofisk field, North Sea, showing some of the principal oil fields. The 34″ oil pipeline to Teesside and the 36″ gas pipeline to Emden are also indicated.

With the discovery of oil at Ekofisk in Norway in 1969, the North Sea has become one of the most active new offshore exploration areas in the world. The location of the North Sea in North–Western Europe and the location of Ekofisk in the center are shown in Fig. 1. The sea covers the offshore portion of a major structural and sedimentary basin which extends from Norway, Scotland and Denmark across Northern Germany and the Netherlands and into Eastern England. The total offshore basinal area, limited northward for the purposes of this paper by latitude 62° North, is approximately 240,000 square miles (621,600 km^2) or nearly the size of France. Fig. 2 shows how the water depths gradually increase northwards, reaching a maximum depth of 2000 feet (over 700 m) in the trench bordering the Norwegian coast. Operations in the middle of the North Sea can be as far as 200 miles (320 km) from land.

In 1959, forty years of previous exploration in Northern Europe had attracted little worldwide attention to the area. Small oil fields had been found in the Carboniferous reservoirs of the East Midlands in the United Kingdom. Gas production had been established in Permian and Lower Triassic reservoirs, and oil in Upper Triassic, Jurassic and Lower Cretaceous reservoirs in Northwest Germany, and small oil and gas fields had been found in the Mesozoic of Holland (Fig. 3). In addition, a small number of wells had been drilled without success in the territorial waters of Holland and Germany.

There were two factors which dampened enthusiasm for the area, and slowed development. One was the lack of success in discovering large fields, which if found offshore would be commercial; the other was the question of

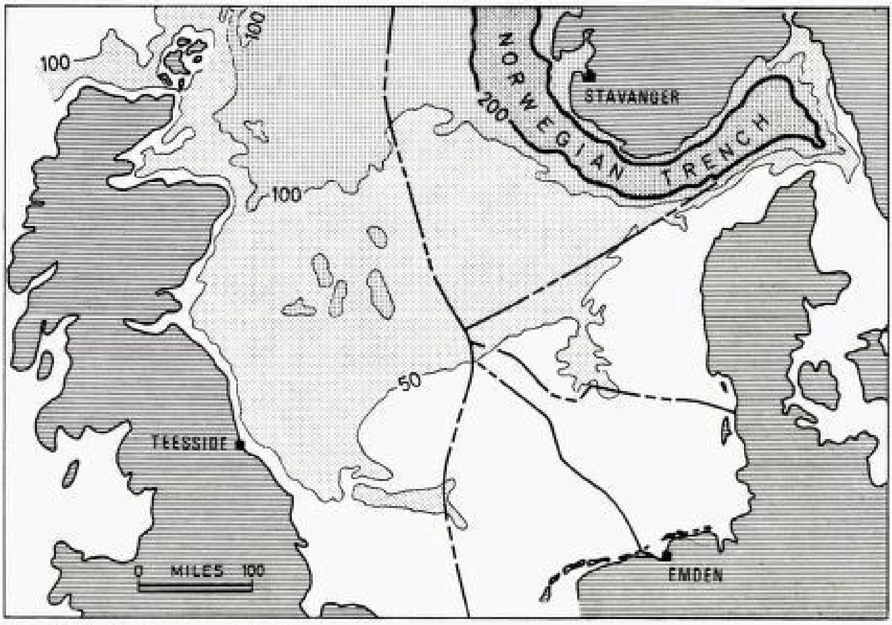

Fig. 2. North Sea water depths.

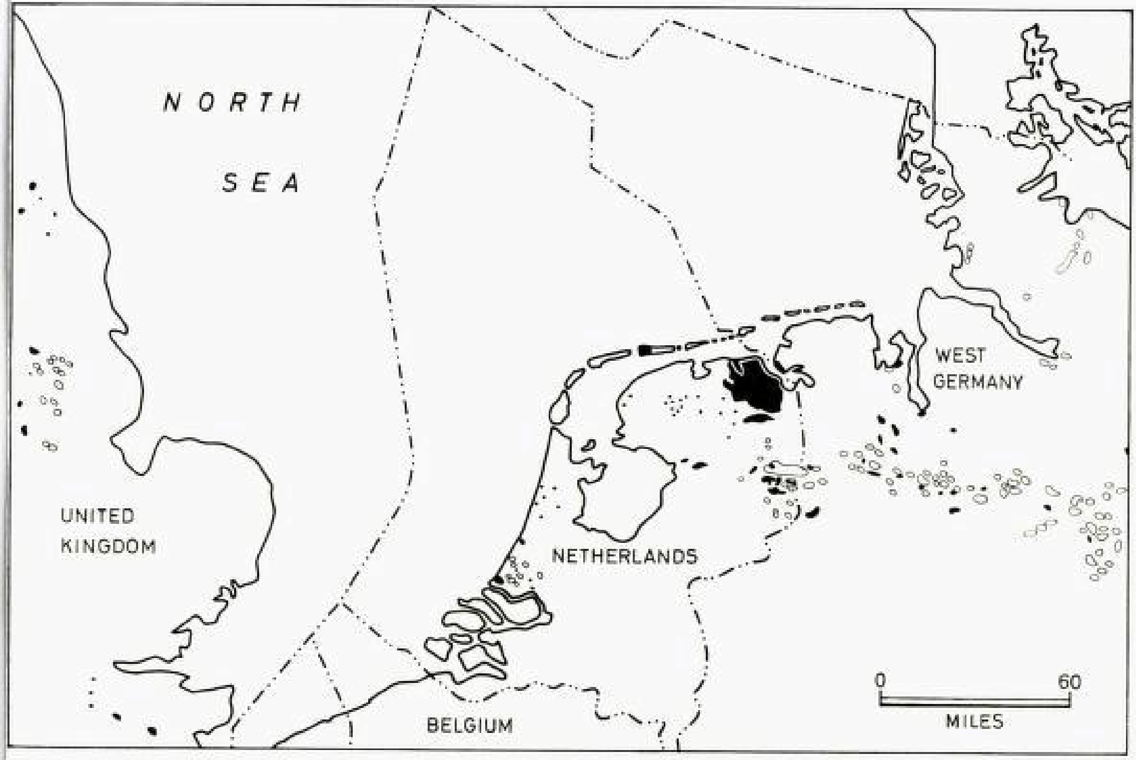

Fig. 3. Onshore oil and gas fields, Netherlands, West Germany and England. Gasfields – black areas and spots; oil fields – unornamented.

ownership of the oil and gas rights outside the territorial waters, or three-mile limit. In 1959, the gas discovery at Slochteren in Northern Holland dramatically changed all this. This discovery, combined with the increasing demand of the European energy market, and the need to diversify supplies in this market, was the catalyst which set off the offshore search.

Information became available over the next few years that this was a major gas field in the Permian Rotliegendes Sandstone. This information, combined with the fact that all the offshore areas of the world were either being actively explored or were seriously being considered for exploration in the early 1960's, focused attention on the oil and gas possibilities under the North Sea. It was not until 1963, however, that it was confirmed that the recoverable reserves at Slochteren were 58 trillion cubic feet, but by that time exploration activities in the North Sea had already started.

At the same time, the concession situation was also being legalized for the waters beyond the territorial limit. In 1958, the Geneva Convention on the continental shelf had established the ground rules for the division of these offshore areas. This convention, however, required ratification by 22 countries, and it therefore did not become effective until 1964. Treaties then had to be negotiated between the individual countries involved, in order to establish the actual boundaries. This was done with dispatch except for a disagreement between the Netherlands, Germany and Denmark which was finally resolved in 1970.

The median lines which now divide the North Sea into five sectors are indicated in Fig. 4. It is important to note that each country established its own rules and regulations for the granting of exploration and production rights. Both Denmark and Germany awarded their entire area to single consortiums, whereas the United Kingdom, the Netherlands and Norway gridded their areas, and put up their blocks for competitive application. These blocks, as illustrated, are of different sizes: a U.K. block covers about 90 square miles (57,600 acres), a Netherland block about 160 square miles (102,400 acres) and a Norwegian block about 210 square miles (134,000 acres). There are further differences in rental costs, terms, royalty, etc., but originally no bonuses were involved, only work or drilling commitments. There have been a total of eleven concession awards in the North Sea outside of Denmark and Germany, five in the U.K. in 1964, 1965, 1969, 1971 and 1972, three in Holland in 1968, 1970 and 1972, and three in Norway in 1965, 1969 and 1973. Each has had either minor or major changes in the basic requirements.

The last concession application award in the U.K. included 15 blocks which were awarded by sealed bid — an innovation for this area. Additional changes can be expected in future U.K. offers possibly similar to the carried interest requirements innovated by Norway in their 1973 acreage offers.

While the legal aspects of ownership were being resolved, exploration work was underway. The period between 1959 and 1963 was confined to a review of the published geological and geophysical data concerning the surrounding land areas. This was supplemented by geological field studies, and a minimum amount of seismic work offshore.

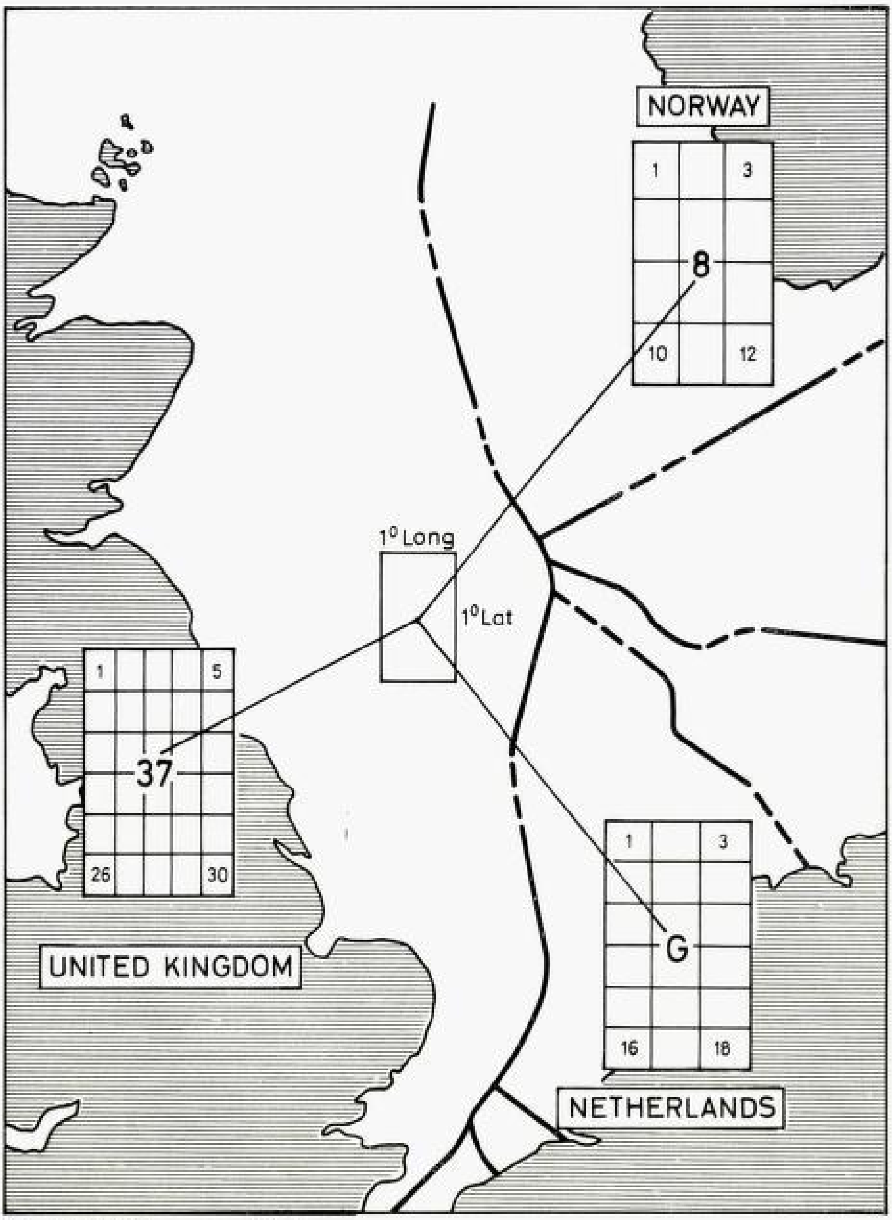

Fig. 4. North Sea average blocks.

The geological picture at that time was assessed by the industry about as follows: It was evident that there was a continuation beneath the North Sea of the Permian (the producing reservoir at Slochteren), the Mesozoic and also the Tertiary sediments which outcropped around its southern margin (Fig. 5). Published geophysical data supported this theory, and drilling in Germany and the Netherlands had revealed, beneath the Tertiary cover, a rock sequence similar to that of the Post-Carboniferous of Eastern England, but of a different facies.

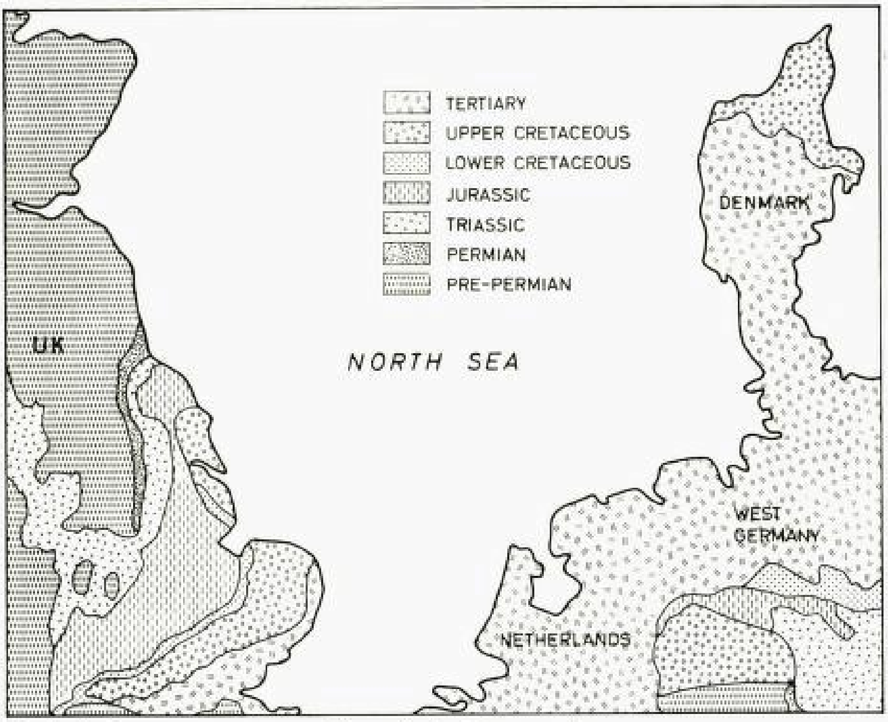

Fig. 5. Onshore outcrop pattern of post-Carboniferous rocks.

From a structural point of view, the general effects of the Hercynian, Kimmerian and Alpine orogenies were well known, and the structural and cap rock potential of the thick Permian and Triassic evaporites could be deduced from the onshore geology. The Pre-Carboniferous stratigraphy was known from the surrounding land areas, and due to the metamorphism resulting from the Caledonian orogeny, these rocks offered little incentive to the oil explorer. Of more interest were the regressive deposits of the Lower Permian, the Early Triassic, the Late Jurassic – Early Cretaceous and, possibly, the Tertiary, as likely reservoir rocks. Transgressive sediments of Carboniferous, Permian, Lower Jurassic, and again Tertiary age, offered promise of likely source rocks.

Features providing hydrocarbon entrapment were known to be of a wide variety: salt-controlled traps, fault traps, stratigraphic traps due to salt movement, anticlines, domes, etc. All the classic traps were present, and these traps could be expected offshore.

Offshore seismic work had started on a large scale in 1973, and in 1964 the first North Sea drilling began off Germany. The same year, petroleum licences were granted by the U.K. with the first exploratory drilling commencing 190 miles offshore in December.

Fig. 6 shows the location of West Sole, the first commercial offshore gas discovery, which was found the next year, 1965. Other major gas fields were quickly found: Leman Bank, Indefatigable and Hewett in 1966, and Viking in 1969. These offshore gas field were tied into two terminals in England,

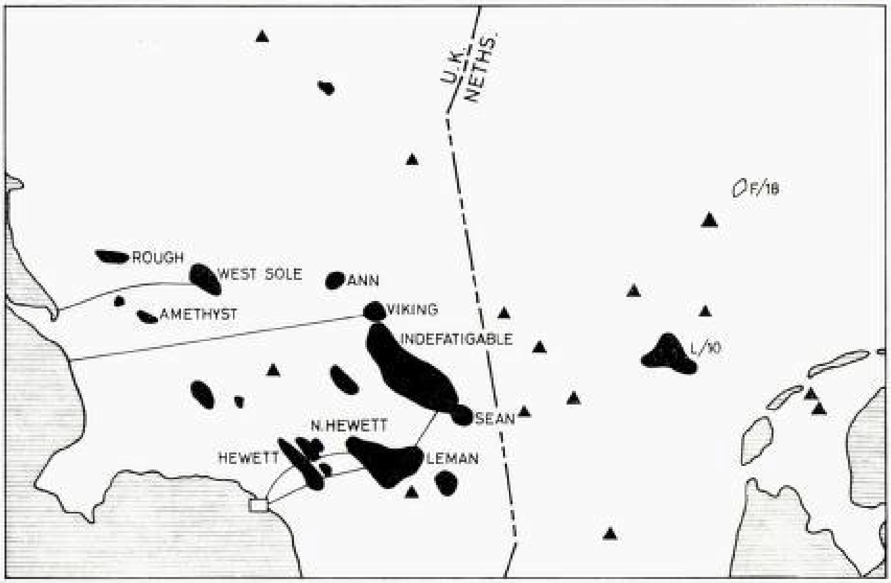

Fig. 6. Southern North Sea, oil and gas fields. Gas–black areas; Oil (F/18)–unornamented

Easington and Bacton, with a third at Maplethorpe, completed last year. Present gas productive capability in the U.K. North Sea is above 2 billion cubic feet of gas per day and it is estimated that it will reach 4 to 4½ billion by 1975.

During this period of success in the discovery of gas in the U.K., exploration operations continued without success in Germany and were discontinued there in 1968 after 17 consecutive dry holes. In the Netherlands, exploratory drilling commenced in 1968 shortly after the award of concessions, and by December gas was discovered in the Permian. Further gas discoveries were made subsequently, and development is underway in Blocks L/10 and L/11. In Denmark, drilling had started in 1966 and significant oil and gas shows had been encountered in the Danian and Upper Cretaceous. Production commenced in 1972 from the 'Dan' Field at the rate of 4000–5000 BOPD. This development was delayed by the boundary dispute with Germany.

Norway awarded concessions in 1965 and in 1969 and drilling commenced in 1966. The Cod Gas Condensate Field was discovered in 1968, but further drilling discouragements led to a decrease in drilling in 1969. About the time when enthusiasm had reached a low ebb, the Philips 2/4–1AX well, on the Ekofisk structure, had oil and gas shows at about 10,000 feet. This well indicated the first commercial oil field in the North Sea and, as it turned out, the first billion barrel or giant oil field in Western Europe. The locations of Ekofisk and other oil and gas fields discovered subsequently in the Northern area of the North Sea are shown in Fig. 7. In the Norwegion sector, Philips has a series of fields in what is referred to as the Greater Ekofisk Area, while 220 miles (350 km) to the north another group has one of the most sub-

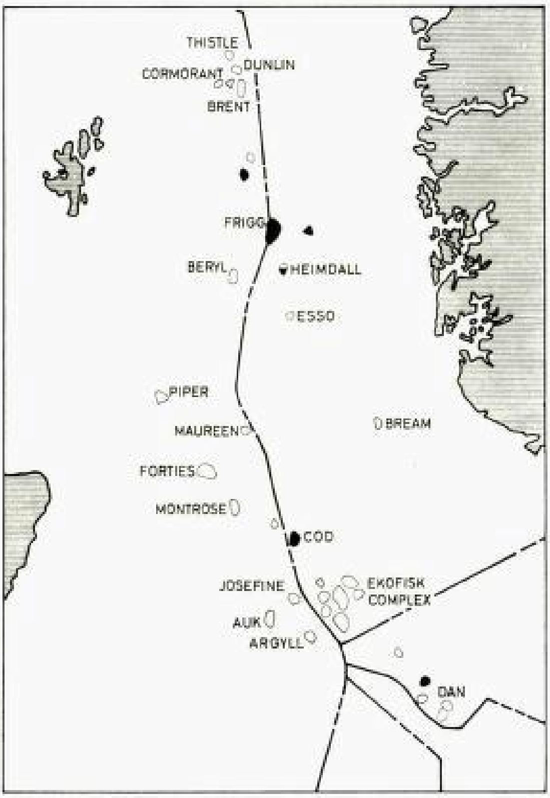

Fig. 7. Northern North Sea, oil and gas fields. Gas – black; oil – unornamented.

stantial gas discoveries to date with over 12 trillion cubic feet of recoverable reserves — the Frigg Field.

To review the oil discoveries of the Northern U.K. basin is an almost continuous task — at one stage there seemed to be a discovery per month. Now, with perhaps 15 or 16 substantial discoveries with productive capabilities varying from 40,000 BOPD at the Auk Field to 400,000 at the Forties Field, the potential production for the U.K. is bright. In all, the total oil reserves discovered to date in the North Sea could probably support a maximum production of 3½ million BOPD in 1980, which is equal to only 15% of Western Europe's predicted demand.

Later in the paper I will return to the Ekofisk area and specifically outline the geology and reservoir information in detail and discuss the present development program. But first, I would like to give you the regional geological framework for Ekofisk and other recent oil discoveries in the northern area of the North Sea.

In 1963, because of the discovery of gas at Slochteren, geological thinking was first directed towards gas and secondly towards the extent of the Permian Rotliegendes sandstone reservoir. Therefore, the primary interest of the industry was centered in the southern portion of the North Sea, where this reservoir had the best chance of being present. Other companies, however,

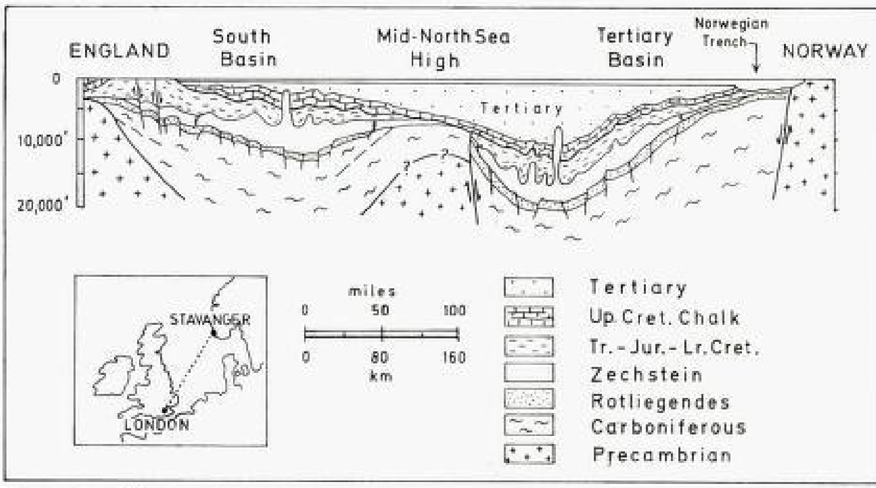

Fig. 8. Simplified geological cross-section, London – Stavanger.

became curious as to the possibilities of the northern portion of the North Sea which lay between the metamorphics and Early Paleozoic sediments of Norway and Scotland.

It was with this background that the companies began serious seismic work in the North Sea in 1961 and 1962. A simplified geological cross-section of about 550 miles from London to Stavanger illustrates some of the data revealed by the initial seismic work (Fig. 8). The main features to note are the Tertiary Basin, which contains the Ekofisk and Forties oil discoveries, the regional unconformity beneath the Upper Cretaceous, which probably plays a significant part in the more recent oil discoveries to the north, and the two Permian sub-basins, in which the original gas was discovered.

Fig. 9 is a generalized composite illustration of the structurally high areas and the intervening basins and troughs; this combines a number of structural features of different geological age.

From south to north we encounter:

1. *The Brabant Massif* of Hercynian origin limiting the North Sea basinal area to the southwest.
2. *The Mid-Netherlands Ridge*, which has been an active high during much of the Mesozoic until Upper Cretaceous.
3. The east-west trending *Mid-North Sea High* or *Northumbrian Arch* extending eastward from the U.K., which has affected Permian through Lower Cretaceous deposition and limits the so-called '*Southern U.K. Basin*' to the northwest, which in turn is an extension of the *Northwest German Basin* to the southeast, which contains huge thicknesses of Permian and Triassic deposits with their accompanying evaporites.
4. To the north the *Mid-North Sea High* is bordered by the shallow, predominantly Mid-Mesozoic Scottish Basin.
5. To the northeast the *Ringkøbing–Fyn High* is a controlling positive feature from Permian through Lower Tertiary sediments.

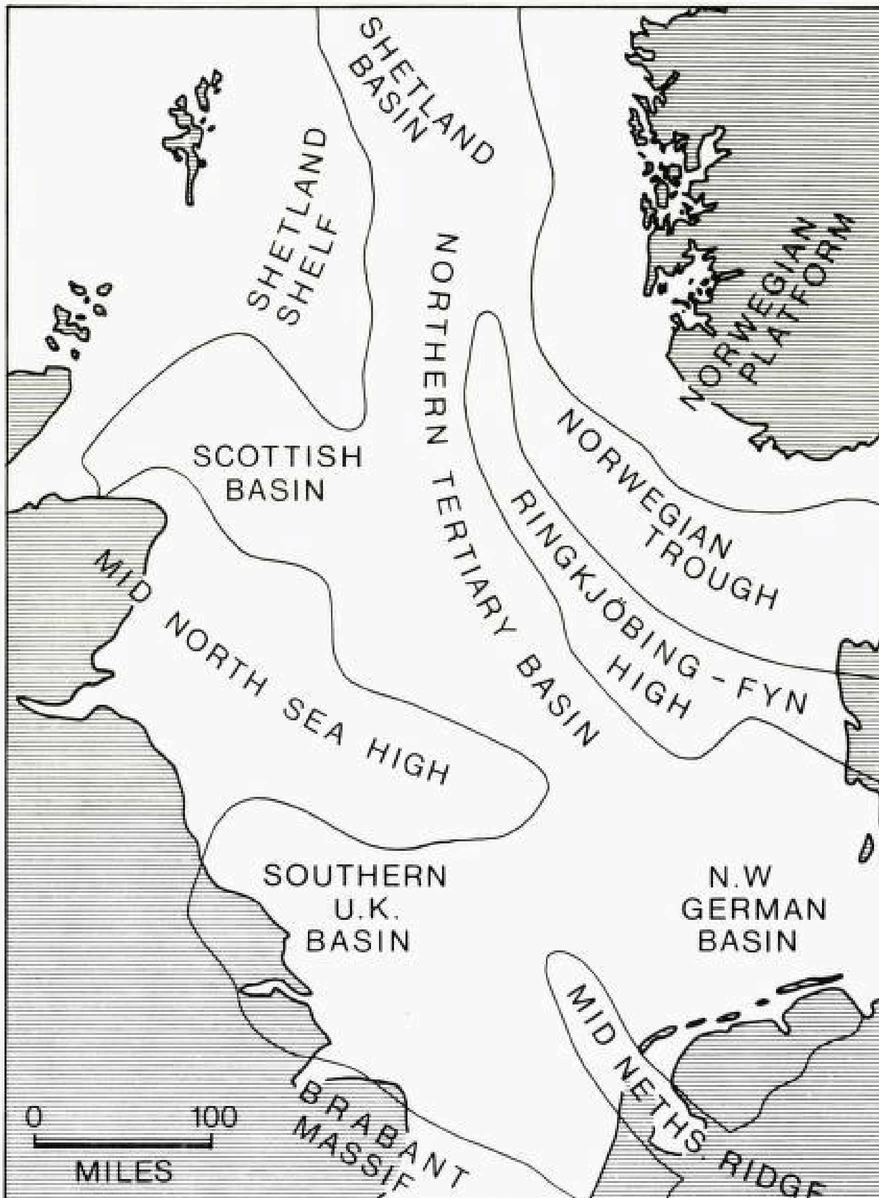

Fig. 9. Major structural features of the North Sea area.

6. To the north of 56°N latitude, the main tectonic features in the Pre-Upper Cretaceous are obscured by the increasing thickness of Tertiary sediments in the Northern Tertiary and Shetland Basins to the east of the Shetland Shelf, and to the west of the Norwegian Trench.

Figs. 10–17 show the present-day distribution of the strata which are most relevant to the oil and gas accumulations, and the relationship to the structural features discussed above.

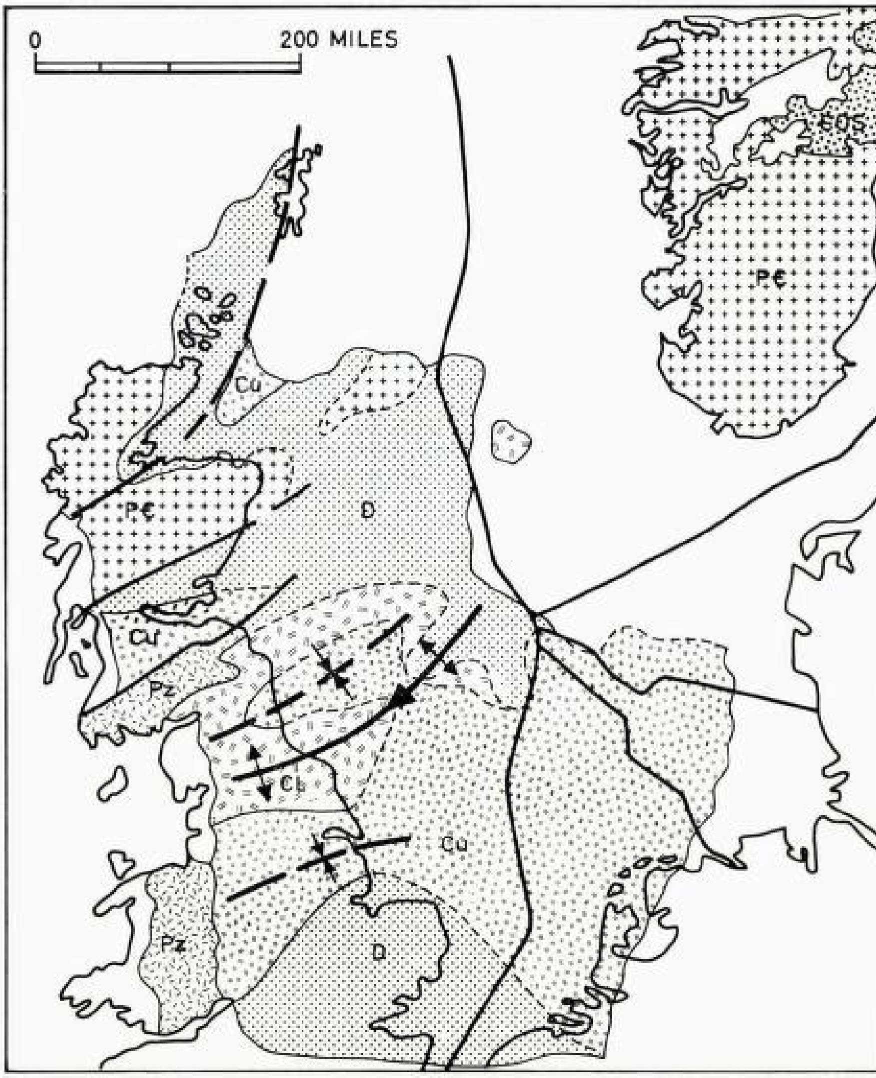

Fig. 10. Pre-Permian sub-crop, North Sea area. Cu — Upper Carboniferous; CL – Lower Carboniferous; D – Devonian; COS – Cambrian–Ordovician–Silurian; Pz – Paleozoic; PC – Precambrian.

Fig. 10 illustrates the Pre-Permian subcrop. Probably the most notable feature is the widespread deposition of the Upper Carboniferous over much of the southern portion of the North Sea. These largely estuarine and swamp deposits, with their high carbon content, play a significant part as a source rock for the gas in that area.

After the Late Carboniferous Hercynian orogeny, the Carboniferous and earlier rocks were eroded and folded, and then covered by the Lower Permian Rotliegendes Formation. This formation is essentially a product of post-Hercynian erosion. It is a continental sandstone in the Southern North Sea

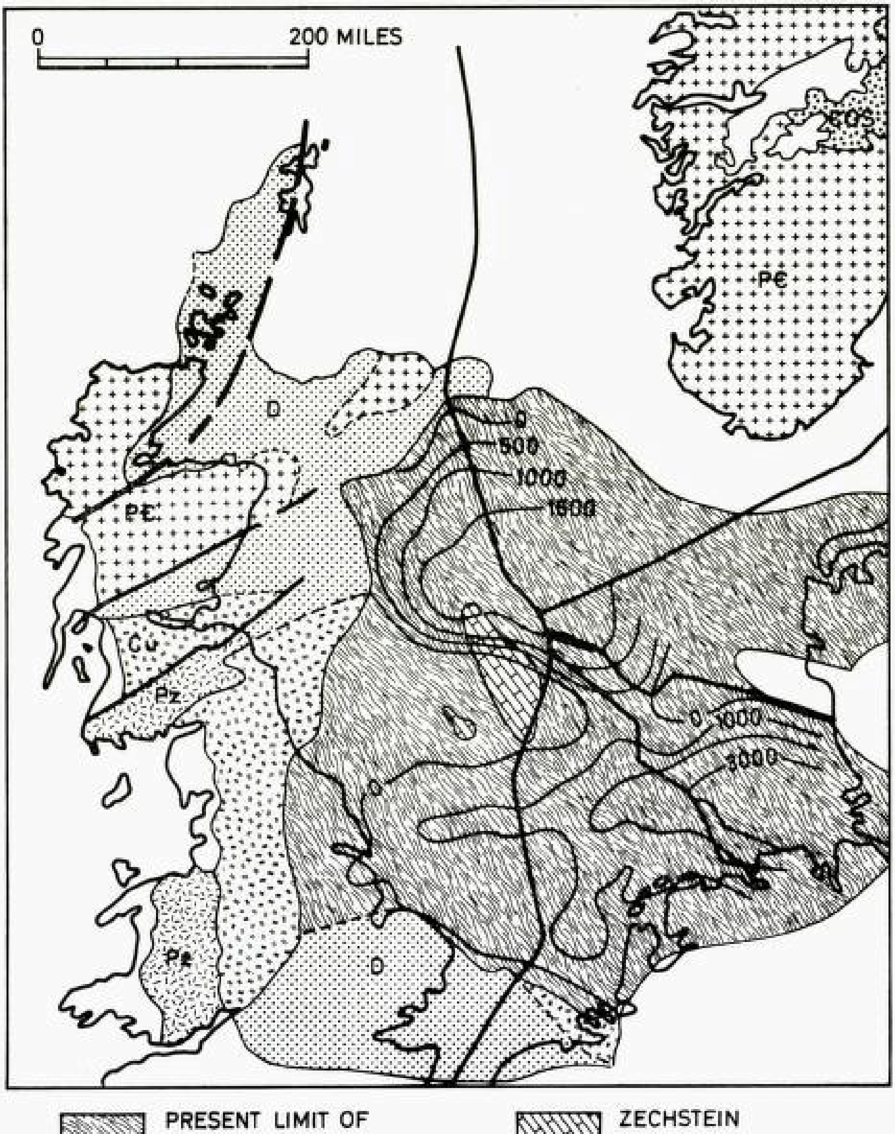

Fig. 11. Present distribution of Permian rocks, North Sea area. Lettering as in Fig. 10. Isopachs in feet.

Basin becoming shaly and marine northward. Fig. 11 shows that the important commercial gas accumulations in the Rotliegendes occupy a marginal position to the south. This is an area of greatest sand deposition where the present-day limit of the Rotliegendes coincides basically with the original southern boundary of the Early Permian deposits.

The Late Permian Zechstein transgression is the first appearance in geologic history of the present North Sea. This sea deposited thick evaporites and basin-margin carbonates. These sediments play an essential role in the subsequent

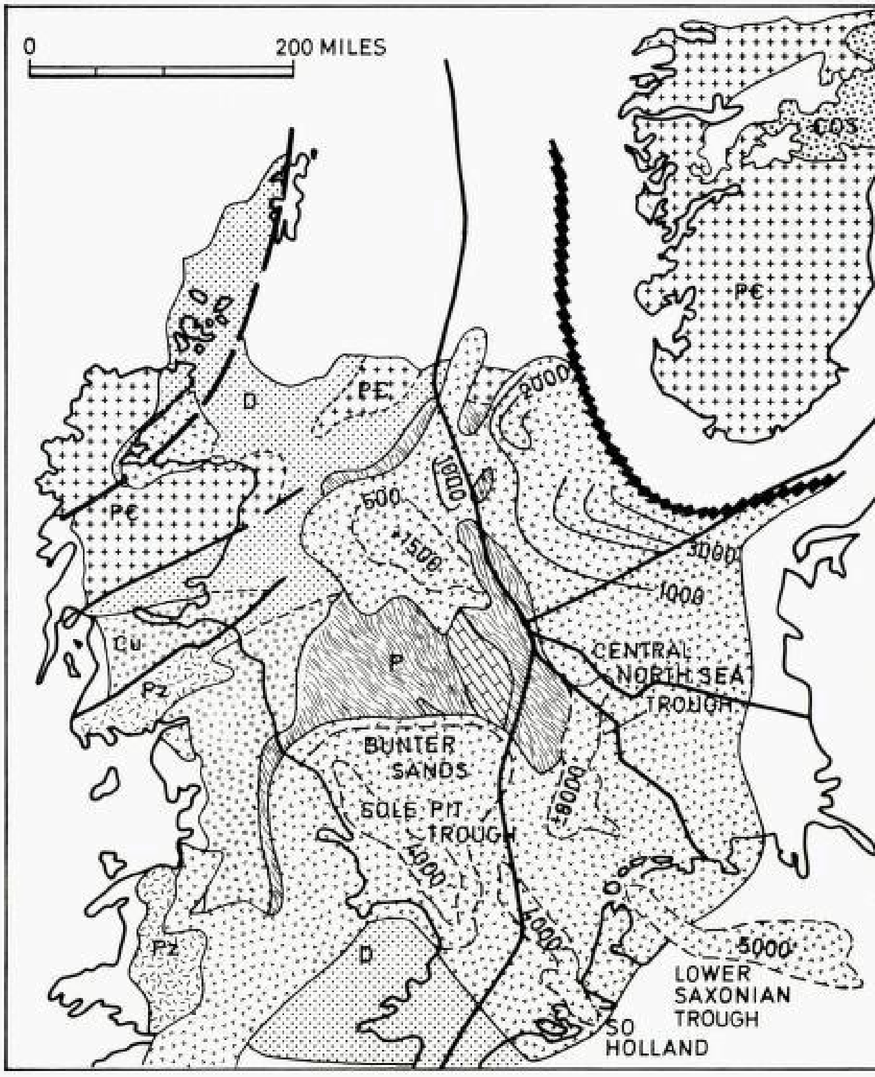

Fig. 12. Present distribution and isopachs of Triassic rocks, North Sea area. Lettering as in Fig. 10. P – Permian. Isopachs in feet.

accumulations of hydrocarbons, as cap rock for the Rotliegendes Sandstone gas reservoirs, as carbonate reservoirs for both oil and gas, and as an active agent for later structural deformation into Tertiary time. Unfortunately, the Zechstein also forms a formidable obstacle for deep seismic exploration, because of the varying salt thicknesses and evaporite composition.

Fig. 12 illustrates the extent of Triassic distribution, and shows that the more important thicknesses already reflect the main basinal areas mentioned before. Conspicuous Triassic troughs such as the Sole Pit, the Central North Sea and the Lower Saxonian troughs formed and were filled with shales and

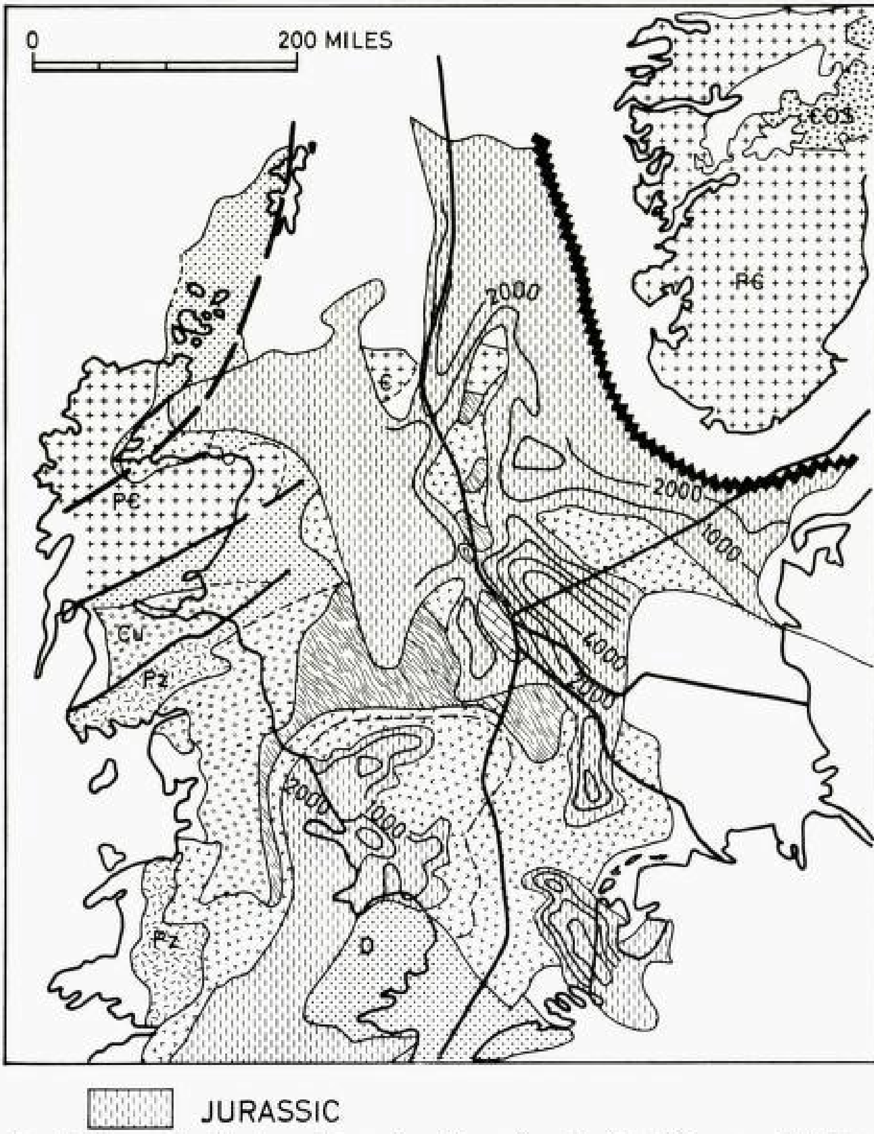

Fig. 13. Present distribution and isopachs of Jurassic rocks, North Sea area. Lettering as in Fig. 10. Isopachs in feet.

clastics. The sandstones of the Triassic serve as reservoirs in the Hewett Gas Field off England and oil shows have been encountered at Josephine in the middle of the North Sea. The sands at Hewett are red-brown and grey-green in colour, and are of continental derivation.

The Jurassic distribution is shown in Fig. 13. Both the Kimmerian movements and intensive salt tectonics show their effects on today's distribution of Jurassic and Lower Cretaceous sediments. The Jurassic and Lower Cretaceous are essential sandstone and shale units, with rapid lateral facies changes. The Jurassic shales are probably the source rocks for oil discoveries and

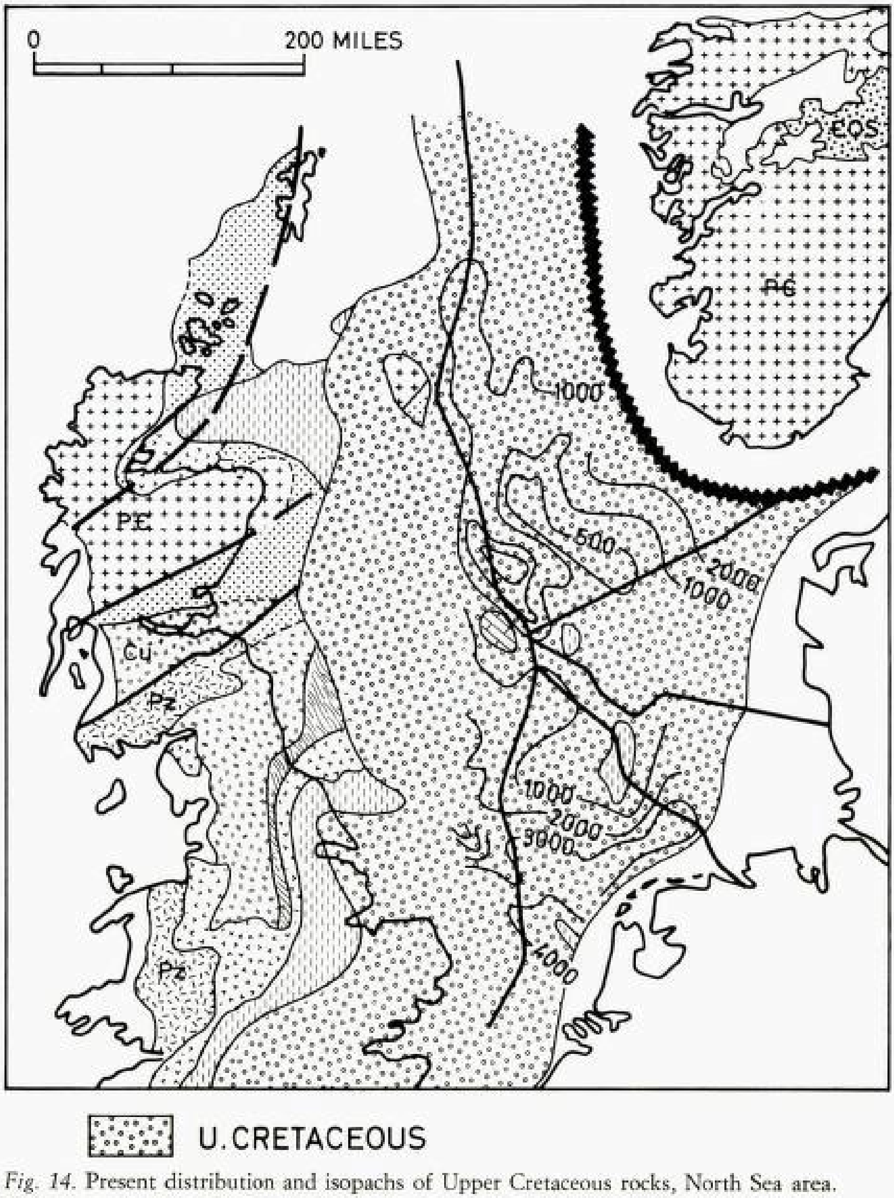

Fig. 14. Present distribution and isopachs of Upper Cretaceous rocks, North Sea area. Lettering as in Fig. 10. Isopachs in feet.

hydrocarbon shows in the Mesozoic, though in the Central and Southern areas, these accumulations occur in rather localized Mid-Jurassic sand reservoirs. To the north, the reservoirs are much more extensive as exemplified by the discoveries in the golden triangle of the Cormorant, Dunlin and Brent fields. Uplift and erosion prior to the deposition of the Upper Cretaceous Chalk caused partial to complete removal of the earlier Mesozoic beds, particularly in the central-southern parts of the basin.

As illustrated in Fig. 14, the homogeneous Upper Cretaceous Chalk is

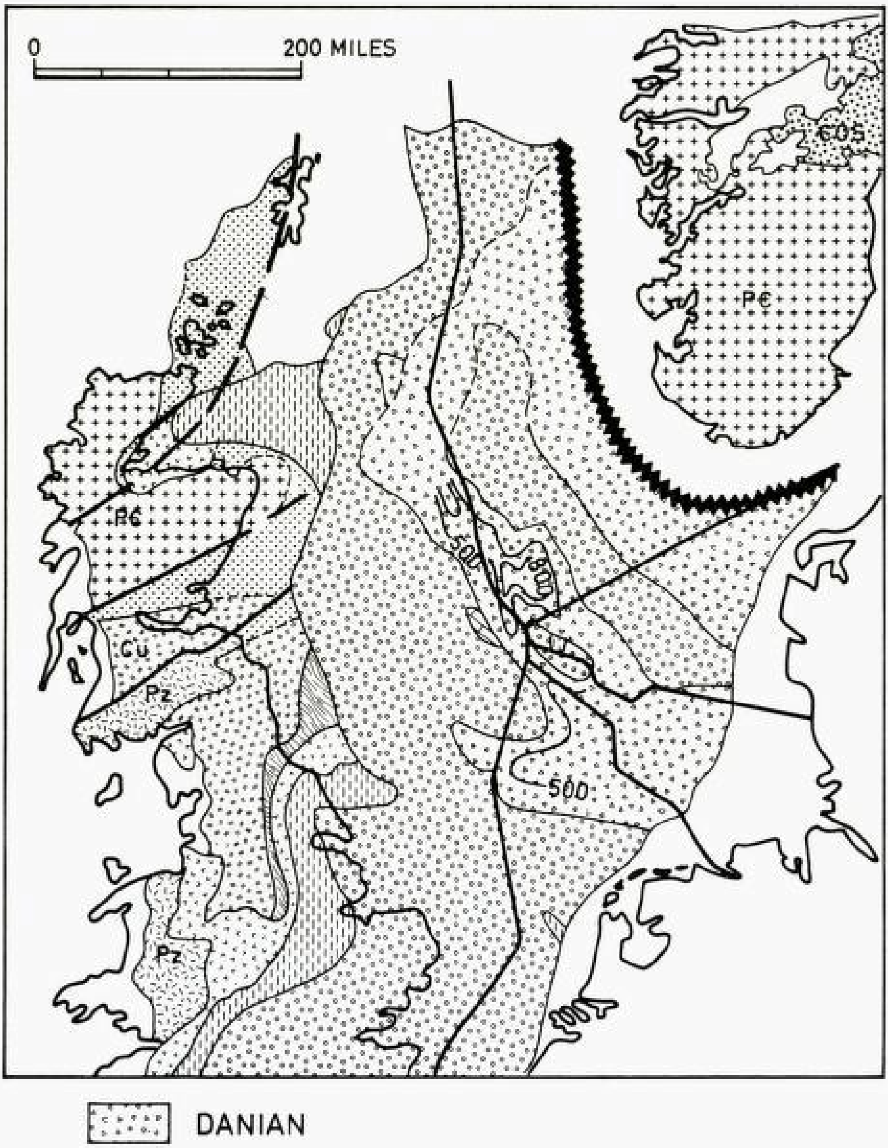

Fig. 15. Present distribution and isopachs of Danian chalk, North Sea area. Lettering as in Fig. 10. Isopachs in feet.

perhaps the most extensive formation in the North Sea. The large structural highs which were already apparent in the Jurassic, persist. While the carbonate deposition continues into the Danian, Fig. 15 shows that its extent becomes restricted to two north-south troughs in the middle of the North Sea which become more pronounced. One of these is the locale of the Ekofisk and neighboring oil fields.

With the Paleocene, another shale - silt - sand sequence begins and continues well into the Pliocene. These shales are excellent source rocks both for the

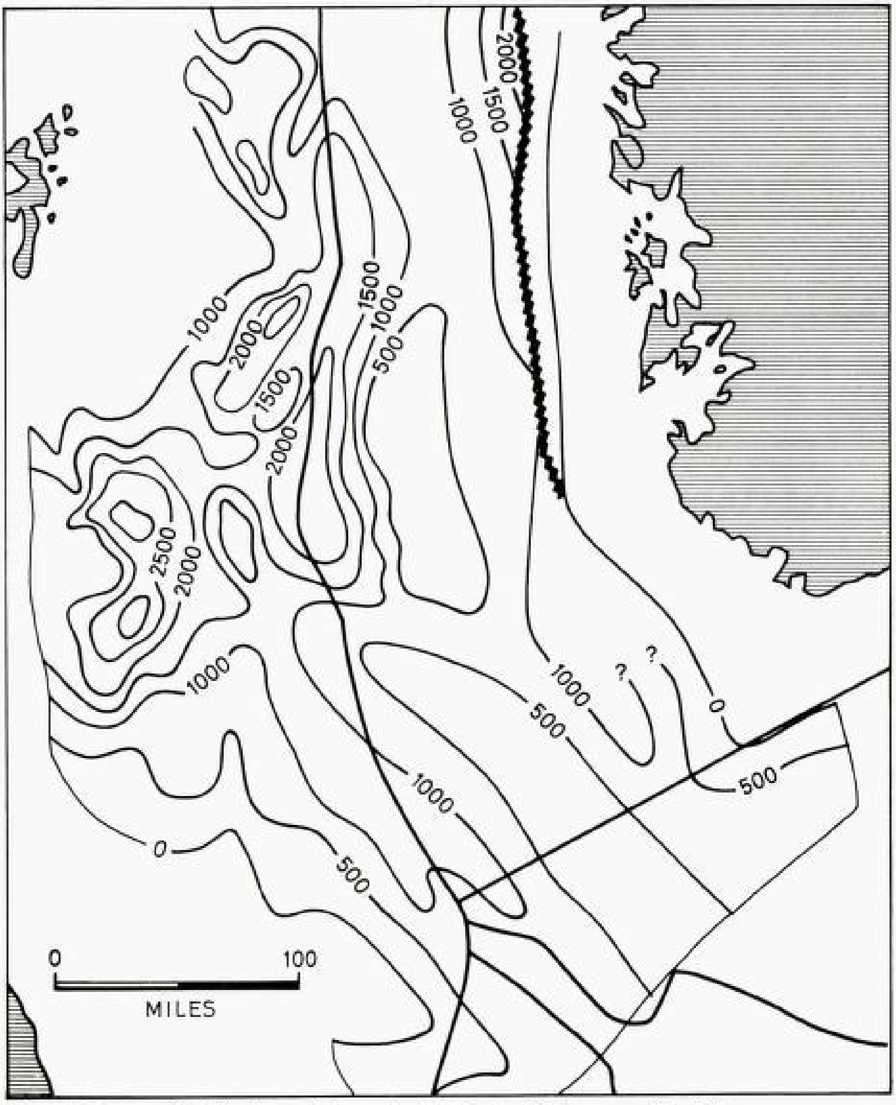

Fig. 16. Present distribution and isopachs (in feet) of Paleocene, North Sea.

underlying Danian/Upper Cretaceous reservoirs, and also for the Paleocene and Eocene hydrocarbon accumulations which occur in relatively localized sand developments such as the Cod, Frigg and Forties fields. It should also be mentioned that gas accumulations have been found as shallow as 5000 feet in localized sand lenses.

Fig. 16 is an isopach map of the Paleocene in Norwegian and northern U.K. waters. Three principal depocenters of total Paleocene (including Danian) can be distinguished. The first is in the southern part of the Northern Tertiary Basin and includes the Ekofisk and Cod fields. The second is a remnant basin flanking the Norwegian coast, west of the Norwegian Trench. This is a continuation of the onshore Danian of Denmark. The third is a possible extension of

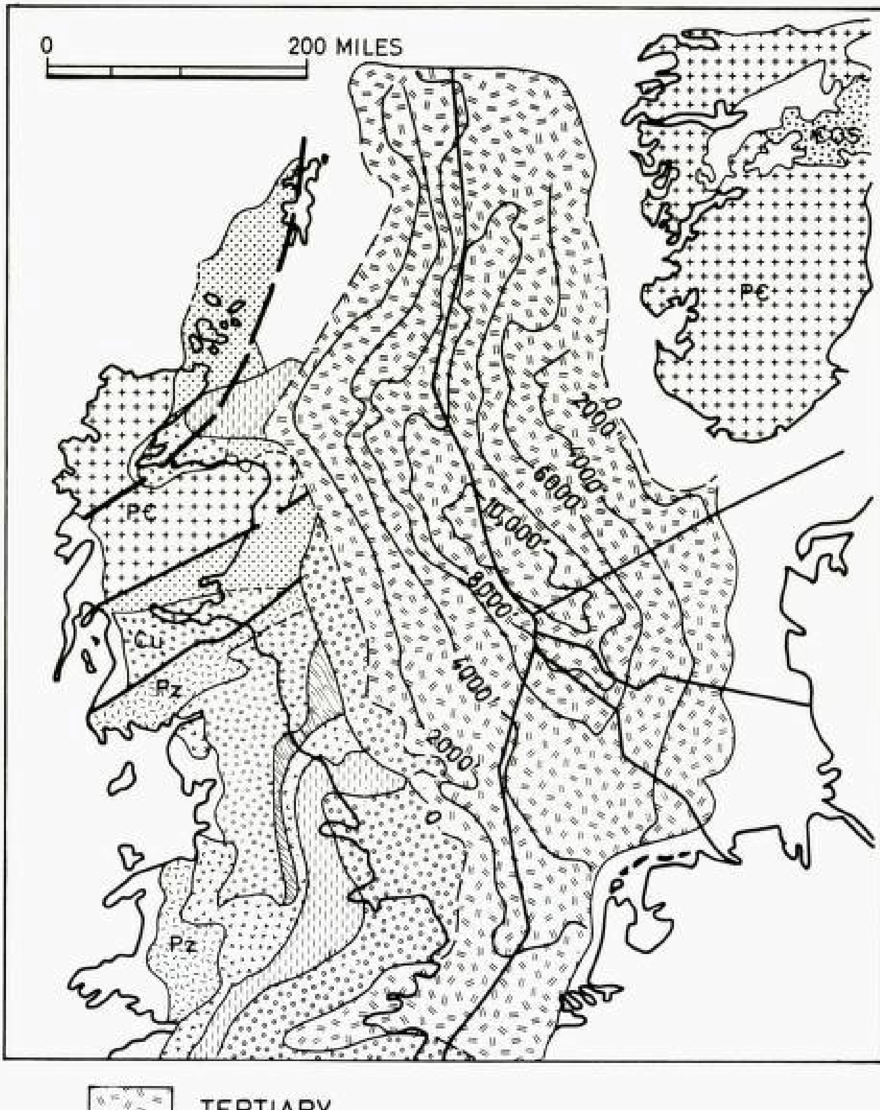

Fig. 17. Distribution and depth to the base of the Tertiary, North Sea area. Contours in feet.

the Northern Tertiary Basin north-westward, which bifurcates into a western depocenter towards the Moray Firth Basin and into a north-trending depocenter immediately east of the Shetland Shelf.

Total Tertiary thickness and its distribution are shown in Fig. 17. Only its southern margins were known at the beginning of North Sea exploration ten years ago. The Tertiary reaches a thickness of 10,000 feet in the center of the basin.

The Ekofisk and neighboring fields are located in the deepest part of the Tertiary Basin in an area containing over 1,300 feet of Danian and Paleocene

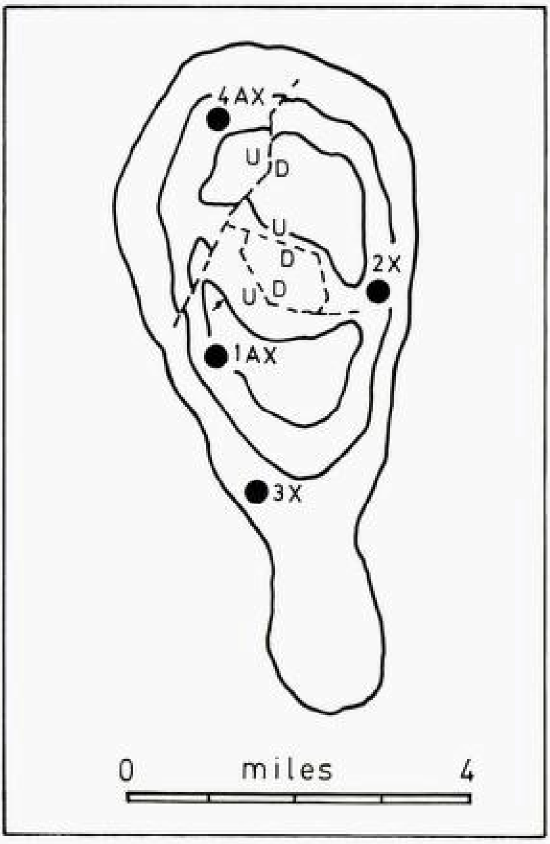

Fig. 18. Ekofisk Field, structural form-lines and positions of the 4 original wells.

sediments, approximately half of which consist of the Danian Chalk section, and the other half of Upper Paleocene clastics. The lower part of the Danian commonly contains reworked Upper Cretaceous fossils which appear also at the top of the Upper Paleocene. It can thus be assumed that at the beginning of the Danian, the Upper Cretaceous Chalk was subjected to submarine erosion wherever it may have been structurally elevated. Isopachous mapping of the Upper Cretaceous suggests the existence of elevated areas which may have been shallow relative to sea level and which were the erosional source areas for part of the Danian section. The bulk of the section, however, was deposited in deep water and consists principally of coccolith remains and lime-muds. The Danian Chalk is the primary hydrocarbon reservoir in the Ekofisk area.

The transition from Danian to Upper Paleocene sediments is usually marl, but in places may be shales and sand. The overlying Upper Paleocene section is predominantly clastic. Shale and silt are characteristic on the margins of the basin and in areas of isopach thins. Sand becomes an important component in areas of thick section, for example in the Cod Field area. Thin marl and limestone beds constitute a minor part. The Upper Paleocene section, like the Danian, is a deep water deposit. The sands in the axial part of the Northern Tertiary Basin and in the Cod Field are turbidites — a feature which is

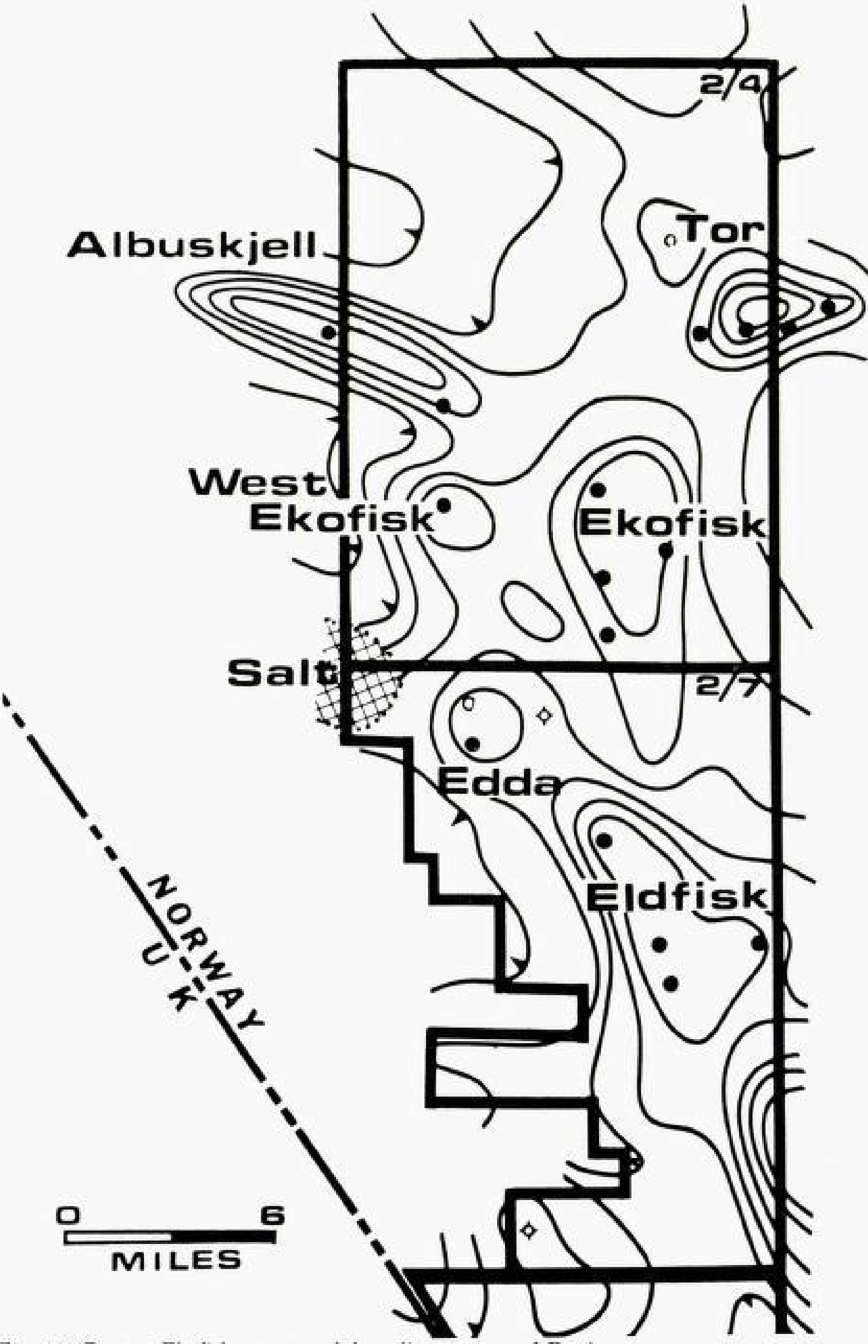

Fig. 19. Greater Ekofisk, structural form-lines to top of Danian.

consistent with the deep water origin of the total section. Sand sources may have been Mesozoic and older sediments eroded from the Mid-North Sea High. Other sources and other environments of deposition are, of course, possible and may be found outside the immediate confines of the Northern Tertiary

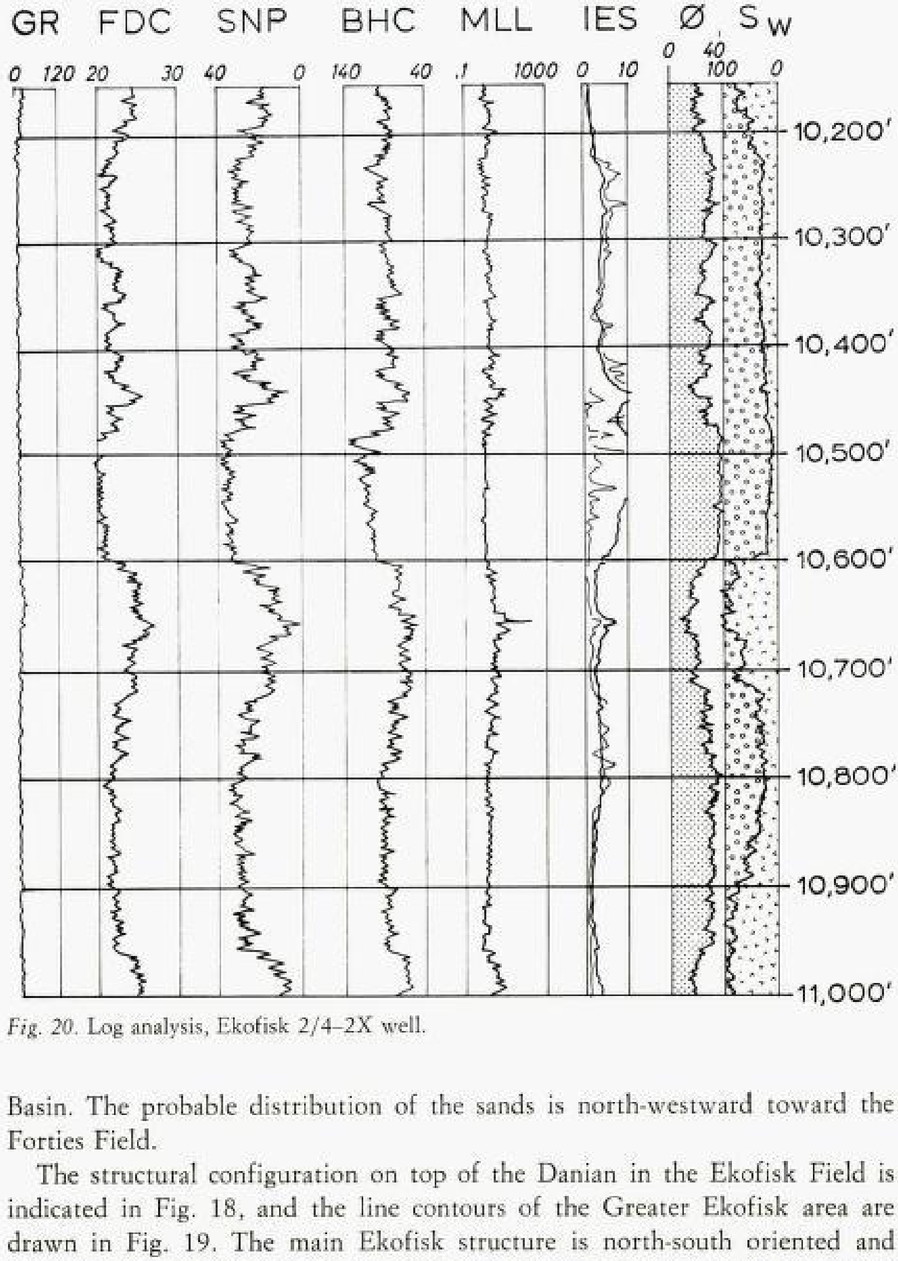

Fig. 20. Log analysis, Ekofisk 2/4–2X well.

Basin. The probable distribution of the sands is north-westward toward the Forties Field.

The structural configuration on top of the Danian in the Ekofisk Field is indicated in Fig. 18, and the line contours of the Greater Ekofisk area are drawn in Fig. 19. The main Ekofisk structure is north-south oriented and 7.5 miles long by 4.5 miles wide. An interesting aspect is that the productive limits do not appear to be entirely structurally controlled. Test information indicates:

1. that in the Ekofisk Field we are dealing with one reservoir only, and
2. that Ekofisk and West Ekofisk are separate reservoirs. Since the productive column exceeds the spill point, there should be continuity between Ekofisk and West Ekofisk. The probable explanation is that the reservoirs are controlled by porosity and permeability.

Fig. 20 shows a suite of logs over the productive interval of an Ekofisk well. Note the zone of reduced porosity from about 10,435′ to 10,510′. This interval can be recognized in all wells drilled to date and was originally thought to separate the upper and lower productive intervals into two reservoirs. Core analysis data would tend to substantiate this contention but test data do not.

The reservoir is a chalky limestone of Danian and Upper Cretaceous age with the high porosities and low permeabilities characteristic of this type of microgranular sediment. Intensive fracturing increases the low primary permeability from less than 1 millidarcy to an average of 10–12 millidarcies. Looking at the lithological characteristics of this carbonate in more detail and in particular at the distinction between 'porous' and 'tight' zones, the fine-grained, homogeneous limestones of the two zones are surprisingly similar in appearance even though their porosities are $> 30\%$ and $< 10\%$, respectively.

It was obvious that conventional thin-section examination would not give the resolution necessary to study such a fine-grained formation, and so a high-powered scanning electron microscope, SEM, was used to investigate the difference between the high and low porosity intervals. In the Fig. 21 photomicrograph, both the porous (32.7%), right, and the tight (8.2%), left, look almost alike. They both basically have the appearance of a foraminiferal micrite, with foraminifera of deep-water facies. Figs. 22 and 23 show with increasing SEM magnification two samples, one of high porosity (32.7%, right) and the other of low porosity (8.2%, left). These Figures illustrate, I believe most spectacularly, the difference between the productive and highly porous section, and the non-productive low porosity section. The highly porous section consists practically exclusively of coccolith fragments and platelets, whereas the low porosity sample clearly shows the secondary calcite growth which reduces the porosity.

In Table 1 the essential reservoir data derived from seismic, drilling and testing are presented. Values obtained from the four producing wells have been averaged. The value 10–12 millidarcies for the permeability should be understood to represent the effective permeability data obtained through testing.

To date, the Phillips Group has established oil and gas production in six fields in the Ekofisk Area (Fig. 24). These six fields plus the Cod Gas–Condensate Field, located 50 miles to the north-west, make up the seven fields currently referred to as the Ekofisk Project.

This project is being developed in four specific phases. The purpose of *Phase 1* was to get the Ekofisk Field on early production to obtain production experience, and reservoir performance information prior to committing to the tremendous investments required for the overall project. This plan was designed for temporary production with conventional equipment, but in an unusual manner. First, the four original wells drilled on the structure were completed using subsea well heads. These wells were then connected to production equipment installed on the 'Gulftide' jackup platform, a former drilling platform. This is done by means of 4½″ flow lines between the wells

Fig. 21. SEM photomicrograph of an Ekofisk Danian limestone.

Fig. 22. SEM photomicrograph magnification of the fields indicated in Fig. 21. Danian limestone, Ekofisk.

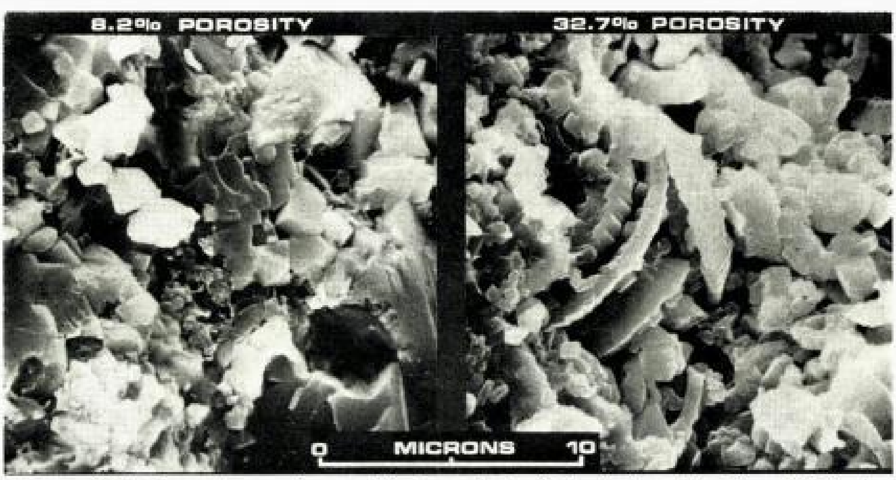

Fig. 23. SEM photomicrograph magnification of the fields indicated in Fig. 22. Danian limestone, Ekofisk. For brief explanation, see text.

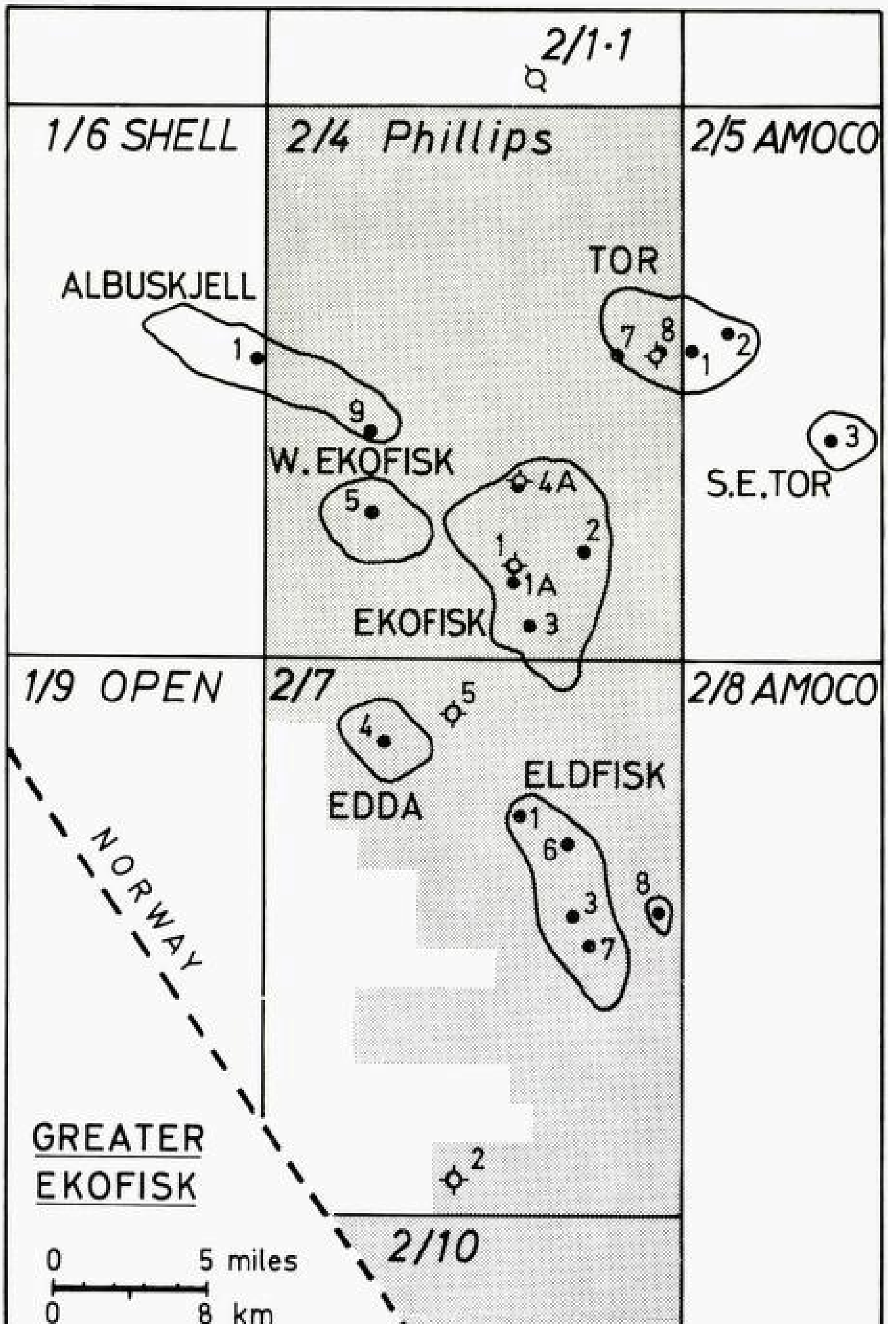

Fig. 24. Greater Ekofisk area, oil and gas fields.

and the platform, where they are then brought in a bundle up through a large caisson to the platform deck. After passing through separation and metering equipment the oil is pumped down the caisson again though one of two 10″

loading lines which go to the mooring buoys. From the buoys, the oil goes directly into a tanker for shipment.

Phase I, the total cost of which has been \$28.5 million, became operational on July 7th 1971, with a production rate of approximately 42,000 BOPD. Since that date, about 26 million barrels have been produced from the field. It has been possible to load tankers 82% of the time; the remaining 18% of the time, severe weather prevents tanker loading, and the field is shut in.

Phase II provides for full development of the Ekofisk field. This phase will require the installation of 6 major platforms, 3 bridge supports, an underwater storage tank, reconnection of the original 4 subsea wells, and the drilling of 38 wells (including 8 injection wells). By March 1974, production will increase to about 80,000 BOPD. The rate will further increase as additional wells are drilled, and by mid-1975 the Ekofisk field will be producing 300,000 BOPD.

The first platform, platform 'B', is located about 1 mile north of the Ekofisk complex (underwater storage). Two drilling rigs began operating from this platform on October 31st 1973, and 17 wells will be drilled from it. Moving south, we come to the most northerly structure in the Ekofisk complex proper, the underwater storage (UWS) tank. An idea of the size of the tank can be gained from Fig. 25. In addition to its use as a storage facility with a one million barrel capacity, the two decks will contain facilities for phase III; these facilities will be discussed later. Platform 'P', which is not installed as yet, will contain 44,000 HP capacity, initiating, crude oil pipeline pumps.

Moving south, the next structure is platform 'C'. Located on this \$23 million platform is a drilling rig with which 8 gas injection and 3 producing wells will be drilled. Drilling commenced on October 3rd 1973. Also to be located on this platform will be high pressure (88,000 HP) gas injection compressors. All gas produced with the 300,000 BOPD phase II will be re-injected back into the reservoir (the maximum volume of gas to be injected is 450 MMCFD).

An adjacent installation is the quarters, a platform which was put in service in July 1973, with services for 185 men. The next installation is the field terminal platform where all Ekofisk production will be processed. The equipment has a capacity of 1 billion CFD gas and 373,000 BOPD. A gas flare platform has also been installed: gas will be flared only in time of emergency.

Drilling platform 'A' (Fig. 26) is located about 2 miles to the south of the flare platform. One rig will drill 9 wells from this platform. Drilling operations started on September 3rd 1973.

Phase II construction is nearly complete and drilling has started on all of the drilling platforms. Concurrently with Phase II, *Phase III* has been proceeding. This consists of developing the Cod, Tor and West Ekofisk fields, completing the installation of the processing facilities on top of the UWS tank, laying oil and gas lines to shore and building shore facilities. Construction of the various platforms is under way. 195 miles (312 km) of the 34" crude line to Teesside, England, have been laid. Only 23 miles (37 km) remain. The

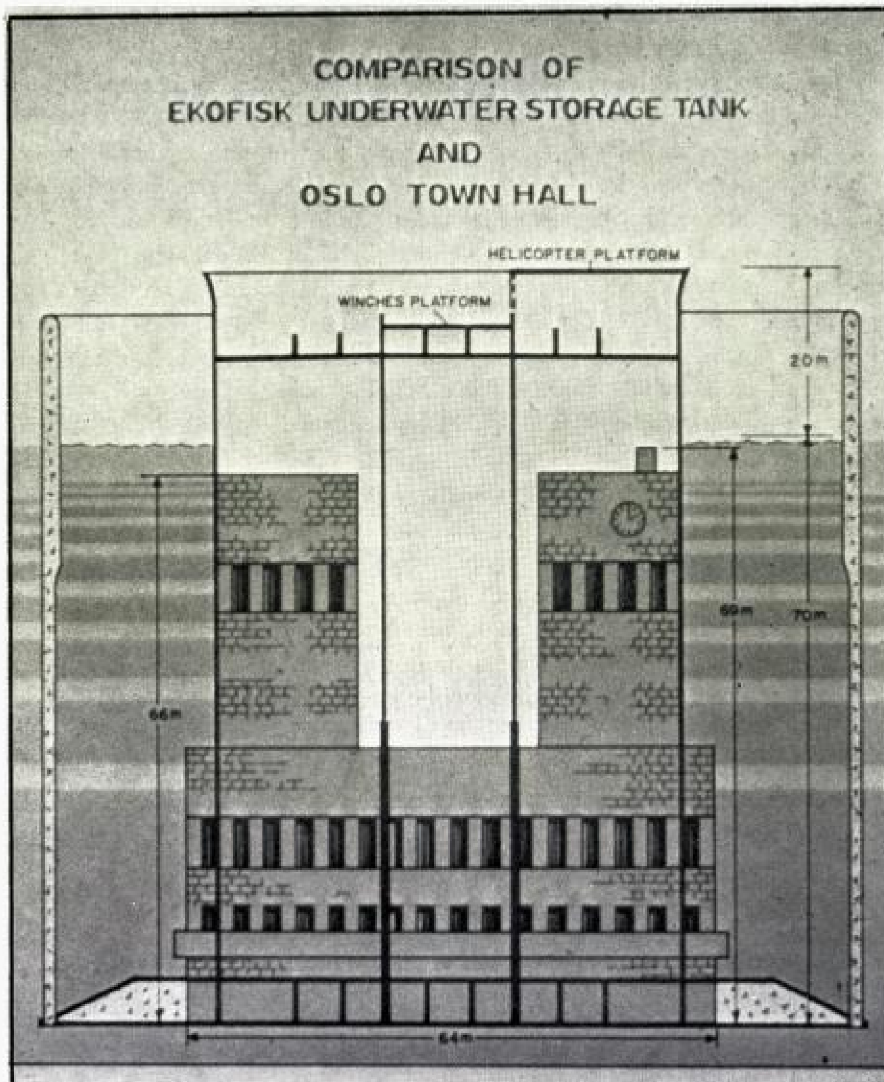

Fig. 25. Comparison of Ekofisk underwater storage tank with the Oslo town-hall.

268 mile (430 km) 36″ gas line to Emden, West Germany, will be laid in the summer of 1974 and the crude line completed. With the pipelines and shore facilities completed in October 1975, the total system will have a capacity of 1 million BOPD and 2.2 BCF gas per day. Maximum production from Phase III is estimated at about 550 thousand BOPD and about 1 BCFGD.

The nerve center of the total Ekofisk project will be the communications, and processing facilities are currently being constructed on the decks on top of the ŪWS tank. The facilities, including Ekofisk center, pipelines and shore facilities ,have been built with a capacity to handle more than Phase III production. Total development will include three additional fields — Albu-

Fig. 26. Drilling platform 'A', Ekofisk fields.

skjell, Edda and Eldfisk — as *Phase IV*. Considerable planning and design have been done and some equipment purchases have already been made in regard to Phase IV, and this is proceeding concurrently with Phase III development. It is difficult to conceive the size and scope of this project, but the investment required to fully develop the project will be more than $2 billion or more than 14 billion Norwegian kroner.

With regard to future possibilities for finding new oil and gas reserves, Fig. 27 shows a generalized stratigraphic column of the North Sea and indicates with symbols the numerous formations in which oil and gas have been encountered. With the large number of targets shown in the geologic column, and the large unexplored areas remaining, there is obviously a large potential for future significant discoveries of both oil and gas.

Also, when one fully recognizes the nearness of this new oil and gas province to one of the world's outstanding energy markets, one can understand why, despite the difficulties of the North Sea operations and the increasing limitations imposed in new concession terms, the area will continue to attract the petroleum industry in its search for new reserves.

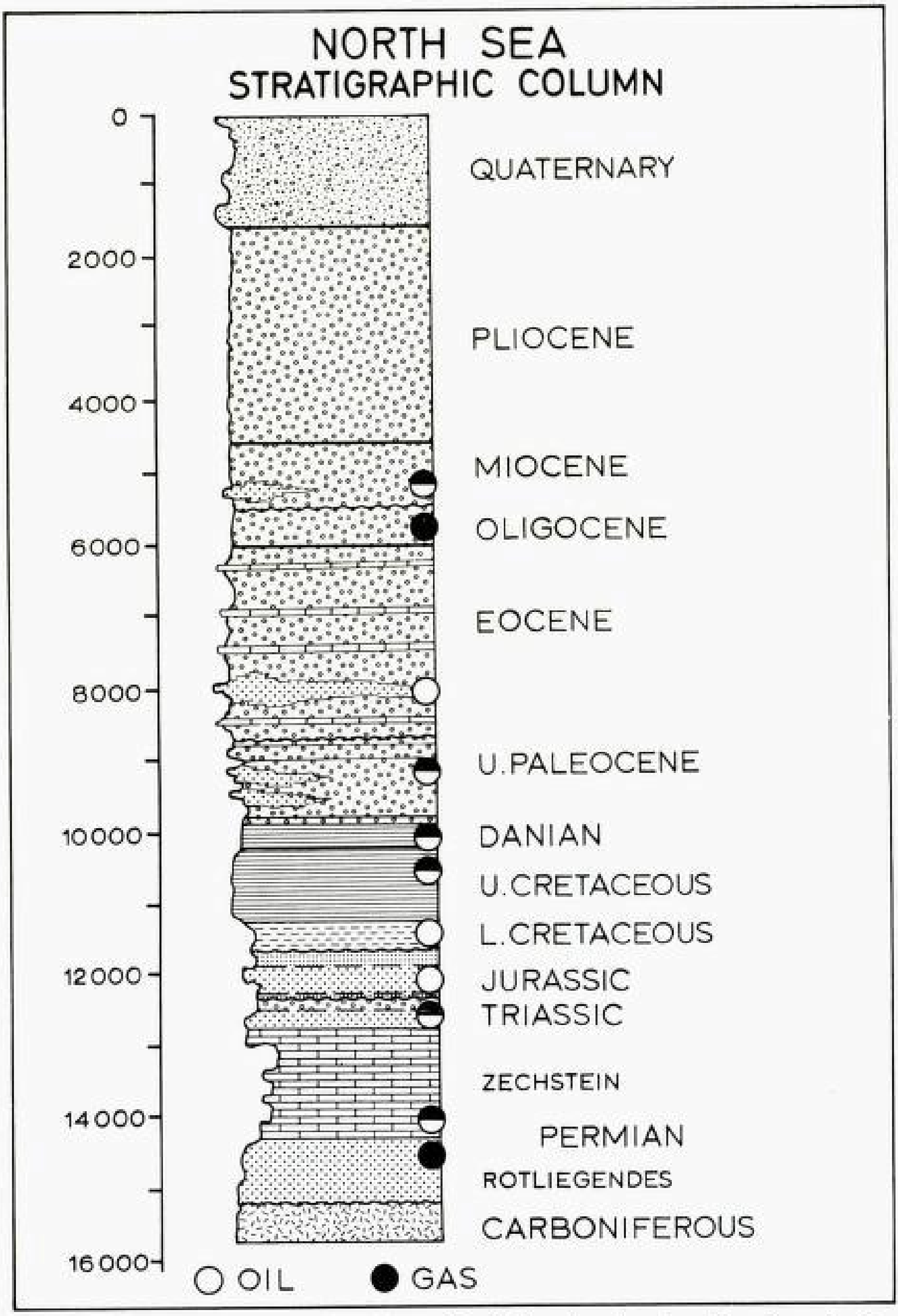

Fig. 27. Generalized stratigraphic column, North Sea, showing the oil- and gas-bearing formations. Thicknesses in feet.

Table 1. Ekofisk field, Danian reservoir data

Lithology	Limestone and chalk
Depth	– 10,400 ft. Mid point
Area	13,625 acres
Net pay	479 ft.
Gross pay	617 ft.
Net to gross ratio	0.776
Porosity	30%
Permeability	10–12 md
Reservoir pressure	7,135 psia (a – 10,400 ft.
Pressure gradient	0.685 psi/ft.
Reservoir temperature	265°F (a – 10,400 ft.
Temperature gradient	2.02°F/ft.
Interstitial water	20%
GOR	1600–1
FVF	1.78
Oil gravity	36° API
Producing mechanism	Solution gas drive

Results of Recent Geological and Geophysical Investigations in Moray Firth, Scotland

M BACON & J. CHESHER

Bacon, M. & Chesher, J. 1975: Results of recent geological and geophysical investigations in the Moray Firth, Scotland. *Norges geol. Unders. 316*, 99–104.

Geophysical, shallow drilling and sea-bed sampling investigations have shown that the main sedimentary basin of the Moray Firth is separated by an ENE-trending, horst-like uplift zone from a secondary basin to the south. Mesozoic sequences are broadly conformable and are dissected by a radiate fault pattern; faulting was most active in pre-Mid Jurassic times. Contrary to previous suggestions, the pattern of outcrop precludes any Mesozoic or Tertiary strike-slip movement on the Great Glen Fault.

M. Bacon & J. Chesher, Institute of Geological Sciences, Marine Geophysics Unit, Edinburgh EH16 6AA, Scotland

The presence of a sedimentary basin in the Moray Firth has long been inferred from the isolated outcrops of New Red Sandstone and Jurassic rocks around its shores (Hallam 1965, Peacock et al. 1968). A shallow seismic and gravity survey carried out by the British Institute of Geological Sciences in 1970, supplemented by shallow drilling with MV Whitethorn, allowed solid geology and gravity maps to be published (Chesher et al. 1972, Sunderland 1972). The geological map showed a basin containing Permian to Lower Cretaceous rocks, delimited in general by the current coastline; the gravity map suggested that the basin was fault-controlled. The Great Glen Fault runs through the mapped area; its nature has been a matter of controversy since Kennedy (1946) suggested a Hercynian sinistral strike-slip displacement of some 105 km. Holgate (1969) has suggested an additional Tertiary dextral movement of about 30 km, and recently Garson & Plant (1972) have proposed a dextral sense for Hercynian movement.

To elucidate the structure of the inner part of this basin, IGS in 1972 commissioned Seiscom–Delta to carry out a deep seismic survey on a 10 by 15 km grid (Fig. 1). Forty-eight-fold coverage obtained during shooting was reduced during processing to an effective 24-fold coverage; the data were processed down to 4 seconds travel time. Several relatively continuous major horizons have been recognised within the area; they have been tentatively identified on the basis of correlation with IGS shallow seismic surveys, IGS sea-bed sampling and shallow drilling results. These horizons range from Lower Cretaceous to ?Permo–Trias in age. Because block 11/30 in the outermost part of the area is currently licensed to a UK operator, this report will concentrate on the inner part of the Firth.

Fig. 2 shows a map of depth to a horizon tentatively identified as base Lower Cretaceous. This is a very strong and continuous event, possibly

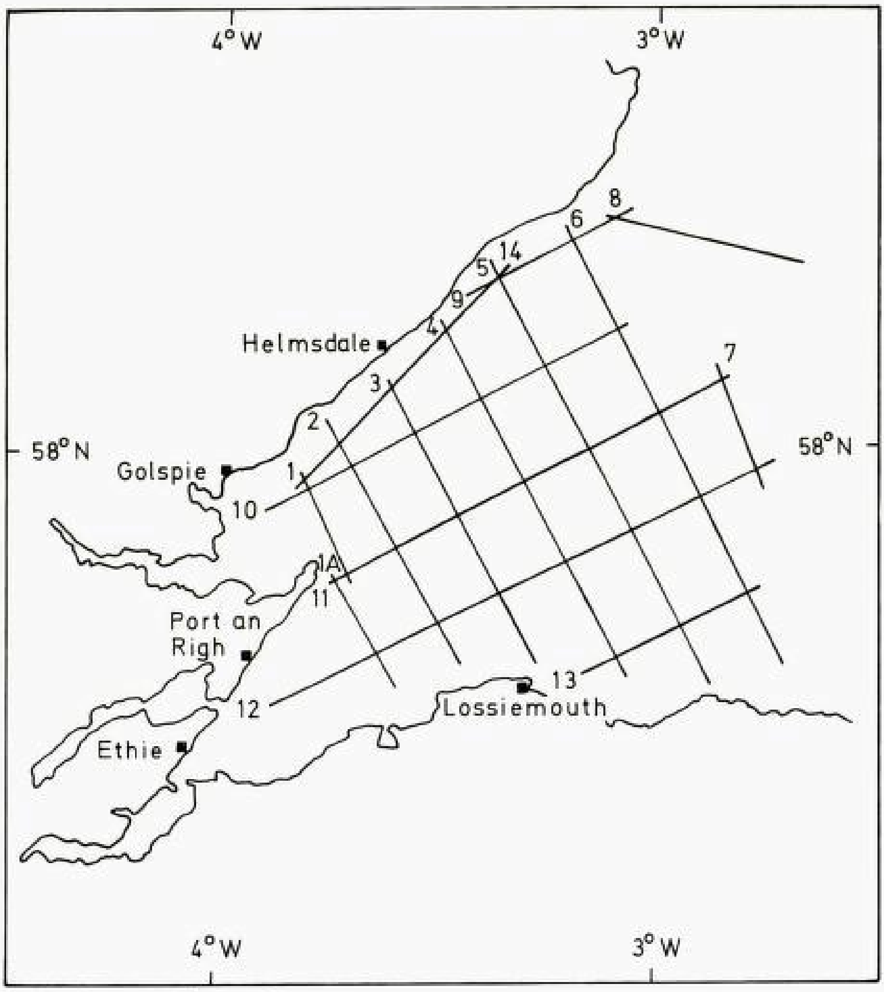

Fig. 1. Location of seismic lines.

representing a widespread unconformity. The horizon follows structural changes in the deeper horizons though it is less affected by faulting. Calculations of RMS velocity to the horizon show high velocities corresponding to regional basin structure and low velocities corresponding to regional highs; thus the apparent topography of the travel-time map is likely to be real.

Fig. 3 shows a map of depth to a horizon thought to be base Jurassic; this is the strongest deep reflector, and can be followed reliably through the central and southern parts of the area, although correlation across the Great Glen Fault is poor. In general the horizon deepens, broken by numerous small and medium faults into the axis of the main basin against the Great Glen Fault zone. In the southern part of the map the most prominent feature is a ridge, fault-bounded on its SE side, to the south of which we see a subsidiary basin.

Fig. 4 summarises the main structural features of the basin. The main basin, the axis of which trends NE–SW, is centred in the NE of the area, and is

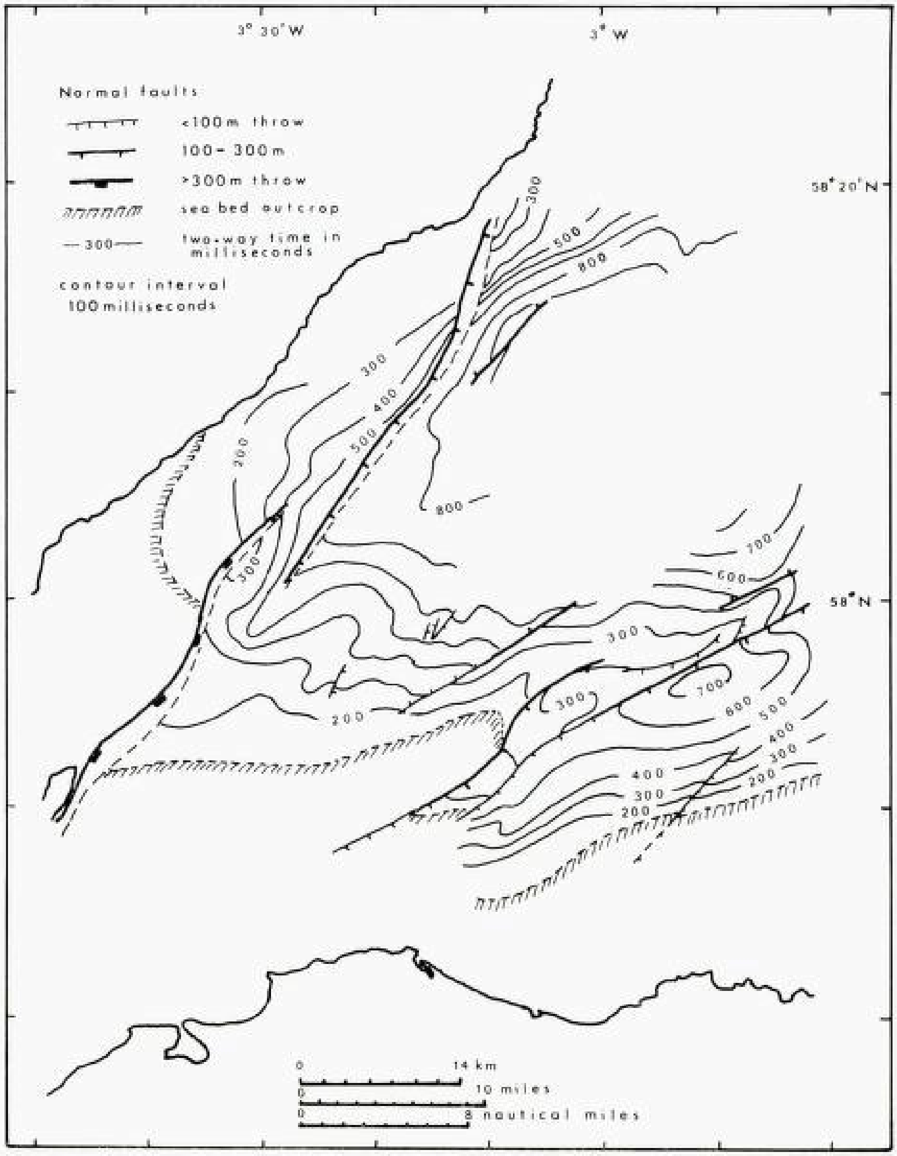

Fig. 2. Map of probable base Lower Cretaceous horizon.

separated by a prominent zone of uplift from a secondary basin to the south. This zone of uplift trends ENE across the centre of the area and seems to be horst-like in structure; the faulting appears to have controlled sedimentation, at least in part. This also appears to be true of the Great Glen Fault.

Horizons within the basin are broadly conformable, though minor discontinuities exist at the base of the Lower Cretaceous and above the base of the Jurassic. There is uniform dip towards the centre of the basin though horizons are broken by normal faulting to form a series of horst and graben structures; they are relatively little affected by folding. The faults vary in

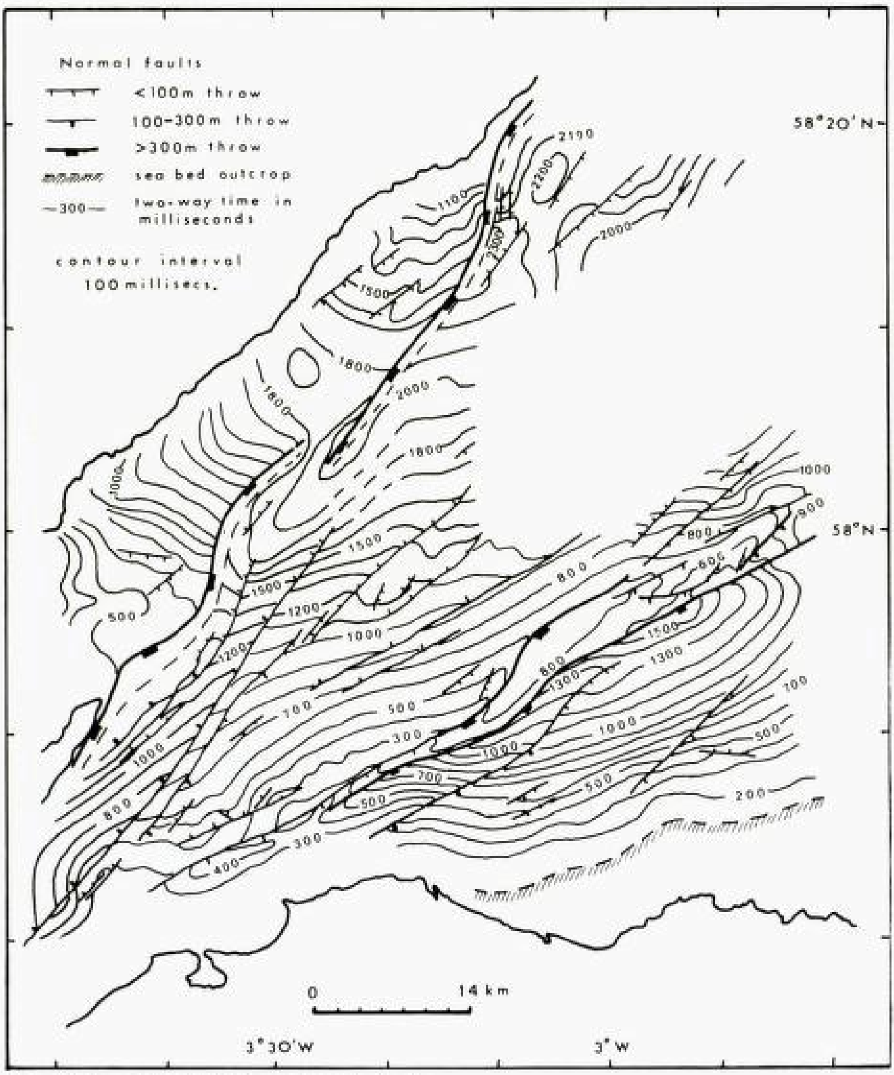

Fig. 3. Map of probable base Jurassic horizon.

trend from NNE in the NW to ENE in the SE of the region, appearing to radiate from a focal point in the SW. Fault activity has been recognised on all the horizons mapped and probably extended throughout the Mesozoic; it was, however, most active pre-Mid Jurassic, and gradually decreased in intensity until during the Lower Cretaceous only the major fault lines remained active.

The Great Glen Fault is undoubtedly the most important fault in the area. The fault zone follows the coastline between Ethie and Tarbat Ness, down-faulting Jurassic against Old Red Sandstone, and continues in a NNE direction with a small northerly kink. Where the zone crosses line 1 it is some 2.5 km wide and has a downthrow to the south of about 800 m/sec for the base–Jurassic horizon. The zone decreases in both width and downthrow to the NE

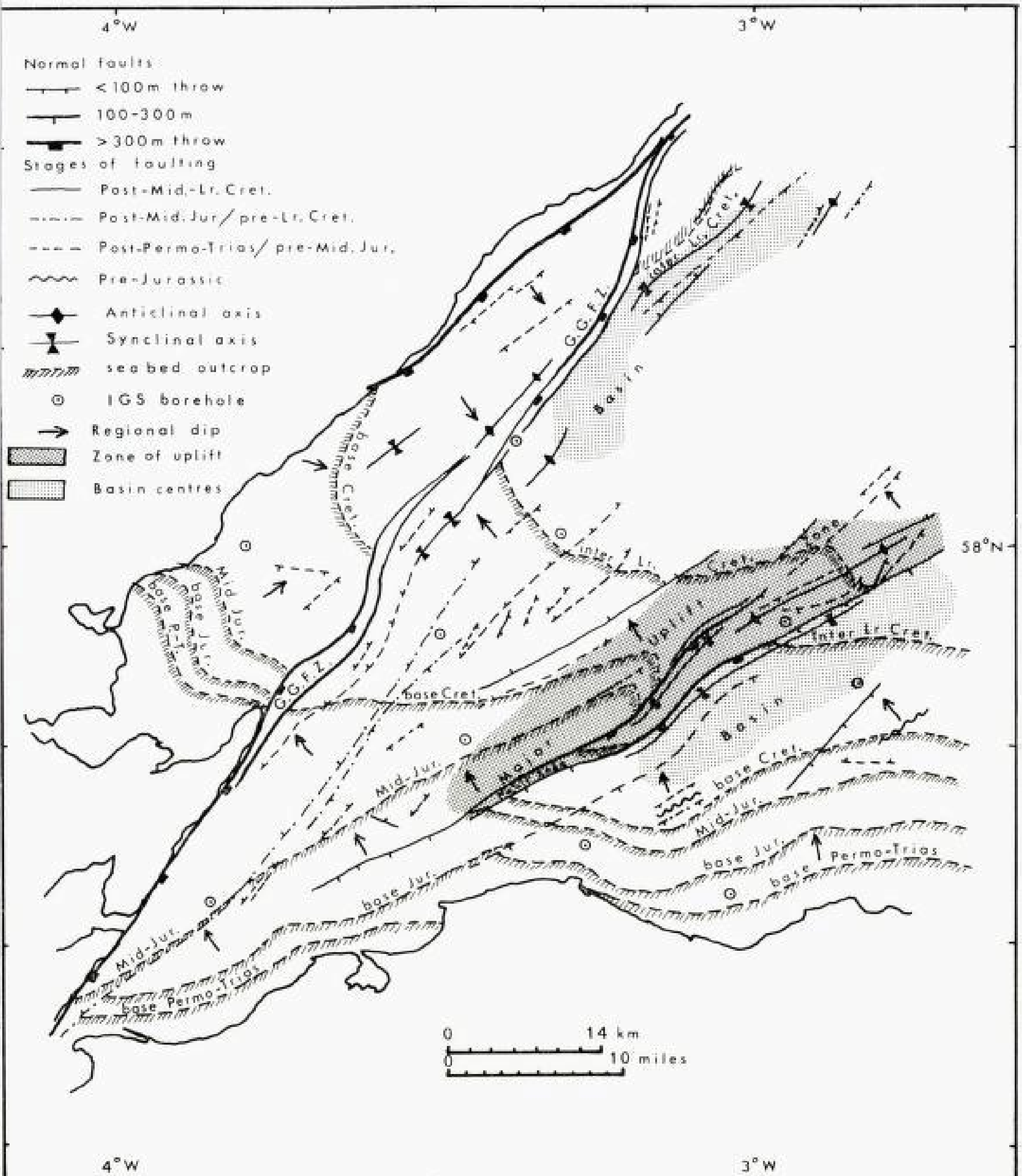

Fig. 4. Structural summary map.

until where it crosses line 3 it is only 8 km wide with a throw of only 500 m/sec on the base Jurassic. Near line 4 the fault line passes into the northern limb of a broad syncline and the fault movement is taken up further south along the axis of the syncline. The fault then continues in a NNE direction with downthrow increasing to 1000 m/sec on the base Jurassic at line 6, where it is 1 km in width. This pattern of outcrop precludes any Mesozoic or Tertiary strike-slip movement on the Great Glen; the fault plexus appears to be entirely normal in character during this period.

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The Forties Field*

A. N. THOMAS, P. J. WALMSLEY & D. A. L. JENKINS

Thomas, A. N., Walmsley, P. J. & Jenkins, D. A. L. 1975: The Forties Field. *Norges geol. Unders. 316*, 105–120.

The Forties Field is a large oil pool which was discovered in 1970 in the northern part of the British sector of the North Sea some 175 km (110 miles) east of Peterhead, Scotland, in water depths of 91 to 131 metres (300 to 430 ft). The reservoir is a sandstone of Paleocene age occurring at a depth of about 2135 metres (7000 ft) at the base of a thick Cainozoic section, consisting primarily of mudstone. The Paleocene is a sandstone/mudstone sequence and is underlain by Danian and Maastrichtian chalk. The trap is a broad low-relief anticlinal feature with a closed area of about 90 km² (35 sq miles). Maximum gross oil column is 155 metres (509 ft). Recoverable oil is estimated at 1.8 billion barrels from an in-place figure of 4.4 billion.

A. N. Thomas, P. J. Walmsley & D. A. L. Jenkins, BP Petroleum Development Ltd., Farburn Industrial Estate, Dyce, Aberdeen, Scotland

Introduction

The Forties Field is located in the British sector of the North Sea. It lies approximately 175 km (110 miles) due east of Peterhead, Scotland (Fig. 1). The major part of the field falls within BP Licence Block 21/10, the eastern end extending into Shell/Esso Block 22/6. Approximate geographical co-ordinates are latitude 57°45'N, longitude 0°45'E. The field is named 'Forties' after the meteorological area in which it was discovered. Weather conditions in this area are severe with frequent gales, particularly in the winter months, and with waves exceeding 5 metres (15 ft) for about one-third of the time.

Water depths across the field vary from 91 metres (300 ft) in the south-east to 131 metres (430 ft) in the northwest. Bottom conditions generally consist of soft clays overlain by a variable thickness of muds.

History of discovery

The Forties Field was discovered in October, 1970, when BP's well 21/10–1 found oil in Paleocene sands at a depth of about 2135 metres (7000 ft). Block 21/10 formed part of a U.K. 2nd Round licence which had been awarded to BP in November, 1965. At that time little was known about the geology of the North Sea, particularly the northern part. Only five marine wells had been drilled in British waters at the time of application for the Forties licence and gas had yet to be discovered in the southern North Sea. It was realised that the North Sea covered a large Tertiary sedimentary basin possibly over-

* This paper was first read at the Annual Convention of the American Association of Petroleum Geologists at Anaheim on 14 May 1973, and was published in the Bulletins of the A.A.P.G. in March 1974.

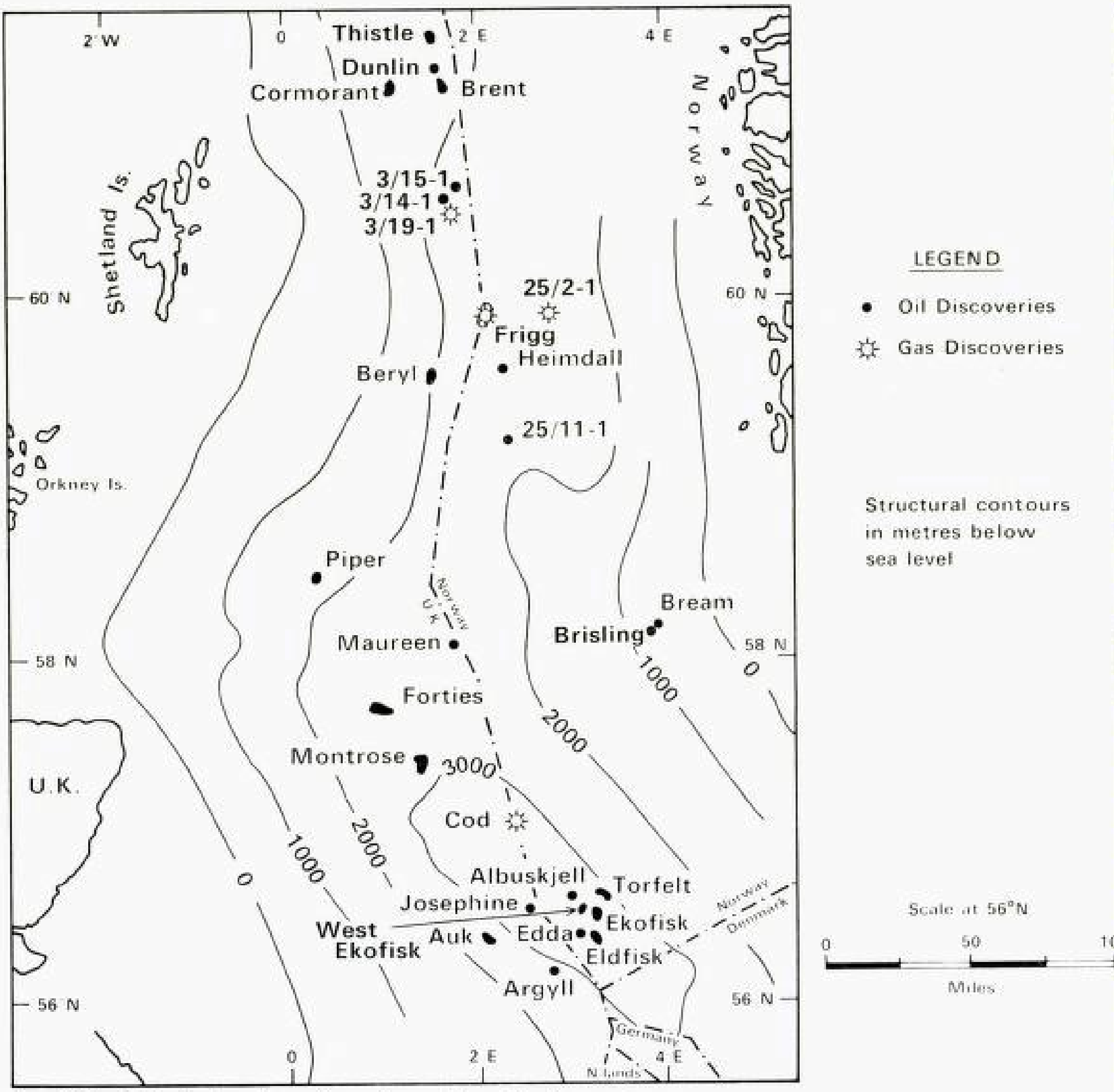


Fig. 1. North Sea hydrocarbon discoveries and base Tertiary contours.

lying in part a thick development of older sediments, but the main interest was centred on the gas area to the south and oil prospects in the northern area were at best thought to be highly speculative. Nevertheless the possibility that oil could occur was recognised as early as 1965, although the subsequent success exceeded all expectations at that time.

During the mid 1960's the main exploration activity was in the southern North Sea. The first indication that the northern area might be an oil province occurred in July 1967 when the second well to be drilled in Norwegian waters, Esso 25/11–1, encountered oil shows. One or two small discoveries were made in Danish waters shortly afterwards, and in June 1968 Phillips made a gas/condensate discovery, the Cod Field, in Norwegian Block 7/11. However, none of these fields was large and by the end of 1969, with over 50 exploration well drilled in northern waters, hopes of making economic oil discoveries had begun to dwindle. It was in December 1969 that Phillips made their important Ekofisk discovery in southwest Norwegian waters (Fig. 1). This, together with the simultaneous, but smaller, Montrose discovery by Amoco/Gas Council in UK Block 22/18, revived the interest of the petroleum industry. The tempo of exploration accelerated and further successes followed rapidly. It was in this climate that BP spudded its exploration well 21/10–1 with Sea Quest in August, 1970.

Early reconnaissance seismic work pre–1965 had indicated a large structural nose plunging southeastwards across Block 21/10 towards the deepest part of the North Sea Tertiary Basin. A 5 × 5 km (3 × 3 miles) seismic grid shot in 1967 had defined this feature and had shown, at base Tertiary, 40 km^2 (16 sq. miles) of closure of low amplitude centred in Block 21/10 upon the axis of the nose. 21/10–1 was drilled upon this feature.

At 2132 metres (6994 ft) below R.T.E. of 34 metres (111 ft) the well entered Paleocene sands, which were indicated to be oil-bearing from mud gas and cuttings analysis. An oil-water contact was clearly established at 2251 metres (7585 ft) b.R.T.E. and subsequent testing produced 37° A.P.I. low sulphur oil at a rate of 4730 bbl/day on a 54/64 in. surface choke. A field of major proportions had been discovered.

A detailed 1.5 × 1.5 km (1 × 1 mile) seismic survey was shot immediately to supplement the 1967 work and the combined data interpreted in the light of the information gained from the discovery well. The new interpretation indicated a larger closed area than had originally been envisaged.

The first appraisal well 21/10–2 was spudded in June 1971 5.5 km (3.4 miles) northwest of the discovery well in order to delineate the field in that direction (Fig. 2). This well found an oil column of 33.5 metres (110 ft) with an oil-water contact at the same depth as that found in 21/10–1. A second appraisal well, 21/10–3, was then spudded 7 km (4.3 miles) west of 21/10–1, but was junked at shallow depth. A replacement well, 21/10–3A, was drilled without incident and found an oil column of 126 metres (413 ft) and once again the same oil-water contact. At the same time Shell/Esso drilled a well in Block 22/6 which was also successful, although the sand development in the upper part of the Paleocene above the oil-water contact showed considerable deterioration compared with the other Forties wells. The same oil-water contact occurred.

By this time plans for the construction of four platforms to develop Forties were well advanced, but some doubts remained as to the structure and sand distribution on the southern flank of the field. In order to select finally a site for the fourth drilling platform a further well, 21/10–5, was drilled 5.5 km (3.5 miles) west-southwest of 21/10–1. This also proved successful and completed the appraisal of the field, (Fig. 2). The first five wells have therefore confirmed the occurrence of a major oil field with an oil column of 155 metres (509 ft) in Paleocene sands, and a closed area of about 90 km^2 (35 sq. miles).

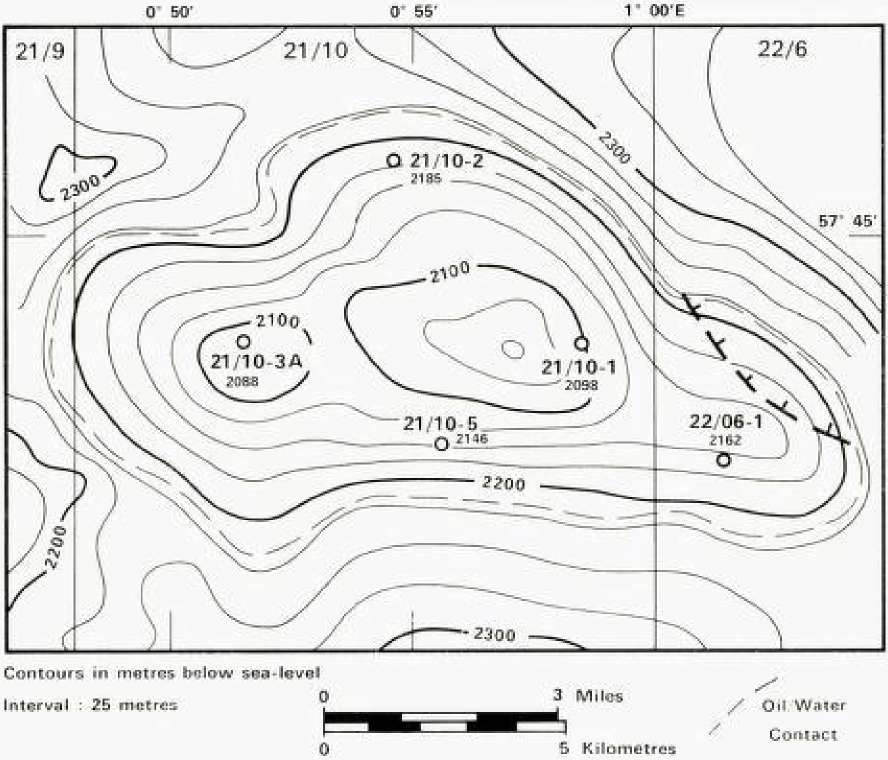


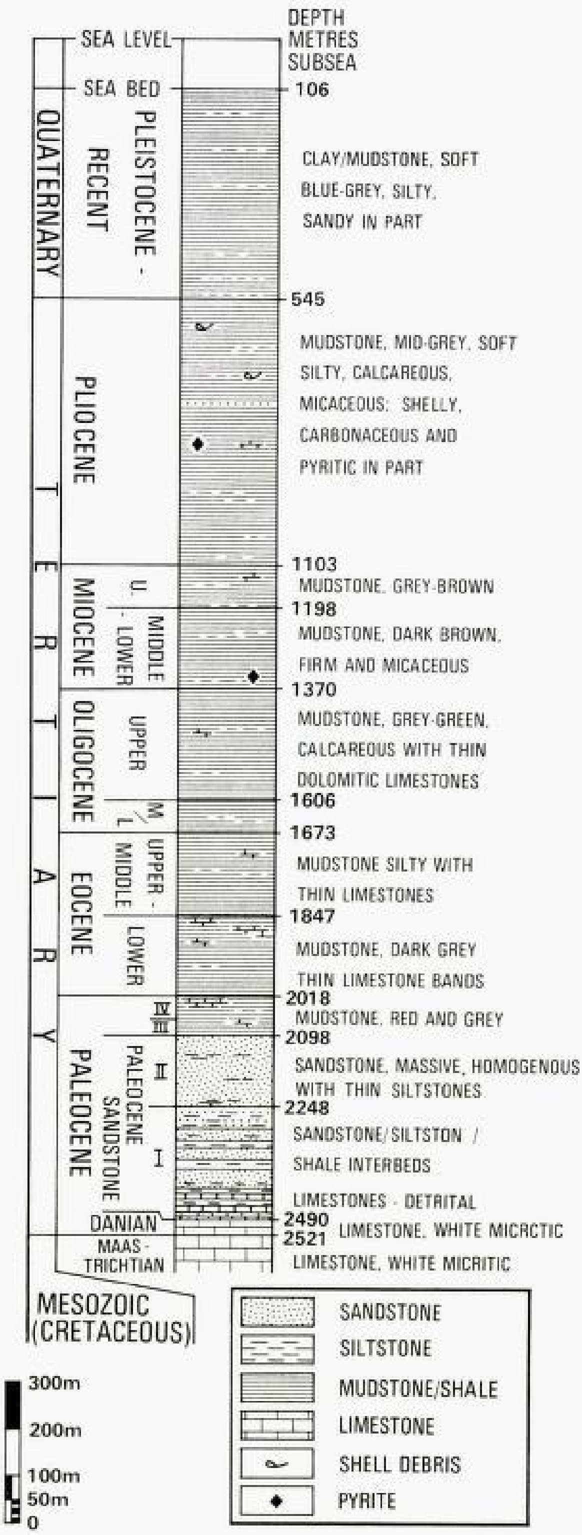
Fig. 2. Structural contours top Paleocene reservoir.

Stratigraphy

The Forties Field lies on the western flank of the North Sea Tertiary basin, the axis of the depositional trough trending approximately north-south along the median line of the continental shelf between the U.K. and Norway (Fig. 1). The Tertiary and Quaternary section within this trough can reach 3500 metres (12,000 ft) in thickness. Apart from the Netherlands and southern England the basin is developed wholly offshore, Tertiary strata being absent from both eastern Scotland and Norway. The Tertiary section in the North Sea basin is almost entirely terrigenous, the predominant lithology being mudstone. Sandstones occur at a number of stratigraphic levels and in some areas can become the major lithotype within the Paleocene and Plio-Pleistocene. Producing sands are primarily of Paleocene age. The Tertiary sequence has been dated using microforaminiferal assemblages.

The central part of the Tertiary Basin overlies a Mesozoic section which also becomes attenuated towards the Scottish coast. The generalised sequence comprises micritic limestones of Upper Cretaceous age underlain by Lower Cretaceous mudstones. The Jurassic, where present, is developed as mudstone with subordinate sands and the Triassic is a distinctive red-brown mudstone. A thin carbonate/anhydrite section, the Zechstein equivalent, occurs at the

Fig. 3. Stratigraphy – Well 21/10–1.



top of the Paleozoic, with salt developed in the central part of the basin. The Zechstein is underlain by the Permian Rotliegendes sandstone.

The stratigraphic column for the Tertiary section of the discovery well in the Forties Field, 21/10–1 is given in Fig. 3. The oil accumulation is in sandstones of Paleocene age, which occur beneath a thick monotonous section of grey to brown variably calcareous and carbonaceous mudstones, ranging from Upper Paleocene to Holocene. Sandstones occur in the Plio-Pleistocene and thin beds of limestone in the Eocene, but the post-Paleocene section is primarily argillaceous. The stratigraphic subdivisions shown for the Tertiary in Fig. 3 are based on micropaleontology. In well 21/10–1 the Paleocene is 502 metres (1647 ft) in thickness. The Eocene/Paleocene boundary is placed on paleontological evidence at 2018 metres (6621 ft) sub-sea, which is close to a prominent peak on the gamma-ray log and a thin limestone band which forms a useful lithological marker. The basal 30 metres (98 ft) of the Paleocene are white to grey micritic limestones of Danian age, but the overlying sequence is predominantly terrigenous. The exact boundary between the Danian and underlying Maastrichtian limestones is not well controlled paleontologically, but is taken at a gamma-ray sonic log marker reflecting a change in limestone character from the clean compact micrites of the Maastrichtian to slightly argillaceous and sandy micrites of the Danian (Fig. 4).

The post-Danian section of Paleocene has been divided by the late Dr. M. J. Wolfe and Mrs. L. Aston of the BP Research Centre at Sunbury into four distinctive lithostratigraphic units, which are shown on the column in Fig. 5. The units are referred to by roman numerals only and no formal terminology is proposed at this stage.

UNIT I 2248–2490 METRES (7375–8169 FT) SUB-SEA

This is divisible into two 'members', the lower (below 2409 metres, or 7903 ft sub-sea) including beds of detrital limestone and the upper comprising argillaceous sandstones interbedded with siltstones and silty shales.

The lower member contains a basal sandstone 4.6 metres (15 ft) thick overlain by a sequence of calcareous sandstones, limestones and mudstones. The unit is characterised paleontologically by reworked Danian and Cretaceous faunas. The limestones are commonly detrital with disseminated quartz of sand grade and indicate contemporaneous erosion of Cretaceous and Danian carbonates following uplift at the end of the Mesozoic. The thickness of the member increases north-westwards across the field, the variations probably reflecting partial infilling of an irregular depositional surface.

The upper member consists of interbedded sandstones, siltstones and shales. Sandstone is the commonest lithology and is typically grey, fine-grained, argillaceous and poorly indurated. The sonic log is characterised by a rapidly fluctuating trace representing thinly interbedded lithologies. Thin bands of limestone and hard calcite-cemented sandstones occur infrequently throughout.

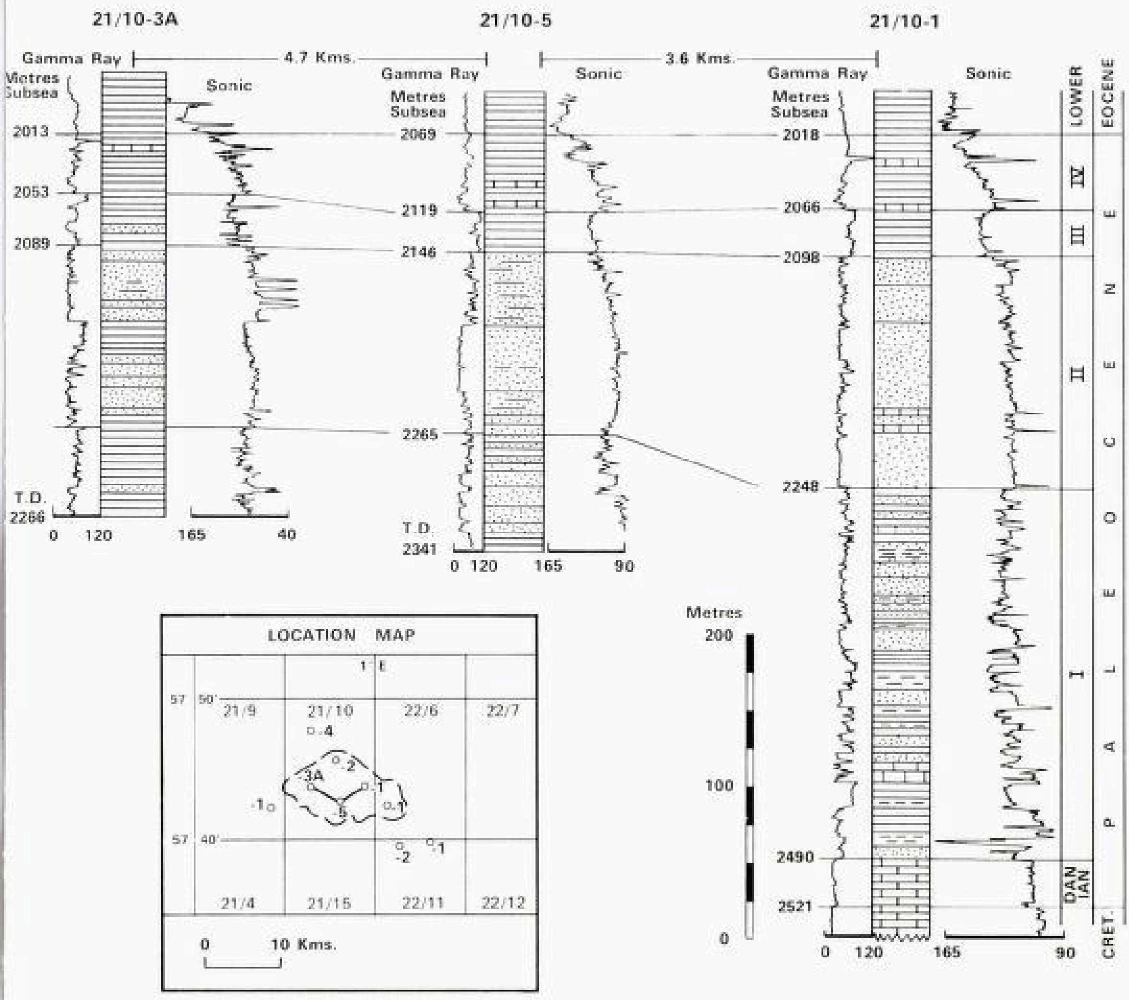


Fig. 4. Well correlation 21/10–3A, 21/10–5, 21/10–1.

UNIT II 2098–2248 METRES (6883–7375 FT) SUB-SEA

Unit II contains all the massive sandstones within the Paleocene and is the producing 'formation' of the Forties Field. It is also productive of oil and wet gas in a number of other structures in the southern part of the northern North Sea. The dominant lithologies are sandstone and mudstone. Studies of core material from wells within the field have allowed the contrasting lithologies to be grouped into four facies, referred to as A, B, C and D.

Facies A is characterised by fine-grained sometimes silty sandstones, frequently with detrital mica and lignite, which are interbedded with laminated siltstones and shales, often as upwards-fining graded units commonly less than 1.5 metres (5 ft) in thickness. The shales are typically kaolinitic and the fauna is sparse. The beds vary from 2 cm to 1.7 metres (5.5 ft) in thickness and the sandstone percentage in the lithofacies varies between 20 and 55%, with the sands having moderate to good reservoir character. Flow and load structures

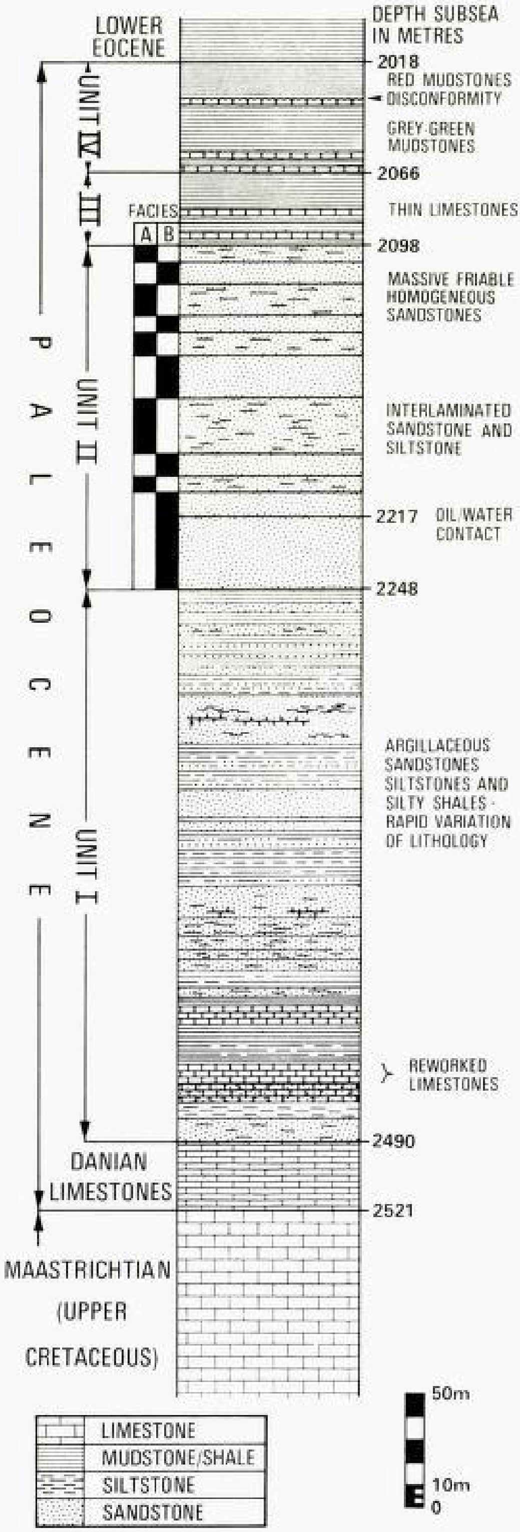


Fig. 5. Paleocene lithostratigraphy Well 21/10–1.

are common, including ripple-drift lamination, load casting, deformed cross bedding, contorted bedding and slump folding, and soft sediment faulting. At a few horizons are intraformational conglomerates with shale clasts up to several centimetres in size.

Facies B comprises the clean homogeneous sandstones. These vary in colour from brown to almost white, in grain-size from fine to coarse grade, and are typically clean and friable. Sorting is poor to moderate, but reservoir properties are excellent. Within thick sequences of sand there are clay laminae and impermeable calcite-cemented zones, and occasional pebbly layers with 50% lithic fragments which are mainly quartzose, phyllite and graphic granite. Fining-upward sections are common, often with convolute and laminar bedding towards the top. Current bedding has not been observed. The sandstones of Facies B in well 21/10–1 vary in thickness up to 35 metres (115 ft), but elsewhere in the field can reach 80 metres (260 ft). The thick sections are believed to be formed of superimposed sand units which are individually about 0.5 to 1.0 metres (1.5 to 3 ft) thick. Fauna is again sparse, as in Facies A.

Facies C is dominated by grey kaolinitic shales and graded siltstone/shale couplets. Occasional thin, dark-brown, fine-grained, sometimes graded sandstone beds occur. The bases of these beds are unusually sharp, erosional and sometimes have sole marks, including rare flute casts. The siltstones show wavy lamination, micro cross-lamination and lenticular bedding. Penecontemporaneous slump structures are common. The shales are rarely bioturbated and fauna is sparse.

Facies D is found only in the eastern part of the field. It is characterised by burrowed green waxy shales which have an abundant marine fauna and flora. Associated with this lithology are purple shales, black limestones, dolomitic mudstones and sideritic concretions.

Over the greater part of the field Unit II is made up of Facies A and B. Facies C occurs in the southern and eastern part of the field and Facies D in the east. The distribution of the four facies appears to be related to thickness variations of Unit II, Facies C occurring within areas of thin development of the Unit and Facies B sands being well represented in areas of thick development. The central part of the field has the highest percentage of clean sands and will have the best production potential.

The change in facies across the field means that a correlation within Unit II is not readily apparent from the logs of the widely spaced step-out wells. The variation in gross lithology is illustrated by the correlation diagram in Fig. 4, which is drawn in a west to east direction from 21/10–3A, through 21/10–5 to 21/10–1. The argillaceous section in the upper part of Unit II in 21/10–5 (Facies C) is probably stratigraphically equivalent to the Facies A and B sands in wells 21/10–1 and 21/10–3A, although an off-lapping relationship could be invoked.

Studies on this and related problems are continuing, but will probably not be resolved until further data from the production wells is obtained. The depositional environment of Unit II is still under study and it will be some time before it is completely elucidated. As additional data become available during the development drilling programme it is expected that critical evidence of the provenance and mode of deposition will be found. At present we consider the most likely source of the sediment to be from Scotland to the west and from locally eroded high areas composed of Upper Cretaceous and Danian chalky limestones. A Scottish provenance for the more arenaceous material is indicated by the predominance of quartzose grains including quartzites and vein quartz, some phyllitic material, hornblende gneiss and pegmatite fragments including much graphic granite, and by the heavy mineral suite consisting predominantly of tourmaline, garnet and zircon, with lesser amounts of staurolite, rutile, sillimanite, hornblende and pyroxene. These features appears to indicate derivation from rocks akin to the Moinian and Lewisian of Scotland.

UNIT III 2066 – 2098 METRES (6778 – 6883 FT) SUB-SEA

Lithostratigraphic Unit III forms the cap rock to the reservoir sands. The contact with Unit II was cored in well 21/10–2 and indicates that the lithological change is transitional. Unit III consists of dark grey, silty, lignitic, shaley mudstones rich in montmorillonite. Thin beds of sand are occasionally developed with the proportion of sandstone increasing westwards. Winnowed pockets of fish remains and several well-preserved skeletons have been found in cores. Planktonic foraminifera are rare, the fauna being dominated by siliceous diatoms and microplankton. The thickness and lithology of Unit III are very constant across the field and the uniformity extends to nearby areas.

UNIT IV 2018 – 2066 METRES (6621 – 6778 FT) SUB-SEA

This is essentially a mudstone unit, the mudstones typically being greenish-grey and slightly calcareous. The upper 10–20 metres (33–66 ft) is, however, red-brown in colour and contains a characteristic red-stained, calcareous, foraminiferal assemblage in which *Globigerina cf. triloculinoides* predominates. These mudstones are separated from the underlying darker coloured beds by a minor unconformity. Several thin, pale-coloured, clay beds with a distinctive mineralogy indicative of degraded volcanic ash are present, together with occasional thin limestone stringers. These lithological features extend beyond the limits of the field and make log markers useful for correlation. The thickness of the unit remains constant across the field and it displays a characteristic sonic log pattern with increase in velocity with depth. This velocity increase serves as a good log correlation feature, and also gives rise to a prominent seismic horizon which is used to map the configuration of the top of the reservoir.

Structure

The regional tilt at the base of the Tertiary sequence in the Forties area of the North Sea is down to the east, towards the axis of the Tertiary Basin. In Block

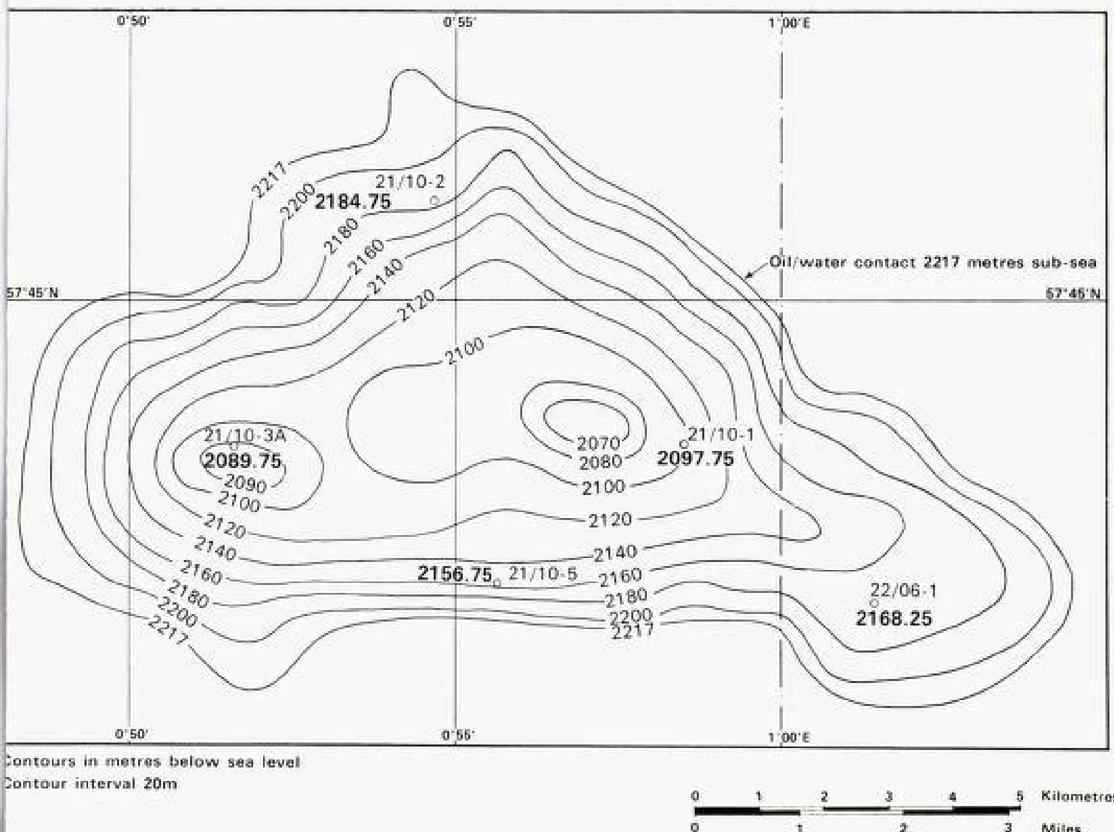


Fig. 6. Isochrons on seismic reflector overlying Paleocene reservoir.

21/10 this regional tilt is interrupted by a large ESE-directed nose which can be followed into Block 22/11. Reversal of dip on this nose provides closure for the hydrocarbon accumulation in the overlying Paleocene sands.

The isochrons on a seismic horizon some 35 metres (115 ft) above the top of the sands are given in Fig. 6, and the depth contours on the top of the reservoir in Fig. 2. A structural cross-section east-west across the field is shown in Fig. 7. Because of horizontal velocity gradients present in the late Tertiary section, the configuration of the structural contours on the top of the reservoir is different from that of the isochrons on the seismic horizon directly above. The velocity gradients are computed from the information from the wells within and adjacent to the field. The most significant change is an increase in closure at the western spill points of the structure, giving an approximate coincidence between structural spill point and the oil water contact at 2217 metres (7274 ft) sub-sea.

The depth contours on the top of the reservoir indicate that the structure is a broad dome elongated east-west with minor faulting affecting only the eastern extremity of the feature. The structure extends 16 km (10 miles) east-

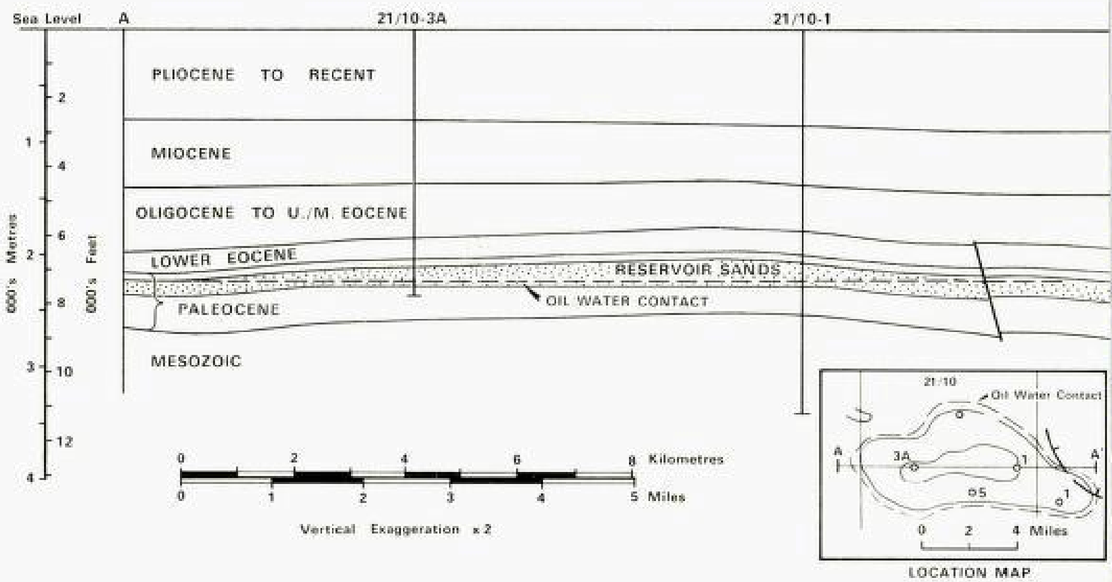


Fig. 7. Structural section E–W across the field.

west by 8 km (5miles) north-south, has a closed area of about 90 km² (35 sq. miles or 22,000 acres) and vertical closure of 155 metres (509 ft).

The Paleocene structure overlies a faulted high at the base Cretaceous level. Seismic penetration below the Cretaceous is poor, but it does not appear likely that the structure in the deeper Mesozioc or Paleozoic is directly related to the base Cretaceous uplift over which the Paleocene feature lies. The base Cretaceous unconformity is a very pronounced widespread regional event affecting the whole of the North Sea basin. The late Jurassic movements which preceded this unconformity were typified by block faulting and a complex uplift of this type underlies the Forties Field. Thickness variations in the Cretaceous and Lower Tertiary indicate that the area of the field remained relatively positive during late Mesozoic and early Cainozoic times, a favourable factor for early migration of hydrocarbons, although at present insufficient data are available to assess the organic diagenesis of the Cretaceous and Paleocene shales. Structural expression above the Paleocene diminishes progressively. Turnover on the nose does not persist above the Lower Miocene and at the base Pliocene level the beds dip uniformly south-eastwards across the field. Hence structural growth was terminated in late Miocene times. Uplift on the Mesozoic block fault may however have ceased during the early Tertiary, with differential compaction allowing structural expression to persist through the Miocene section.

Reservoir characteristics and reserves

Well 21/10–1 was tested over the interval 2108.5–2113 metres (6918–6933 ft) sub-sea and produced 37° API oil at a rate of 4730 bbl/day on a 54/64 inch surface choke with an estimated solution gas-oil ratio of 250 scf/bbl. The crude was low sulphur (0.3%) and medium wax at 8.5%. 21/10–3A was tested over the interval 2092–2137.5 metres (6865–7013 ft) sub-sea at rates up to 3260 bbl/day, the production being limited to this rate by equipment restrictions. Bottom hole PVT samples during the test established the GOR at 330 scf/bbl.

Core analysis of the five wells, particularly 21/10–5, where extensive coring of the reservoir was carried out, indicates high porosities and permeabilities. Average well porosities range from 25 to 30% and permeabilities vary up to 3900 md. With the exception of the tight calcite cemented layers the permeability in Facies B varies from 1000 m to 3900 md. In Facies A and C the sandy layers have a spread of permeabilities from 0.1 to 1000 md with the most common permeability measurements lying in the range of 100–200 md.

Analysis of the flow test indicates that 21/10–1 and 21/10–3a are capable of producing rates in excess of 15,000 bbl/day. Both wells were, however, drilled in the crestal position where virtually the full oil column is present. An initial average well rate of 8000 bbl/day for development wells has been assumed. The initial reservoir pressure is about 3200 p.s.i. and with the oil being undersaturated there is no original gas cap.

Oil-in-place and recoverable oil calculations are based on the whole accumulation including that part which falls in Shell/Esso Block 22/6. The calculations yield an average oil-in-place figure of about 1400 bbl/acre ft which in turn leads to a figure of about 4.4×10^9 bbl stock tank oil initially in place. A recovery of 40% would yield a figure of about 1.8×10^9 bbl recoverable oil.

Production plans

Development drilling will take place from four fixed platforms which will also serve as production platforms. The design of each platform allows for 36 drilling slots, although only 27 wells can be completed from each on the required spacing. It is planned to drill wells with a horizontal displacement of up to 2195 metres (7200 ft), at a maximum of 55° from the vertical.

120 acres has been selected as the maximum spacing required to ensure full recovery of the reserves over a 20–25-year period for an initial well rate of 8000 bbl/day. Drilling is to take place over those parts of the field which are 100 ft above the oil-water contact, i.e. within an area of 15,500 acres. On this basis the field can be drilled by 106 wells deviated from the four platforms (Fig. 8).

The field will be developed as follows: the initial phase includes the emplacement of two drilling platforms with ancillary production equipment, a 32 in. submarine pipeline to Cruden Bay near Peterhead, a 36 in. land line from Cruden Bay to a gas separation plant at Kerse of Kinneil adjacent to BP's Grangemouth refinery where about half the production will be refined, and a marine terminal on the Firth of Forth for export of the balance. Fifty-four wells will be drilled from the first two platforms and it is estimated that this will lead to a peak production rate of 250,000 bbl/day. The emplacement of

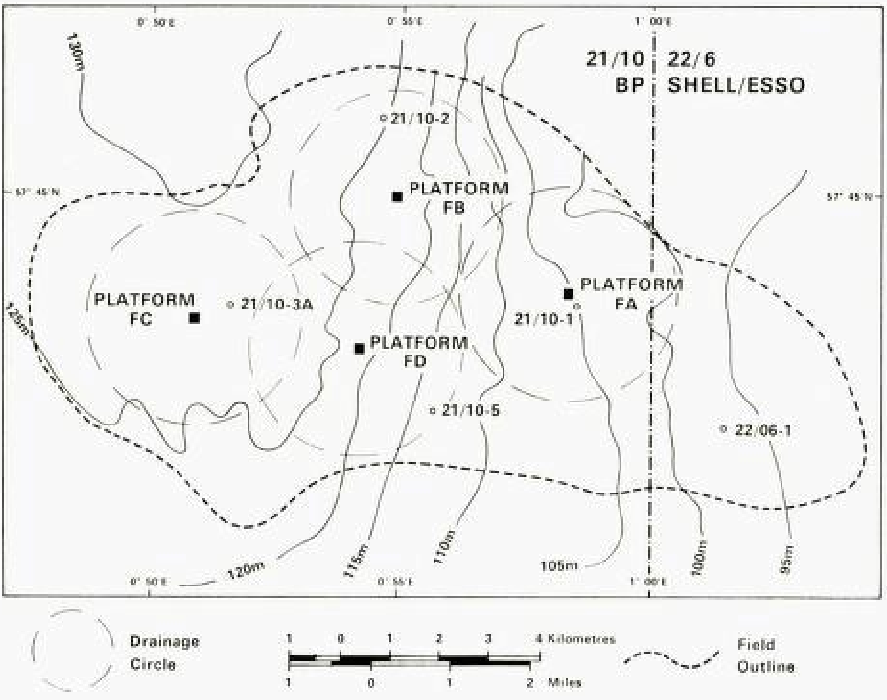


Fig. 8. Platform locations and bathymetry.

two additional platforms and the extension of the storage capacity at the Firth of Forth will follow. Fifty-two wells will be drilled from the third and fourth platforms increasing the production from the field to an average rate of 400,000 bbl/day.

It is hoped that this ambitious programme can be achieved by 1977 when the field should be on full production. Based on the well potentials assumed, this peak production of 400,000 bbl/day can be maintained for about three to four years before decline sets in.

Because of the under-saturated nature of the crude, pressure maintenance at about 2500 p.s.i. will be effected in order to maintain well efficiency. This will be commenced early in the life of the field in order to delay the need for artifical lift as water cuts increase.

Initially the oil and gas will be separated on the platforms at a pressure of 125 p.s.i.g. and about a third of the solution gas will be piped ashore dissolved in the oil. Some 20–25% of the remainder will be required as fuel on the platforms and thus, at peak throughputs, about 15–20 mmscf/day will remain for disposal from each of the first two platforms. During the drilling phase this must of necessity be flared, but subsequently it will be refrigerated for the recovery of natural gas liquids which will be shipped to shore mixed with the crude. The gas balance at this stage indicates that about one half of the total gas will be recovered, some 20% used as fuel and the remaining 30% flared. Of the gas recovered about 50% will be in the form of dry gas and the balance will comprise condensed liquids in the C_3–C_5 range.

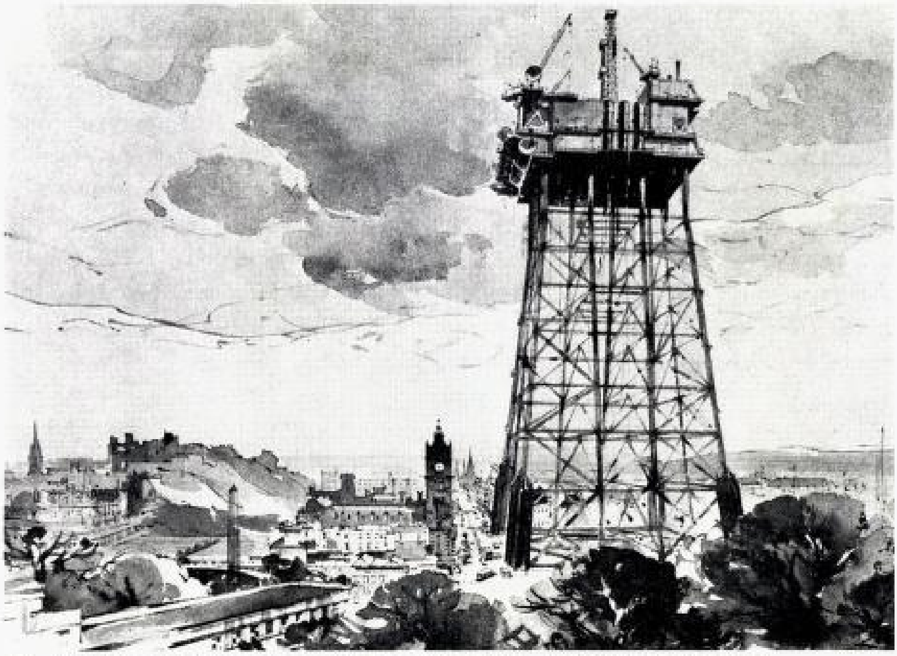


Fig. 9. Forties platform, as it would appear if standing in Edinburgh, Scotland.

The first two platforms are currently under construction, one at Nigg Bay on the Cromarty Firth, and the other at Middlesborough in Yorkshire. These mammoth structures will be larger and in greater water depths than any presently installed anywhere in the world. Conventional jacket templates will be floated to location, righted by controlled flooding and pinned to the sea bed with piles. The largest, which is to be installed in 128 metres (420 ft) of water will be 146 metres (480 ft) high, 61 × 76 metres (200 × 250 ft) in plan and will weigh 17,000 tons. Each template will be topped by a three level deck unit to accommodate drilling and production equipment. These will measure 43 × 52 metres (140 × 170 ft) and each weigh 15,000 tons (Fig. 9).

The North Sea is notorious for its severe weather conditions and design criteria have to cater for possible wind speeds of 114 knots and wave height of 29 metres (94 ft). The 32″ submarine pipeline to Cruden Bay will be about 175 km (110 miles) long and, apart from the shoreward 24 km (15 miles) will be laid in water depths of 91–130 metres (300–425 ft). The pipe will have a protective wrapping of fibreglass and coal tar enamel and an overall coating of 2½″ reinforced concrete. It will be laid from a conventional lay barge during 1973/1974 and subsequently buried by removing the sea bed from under the pipe with high pressure water jets. At Cruden Bay the oil will enter a sealine receiver trap and then pass directly into the land line. An emergency flow tank will be provided in case of shutdowns.

The 36 in. land line will be 203 km (127 miles) long from Cruden Bay to the gas separation plant at Kerse of Kinneil. Here the crude will be split, part going to BP Grangemouth Refinery and part going to the Firth of Forth terminal for export.

The foregoing account describes the discovery in 1970 by BP of the first major oil field in the British sector of the North Sea. The British regulations do not provide for the early release of well data and it is therefore not possible to make public technical inofrmation without prejudice to one's competitive position in the industry. In this respect the authors' wish to apologise for the lack of discussion of certain aspects of the geology of the Forties Field. Nevertheless we hope that this general review will prove to be a useful contribution to the understanding of this new and exciting petroleum province and stimulate others to release information for the benefit of all.

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Seismic Interpretation of the West Sole Gas Field

J. T. HORNABROOK

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The first seismic data across the West Sole gas field were shot during 1962, the field discovery well (48/6–1) was drilled in 1965 and production well 48/6–22 and 23 were drilled during 1973. This paper demonstrates the gradual evolution over those eleven years of our knowledge of the structural configuration of the field, from the earliest simple time map to the current picture obtained by multi-layer, three dimensional migration using sophisticated velocity functions. The formulation of these functions using all available seismic and well information for each of the major lithologic intervals is discussed, and the overall accuracy of the structural picture is analysed. West Sole is of particular geophysical curiosity because the final depth map on the Base Zechstein bears no direct resemblance for the time map, in fact the shortest reflection times are associated with the deepest part of the feature.

J. T. Hornabrook, British Petroleum Co. Ltd., Britannic House, London EC 2Y 9BU, England

Introduction

This paper tells the geophysical story of just one of the many hundreds of licence blocks in the UK sector of the North Sea. Block 48/6 contains part of just one of the hundreds of salt features which litter the Zechstein Salt Basin and lying largely beneath the south-western flank of this salt pillow in the West Sole Gas field, just one, and a modest one, of the several gas fields discovered so far in the North Sea.

Sole Pit is a sea bed depression approximately 50 miles east of the mouth of the River Humber. The area is always crowded with trawlers from Grimsby and Hull in search of the fish which give the feature its name. The West Sole Gas Field was so named, as it lies at a depth of approximately 3000 metres beneath the western edge of the Sole Pit.

The first seismic data were recorded in the West Sole area in the summer of 1962 using 50 lb. dynamite charges shot single cover into a straddle spread of 400 metres on each side of the shot. The analogue tapes thus recorded were plagued with heavy ringing and very strong multiples and the seismic sections produced were almost unreadable at all but the very shallow levels. Despite the general poor quality of the data, however, it was possible to demonstrate that the Zechstein evaporite series which was known to extend from Poland, through North Germany and Northern Holland, also extended across the North Sea practically to the English Coast.

The reflection from the top of the Zechstein Salt is easily recognised as being immediately above a series of very pronounced broad diffraction patterns. This is a characteristic and therefore diagnostic feature of the Permian Salt reflection over most of the Southern North Sea, and is probably caused by the diffraction of energy from the edges of fractured rifts of the brittle Z3 anhydrite, which is usually present near the top of the salt section.

The Basal Zechstein anhydrite was readable intermittently and this was mapped where possible in the hope that the Rotliegendes Sands, which lie beneath this unit at Groningen, also extended across the North Sea.

Although it was recognised in the early stages of exploration that there were various potential hydrocarbon bearing targets in the North Sea, the initial search in the south was of course for duplication of the conditions present at Groningen, where several thousand feet of Upper Permian (Zechstein) salt form the cap to the gas reservoir of high porosity Lower Permian (Rotliegendes) sand. The underlying, unconformable Carboniferous Coal Measures are probably the source of the gas. The presence of all these conditions cannot be determined by reflection seismology alone. In the reconnaissance Seismic Surveys of 1963 and 1964 the Base Zechstein anhydrite was the deepest reflector that could be reliable indentified, the existence of the Rotliegendes sand and Coal Measures could only be postulated.

The seismic problem

The principal seismic objective therefore became that of finding closed Base Zechstein Structures, and the seismic section in Plate 1 (a) illustrates the difficulty of achieving this object with the data available in 1962.

Subsequently the seismic method developed rapidly as illustrated in the historical summary in Fig. 1, and with the most recent seismic data in the area it is now possible to read the Basal Zechstein anhydrite reflection with considerable confidence and precision. The two lines illustrated in Plate 1 demonstrate the improvement in seismic quality which can be attributed to the improved recording and processing techniques summarised in Fig. 1. It is clear therefore, that the evolution of the structural interpretation of the field has been directly related to this improving seismic quality. As more seismic data were shot, not only did the quality improve, but the density of data increased, more and more well information became available and greater understanding of the velocity problems developed.

In some parts of the southern North Sea, reliable reflections could be obtained from several horizons. Studying the scanty data obtained from the velocity profiles that were shot in the reconnaissance stages of 1963 and 1964 it was soon clear that there were considerable variations in interval velocities with depth, and that many more data were required in order to determine velocity-depth functions that would be adequate to convert the reflection sections into depth profiles.

If all reflection horizons were conformable, erroneous velocity functions would have little effect on the structural picture. Isochron highs would still exist as highs when converted to depth; the actual depth to any point would be wrong but the general shape of the structure would still be valid, and the choice of a drilling location would not depend on the velocity function used.

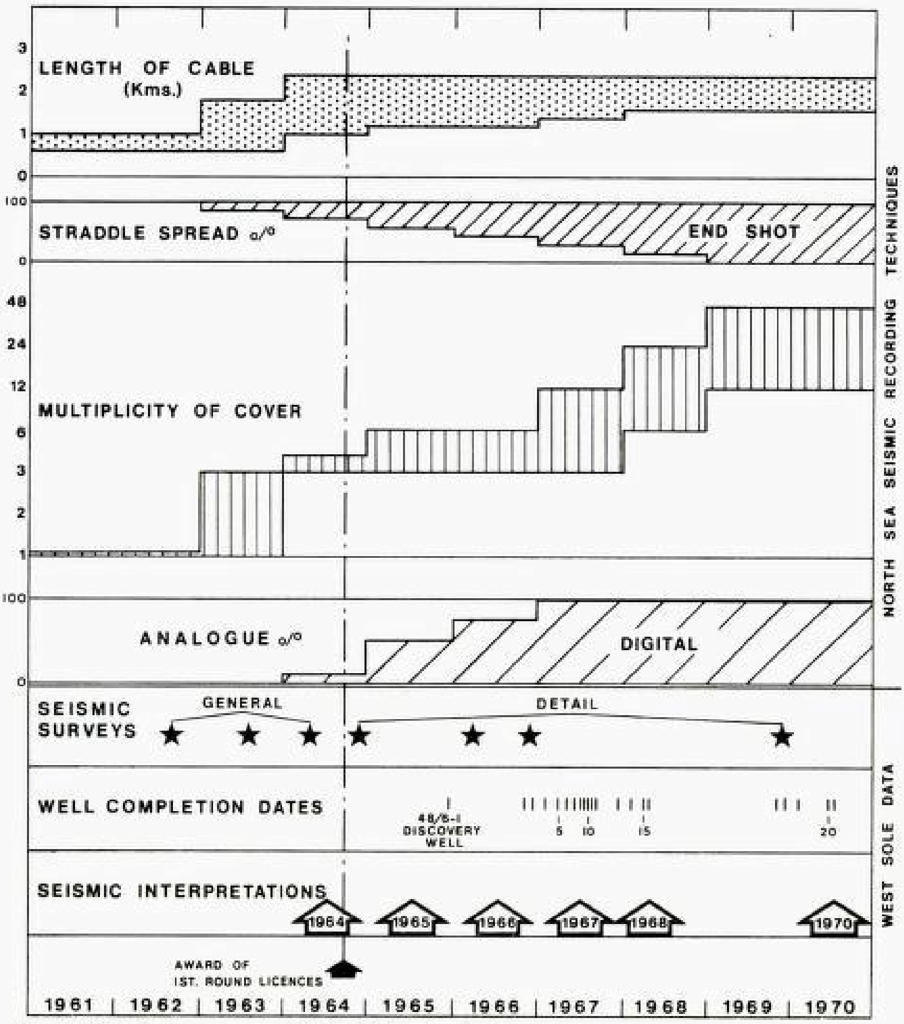


Fig. 1. Historical summary of West Sole seismic data.

In the southern North Sea, it is an understatement to say that the horizons are not always conformable! Due to the diapiric nature of the Permian salt and the subsequent rapid lateral changes in thickness, depth and interval velocity of the overlying Mesozoic and Tertiary, closed isochron features on the Base Zechstein cannot necessarily be assumed to exist as genuine closed structures. In fact there are many cases in which the use of inaccurate velocity functions can create apparent structure where none exist, and closed, attractive looking isochron features can often be attributed to velocity 'pull-up' due to overlying, high-velocity salt pillows or domes. It was very soon recognised that accurate interval velocity information is as important as the reflection sections themselves, and although we may be primarily interested in the Base Zechstein, it is usually essential to follow a number of shallower horizons merely to facilitate the conversion of the Basal Zechstein reflection times into depths.

Many wells have been drilled during the last few years, but these are generally on or near the top of Mesozoic highs and so yield very limited data on the variation of interval velocity with depth. The bulk of velocity data must therefore come from move-out studies or from development wells on Basal Zechstein highs that are overlain by offset salt wells. This entails penetrating the Mesozoic intervals at a wide range of depths whilst drilling crestal production wells in the Rotliegendes reservoir.

The West Sole field provides an ideal example of this, and in order to be able to see the effect of the growth of data on the evolving picture of the structure, we will consider the interpretation year by year.

INTERPRETATION 1964

The first isochron map on the Basal Zechstein in the West Sole area was completed in July 1964 and showed (Plate 2) a loosely defined high with a closure of at least 100 millisec. lying with a N.N.-Westerly axis across the 48/6 block. The block was applied for on the basis of this map, it being considered at the time that the problems of depth conversion were too complex for a simple method to give a valid conversion, and the data too poor to justify a more complicated approach.

INTERPRETATION 1965

After the allocation of concessions by the Ministry of Power in September 1964, four lines were shot within the block during October and November, 1964. All the 1962, 1963 and 1964 data were used in the construction of the isochron map (Plate 2) which, although using data which were largely fragmentary, were considered reasonably reliable.

Conversion to depth presented a problem as, at this time, there had been no wells drilled in the North Sea and most of the seismic data were too poor to yield reliable velocity data from move-out analysis on the Basal Zechstein reflection. The top and bottom of the salt, however, could be mapped with confidence, and various wells drilled through the Zechstein Salt in Germany and Holland had shown a reasonably constant salt velocity of approximately 4,400 m/sec.; the main problem, therefore, was the identification of the Mesozoic reflection (Tertiary being absent locally) and the selection of suitable velocity functions for the Mesozoic. Papers by S. M. Wyrobek (1959) and Muhlen & Tuchel (1953) yielded interval velocity data for the Mesozoic in both England and N.W. Germany, but without positive identification of the various Mesozoic intervals, and the knowledge that their lithological characteristics remained constant across the North Sea, this information was of no great help. It was clearly necessary therefore, to build up local velocity information on the interval between sea-bed and top Zechstein Salt. Fortunately, although the top Salt reflection is not generally of good quality it was possible to develop the velocity function from move-out studies over a limited range of reflection times. Corrections were applied to this velocity function for the effect of dip but, unfortunately, the feathering angle was not measured

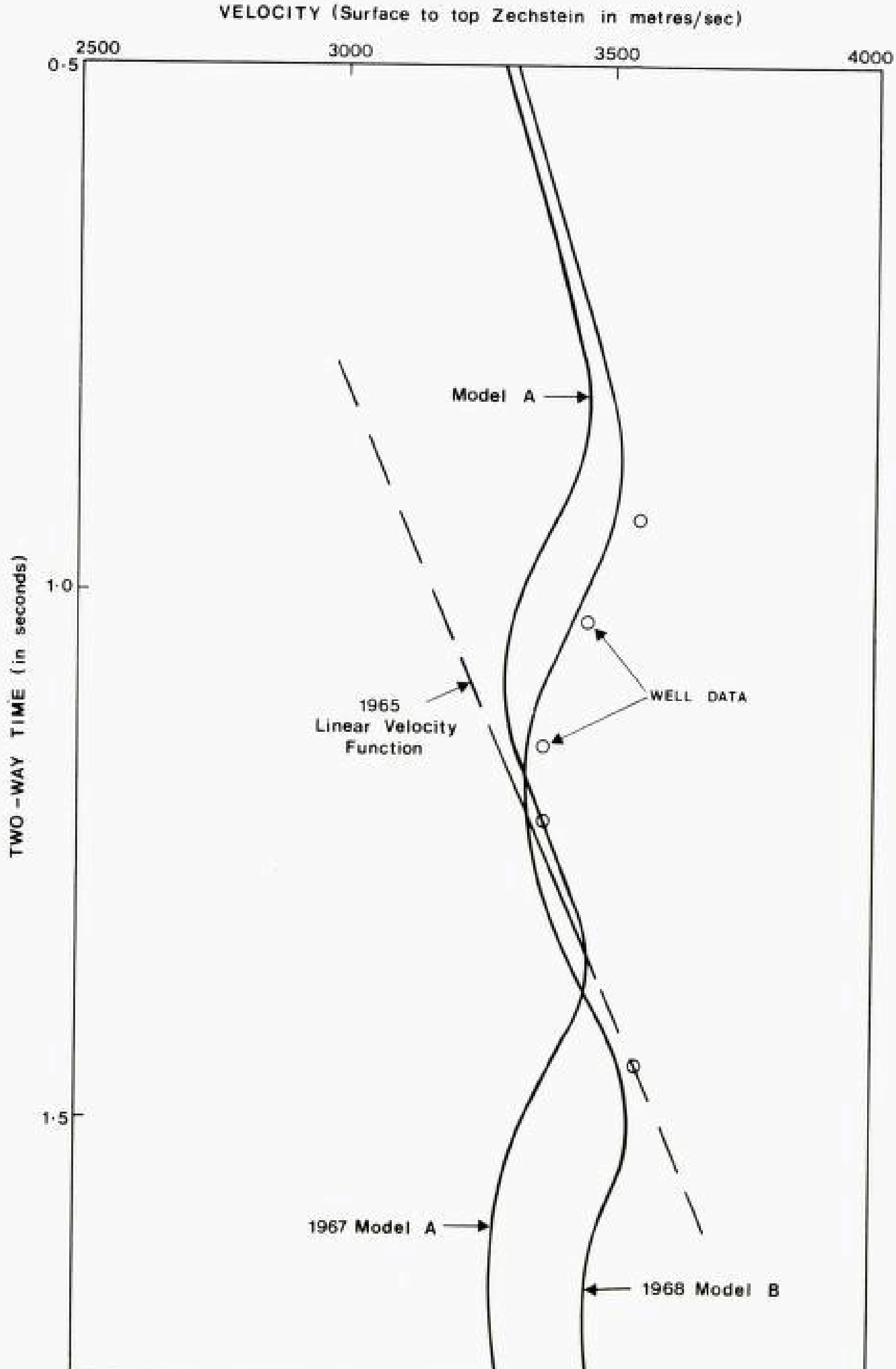


Fig. 2. Velocity functions, 1965–1968.

in these early surveys, so could not be accounted for. This analysis yielded a surface to Top Zechstein Salt average velocity curve, but only for a very limited range of reflection times (1.1 to 1.4 sec see Fig. 2) as reflection quality decayed beyond these limits.

It has been generally found in other areas that over quite large depth ranges, the interval velocity-depth curves approximate to straight lines. Since the surface to Top Permian average velocity curve was approximately linear over the observed time range, it was assumed, for the purposes of depth conversion, and in the absence of any contrary data, that this straight line could be extrapolated in both directions. This velocity function was then used to convert reflection times to the Top Salt depth and a constant salt velocity (4400 m/sec.) was again used for the incremental interval to Basal Zechstein. It was decided to test the Base Zechstein structure at the highest point indicated on this interpretation and the well 48/6–1 was spudded in by the drilling platform Sea Gem.

INTERPRETATION 1966

Following the discovery of commercial quantities of gas in the 48/6–1 well (December 1965), a detail seismic survey was shot over the structure in the summer of 1966. The general shape of the time high was similar to that of 1965 but the faults which could not be correlated in 1965 because of insufficient coverage and poor quality data have now been mapped.

Since the Mesozoic events could still not be followed with confidence, a velocity model was constructed using the 48/6–1 well data. This model (Model 'A' in Fig. 2) was developed by creating interval velocity versus depth functions for all the major stratigraphic units in the well using move-out data, UK and Continental European published data, adjusting them to 'best fit' the move-out data for the Top Zechstein Salt (Hornabrook 1967).

The model was extended up and down flank from the 48/6–1 well using what scanty seismic data were available to control the thickness variations in the Mesozoic intervals. It was then assumed that the Mesozoic intervals maintained a constant thickness in the Mesozoic strike direction. This enabled a velocity function for surface to Top Salt to be constructed and used even where the individual Mesozoic units could not be followed.

Again a constant salt velocity was used. The resulting depth map has clearly reduced the crestal area of the main structure.

INTERPRETATION 1967

Additional seismic detail shot towards the end of 1966 was incorporated into the 1967 interpretation. The same velocity function (Model A) was used as for the 1966 interpretation.

In the depth map we see for the first time a third lobe developing in the north west corner of the block; in the interpretations of 1964, '65 and '66 only two separate culminations appeared on the structure.

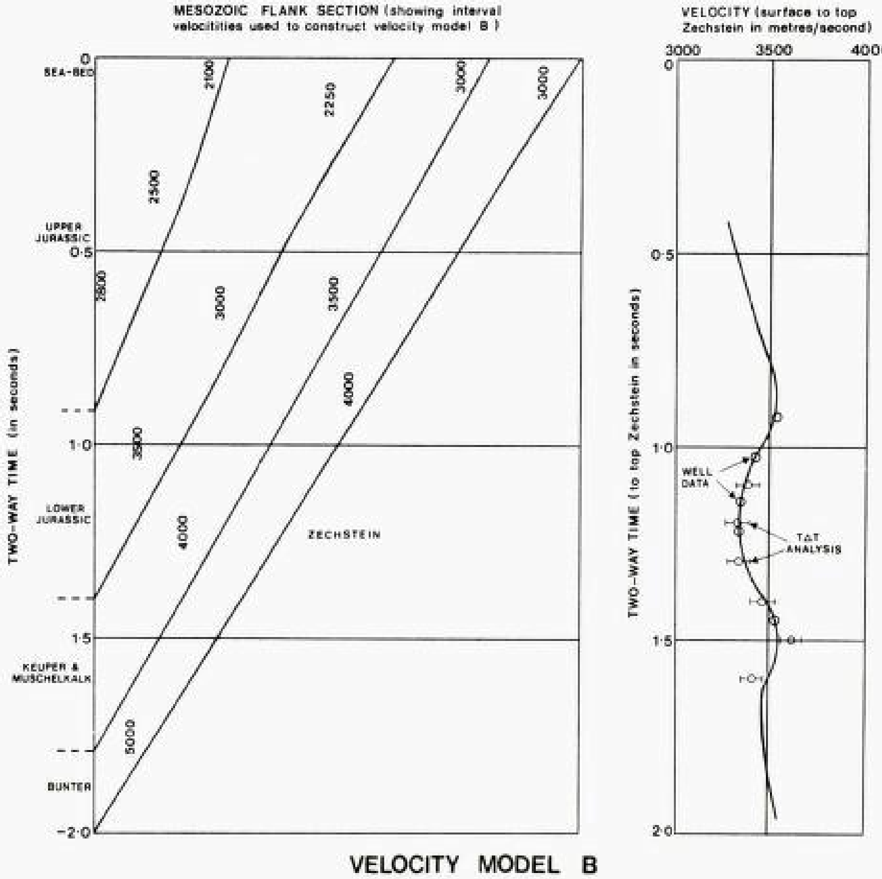


Fig. 3. Velocity model B.

INTERPRETATION 1968

This interpretation used the same time map as the 1967 version but used a more sophisticated velocity treatment, which we shall refer to as velocity model B (Figs. 2 and 3). This was constructed in a similar fashion to model A, but had access to stratigraphic data from the various wells (up to 48/6–7) on the West Sole Field and from several other wells in the vicinity. The individual interval velocity functions were modified outside the limits of well control to give a best fit to the shapes of an extensive move-out velocity analysis (from surface to Top Permian) by an iterative process.

Model B fitted the available well data to better than 2%, but the bulk of these wells lay in the southern portion of the field and it was important to estimate the validity of the model further north. By its method of construction the model shows the postulated velocity situation along a section perpendicular to the salt axis. There could possibly be minor variations in both the thicknesses and interval velocities of the various stratigraphic intervals in the direction of the salt axis, but a comparision of the move-out data from both ends

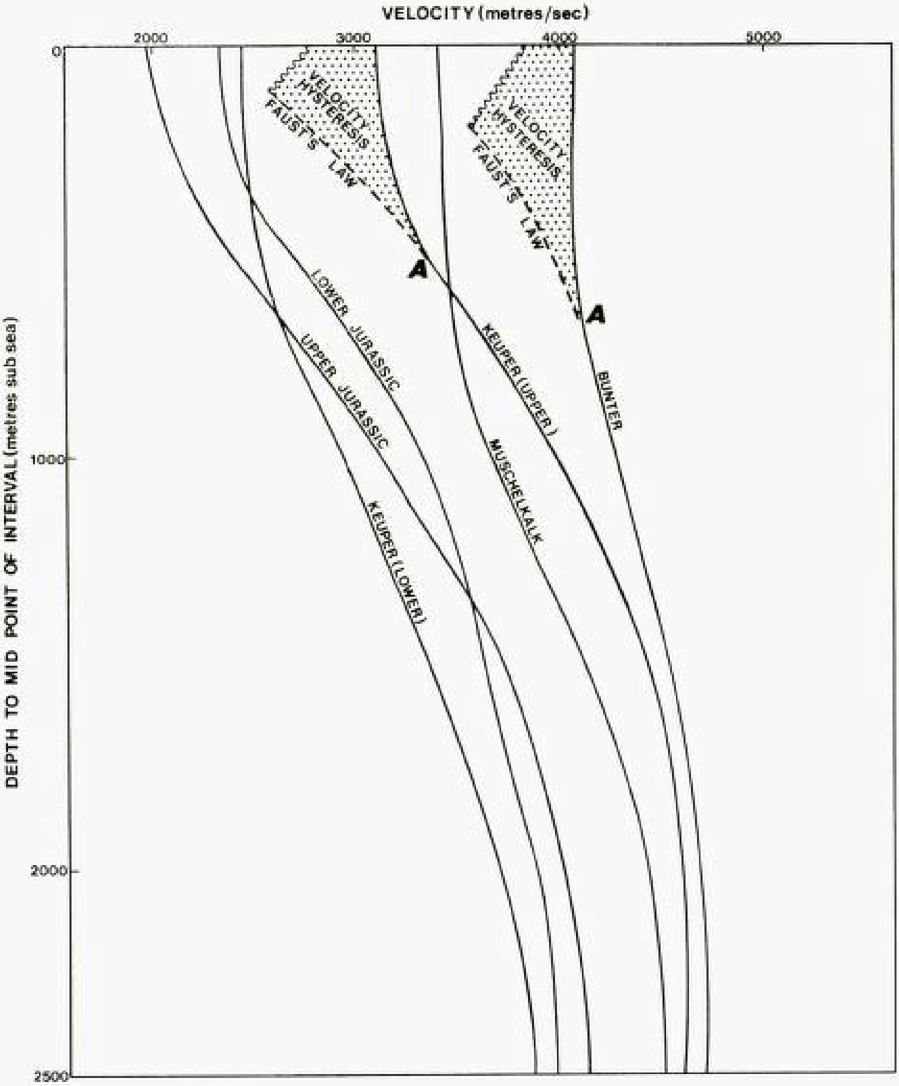


Fig. 4. Interval velocity functions, 1970.

of the salt axis showed that any variation in the model from south to north was less than the statistical error of the analysis (i.e. less than 3%).

From the first seven West Sole wells, and many others in the North Sea and onshore Holland, 4400 m/sec. appears the best assumption for salt velocity, but the Zechstein velocity problem lies in defining the thicknesses and velocities of the various non-salt layers within the Zechstein column. If these were always present, and of a constant velocity and thickness, their effects on Base Zechstein depths could easily be calculated. Unfortunately, when the salt starts to flow the anhydrite is broken into a number of blocks or rafts, and some of these are carried by the salt flow, and often deformed beyond seismic recognition. Generally, therefore, we cannot define the limits of these beds, and so a constant velocity must be used until sufficient Zechstein data are available for a more elaborate treatment.

Locally we find that the Z3 anhydrite is sometimes a very good reflection, and in the southwestern part of the 48/6 block it can be followed on all the lines. On most of the lines in this area, the limit of the Z3 can be seen (Plate 3), and the zone of unfractured Z3 is shown in Fig. 5. To the northeast of this zone the anhydrite has been broken up and carried by the salt. There are various fragments of strong reflection and diffraction which are probably derived from the Z3 anhydrite, but it cannot be plotted with any certainty.

Where the anhydrite exists in its unfractured state, its thickness is of the order of 120 metres, and velocity approximately 6000 m/sec.; when it is fractured and carried away by the salt flow, it is replaced by salt with a velocity of 4400 m/sec., and this has the effect of apparently lowering the Base Zechstein depth profile by 32 metres (just over 1%). Over most of the area we cannot say whether the Z3 is present in the salt column or not. There is therefore a possible error at any point on the Base Zechstein depth profile owing to this uncertainty of the order of 1%.

The depth map resulting from this more elaborate velocity treatment looks markedly different, especially in the central part of the field, than the 1967 interpretation of the same time data. A typical section across this part of the structures is shown in Fig. 7 and illustrates the difference between this version and that of 1967.

INTERPRETATION 1970

The survey of late 1969 was designed to look for possible extensions to the productive area of the West Sole field. At this stage 21 wells had been drilled of which 19 were production wells.

With the improved quality of the 1969 data it was possible to follow several Mesozoic horizons over the entire field area; it was therefore worth considering the construction of individual interval velocity functions for all the mappable units.

Interval velocity function

Interval functions were built up using the same general approach as for models A and B, but there were now 21 wells available over the feature, and although they were generally on the Crestal area in the Basal Zechstein they straddled quite a wide range of depths through the Mesozoic intervals. It was still necessary, however, to extrapolate the function beyond the limits of well control and the simple straight line approach gives in many cases ridiculously low velocities at shallow depths and improbably high velocities at large depths.

If the change of interval velocity with depth is due only to the increased compaction due to the greater weight of overburden then Gassman (1951) and Faust (1953) have both produced laws that allow a gradual increase in velocity with depth of burial. When we have subsequent uplift of a deeply

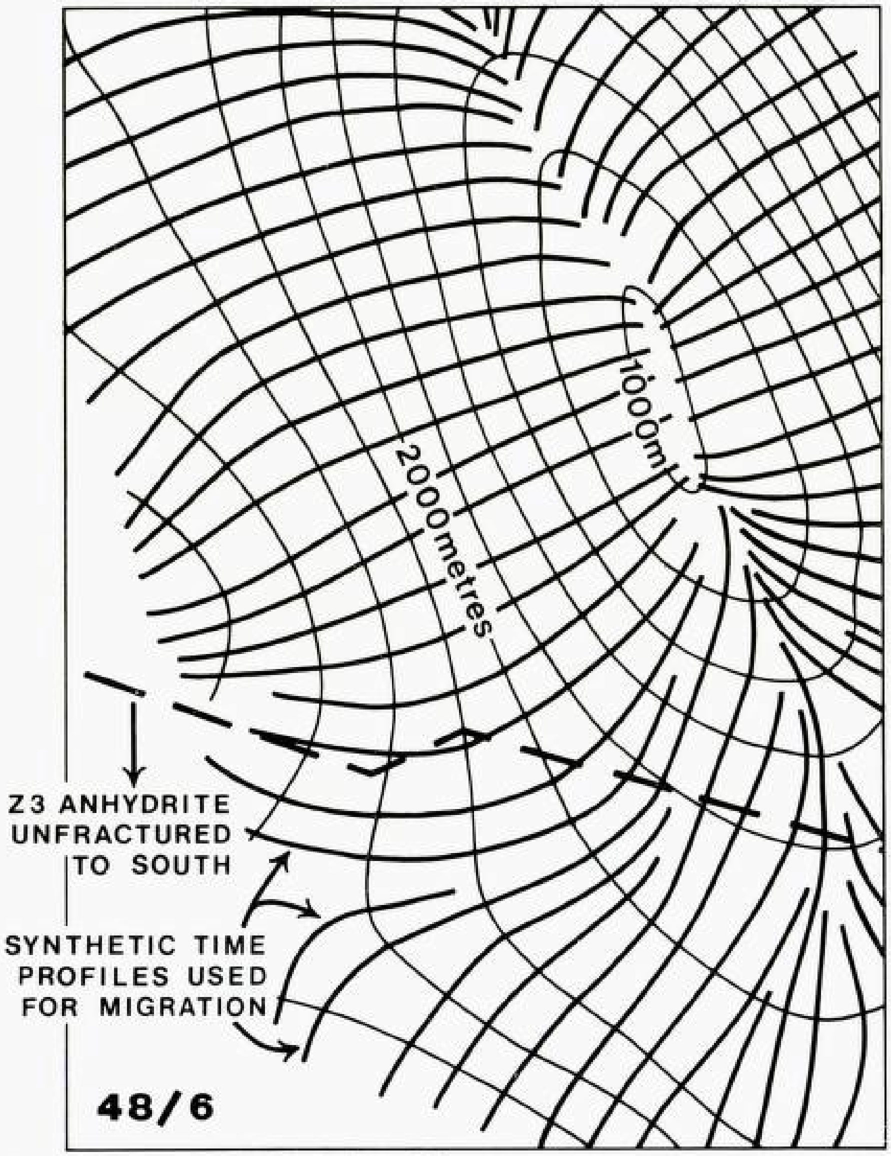


Fig. 5. Depth contours on Top Zechstein Salt.

buried interval, however, we would not generally expect that its interval velocity will decrease to fit these simple laws. This would assume perfect elasticity over very long periods of time, and ignores any effects of in situ cementation. So, we might expect the interval velocity for, say, the Bunter Sandstones or Upper Keuper over the top of a Salt swell, to be controlled more by the greatest depth of burial during its geological history than by its present depth. The Faust or Gassman curve probably still applies for intervals now lying at their historically greatest depth but must be modified as suggested in Fig. 4 for intervals subsequently uplifted. The choice of starting point for this modification is complicated by continuing depression and deposition regionally in addition to the localized salt uplift. The extent of the velocity relaxation, or hysteresis above point A (on the Upper Keuper curve of Fig. 4) depends on the lithology, extent of cementation and time scale of the movement. If well data fail to give control over the shape above A it is probably better to assume a constant or slightly decreasing velocity at shallow depth than the values given by either a straight line extrapolation or Faust or Gassman laws.

It is clear, therefore, that this velocity 'hysteresis' effect will be influenced by the entire burial history of the interval. Since the well velocity data we have available only cover the middle range of the curve, and we wish to study the Base Zechstein structure beneath the salt swell, where all the intervals are much shallower, an attempt had to be made to extrapolate the functions through this 'hysteresis' zone.

A study of the well data suggests the salt flow was initiated in the Upper Bunter, but was of small vertical extent at this stage (20–30 metres). As the entire area progressively depressed, and the weight of sediment increased, the velocity of Salt movement also increased, so that 100 metres or so of movement had taken place by the end of the Keuper. The bulk of the movement took place during the Jurassic (2000 metres).

The velocity function used in the depth conversion attempted to use this history in estimating the 'hysteresis' effect on the velocity curves. The functions actually used for depth conversion and migration are indicated on Fig. 4, each function being approximated to three straight lines for presentation to the computer. Although we believe these functions accurately define the velocities over the West Sole field, they clearly should not be used over other features in the area, as apart from regional variations in the individual lithologies, which affect the entire velocity curve, the 'hysteresis' effect will be different over each Salt uplift.

Depth conversion

Seven horizons were mapped and these are indicated on the seismic section of Plate 3; the main purpose of interpreting these horizons was of course to obtain a more accurate depth conversion to the Basal Zechstein. By using the individual interval velocities allowance could be made for outcropping beds and varying formation thickness.

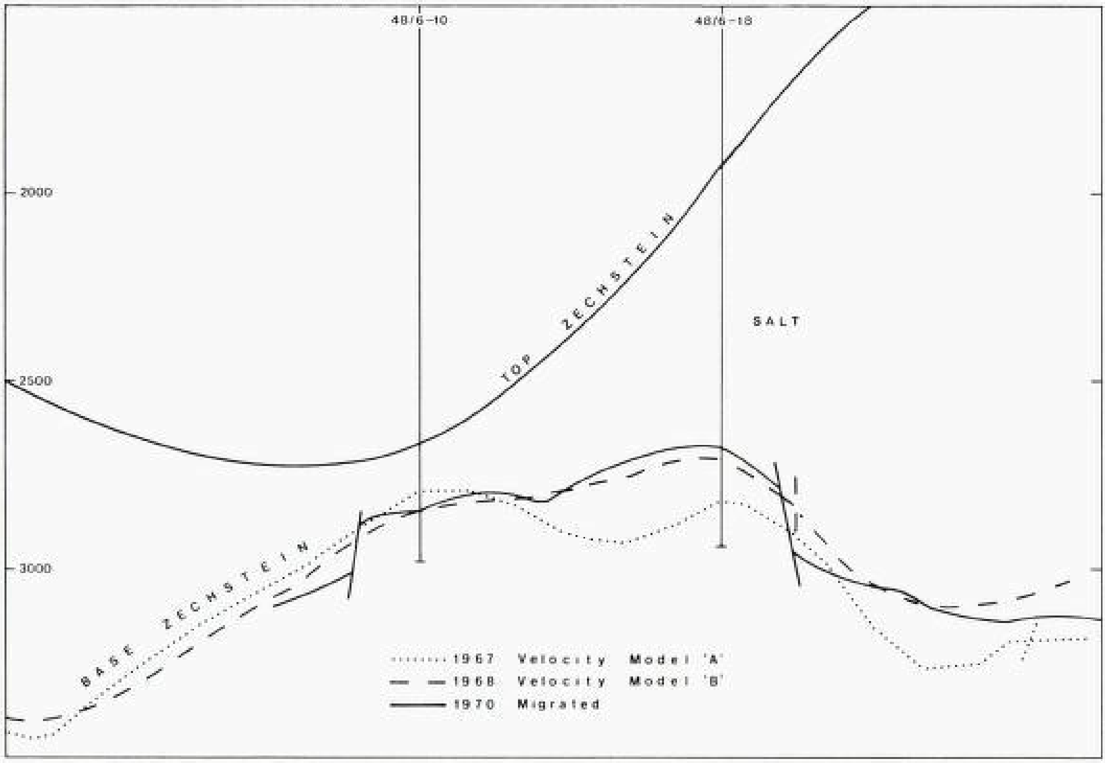
The time map produced shows the same general high area as previously, but is further complicated by additional faults.

In order to convert to depth and to account for migrated travel paths a full migration using all seven intervals mapped was performed. Isochron maps were drawn on these horizons and synthetic time profiles prepared. Because Zechstein salt flows are responsible for most of the Mesozoic structure, which exhibits dips up to 20%, compared with less than 4% for the Basal Zechstein level, these profiles were constructed along dip lines on the Top Zechstein

Fig. 6. Comparison of time and depth sections, 1970.

Salt horizon (Fig. 5). In this manner, an attempt to treat a three-dimensional migration has been made using only two dimensions. Care was taken to ensure that the horizontal curvatures of the synthetic time profile are not too great to introduce distortion.

The migration method, once these synthetic time profiles were created, was to take the first layer and to construct a fan of possible wave paths for each reflection point. The envelope to the series of wavefronts thus generated was taken as the depth profile. The same process was repeated for the next layer using the depth profile already obtained for the first. This was repeated



EFFECT OF VARIOUS VELOCITY MODELS ON DEPTH PROFILE OVER WEST SOLE FIELD

Fig. 7. Depth profiles based on various velocity functions.

for all seven layers. An example of time and migrated depth profile is shown in Fig. 6, and a comparison with earlier velocity models is shown in Fig. 7. Since the synthetic time profiles used for the migration were dip lines within the Mesozoic, this method produced an accurate three-dimensional migration down to the top Salt. For the Basal Zechstein reflection the component of dip perpendicular to the synthetic time profile was considered; and a comparatively crude attempt was made to allow for migration in this direction. Since these dips are much smaller than the Mesozoic dips, this 'second order' effect was small and caused only slight modification of the Basal Zechstein contours. After comparing the depths calculated using the initial velocity functions with the known well depths, some of the functions were slightly modified; also, a correction was applied to the Basal Zechstein depths in an attempt to account for the amount of Z3 anhydrite/dolomite present in the Zechstein. As an indication of the accuracy of the final map, it is found that the calculated depths are within 40 metres of the well depths for the 21 wells on the field. Residual differences, such as these, are bound to remain since no model can consider all the parameters; inevitably errors due to unaccounted small thickness variation and velocity inhomogeneities within the formation must remain.

One of the problems that we have looked at but have insufficient data to reach sensible conclusions on is that of anistropy. With dips in the Mesozoic overburden of up to 20° and with deviated holes both up and down dip of up to 40° from the vertical, it has not been possible to separate out the contribution of anistropy from the depths effect on the velocity variation.

Accuracy of final maps

The further we stray from well control, the more uncertain our velocity functions become, but we believe that within field area the velocity functions used are in error by no more than 2%.

There are several possible minor sources of error in the reflection travel time but generally we believe they are accurate to within 10 msec (0.7%). The plotted positions of the seismic shot points themselves are not beyond some doubt; various survey chains were used including Seashell, Seasearch I and Humber Hifix and the first two were used at night beyond the hours of reliability. Quite large positional errors (several hundred metres) undoubtedly occur on some of the early lines, but these lines, also of poor seismic quality, have not been used in the 1970 interpretation. The bulk of the data used in this last interpretation used Humber Hifix which were considered to give positional accuracy of better than 30 metres.

Even if we accept the plotted positions of the shotpoints, however, the depth estimate attached to it does not necessarily refer exactly to that point. With 2400 metre cables, feathering angles of up to 15° can easily occur, and for angles of this magnitude the reflection time picked on the section is the mean value along a line 300 metres long at right angles to the seismic line, not the actual value at the shotpoint. An attempt has been made to take account of this in the contouring, but in the earlier data the feathering angle was not recorded, so an unresolved uncertainty persists. Since the depth of an horizon is the product of its reflection time (one-way) and the overburden velocity, the accuracy of the depth map for the Basal Zechstein depends on the accuracy of both time and velocity, and, since the probable errors of the time data are considerably less than that of the velocity function, the shooting of additional lines, although improving the time map, will not greatly influence the accuracy of the depth map, for although there would be additional depth estimates, their accuracy will be dominated by the accuracy of the velocity function.

The accuracy of depth estimates at points which are not on seismic lines will depend on all the above factors, but the effect of contouring would also need to be considered. The amount of interpolation involved varies enormously over the area, and the general trend, local complexity, possible faults and distance from seismic control would all need to be considered; the question is too involved to discuss in general terms, and each case would have to be considered separately.

Summarising the factors which influence the accuracy of the Base Zechstein depth map, we have:

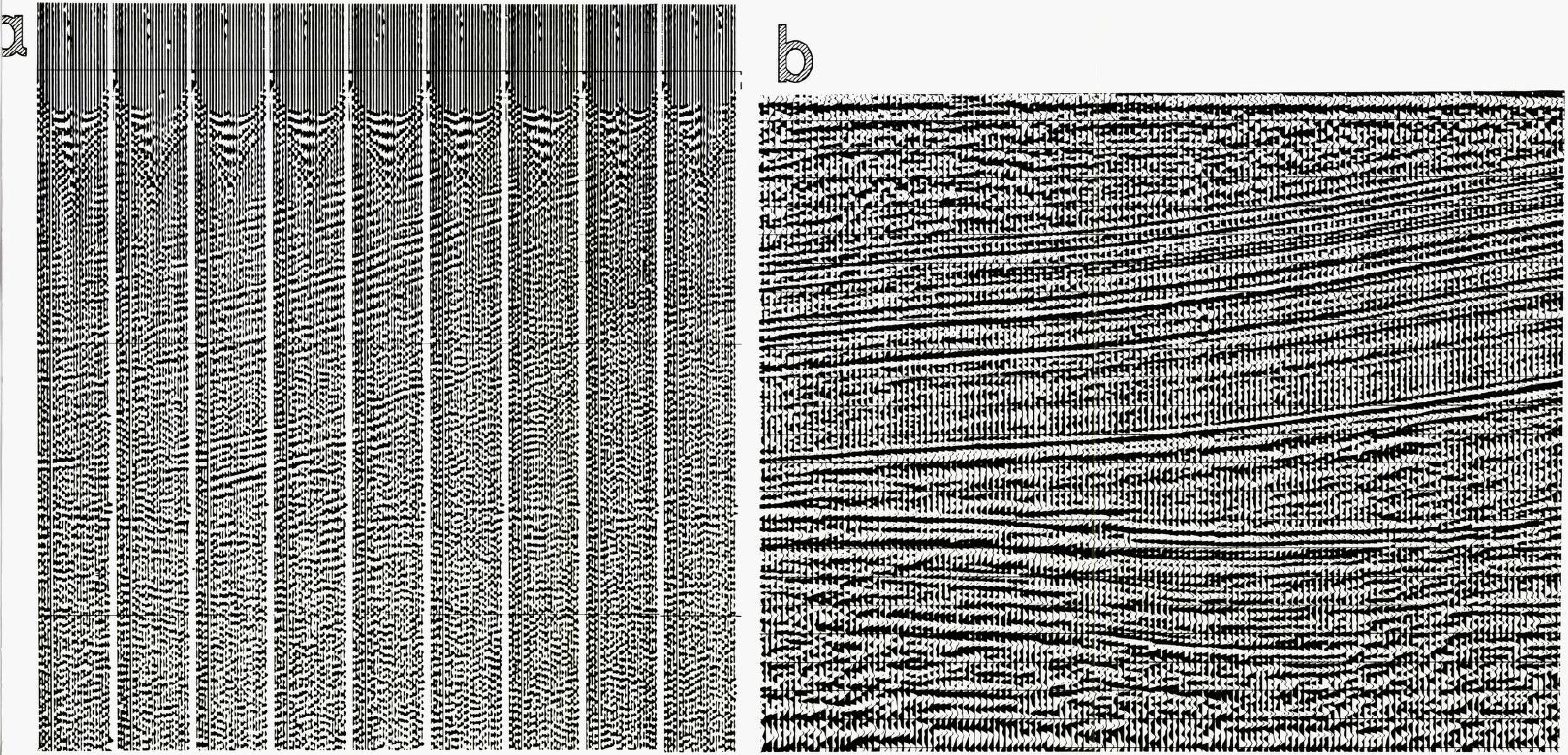


Plate 1. Seismic sections across the West Sole Field. (a) 1962; (b) 1970.

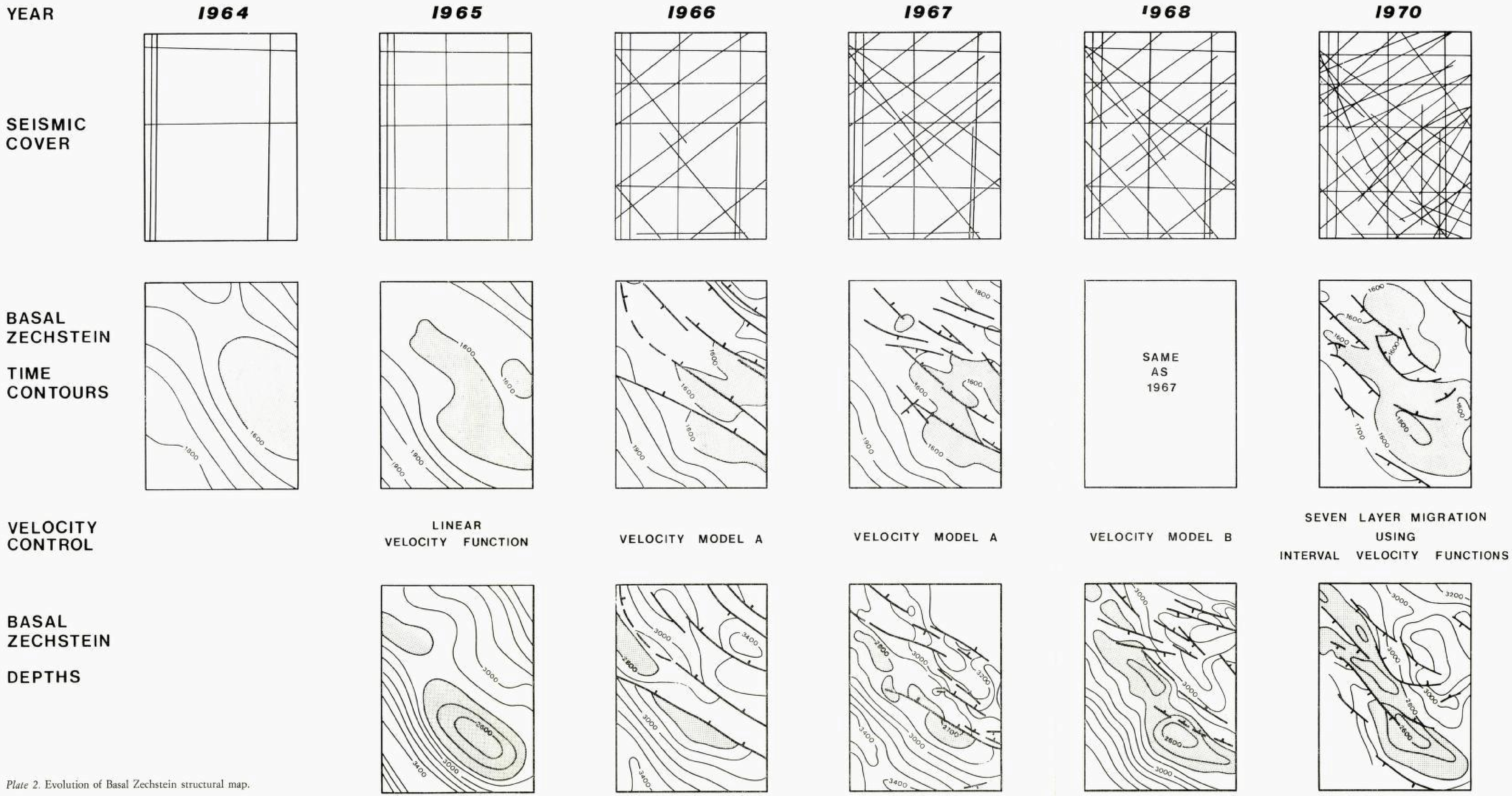


Plate 2. Evolution of Basal Zechstein structural map.

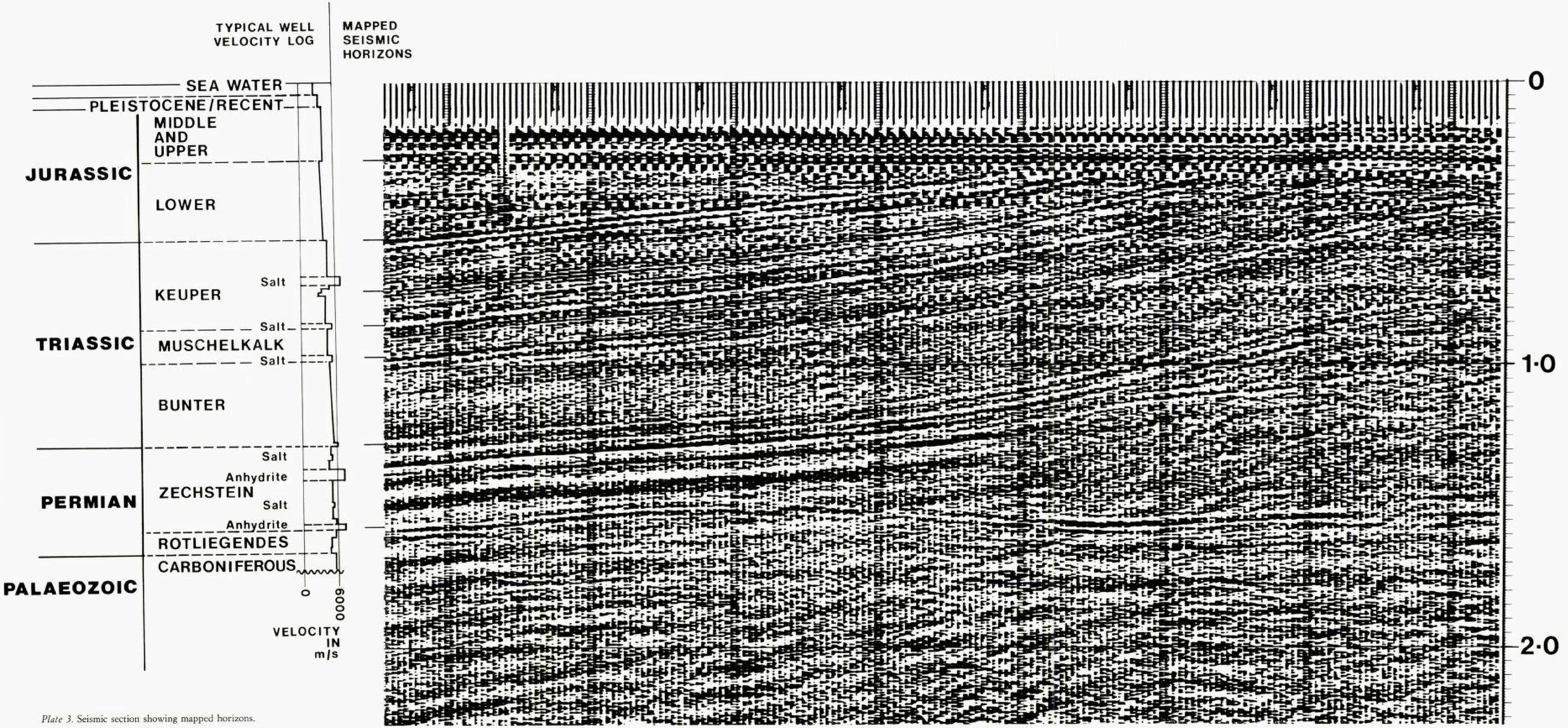


Plate 3. Seismic section showing mapped horizons.

	Probable error of Base Zechstein Depth
1) Reflection time to Top Salt	0.6%
2) Overburden velocity for Top Salt	2.0%
3) Interval time through Salt	0.7%
4) Interval velocity in Salt	1.0%
5) Migration and contouring effects	1.0%
6) Survey errors	0.5%

The probable error for the Base Zechstein depth estimate at any particular shot point on the West Sole structure is therefore less than 3%, i.e. less than 75 metres.

Conclusion

The series of maps show a progressively more complicated picture emerging as more and better data are added, until the final depth conversion (1970), which suddenly simplifies what has become a rather grotesque looking time feature. It is apparent, therefore, that much of the apparent structural complication on the time maps is due to the distortive effects of variable overburden velocities, and it is also clear that the only way to eliminate these effects is to develop local velocity functions which are related to the structural history of the feature.

There is little doubt that many of the early exploration holes in the Rotliegendes gas area were drilled on apparent, but not real structures, and that several genuine structures are so distorted by velocity effects that they have yet to be recognised.

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North Sea Troughs and Plate Tectonics

A. J. WHITEMAN, G. REES, D. NAYLOR & R. M. PEGRUM

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Interlinked roughly trilete-shaped predominantly sub-Upper Cretaceous trough systems extending north-south for more than 1200 km are described for the North Sea Basin and the Northeast Atlantic. Palimpsest tectonic controls are rejected as a major explanation of their development and an explanation in terms of lithospheric plate development is offered. The trilete trough patterns are seen as failed arms, superficial manifestations and consequences of plume or hot spot generated crestal uplifts initiated mainly in Late Carboniferous and Early Permian times over an area extending from Hatton and Rockall Banks in the west to the Skagerrak in the east. The Tertiary and Late Cretaceous broad basinal development of the North Sea Basin is seen as an inner continental margin development of the Bott and Watt type related to Cenozoic spreading of the North Atlantic arc development of the Cenozoic continental margin. The Mainz trilete system may be part of the overall pattern but data are inconclusive. Close relationships exist between trough and trap formation, geothermal history and the generation, maturation and accumulation of hydrocarbons. The role of mantle plumes – hot spot activity in the formation of the rift network is discussed.

A. J. Whiteman, Petroleum Exploration Studies, Department of Geology, University of Aberdeen, Aberdeen, Scotland

G. Rees, D. Naylor & R. M. Pegrum, Exploration Consultants Ltd., 31/33 Grosvenor Hill, London W1X 9HG, England

Introduction

If publicly available data on the Mesozoic structure, isopachs and facies distributions for the North Sea are plotted together with similar data from the adjoining land and sea areas, a highly distinctive and striking pattern emerges (Fig. 1). The pattern is necessarily generalised as only part of the information concerning bounding and internal structures, thickness variations, etc. is available in the public sector. Nevertheless, long linear troughs, some of which are over 50 km wide and over 300 km long, infilled by thick wedges of sediment of up to 10 km thick and ranging in age from Permian to Cretaceous, can be delineated. The Mesozoic troughs are separated by horsts and platforms upon which the sedimentary sequences are much thinner.

The trough and platform systems extend from Hatton Bank in the west to Poland in the east; and from the Pre-Alps in the south to the Lofoten Islands in the north (Figs. 1 and 2).

The trough margins are either faulted or monoclinal flexures, or combinations thereof. The interlinked trough system in the North Sea area roughly follows the median line between offshore U.K. and Norway and extends from the continental margin northeast of the Shetland–Orkney platform at approximately 63°N to the Broad Fourteens Trough in the Dutch offshore waters. This trough appears to continue into the West Netherlands–North Rhine

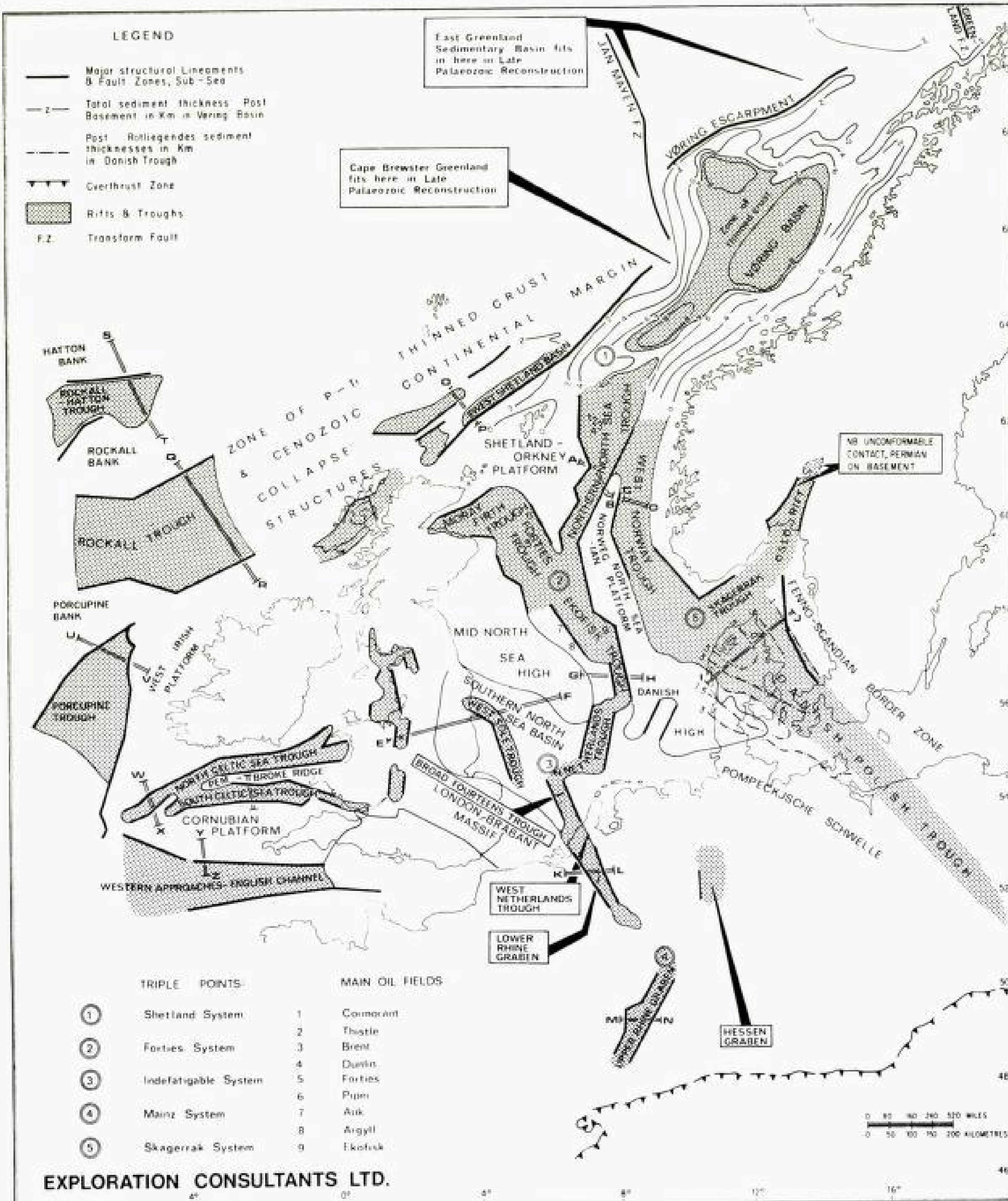


Fig. 1. General map showing troughs and platforms. North Sea and adjacent areas.

Trough which in its turn may be linked with the Upper Rhine and Hessen grabens. The latter apparently converge in the Mainz district.

A trough system extending from the continental margin between Shetland and Norway into the Netherlands is almost 1200 km long and constitutes a first order structural feature on the geological map of Europe (Fig. 1). Including the Upper Rhine and Hessen grabens as an integral part of the trough system, this structural entity is over 1800 km long.

The Mesozoic trough systems are extremely important economically because in the northern North Sea the large oil and gas fields appear to be genetically related to the formation and sedimentary history of these troughs.

Another trough system of similar size, exhibiting a marked asymmetrical trilete shape in plan, is present between southern Norway, Scania, Denmark, Germany and Poland. It consists of the West Norway, Oslo–Skagerrak and Danish–Polish troughs. The latter extends from the Skagerrak into East Germany and Poland and is more than 1000 km long (USSR International Tectonic Map of Europe 1 : 2,500,000 scale). In northern Denmark it contains more than 9 km of post-Palaeozoic sediments.

West of the British Isles there is a further system of sedimentary platforms and troughs approximately defined by the major elements of sea floor topography. These include the Hatton Bank, Rockall–Hatton Trough, Rockall Bank, Rockall Trough, West Irish Platform, Porcupine Bank, Porcupine Trough and the Southwest Irish Platform. Sea floor spreading and continental margin evolution (consequent on spreading) has clearly been involved in the formation of these and adjacent features. Further north McQuillin & Binns (1973) have shown that the structural history of the Sea of the Hebrides is consistent with the idea that the Sea of the Hebrides and the Inner and Outer Hebrides Troughs evolved as the Northeastern Atlantic opened.

In addition a linear fault-bounded basin occurs in the Celtic Sea south of Ireland and north of the Cornubian Platform (Fig. 1). A northern arm extends into the Cardigan Bay area and a southern arm into the Bristol Channel between the Pembroke Ridge and the Cornubian Platform. Both troughs are fault-bounded beneath the Cenozoic and contain thick Mesozoic sequences beneath a thinner Cenozoic and Upper Cretaceous cover (Pegrum & Rees 1973) (Fig. 2).

The Vøring and Stadt basins (Sellevoll & Sundvor 1973; Talwani & Eldholm 1972; Exploration Consultants Limited Report 1972) situated north of 62°N mainly on the Norwegian Continental Margin (although much broader apparently, if we are to judge from various generalized isopach patterns available, than the other troughs mentioned above), may also belong to this northwest European trough system (Fig. 1) and are associated with crustal thinning (Hinz 1972).

To date, the greater part of the commercial oil and gas found in N.W. European shelf has been associated with the fault-bounded troughs described above. A common origin, similarities of basin architecture and sedimentary development, etc., may well point to favourable prospecting conditions existing in other troughs and on the flanks of other platforms.

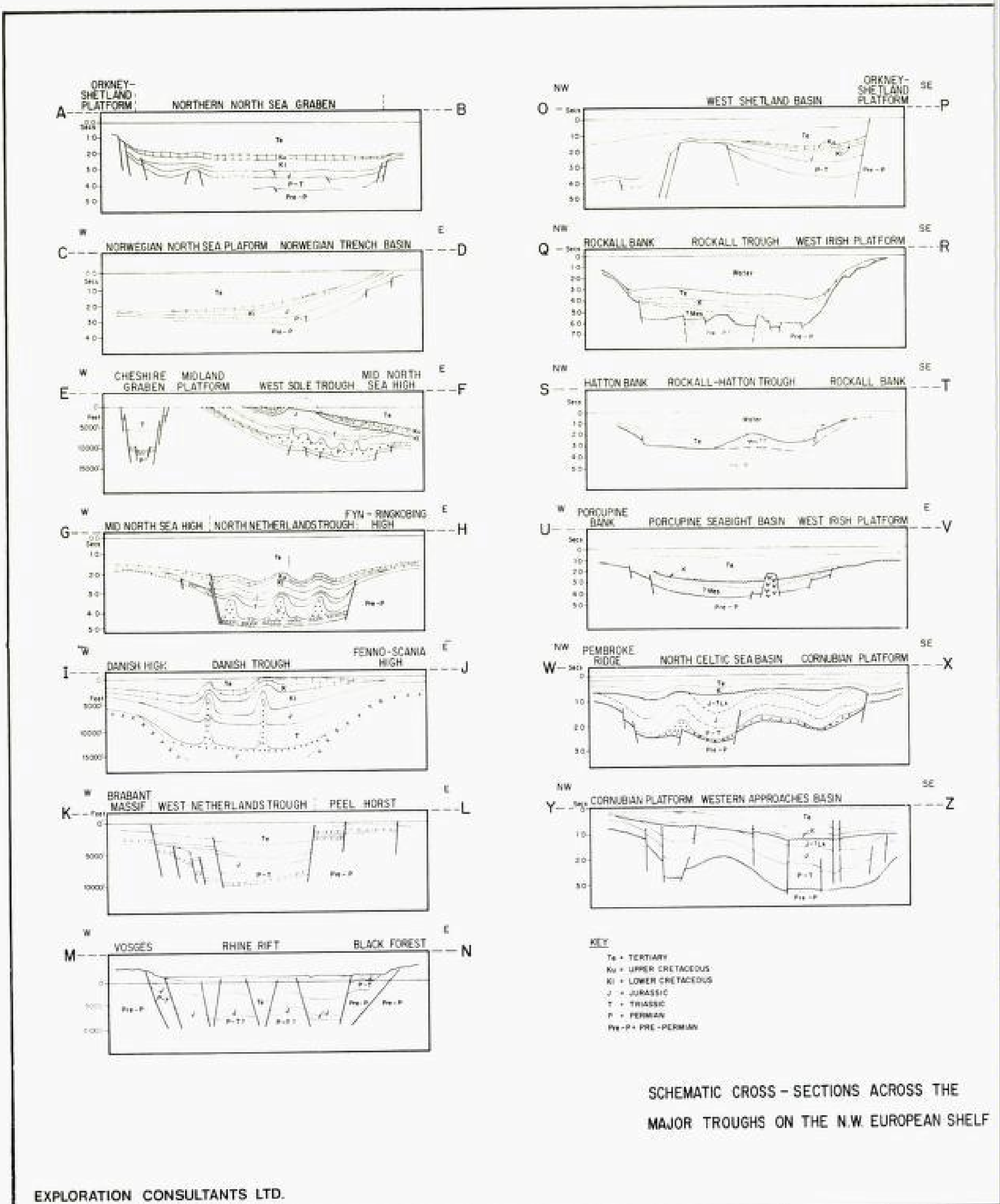


Fig. 2. Geological interpretations based on seismic profiles, North Sea and adjacent areas.

Origin of the Northwest European and Northeast Atlantic trough-platform systems

All these troughs are believed to have deep-seated controls and a common origin, 1) because of the shape, size and structure of the pre-Cenozoic North Sea Troughs and their close association with the Rhine and Oslo rifts, which are known to be underlain by thin crust. (Mueller et al. 1969; I. Ramberg 1972) and 2) because similarly shaped troughs (some complete and some now fragmented and modified) are closely associated with oceanic and thinned crust generated in Triassic time (Bott & Watts 1970 & 1971) or earlier in the troughs are developed over zones of new oceanic crust as in the case of the Hebrides, Orkney and Shetland.

Certain features can be recognized in trough-platform development: *either* the troughs are developed over zones of new oceanic crust as in the case of the Rockall–Hatton and the Rockall troughs (Fig. 2); *or* they evolved over zones of thinned crust (Figs. 2 & 3) as in the Oslo section of the Skagerrak system. They are 'failed arms' in the terminology developed by Burke & Whiteman (1973) and Burke & Dewey (1973).

The sections (Fig. 2) show that sedimentation must have proceeded contemporaneously with crustal development involving: either (i) thinning, spreading and then cessation of spreading, or (ii) crustal thinning, no spreading and failure. The sections also show that the sediment-filled troughs under discussion may be thought of as the near surface expression of deep-seated crustal structures which developed at different rates through Mesozoic time. Troughs then may be used to identify crustally thinned zones.

The majority of the troughs were formed as the Late Palaeozoic Laurasian super-continent began to break up and some of the trilete patterns of the North Sea and Skagerrak (Centres 1, 2 and 5, Fig. 1) may have originated in plume-generated crestal uplifts (Burke & Whiteman 1973) initiated in Late Carboniferous and Early Permian times. The Indefatigable and Mainz systems (3 and 4, Fig. 1) appear to have developed later. The Mainz system (4, Fig. 1) involving the Rhine and Hessen rifts developed much later in the Late Mesozoic and Cenozoic but nevertheless it may be considered as a trilete break-up feature affecting the western part of the Eurasian continent under which plume systems and plume-derived phenomena may have been active in some places as late as Recent (Burke et al. 1973). Troughs forming part of the Indefatigable system may have originated in the Triassic and have been superimposed on Permian basins whose architecture had been blocked out by Hercynian earth movements.

According to crestal uplift hypotheses (Burke & Whiteman 1973, Burke & Dewey 1973) the early stages of continental break-up (as in Mesozoic and Cenozoic Africa) are marked by the development of plume-generated crestal uplifts formed as large quantities of magma are generated near the lithosphere–asthenosphere boundary. During this process rift arms form as tensional structures and meet in triple junctions (rrr – rift, rift, rift) on the crestal uplift; the pattern probably being a least work configuration. Further evolution (without large-scale crustal melting; Burke & Whiteman 1973) involves the intrusion of axial dykes in the r-branches throughout the lithospheric break. The Gregory Rift has just reached this stage. Once dyke intrusion has started then continental rupture can begin as the system develops both longitudinally and laterally, connecting up adjacent plume centres and establishing new ones.

Spreading results only if the movements can be accommodated in the world-wide plate system. If this occurs then the rrr's pass into the RRR's or other combinations of spreading and transform systems as described by McKenzie & Morgan (1969).

A system may then proceed by gravity drive with the plates moving away from the topographically high ridges (Hales 1969, and others) or by the plates being 'pulled down' or subducted by the leading edge (Jacoby 1970).

In some rrr systems dykes are intruded but spreading only takes place on two arms e.g. (Benue Depression, Gulf of Guinea and South Atlantic in pre-Santonian times; Burke et al. 1971) or the system spreads on one arm, transforms on another and the third arm fails (Northern Red Sea, Gulf of Aqaba – Dead Sea and Gulf of Suez). The failed arm becomes a major depocentre infilled with clastics derived from the shoulders of the rift, and from within the rift depression itself, and with physicochemical sediments formed within the rift. Sediments may accumulate to thicknesses around 10 km and depositional conditions frequently approach optimum for the generation, maturation, migration and accumulation of hydrocarbons. Hence failed arms are of considerable interest to the oil industry (e.g. Benue Depression, Niger Delta, Gulf of Suez, North Sea, Barents Sea.) Examples of rrr development occur in many parts of the world and Burke & Dewey (1973) deal with the evolution of 45 selected triple junctions.

It is also conceivable that some (rrr) trilete systems do not reach the lithospheric dyke injection phase in all three arms and that therefore all three arms fail. This could be brought about by rapid plate movements immediately after formation, so carrying the upper crustal part of the structure 'off plume'. Failed rift arms generated in this manner would not remain static but develop into first order depocentres, as would single failed arms which evolve at different rates and in different ways, as the thermal energy derived from the plume system is dissipated.

The trilete structural and depositional pattern which is proposed herein for the North Sea and adjacent areas (Fig. 1 and Table 1) appears to be accountable in terms of the plate development, crestal uplift hypothesis, and the waning history of failed arms. Alternative hypotheses are 1) continental margin development (Bott & Dean 1972) and 2) palimpsest tectonics: in which old lineations find expression in successively younger sediments.

Table 1. Five trilete trough systems postulated for North Sea and adjacent areas

1. The Shetland System consisting of:
   1.1. Northern North Sea or Viking Trough.
   1.2. The Rockall Trough–Faeroes–Shetland Channel.
   1.3. Vøring and Stadt Basins.
2. The Forties System consisting of:
   2.1. The Moray Firth–Forties Troughs.
   2.2. The Ekofisk Trough.
   2.3. The Northern North Sea Graben.
3. The Indefatigable System consisting of:
   3.1. West Netherland and Broad Fourteens Trough.
   3.2. The English Sub-Basin and the West Sole Trough.
   3.3. The North Netherlands Trough.
4. The Mainz System consisting of:
   4.1. The Upper Rhine Graben.
   4.2. Hessen Graben.
   4.3. North Rhine Graben, West Netherlands Trough and Broad Fourteens Trough.
5. The Skagerrak System consisting of:
   5. The highly asymmetrical Oslo Rift and Skagerrak Basin.
   5.2. The Danish–Polish Trough.
   5.3. The West Norway Trough.

(See Fig. 1 for locations).

The Skagerrak trough system

Soon after the Asturian and Saalian climactic phases of the Hercynian orogeny had ended, the Laurasian super-continent began to break apart. It is proposed that a rrr crestal uplift (a potential RRR triple junction) developed under what is now the Skagerrak–Oslo rift trough, the Danish–Polish trough and the fault- and monocline-bounded West Norway Trough (Fig. 1).

In the Oslo Rift, crustal thinning (Ramberg 1972) took place in Late Carboniferous and Early Permian times when a variety of highly complex high-level igneous intrusives were emplaced and lavas were erupted (Oftedahl, *in* Holtedahl 1960).

Rapid easterly plate movement carried the crestal uplift rrr system off the plume energy source so initiating a waning pattern of sedimentary, igneous and structural development. In the Skagerrak area the waning phase lasted through Mesozoic and part of Cenozoic time with igneous activity apparently restricted mainly to the Permian in the Oslo rift. Some igneous activity may have been recorded in Rødby –1 area in Denmark (Sorgenfrei 1969).

Distension tectonics, consequent on thinning of the crust (clearly demonstrable in the Oslo area) resulted in faulting which operated during deposition of sediments and igneous intrusion. Faults cut the youngest Permian lavas and subsidence in the Oslo Rift may have continued well into the Mesozoic and perhaps into the Tertiary. Igneous and sedimentary rocks of these ages are not now known onshore in the rift, but they may well have been intruded and deposited and subsequently eroded since they are present in the Skagerrak section of the rift arm. The present-day Oslo Rift may be regarded as the partially stripped-out section of a southwesterly pitching irregularly floored rift depression.

The Danish Trough, outlined by Mesozoic isopachs (Sorgenfrei 1969; Voigt 1963; Schott 1969) and locally containing as much as 9 km of post-Rotliegendes sediments, is situated between the Danish High and the Basement Complex of the Fennoscandian Border Zone (Fig. 1). Its long linear form may be due to the trough being underlain by thin crust, and subsidence being controlled by crustal thinning tectonics. Because of the great thickness of sediment overlying the 'Basement' and because of the complexities of halokinesis, this trough does not show the same gravity pattern as the Oslo Rift and the Skagerrak Trough, where there are compounded effects of thinned crust, igneous intrusions of Permian and Tertiary age, troughs with thick sediments and platforms with thin sediments.

Marked subsidence took place in the Danish Trough in Permian times. The scale of Rotliegendes (early Permian) subsidence is not known but the Zechstein is thick enough for the salt to have been mobilized. Whether the trough is fault- or monocline-bounded, or both, has never been stated publicly except for northern Jutland (Sorgenfrei 1969). The linearity of the Zechstein boundaries plotted by Heybroek et al. (1967) point to strong, structural, probably fault but possibly monoclinal controls. The Danish trough continued to subside in Triassic and Jurassic times but by Early Cretaceous times subsidence had been greatly reduced and in Tertiary times was comparatively small.

The Polish section of the trough which extends in a southeasterly direction from Høllviken is more than 800 km long and was active from Permo–Triassic, through Jurassic into Cretaceous times. Again the amount of subsidence during the Cenozoic appears to have been much less than in the earlier Mesozoic (Voigt 1963).

Preliminary seismic evidence acquired by the oil industry indicates that thick Permian and Mesozoic sediments occur in the West Norway Trough, which is fault-bounded on the east and probably fault- and monoclinally-bounded on the west (Figs. 1 & 2). It is suggested that subsidence here was controlled by 'necking' associated with the development of the third arm of a Skagerrak rrr system. Like the Oslo Rift this could have been initiated in Late Carboniferous times.

Considering the West Norway, Danish and the Skagerrak–Oslo troughs together, and assuming that plume activity may have needed approximately 30 million years to run from inception to the emplacement of shallow igneous intrusions and surface vulcanicity (in the Cameroon Zone onshore plumes have been operating for more than 30 million years, while in the East African Gregory Rift vulcanicity started 30 million years ago in the Miocene), then plume activity and crustal thinning must have started in Carboniferous time. Surface volcanic activity appeared in the Oslo section in Early Permian times and apparently subsidence started in the Permian, continued through the Triassic, Jurassic and Early Cretaceous, tailing off in the Late Cretaceous and Tertiary in the Skagerrak section of the trough.

The median North Sea trough systems

As three possible systems may be involved in the North Sea area their developments are necessarily more difficult to decipher than the Skagerrak system. Also much wider issues are involved at the Atlantic and Alpine ends of the trough systems and the present data need integrating with the existing knowledge of these areas. Most of the data are derived from the offshore subsurface and only a limited amount are available in the public domain. This is turn prevents us from accurately drawing in our trough and platform boundaries.

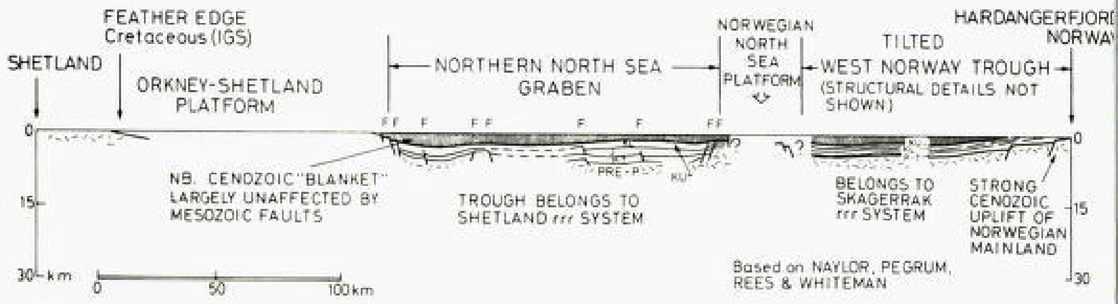
NORTHERN NORTH SEA GRABEN – VIKING GRABEN

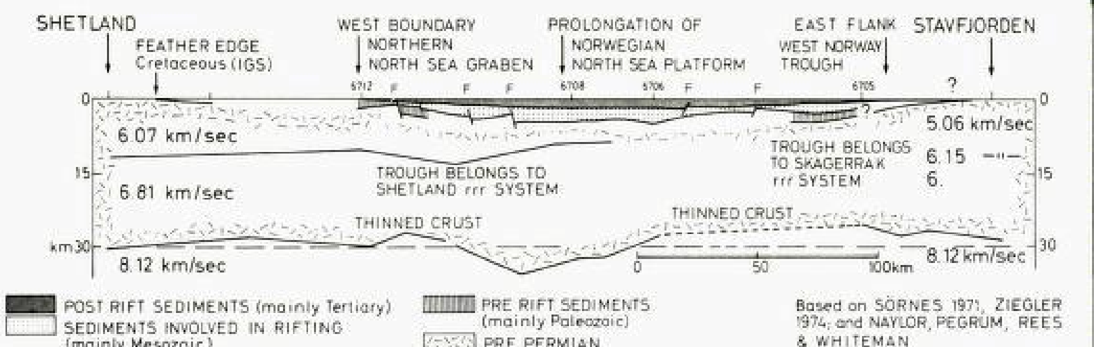
The northern limit of this trough has still to be defined. It appears to extend to the Faeroes–Shetland section of the continental slope lying south and west of the Jan Mayen Fracture Zone (Fig. 1). Whether the structure continues into the Cape Brewster–Scoresby Sound area in Greenland, which in a Late Palaeozoic–Early Mesozoic reassembly lies in juxtaposition to the northern North Sea is not clear. It is unlikely in our view. Much of the area between the continental margins of Norway and Shetland and Greenland, and between the Jan Mayen and Iceland–Faeroes fracture zones consists of 'new' crust.

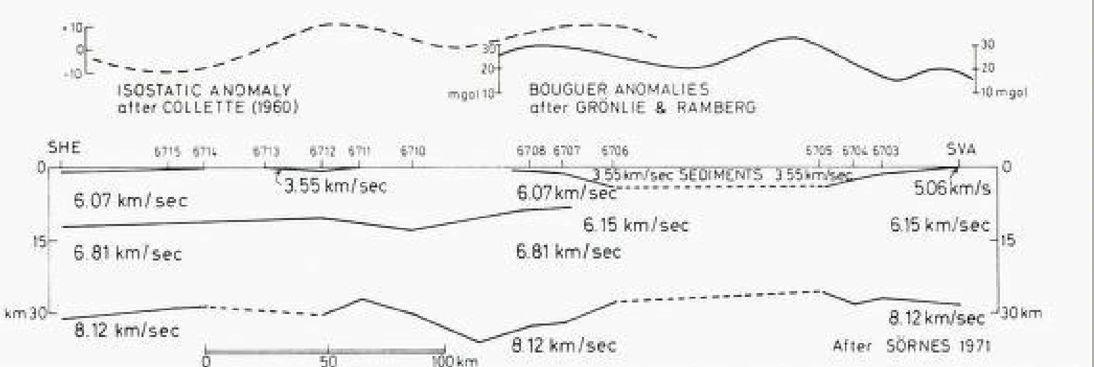
The spreading history of the Norwegian Sea (Johnson & Heezen 1967; Talwani & Eldholm 1972) clearly indicates that the Northern North Sea pre-Late Cretaceous Trough was formed before the 'dormant' or 'dead' spreading ridges which occupy the southeastern Norwegian Sea. These were actively spreading from 42 to 60 m.y.a. (Anomalies 18–24) and are not connected with the formation of the Northern North Sea Trough. Indeed at this time (Early Eocene) the North Sea Basin overall was steadily subsiding under a thick sequence of marine sediments while interconnected uplift was taking place around the margins of the basin in Britain, Norway and elsewhere. The marine transgression which started in the Late Cretaceous with the widespread deposition of the Chalk blanket continued into the Tertiary and the basin development clearly shown by isopachs (Heybroek et al. 1967; Pegrum 1970) is believed to be related to the rapid opening of the North Atlantic and regional adjustment of the newly developed continental margins of which the North Sea is a part.

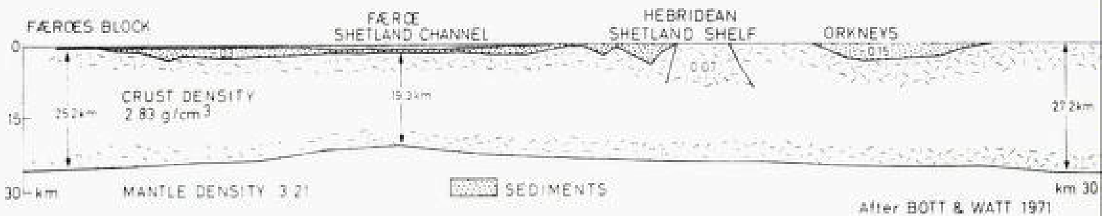
Other dimensions of the Northern North Sea trough have been described publicly only in very general terms (Armstrong 1972; Hinde 1972; Pegrum 1970; Dunn et al. 1973). The structure may have a width of 150 km in the north but narrows considerably in the south, where it merges with the Forties Troughs and the Ekofisk Trough.

A shallow seismic profile across the northern end of the trough extends from the Northern Shetland shelf to the Norwegian coast off the mouth of Hardanger Fjord (Talwani & Eldholm 1972). The Tertiary appears to be greater than 1.5 km thick on their maps. Other published isopachs (Dunn et al. 1973) indicate thicknesses of between 6000 and 8000 ft. (1.82–2.43 km)


FEATHER EDGE
Cretaceous (IGS)
SHETLAND
ORKNEY-SHETLAND PLATFORM
NORTHERN NORTH SEA GRABEN
NORWEGIAN NORTH SEA PLATFORM
TILTED
WEST NORWAY TROUGH
(STRUCTURAL DETAILS NOT SHOWN)
HARDANGERFJORD NORWAY
NB. CENOZOIC "BLANKET" LARGELY UNAFFECTED BY MESOZOIC FAULTS
PRE-P
KU
TROUGH BELONGS TO SHETLAND rrr SYSTEM
BELONGS TO SKAGERRAK rrr SYSTEM
STRONG CENOZOIC UPLIFT OF NORWEGIAN MAINLAND
0
15
30 km
0
50
100 km
Based on NAYLOR, PEGRUM, REES & WHITEMAN


SHETLAND
FEATHER EDGE
Cretaceous (IGS)
WEST BOUNDARY NORTHERN NORTH SEA GRABEN
PROLONGATION OF NORWEGIAN NORTH SEA PLATFORM
EAST FLANK WEST NORWAY TROUGH
STAVFJORDEN
6712
6708
6706
6705
6.07 km/sec
6.81 km/sec
8.12 km/sec
TROUGH BELONGS TO SHETLAND rrr SYSTEM
THINNED CRUST
TROUGH BELONGS TO SKAGERRAK rrr SYSTEM
5.06 km/sec
6.15
6.
8.12 km/sec
0
15
km 30
0
50
100km
POST RIFT SEDIMENTS (mainly Tertiary)
SEDIMENTS INVOLVED IN RIFTING (mainly Mesozoic)
PRE RIFT SEDIMENTS (mainly Paleozoic)
PRE PERMIAN
Based on SÖRNES 1971, ZIEGLER 1974; and NAYLOR, PEGRUM, REES & WHITEMAN


+10
0
-10
ISOSTATIC ANOMALY after COLLETTE (1960)
30
20
mgal 10
BOUGUER ANOMALIES after GRÖNLIE & RAMBERG
10 mgal
SHE
6715
6714
6713
6712
6711
6710
6708
6707
6706
6705
6704
6703
SVA
3.55 km/sec
3.55 km/sec SEDIMENTS
6.07 km/sec
6.15 km/sec
5.06 km/s
6.81 km/sec
8.12 km/sec
30 km
0
50
100 km
After SÖRNES 1971


FÆRŒS BLOCK
FÆRŒ SHETLAND CHANNEL
HEBRIDEAN SHETLAND SHELF
ORKNEYS
0.15
0.07
CRUST DENSITY 2.83 g/cm3
25.2 km
19.3 km
27.2 km
MANTLE DENSITY 3.21
SEDIMENTS
0
15
30 km
km 30
After BOTT & WATT 1971

for the Cenozoic. The broad regional relationships of the Northern North Sea Graben, the Norwegian North Sea Platform and the West Norway Trough and related crustal thinning are shown in Fig. 3, which is based on data from Fig. 2 (this account); Zeigler (in press) and Sørnes (1971).

The lowermost Tertiary is partly fault-bounded on the west against the Orkney–Shetland platform but the higher Tertiary rocks overlap and rest nonconformably on the crystalline and metamorphic Basement complex and the Devonian (O.R.S. facies). The eastern boundary of the Northern North Sea Trough is also faulted along the western margin of the Norwegian North Sea Platform (Figs. 1, 2 & 3). The 'graben' is some 500 km long, extending from the Faeroes–Shetland 'Escarpment' in the region of 67°N, to the trilete junction with Forties Embayment and the Ekofisk Trough (Fig. 1), which may be a failed rrr triple junction.

The floor of the graben is block-faulted, schematically shown in profile (Fig. 1) and the trough is thought to contain a thick sequence of Permo–Triassic, Jurassic and Cretaceous rocks which have been proven to be highly prospective for oil and gas.

Within the Northern North Sea Trough the structure is much more complex than we are able to show, and there may be several Mesozoic unconformities present, including the important Late Jurassic (Kimmerian) phase of epeirogenic uplift and unconformity. This phase of widespread uplift has been documented further south in the Southern North Sea (Heybroek et al. 1967; Kent & Walmsley 1970; Pegrum 1970). The total post-Palaeozoic sediments in this trough must exceed 6 km in thickness, of which the Cenozoic may account for 1.5 km and the Mesozoic 4.5 km.

Regarding the structural history of the Northern North Sea Trough the following observations can be made: 1) Bott & Watt (1971) believe that the Rockall Trough was formed by ocean floor spreading in Permo–Triassic times; 2) It has been suggested that a zone of thinned crust underlies the Shetland–Faeroes Channel (Fig. 1); 3) We postulate that this zone of crust thinned in Permo-Triassic time (now considerably modified by continental margin adjustments of the type described by Pegrum & Rees 1973; Bott & Watts 1971; Hall & Smythe 1973; McQuillin & Binns 1973) and may have extended north-eastwards to meet the Northern North Sea Trough. It seems that there was a Late Carboniferous–Permian–Triassic Northern North Sea Trough which formed part of a Shetland rrr triple junction (Fig. 1), of which the third arm lies under the Vøring Basin and the (?) deeper but smaller Stadt Basin. Such a trough system could have provided the Southern North Sea–German–Dutch–British Rotliegendes and Zechstein basins with the connection to the open Permian ocean to the north.

Again the lack of available data (Sellevoll & Sundvor 1973; ECL Report 1972; Talwani & Eldholm 1972) prevents us from saying very much about

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Fig. 3. Geological, seismic and gravity cross-sections, Northern North Sea and Faeroes–Orkney, showing crustal thinning and relationship to trough development.

the structural history of the Vøring and Stadt basins, but the shape of the basins, the presence of salt diapirs, and the fact that the basins have developed on continental crust may point to some unusual deep-seated structural control (Fig. 1). Hinz (1973 has recently postulated that the crust is thin under the Vøring Basin.

In conclusion, then, we think that there is a possibility that a trilete trough system may have existed between Norway and Scotland and Greenland in Late Carboniferous and Early Permian times and that the Northern North Sea graben began to develop as long ago as the Late Carboniferous.

THE FORTIES-MORAY FIRTH TROUGHS

The westward and landward end of the Moray Firth Trough was described by Dunham (1972). The northern and southern boundaries are, in part, fault-bounded, enclosing a trough containing Devonian, Permo–Triassic, Jurassic, Lower and Upper Cretaceous sediments (Fig. 1). Permo–Triassic and Jurassic rocks crop out on the margins of the Moray Firth Basin and are extensively faulted, whereas the Permo–Triassic at the south end of the Midland Valley is unaffected by faults which clearly pre-date these rocks.

A trough with locally thick Mesozoic sediments extends from the Ekofisk Trough into the Forties and Moray Firth Troughs and contains the Forties, Montrose and Piper fields. The Permo–Triassic rocks in 'red bed facies' and Jurassic and Lower Cretaceous rocks in marine and deltaic facies are over 1.5 km thick in this trough. In the Forties Field area, the trough is more than 105 km wide but it narrows considerably at its western end due to the south-easterly downthrow of the Great Glen Fault (Fig. 1). Howitt & Aston (in press) have described extensive Forties–Piper area basalts from trough intersection.

EKOFISK TROUGH

The Moray Firth Trough merges with the Ekofisk Trough (2, Fig. 1), which is the least documented of the North Sea troughs. Its eastern margin is possibly fault-bounded beneath the Cenozoic and constitutes the western boundary of the Norwegian North Sea Platform (Figs. 1 & 2), the narrow northward extension of the Fyn–Ringkøbing High or Danish High (Hinde 1972; Dunn et al. 1972).

The western boundary is also fault-bounded (en échelon) with the main fault zone lying to the east of the Shell Auk 30/16 Field. The Auk field has a Zechstein reservoir (Armstrong 1972) and may be located on the upthrow side of the fault zone. The Hamilton Argyll Field in Block 30/24 is situated along strike with a Lower Permian reservoir. This suggests that the western margin of the trough is locally a Mesozoic platform or ridge extending north-westwards from the Mid–North Sea High in a similar manner to the Norwegian North Sea Platform, which is an extension of the Danish High (Fig. 1).

Beneath the thick Cenozoic cover (up to 3.5 km according to Armstrong 1972) there are thick Triassic and Jurassic sediments often updomed by halokinetic Zechstein salt structures, The Cod, Ekofisk, Torfield and Josephine discoveries are located within the trough. At its northern end (Fig. 1) it is about 60 km wide between the Mid–North Sea High and the Danish High.

Again, the trilete trough pattern formed by the Moray Firth, Northern North Sea and Ekofisk troughs may have had its origin in a Late Carboniferous –Early Permian crestal uplift. We would, on the other hand, be dealing with the chance interlocking of faults, the trends of which were determined by structures generated along Charnian, Caledonian and Hercynian lines (palimpsest tectonics). Many more data are needed to resolve the question; although we strongly favour the trilete explanation, especially because of the existence of a large Jurassic (Late) volcanic basalt pile (Howitt & Aston, in press).

NORTH NETHERLANDS TROUGH

The Ekofisk Trough connects with the North Netherlands Trough, (Figs. 1 & 2) a fault-bounded structure situated beneath Cenozoic and Upper Cretaceous strata; bounded on the east by the Danish High (Fyn–Ringkøbing High) and on the west by the Northumbrian Arch–Mid–North Sea High. The trough links with the Broad Fourteens Trough along the western margin of the Texel or Mid–Netherlands ridge but as with the link with the West Sole Trough, relationships are difficult to decipher because of a complex fault pattern, strong halokinesis and lack of detail.

The North Netherlands Trough is infilled with sediments of Permian Triassic, Jurassic and Lower Cretaceous age. Upper Cretaceous and Cenozoic sediments extend well beyond the graben and bury the flanking platform highs. Clearly there was marked differential subsidence in the Mesozoic brought about by the so-called 'rim trough' mechanics (Voigt 1963). This was followed by broad regional subsidence in Late Cretaceous and Early Cenozoic times. The trough contains large sub-linear, north-south trending Zechstein salt walls and internal block faults. Halokinetic structures here clearly deform the Tertiary sediments (Fig. 2). Locally the northern part of the North Netherland Trough is called the West Danish Trough and is noted for its thick Jurassic sequence.

Several kilometres of pre-Late Cretaceous sediments must have accumulated in the North Netherlands Trough, which was probably initiated in Late Carboniferous and Early Permian times and formed part of the link (1 & 2, Fig. 1) between the northern Permian ocean and the English and German Permian Basins. Such a link has been sketched or is implied in maps produced by Wills (1951), Heybroek et al. (1967) and Pegrum (1970). This trough system remained a main connecting seaway with the open ocean throughout Triassic, Jurassic and early Cretaceous times, frequently marked by deeper water sediments.

WEST SOLE TROUGH

A broad east-west trending basin known generally as the Southern North Sea Basin is outlined by Mesozoic isopachs and lies between the Mid-North Sea High to the north, the Pennine High on the west and the London–Brabant

Massif or Platform to the southwest (Heybroek et al. 1967; Kent & Walmsley 1970; Pegrum 1970) (Fig. 1). In the deeper part of this basin there are more than 2–3 km of Mesozoic sediment.

The West Sole Trough lies in the eastern half of this basin and is situated en échelon with the Broad Fourteens Trough described below. These two troughs are separated by an area which was uplifted in Late Jurassic times (Late Kimmerian epeirogenic movements). The Viking and Indefatigable gas fields are located on this uplift. The West Sole Trough extends onshore into Yorkshire and Durham and across the Southern North Sea (Fig. 1). The trough is a 'half graben' flanked to the southwest by the East Midland and London–Brabant Platforms. Its north-eastern margin is ill-defined.

The interpretations of the structure and stratigraphy of the West Sole Trough are difficult because of the complications produced by strong halokinetic deformation of the Zechstein salt. The trough is infilled by Triassic, Jurassic and probably Lower Cretaceous sediments and its axial zone has been deeply eroded consequent on Late Cretaceous uplift (inversion). Data illustrating the regional structure of the trough have not yet been published but the Hewett, West Sole, Leman and the Viking–Indefatigable gas fields associated with it have been described (Kent & Walmsley 1970).

Again we believe that the trilete pattern (3, Fig. 1) is highly suggestive of deep structural control but many more data are required to prove the point.

From published data it is not clear to what degree the margins of the Permian English sub-Basin and German Basin are fault- and monoclinally controlled. The data which have been released indicate that the basin framework was strongly controlled by Hercynian structures. Certainly the limits of the Rotliegendes sandy facies and the porous Zechstein facies show a strong WNW–ESE Hercynian trend, and palimpsest tectonics probably provides the best explanation of basin architecture for the English and German Permian Basins. Available isopach evidence certainly does not show a trilete pattern in the Permian.

If our ideas are accepted then the Indefatigable system must be younger (Triassic–Jurassic–Early Cretaceous) than the Shetland and Forties systems where the stratigraphy points to a Late Carboniferous and Permian existence of the troughs.

BROAD FOURTEENS TROUGH–WEST NETHERLANDS TROUGH–LOWER RHINE GRABEN

The Broad Fourteens Trough is a fault-bounded linear structure flanked to the southwest by the London–Brabant Platform and to the northeast by the Mid-Netherlands or Texel Ridge (Fig. 1). Considerable subsidence took place in Late Jurassic and Early Cretaceous times with almost 2 km of sediments accumulating in the trough. Sediments of this age are either thin or absent on the flanks. The northwestern end of the trough is structurally complicated because the Zechstein salt is halokinetically deformed. Again, published data on this basin are lacking.

A schematic section across the West Netherlands Trough (Fig. 2) indicates that the majority of the faults associated with the structure are clearly of pre-Tertiary age but that others affect the Tertiary. The existence of Tertiary faults taken together with the development of the Rhine Delta complex indicates that the trough is still actively subsiding. Subsidence appears to have taken place more rapidly in parts of the Jurassic, Lower Cretaceous and Tertiary, all of which are considerably thicker in the trough than on the flanks (Haanstra 1963).

The West Netherlands Trough is continuous with the Lower or North Rhine graben. Together they are more than 350 km long and both have had similar sedimentary histories.

DEVELOPMENTS OF SOUTHERN NORTH SEA TROUGHS

From the sparse published information on the Southern North Sea a preliminary systematic account of trough development can be presented, but a detailed explanation of the origin of the pattern certainly cannot be made at this time.

Permian isopach and facies patterns point to important Hercynian control of the Rotliegendes and Zechstein basin frameworks. Rotliegendes isopachs (Dunn et al. 1973) show elongate eastward trends swinging to south of east in south Denmark and northwest Germany. The 1000 ft isopach shows a marked linearity, which may reflect some structural control.

A well-defined north-south trending trough bounded by the 1000 ft isopach extends into the Broad Fourteens area towards the West Netherlands graben, again perhaps indicating some degree of structural control.

The area enclosed by the 1000 ft isopach (Dunn et al. 1973) broadens eastwards and encloses the area of deposition of the Haselgebirge (halite) facies of North Germany. This facies exceeds 1 km in thickness.

Permian volcanics occur on the northern and southern sides of this trough, east of Groningen and beneath eastern Denmark, at Rødby, for instance, and are believed to be present on the margins of the Northern North Sea Trough, but hitherto the structural setting of these volcanics has not been discussed in relation to the Permian break-up of the Northwest European plate.

It is only in the northern part of the Rotliegendes and Zechstein depositional basins that there is any obvious indication of structural control of sedimentation (Heybroek et al. 1967; Sorgenfrei 1969; Pegrum & Rees 1973). Here the zero isopach in the Norwegian, Danish and German sectors clearly defines a trough centred on the North Netherlands Trough which was later filled with thick Triassic, Jurassic and Cretaceous rocks. The Rotliegendes thicknesses are less than 150 m and the trough opens northwards into the Northern North Sea.

Dunn et al. (1973) did not present Zechstein isopachs and the only reliable information available has been supplied by Heybroek et al. (1967), who gave general spot depths for these base of the Zechstein and the 3.5 km structure contour on the base of the Zechstein. The English and German basins are clearly defined by this contour.

The limits of the Zechstein shelf carbonate and anhydritic facies show some degree of structural control in the West Netherlands Trough, but the Zechstein facies boundary transgresses the limits the Broad Fourteens and West Sole troughs, which are mainly Kimmerian and Laramide features, i.e. Late Jurassic and Late Cretaceous.

There are strong linear Zechstein salt piercement structures and pillows in the North Netherlands Trough, which links with the Ekofisk Trough.

Two different sets of isopachs are available for the Triassic rocks of the Netherlands (Haanstra 1973; Dunn et al. 1973). Haanstra shows a thin Triassic in the Netherlands thickening northeastwards and ranging from 250 to 500 metres. Marked thickening also occurs along the German–Dutch frontier where strata of this age are from 500 to 1300 m thick. Dunn et al. (1973) do not show this embayment near the Dutch–German border, but show a 5000 ft thick sandy Trias in the West Netherlands–Broad Fourteens troughs. Similarly, the Sole Pit Trough is defined by a 4000 ft. isopach. The North Netherlands Trough (Central North Sea Trough of Dunn et al.) is shown as containing more than 8000 ft of Triassic sediments. The axis is generally coincident with the Zechstein trough.

By Jurassic time more than 1 km of sediments had accumulated in the West Netherlands–Broad Fourteens troughs, and the North Netherlands Trough is shown as a major depocentre with more than 4000 ft of sediments (Dunn et al. 1973). The Jurassic sediments are predominantly marine and mark the first *major* marine incursion into the area since Early Carboniferous times. Sedimentation was complex and was strongly influenced by structure. Thick sediments were deposited in the Northern North Sea, the Ekofisk and the North Netherlands troughs at this time. Markedly differential movements took place. Sole Pit Trough (Heybroek et al. 1967; Kent & Walmsley 1970) developed between the London–Brabant Massif and the Mid-North Sea High; and the West Netherlands and Broad Fourteens troughs between the London–Brabant Massif and the Mid-Netherlands Ridge. The North Netherlands Trough (Heybroek et al. 1967) developed between the Danish (Fyn–Ringkøbing) and Mid-North Sea highs (Fig. 1). The Groningen structure (Stauble & Milius 1971) and probably the Leman and Viking–Indefatigable structures (Kent & Walmsley 1971) were probably formed at this time.

Widespread erosion followed the strongly differential epeirogenic Late Jurassic movements, and over large areas the Jurassic was stripped off and erosion extended deep into the Triassic and locally as deep as Zechstein (Brunstrom & Walmsley 1969; Kent & Walmsley 1970; Pegrum 1970). Voigt (1963) has described the effects of these movements on land.

A very significant feature is the area extending from the eastern margin of the Mid-North Sea High into the Northern North Sea, in which the Upper Cretaceous directly overlies Zechstein carbonates. This area is believed to be an asymmetrical fault-controlled block, which Dunn et al. (1973) suggest was emergent for much of the Upper Mesozoic and/or suffered deep Kimmerian erosion. This ridge flanks the deep Ekofisk trough and could be regarded as one of the horst blocks bordering the Ekofisk graben, which has been buried only by Tertiary subsidence and sedimentation.

Some of the troughs continued to subside during the Early Cretaceous and all appear to have been affected by epeirogenic movements during the Late Cretaceous and Early Tertiary (Laramide) phase (Voigt 1973). As elsewhere in the North Sea the Tertiary was a period of overall subsidence (Pegrum 1970; Heybroek et al. 1973) with minor transgressive and regressive phases recognizable (Brouwer 1963).

Upper Rhine Graben and Hessen Graben

As mentioned above, the West Netherlands Trough continues into the Lower or North Rhine graben and meets with the Hessen and Upper Rhine grabens in the Mainz area (Cloos 1939). The Upper Rhine Rift, which breaks across the Odenwald, Schwarzwald and Vosges, is the best known of these structures (H. Cloos 1939; Illies 1962; Picard 1968, Mueller et al. 1969; Sittler 1969).

The Upper Rhine graben is more than 300 km long and extends from the folded Jura Mountains to the Mainz area. It is generally in excess of 40 km wide and trends north-northeast into the Mannheim region, where it becomes more or less north-south. The average throw on the eastern side of the northern section of the graben is 4 km and in the Heidelberg area it is perhaps as much as 5 km (Andres & Schad 1959). On the western border the throw is estimated at 2–3 km.

The initial movement of the principle Upper Rhine graben faults may have started in the early Mesozoic and probably resulted in the deposition of distinctive Jurassic facies in the trough (Bruderer & Louis 1958). However, others (e.g., Sittler 1969) maintain that the deposition of the Mesozoic sediments was not controlled in any way by the Rhine graben. Liassic isopachs do not show a correlation with graben structure, for example. Generally speaking total thickness decreases regionally southwards (Sittler 1969).

The Cretaceous is not present in the graben (Bruderer & Louis 1958; Sittler 1969) and Middle Eocene deposition was restricted to small lacustrine basins. The trough deepened in Late Eocene and Oligocene times. Early Oligocene (Lattorfian) subsidence was greatest in the south and 1,500–2,000 metres of saliferous marls were laid down in the Mulhouse potash Basin. About 500 m of Pechelbronn oil-bearing marls accumulated in the north.

In Rupelian times (Middle Oligocene) thick grey marls were laid down in an arm of the sea which extended from the North Sea, and in Late Oligocene and Miocene times the trough subsided mainly in the north. The sea gradually withdrew in this direction. Some 2,000 metres of marine, brackish to fresh water formations accumulated and are overlain by 300 metres of Quaternary alluvium.

Younger volcanism is concentrated mainly away from the rift in the Eifel, Laaken, Neuweld, Siebergebirge and Westerwald areas, as it is in some other rift areas. The Kaiserstuhl is the main centre of volcanism in the southern part of the Upper Rhine graben and was active some 20 million years ago.

Opinions vary concerning the deep structure and mode of formation of the Mainz system, especially concerning the Upper Rhine arm. A 'rift cushion' with a compressional velocity of 7.6–7.7 km/sec. is 'considered to be the driving mechanism of the rift process' (Mueller et. al. 1969). Exactly how this operated is not made clear, especially as the rift cushion has a thickness of 15 km and an east-west extension of 200 km and the rift is asymmetrical with respect to the anomalous zone. The fault-bounded zone is nearly everywhere only 50 km wide. Whatever the detailed geophysical interpretation, it appears that some crustal thinning and modification have taken place and the Upper Rhine graben has resulted from crustal thinning tectonics. The date of initiation of the structure is less clear from the geological evidence.

Also, there are fundamental differences between the deep structure of this graben system and the Northern North Sea graben resulting from age differences and patterns of development (Fig. 3). The Northern North Sea graben is an old (initiated in Late Carboniferous and Permian) failed arm, whereas the Rhine graben appears to be in the early uplift stage of the sequence proposed by Burke & Whiteman (1973), having been initiated in the Late Mesozoic & Early Cenozoic. There is, however, possible evidence of a Permian age for some of the bounding faults.

The gravity pattern is also not readily explainable. Unlike the Oslo Rift, it is underlain by negative anomalies which appear to be basin-located; the main gravity low being situated near the eastern side of the rift, west of the Odenwald, where the sedimentary rocks are thickest.

The northeastern arm of the Mainz triple trough is the Hessen graben, which is more than 200 km long. The Vogelsberg shield volcano separates the Frankfurt section of rift from the Kassel section. Immediately north-northeast of the Vogelsberg the fault belt is narrow and is only 30 km wide (*International Geological Map of Europe* 1:1.5,000,000 Ed 1972) but widens markedly towards Hamelin, and east of the Teutoburgerwald it is more than 80 km wide.

Considering the three arms of the Mainz structure it appears that separation of the French and German blocks has not taken place and that axial dykes have not been intruded. Crustal modification and thinning have taken place, however, and the long narrow isopach patterns and the faulted and monoclinal margins of the troughs point to crustal thinning controls.

Regional uplift associated with the Mainz structure has been considerable (Cloos 1939; *International Geological Map of Europe* 1972). This is evident from the dispositions of the erosional boundaries of various Mesozoic formations. The uplift and trilete rift pattern was tersely described by Cloos (1939) as the result of 'Hebung, Spaltung und Vulkanismus'. Burke & Dewey (1973) believed that it is plume generated.

West Britain grabenal system

A system of grabens is developed on the continental shelves around the western shores of the British Isles. Whereas the linear troughs of the Western Approaches and Celtic Sea bear striking resemblance to those of the North Sea, the troughs along the Atlantic margin are less obviously part of an interconnected grabenal system. The evolution of the Late Mesozoic–Cainozoic continental margin has clearly influenced this latter group of troughs.

WEST SHETLAND BASINS

Reconnaisance geophysics conducted by the Institute of Geological Sciences and by several universities in the late 1960's revealed a series of elongate SW–NW trending basins on the continental shelf west of Scotland and the Orkney–Shetland islands. These basins — the West Shetland Basins — are filled with low velocity sediments which contrast strongly with the adjacent basement rocks. Bott & Watts (1970, A & B) outlined four main sedimentary troughs separated by areas of shallow basement. A marked north-easterly trending feature — the West Shetland Platform — forms a gravity high and is interpreted as one of the basement highs. Magnetic anomalies and sparker profiles confirm that basement metamorphic rocks crop out on the sea floor in this area. To the west of this feature are several sedimentary basins characterized by gravity lows. Two of these basins, D & E of Bott & Watts (1970), have been confirmed by gravity, magnetic and sparker records. Basin E may be 5.2 to 7.7 km deep, assuming a density contrast of 0.4 to 0.5 g/cm^3 while Basin D may be 3.4 to 5.0 km deep using the same density contrast. Seismic profiles across Basin E (Fig. 2) show landward-dipping sediments unconformably overlain by seaward-dipping strata which thicken to at least 1500 ft on the continental slope. The sediments occupying the deeper part of Basin E and lying beneath the unconformity are likely to be Mesozoic in age. Also, it is worthwhile noting that the Faeroes–Shetland Channel is underlain by thinned crust (Fig. 3).

The faults outlined by Feir (1971) from a reconnaisance reflection seismic programme show a remarkable agreement with the margins of the basins outlined by Bott & Watts (1970 & 1971). The faults bounding Basins C, D and E are thought to have displacements in the order of 3000–5000 m. In general, the more prominent faults downthrow to the west but there are a number of lesser faults downthrowing to the east.

Corroboration of the presence of Mesozoic rocks in these basins is provided by samples dredged from the sea floor just north of the Minch and by the likelihood of the 13,000 ft Stornoway Formation being Permo–Triassic in age, rather than Torridonian (Steel 1971 and in press).

ROCKALL PLATEAU AND TROUGHS

The structure of Rockall Plateau has been discussed by Roberts (1971), Scrutton & Roberts (1971) and Scrutton (1972). There is a general agreement that the Rockall plateau is a fragment of continental crust, laterally displaced from N.W. Europe by the mechanism of sea-floor spreading. Reflection and refraction seismic work conducted by Scrutton & Roberts (1971) established velocities of 3.8 km–6.36 km/sec, compatible with that of continental crust. Only a thin veneer of sediment is indicated on Rockall and Hatton Banks but the reflection profiles reveal a thick sequence of layered sediments in the Rockall and Rockall–Hatton Troughs. Subdued magnetic lineations in the Trough indicate spreading which appears to have occurred during the Lower Cretaceous. The trough was certainly in existence in Mid to Late Cretaceous times and crustal thinning and spreading may have begun much earlier — in the Jurassic or Triassic.

At least 10,000 feet of sediments are present in the Rockall Trough (Scrutton 1972). Refraction profiles in the Hatton–Rockall Basin show 1700–2350 feet of relatively unconsolidated sediments (5600 ft/s to 7880 ft/s), which can be correlated with Middle–Upper Tertiary sediments of the Joides Borehole No. 116. The underlying 4000–5000 feet of strata with a velocity of 9450 ft/s corresponds to the Eocene/Paleocene of the Joides 116 and 117 sites (Laughton 1972) and an even deeper layer some 10,000 ft thick has a velocity of 14,500–15,000 ft/s. The latter interval, which could be sedimentary or volcanic lavas, rests on a high velocity 22,000 ft/s layer that can be correlated with the crust beneath Rockall Bank.

To summarize, both Hatton and Rockall Banks have a shallow acoustic basement of continental crust, and are covered by a relatively thin veneer of Upper Tertiary sediments. Rockall Bank has a number of fault-bounded sedimentary basins with up to 1 sec of Lower Tertiary strata. Rockall Trough has a much thicker sedimentary sequence and may have been initiated as a graben in the early Mesozoic infilled in part with Triassic. A later marine incursion in the Upper Late Mesozoic was related to the spreading of the trough. At the same time there may have been an abortive attempt at opening the Rockall–Hatton Basin, resulting in crustal sag. Although there is considerable crustal thinning there is no indication of oceanic crustal material (Roberts 1971).

PORCUPINE BASIN

Porcupine Banks and Sea-bight have attracted considerable attention in recent years. (Gray & Stacey (1970), on the basis of some 500 km of seaborne gravity and magnetic work, suggest that later displacement of the Porcupine Bank relative to the main continental mass of Ireland has resulted in crustal thinning.

Porcupine Bank is a large shallow water area on the continental margin of western Ireland about 250 km from the mainland. The Bank is separated from the Irish shelf by a north-south trending trough, the Porcupine Sea-bight. Porcupine Bank is characterized by NE–SW Caledonoid magnetic lineations and the recovery of metamorphic and ancient igneous rocks (Stride et al. 1969) indicate shallow Caledonian basement. The Slyne Ridge linking Porcupine Bank to the mainland probably has the same structure. Gray et al. (1971), from seismic reflection and refraction and magnetic data, conclude that the Sea-Bight basin contains about 5 km of sedimentary section. From the refraction data they define six velocity layers. Layer 5 (5.1 km/sec) may represent Older Paleozoic or metamorphic rocks and the underlying and deepest recognizable Layer 6 (7.7 km/sec) an intermediate layer of the earth's crust. Layer 4 (4.4 km/sec) probably represents Permo–Triassic to Lower Cretaceous strata and the overlying layers (Layers 1–3) successively younger rocks.

At the northern end of the Sea-Bight Trough the sedimentary fill is affected by folding and marginal faulting (Bailey et al. 1970). The faults propagate from acoustic basement, show an upward diminution in throw and fail to affect the younger reflectors. They show displacement down into the axis of the trough.

A sediment-filled trough, the Slyne Trough, forms a north-eastward shallower arm of the Sea-Bight Trough, cutting across the east-west trending Slyne Ridge. The relationship of the Slyne Trough to the shelf margin basins farther north is not known.

WORCESTER GRABEN, CHESHIRE AND IRISH SEA BASINS

A complex of connected rift valleys formed on the west side of the Pennine axis in England during Lower Permian times. This development has been documented by Wills (1956), who refers (p. 107) to a 'large-scale rift valley trending from the present Irish Sea and West Lancashire . . . south southeastwards to Worcester and Gloucester'. The inferred great thickness of Permo–Triassic strata has recently been confirmed by the drilling of the Prees–1 well in the Cheshire graben. The well spudded on an outlier of Lias and is the deepest boring, to date, in Britain. A considerable portion of the 12,500 feet penetrated was most likely in beds of Permo–Triassic age.

The Cheshire–Worcester graben system continues northwards into the Irish Sea and may be part of the Solway–Stranraer–Minches grabenal systems. To the south, Permian sediments occur in the graben only as far south as Worcester. During the Triassic, however, sedimentation extended further south, probably as the topographic expression of the grabenal structure became more marked.

WESTERN APPROACHES–CELTIC SEA BASINS

The Western Approaches and Celtic Sea margin has been the subject of several studies following the pioneering work of Bullard & Gaskell (1941), Hill & King (1954) and Hill & Vine (1965). Again there appear to be a number of linear basins or grabens with sedimentary fill of probably Upper Paleozoic (Day et al. 1956) to Lower Cretaceous age. Upper Cretaceous and Tertiary sediments blanket the older rocks with marked unconformity, transgressing both basin and platform areas. In the Western Approaches basin sediments of post-Carboniferous age may be up to 10,000 feet thick (Day 1959). This basin extends eastwards into the Channel and is confined to the north by the Cornubian Massif and to the south by the Brittany Massif. North of the Cornubian Massif a further deep linear basin — the Celtic Sea Basin — extends north-eastwards from near the shelf edge south of the Irish mainland.

It maintains a rough parallelism to the southern coast of Ireland, and bifurcates around the Pembroke Ridge. One arm passes northwards through St. George's Channel into Cardigan Bay, the other into the Bristol Channel between Wales and Cornubia.

It seems probable from the evidence of Blundell et al. (1971) and from the Mochras Borehole (Woodland 1971) that the Cardigan Bay Basin fill ranges in age from Permo–Trias to Quaternary, although Upper Cretaceous (Chalk) strata are notably absent.

It seems reasonable to conclude from this evidence and from a consideration of the geology of Portuguese and Newfoundland coastlines, given their proximity in the pre-drift stage, that Permo–Trias and Jurassic rocks fill the grabenal basins of the Celtic Sea.

Conclusions

We believe that most of the trilete structural trough pattern for the North Sea and adjacent areas (Fig. 1) was initiated in Late Carboniferous and Early Permian times. This could have been due to relative movement of the lithospheric plate with respect to the asthenosphere, so enabling plume-generated crestal uplifts to develop in the area extending from Rockall Bank to the Skagerrak, as happened in Miocene and later times for Africa (Burke & Wilson 1972; Burke & Whiteman 1973).

Rapid plate movement then carried the crestal uplifts off plume and so prevented the rrr systems from developing by lithospheric dyke injection into RRR spreading systems. On this hypothesis the Skagerrak and Forties trough systems, separated by approximately 300 km, may have developed over the same plume system due to the rapid plate movement (5 & 2, Fig. 1). Trilete trough system 1 may also have been formed over a plume system, which because of plate movement gave rise to volcanic activity in the Faeroes and which is now the Icelandic Plume.

Trough systems 1, 2 & 5 (Fig. 1) and the Rockall–Hatton and Rockall Troughs began to develop as major depocentres in Early Mesozoic time with individual arms (troughs) showing different rates and patterns of sedimentation and structural development (Figs. 2 and 3). In general, the North Sea troughs became infilled in Permo–Triassic times with predominantly continental arid sequences with evaporites, and with marine-paralic sequences in Jurassic and Lower Cretaceous times.

A large intracratonic salt basin (the architecture of which was strongly Hercynian dominated) was established in the southern North Sea in Permian times extending from England to Poland and it is not clear whether the Indefatigable trough system (3, Fig. 1) was initiated in Permian or Triassic times. A later Mesozoic history is decipherable from the published data for this trough.

During Jurassic and Early Cretaceous times several stages of active block faulting associated with the North Sea central graben system have been defined.

These movements may have occurred because of adjustment beneath the continental margin consequent on new crust being generated in the Atlantic. Marine incursions entered the area from the spreading Atlantic and from the Tethyan ocean. Most of the North Sea troughs had become inactive by Late Cretaceous time but some differential activity persisted into the Paleocene.

The Rhine and Hessen grabens developed in Late Mesozoic and Early Tertiary times and may be considered part of the fragmentation sequence. The West Netherland Trough appears to be currently actively subsiding (Rhine Delta Complex) and the Rhine graben may also still be active.

During most of the Cenozoic time much of the North Sea generally subsided. This was coupled with uplift of the adjacent land masses of Norway, Scotland, England, Wales, etc.

There is a close relationship between oil field distribution and the trough system, especially in the Northern North Sea area (Naylor et al. 1974).

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Geological History and Exploration, North Sea

D R. WHITBREAD

Whitbread, D. R. 1975: Geological history and exploration, North Sea. *Norges geol. Unders. 316*, 163–64.

D. R. Whitbread, Cluff Oil Ltd., 13 Park Place, St. James's Street, London SW1, England.

History of exploration

The history of exploration of the North Sea and the areas West of Britain was briefly described by reference to a series of slides showing the successive allocations of acreage in the U.K., Norway, Germany, Denmark and Ireland, since the first allocation in 1964 in Offshore Germany. This series of slides showed the initial exploration emphasis on the Southern North Sea following the discovery of the Groningen Gas Field in Northern Holland in 1959. The slides also illustrate the initial exploration emphasis on the extension of the Mid-North Sea High in British Waters which attracted much of the attention in the first round of licensing in U.K. Waters. This attention was eventually proved to be unjustified since the Rotliegendes Gas Fields were all discovered in the Southern North Sea south of the Mid-North Sea – Fyn–Ringkøbing Highs. The slides also illustrate the gradual extension northwards of exploration interest into the Northern North Sea and especially into the Viking Graben. The present U.K. licensing situation, including the various rounds of relinquishment, was shown illustrating clearly the present-day concentration on the five main basins of interest in U.K. Waters, that is (i) South North Sea, (ii) Northern North Sea (including the Viking Graben), (iii) West of Shetland Basins, (iv) Irish Sea Basins, and (v) Celtic Sea Basins.

Geological history

The geological history of the North Sea Basins was briefly described and the geological situation of the oil and gas fields discovered to date was briefly illustrated by reference to two slides showing respectively the gas fields of the Southern North Sea and the oil and gas fields of the Northern North Sea. A brief description of the age of the reservoirs together with an analysis of the regional distribution of the facies was given. Representative seismic sections of the Southern North Sea and of the Northern North Sea were briefly shown and discussed.

North Sea drilling activity

The build-up in activity over the years 1964–1972 was discussed and the number of wells drilled was analysed as between exploration, delineation and development wells. Clearly the discovery of the Ekofisk Complex in Norway had a profound effect on exploration activity in the North Sea in 1970 and 1971. The activity since that date has built up steadily to its present remarkable level. A comparison was drawn between the North Sea and the Gulf of Mexico illustrating the large area at present licensed in the North Sea (some 130,000 sq miles under lease as opposed to only 7,000 sq miles under lease in the Gulf of Mexico). On the other hand the reserves discovered to date are approximately two and a half times those discovered over a much longer time span in the Gulf of Mexico. Finally, exploration objectives over the next decade were briefly discussed and a map was shown of the area North of 62°N in Norway and of the West Coast of Britain and Ireland showing the deep water prospects in these areas.

Since the events over the last year have rendered most of these slides out of date they are not published with this brief abstract.

Some Geophysical Profiles in Areas of Interest in the North Sea

W. DOMZALSKI

Domzalski, W. 1975: Some geophysical profiles in areas of interest in the North Sea. *Norges geol. Unders. 316*, 165.

The North Sea and particularly the northern part of it is an area attracting at present one of the greatest exploration efforts currently undertaken. However, due to the confidential aspect of most data, little is available for free dissemination.

It is felt therefore that geophysical information which can be discussed, particularly in areas of discoveries, may be of considerable interest to all connected with North Sea activities.

This paper presents and discusses the interpretation of the following data:

(i) Aeromagnetic profile running east of the Shetlands and passing over areas close to the Cormorant and Thistle discoveries.
(ii) Aeromagnetic profile passing over the area of the Josephine Field.
(iii) Seismic and marine magnetic profiles and aeromagnetic data north-west of the Orkneys.
(iv) Aeromagnetic profile over Norwegian Sector east of the Brent, Dunlin and Thistle discoveries.

W. Domalski, 7 Green Lane, Watford WD1 4NL, England

PART II

Northeast Atlantic Continental Margin; Geology and Geophysics

Tectonic Evolution of the Northeast Atlantic Ocean; a Review

A. S. LAUGHTON

Laughton, A. S. 1975: Tectonic evolution of the northeast Atlantic Ocean; a review. *Norges geol. Unders. 316*, 169–193.

The hypotheses of sea floor spreading and plate tectonics enable the evolution of ocean basins to be understood provided that there are adequate data on the present and past plate boundaries from the magnetic anomaly patterns and some geological control of the nature and age of critical parts of the ocean floor. From such data, the following main phases of evolution of the north-east Atlantic from a Triassic unrifted continental mass have been identified.

1. Early Jurassic (80 my). Initial split of Africa from North America bounded to the north by a sinistral transcurrent fault from Newfoundland to southern Spain.

2. Early Cretaceous (120 my). The Iberian peninsula, rotating anticlockwise, started to split from the Grand Banks of Newfoundland and from the Celtic Sea opening the margin of the NE Atlantic and the Bay of Biscay. This was accompanied by shearing along the North Pyrenean fault. Spreading between Newfoundland and Spain may also have been coupled, via a transform fault NE of Newfoundland, to the separation of Rockall Plateau from Greenland to form the proto–Iceland basin and from Europe to form the Rockall Trough.

3. Late Cretaceous (80 my). Biscay spreading ceased as Spain stopped rotation with respect to Europe. At about this time spreading started in a new direction along a new axis running NW from Biscay to the Labrador Sea along the old transform fault.

4. Palaeocene (60 my). A change in stress pattern below the plates resulted in a split between Greenland and Rockall Plateau to the east of the Cretaceous trough axis, which grew into the Reykjanes Ridge, and further NE into the Aegir and other ridges north of the Faeroes. At this time the Vøring Plateau became separated from the Greenland margin, and volcanoes and dykes appeared in E. Greenland and N.W. Scotland. A triple junction developed south of Greenland, which now became a separate plate.

5. Middle Eocene (45 my). Spreading in the Labrador Sea virtually ceased and Greenland joined the North American Plate. The northward movement of the Iberian plate, once more moving independently of Europe, resulted in the compression phase of the Pyrenean orogeny, the subduction beneath Spain of some of the Bay of Biscay, the uplift of the central region of Biscay and possibly the formation of King's Trough further west.

6. Middle Oligocene (30 my). North of the Faeroes, the axis of spreading shifted from the Aegir Ridge to a position along the east coast of Greenland separating off a sliver of continent now preserved as the Jan Mayen Ridge.

7. Miocene (15 my). At about this time, Iceland began to appear as an area of abnormal magma output related to the development of a mantle plume, although the existence of the transverse ridges either side of Iceland suggests that abnormal activity in this region may have occurred throughout the Tertiary.

8. Present (0 my). Spreading continues along the Mid-Atlantic Ridge and the Reykjanes Ridge, separated by the Charlie–Gibbs Fracture Zone. North of Iceland, the Kolbeinsey Ridge is the site of asymmetrical spreading since the Late Miocene, and this connects northwards with the Mohns and Atka Ridges.

A. S. Laughton, Institute of Oceanographic Sciences, Wormley, Godalming, Surrey, England

Introduction

The theory of sea floor spreading and the subsequent global theory of plate tectonics have enabled the oceans and continents to be studied in an entirely new way, and now allow precise statements to be made about the stage by stage evolution of the oceanic crust and lithosphere. The widespread success of these theories in explaining the main features of the oceans, as well as in providing a rational basis for understanding orogenic processes on the continents leaves little doubt that in their essentials the theories are correct.

In the oceans, reversals of the earth's magnetic field through the Mesozoic and Tertiary have imprinted magnetic patterns on the upper layers of the crust during sea floor spreading and where these have been mapped, the crustal ages can be deduced by comparison with the reversal time scale (Heirtzler et al. 1968, Larson & Pitman 1972). Thus isochrons can be mapped wherever adequate magnetic data exist.

In the northeast Atlantic many surveys of magnetic anomaly patterns have been carried out and sufficient age identifications have been made to enable a fairly detailed sequence of plate movements to be deduced. Many of these age identifications have been confirmed by direct sampling of sediments overlying the basement rocks, or by sampling reflecting horizons in a sediment sequence which can be traced to basement contacts.

To reassemble the continents into their pre-split configuration, it is also necessary to determine the position of the edges of the continental blocks and to locate any continental fragments which may have subsided following the break-up of the main continental mass.

On the assumption that major distortions have not occurred within plates, palaeogeographic reconstructions for any age can be prepared by removing all younger crust, and by closing up the plates in the directions dictated by the fracture zones until the continental edges are adjacent. The process becomes one of geometry on a spherical earth, bearing in mind always that the plates that are moved have another edge usually outside the area being described, and that there may be constraints there on their motion.

Morphology of the northeast Atlantic

For the purposes of this paper, the northeast Atlantic lies north of a line from southern Spain to the shelf south of Newfoundland. It includes, therefore, the Labrador Sea, the Bay of Biscay, the Norwegian Sea and the Greenland Sea (Plate 1).

The mid-ocean ridge is the dominating morphologic feature in the area running approximately north from the Azores Plateau as part of the mid-Atlantic Ridge. It is offset westwards by a series of east-west oriented fracture zones of which by far the longest is the Charlie Gibbs Fracture Zone which can be traced from 22°W to 47°W along 52°N. The mid-ocean ridge (in this place called the Reykjanes Ridge) runs northward to 57°N and then bends to the northeast as a remarkably linear ridge as far as the southwest corner of Iceland. Iceland is entirely volcanic, having been derived from differentiation of the mantle in the same way as the ocean ridges, but it is anomalous in that the quantity of magma produced exceeds that of a normal mid-ocean ridge by a factor of three or more (Piper 1973). A spreading axis can be identified in Iceland although in central and northeast Iceland it is offset eastwards from the line of the Reykjanes Ridge and its northward continuation.

North of Iceland the morphology of the sea floor becomes considerably more complex. The axis of present spreading lies along the Kolbeinsey (or Iceland–Jan Mayen) Ridge as far north as Jan Mayen Island. The Jan Mayen Fracture Zone offsets the spreading axis eastwards from the Kolbeinsey Ridge to the Mohns Ridge and then continues to the Atka Ridge, west of Spitsbergen. The Spitsbergen Fracture Zone links the Atka Ridge with the Nansen Ridge in the Arctic. Earthquake epicentres lie along the axis of this mid-ocean ridge system and along the linking transform faults indicating that this is the locus of current sea floor spreading and thus identifying the present boundary between the Eurasian and North American plates.

Other older and discarded spreading axes in the Norwegian Sea, Labrador Sea, Rockall Trough and Biscay, which are indicated from the magnetic data or from the geometry of the reconstructions, often have no topographic expression and lie buried beneath thick sediments.

A major ridge crosses the northeast Atlantic through Iceland linking the Greenland continental shelf with the Faeroes shelf. Little work has been done on the western part of this ridge, but studies on the Iceland–Faeroes Ridge indicate that the crust is oceanic in character but is anomalously thick due to unusually active differentiation of basalt from the upper mantle prior to the formation of Iceland (Bott et al. 1971).

Continental boundaries

The edge of a continental block is usually assumed to be close to the base of the continental slope, since on the slope itself, continental rocks can often be sampled. However, the intitial stages of rifting which lead to a split and subsequently to the emplacement of oceanic crust, may involve a period of crustal thinning by stretching, block faulting and tilting, with subsequent downwarping or subsidence during which time the original continental edge becomes blurred. The transition from oceanic to continental crust may be buried beneath thick sediments at the foot of the slope and may be scarcely detectable by geophysical means. Prograding sedimentation on the shelf may on the one hand advance the shelf break beyond the transition, or erosion may cut it back into the continent. There is therefore a measure of doubt about the exact reconstruction of continents due to uncertainties about the true position of the original break. In the reconstructions in this paper the top and bottom of the continental slope are shown and fits are made generally using the base of the slope.

However, there are many areas where the conventional shelf, slope and rise sequence are absent and where detached and isolated continental fragments are found. East of the Grand Banks off Newfoundland, Flemish Cap is believed, from bottom samples and from geophysical data, to have a central basement area of an eroded complex of intruded and metamorphosed sedimentary rocks of late Precambrian age, covered in part by Cretaceous limestone formations, and is isolated from the Grand Banks by Jurassic faulting (Grant 1972).

450 km north of Flemish Cap, the continental fragment of Orphan Knoll was drilled on Leg 12 of the Deep Sea Drilling Project, and its subsidence from sea level to 2000 metres in the Palaeocene was determined from the sediments recovered (Hole 111 in Laughton et al. 1972). Grant (1972) believes that the deep water col between Orphan Knoll and the continental slope is also subsided continent, and that the latest possible date for the isolation of the Orphan Knoll and Flemish Cap as a result of sea floor spreading is Early Cretaceous.

Baffin Bay has been shown by seismic refraction measurements to be floored by rather thin oceanic crust in the central basin (Keen & Barrett 1972). Under the Davis Strait sill, linking Baffin Bay to the Labrador Sea, a seismic structure was found similar to that beneath Iceland, suggesting that the sill was formed by excessive outpouring of oceanic basalts.

There is little direct evidence of the crustal structure between Spitsbergen and northeast Greenland. Magnetic data and the evidence of sea floor spreading geometry suggest that the Lena Trough is a fracture zone in which some oceanic crust has subsequently developed and that the Yermak Plateau northwest of Spitsbergen is continental (Johnson & Heezen 1967, Vogt & Ostenso 1970). Certainly Spitsbergen and the Barents Shelf are continental in structure and, together with Norway, form the northeastern boundary of the oceanic crust in the Norwegian Sea (Harland 1969; Emelyanov et al. 1971).

The Vøring Plateau, off the west coast of Norway, is a sediment-filled basin bounded on the western margin by a basement high. Whereas it is generally agreed that the sediments overlie subsided or downwarped continental crust, Hinz (1972) believes that the basement high is also continental whereas Talwani & Eldholm (1972) believe it to be oceanic. Depending on which view is correct, the continental boundary either cuts through the Plateau or swings out westward from the continental slope off Norway. An accurate reconstruction of the Norwegian Sea based on a magnetic survey over the Mohns Ridge might indicate whether there was room for the outer part of the Plateau in its fit against the East Greenland shelf edge.

South of Jan Mayen Island, there is some evidence that the Jan Mayen Ridge is a continental sliver detached from the continental margin of Greenland during a mid-Tertiary shift of spreading axis (Johnson & Heezen 1967; Johnson et al. 1972; Eldholm & Talwani 1973), although towards its southern end the ridge appears to have been built up of current-carried oceanic sediments (Hinz, this volume).

Between the Vøring Plateau and the Faeroe Islands, the continental margin has been determined from seismic and magnetic data to lie along the northeast extension of the Faeroes–Shetland Escarpment on the gentle continental slope off Norway (Talwani & Eldholm 1972). Structurally this escarpment continues SW into the Faeroes–Shetland Channel and may mark the continental edge of NW Scotland. There is evidence from gravity and seismic data that continental crust underlies the Faeroe Islands (Bott et al. 1971; Casten 1973) and that this, together with Rockall Plateau is a large detached continental fragment.

Rockall Plateau is linked to the Faeroe Islands by the Faeroe Rise, a rather shoal region on which there are a number of shallow banks and which is interpreted by Bott & Watts (1971) to be continental in structure. The geophysical evidence of the continental nature of Rockall Plateau itself is clearer (Roberts 1971; Scrutton & Roberts 1971; Scrutton 1972) and this has been confirmed by direct sampling (Roberts et al. 1972, 1973; Miller et al. 1973). Data from the JOIDES drill holes showed that subsidence of Rockall Plateau occurred during the Palaeocene (Sites 116 and 117, Laughton et al. 1972).

Porcupine Bank is a spur of shoal water running southwest from western Ireland. Sampling, gravity, seismic reflection and seismic refraction data all indicate that it is continental (Stride et al. 1969; Gray & Stacey 1970; Clarke et al. 1971; Whitmarsh et al. 1974). It is not certain, however, whether the crust underlying Porcupine Sea Bight separating it from the continental shelf is oceanic or subsided continental, and thus whether the Bank can be closed into the shelf during a reconstruction (Scrutton et al. 1971).

Finally, Galicia Bank and the smaller Vigo Seamount off the northwest coast of Spain have been shown by dredging and by seismic and gravity measurements to be fault-bounded and subsided continental blocks (Black et al. 1964).

We have, therefore, a number of more or less totally submerged continental fragments that must have become detached from their parent landmasses during or after the split of the North Atlantic and which subsequently subsided. In any reconstruction these pieces have to be included in the geometry. However, apart from Rockall Plateau (see below), there is little or no geological control to suggest how far they have moved horizontally or in what direction since they have become detached. On a somewhat arbitrary basis in the predrift reconstruction Porcupine Bank has been rotated back to close Porcupine Sea-bight, and Orphan Knoll, Flemish Cap and Galicia Bank have been moved northward in order first to allow the Iberian continental slope to lie adjacent to that off the Grand Banks, and second to fill in a gap between Ireland and Labrador.

Palaeogeographic reconstructions

In deriving palaeogeographic reconstructions it is necessary to start from the present, identifying active plate boundaries from the epicentres. An older plate boundary is determined from the magnetic data and the new oceanic crust is removed from the map by rotating the plates together towards the active centre using the fracture zones or trends in the anomaly pattern to determine the direction of relative movement. Although there appears to be a general tendency for spreading directions to be perpendicular to the spreading axis (or perhaps more significantly, for the spreading axes to become perpendicular to the spreading directions), nevertheless, in the North Atlantic there are many clear cases of oblique spreading, of which the Reykjanes Ridge is the best documented (Talwani et al. 1971).

A reconstruction thus derived must then undergo the same treatment to go back another stage, bearing in mind that features such as segments of fracture zones which appeared to be unrelated before the first stage may now show a clear relationship to guide the second stage of reconstruction, and that spreading axes or poles may have changed during the evolution.

This stage by stage reconstruction is possible so long as the magnetic anomalies are clearly identifiable or until continental edges collide. Unfortunately, near to the continental edge, the magnetic anomalies are often ill-defined, weak or absent. Various theories have been put forward to explain this magnetic quiet zone (Vogt et al. 1970; Larson & Pitman 1972; Poehls et al. 1973) some of which state that at the time of formation of these marginal regions (in the middle to late Cretaceous between 110 and 85 my ago) there were no reversals of the earth's magnetic field, whereas others attribute the lack of anomalies to the downwarping of the oceanic crust under the influence of marginal subsidence and sediment accumulation, and the consequent higher temperatures which may exceed the Curie point or at least speed the thermally induced decay of magnetisation. The Cretaceous history is therefore more speculative although the size of the step from the better determined Late Senonian (78 my) reconstruction to a pre-split configuration is not too large.

In the region between Labrador, Greenland, Iceland and U.K. several detailed magnetic surveys have been made both by air and by sea. The magnetic anomalies identified from these surveys and which are used in the subsequent set of reconstructions are shown in Fig. 8. The major part of the data from Iceland southwards across the Reykjanes Ridge to the southern side of Rockall Plateau are derived from USNOO data (Avery et al. 1969; Ruddiman 1973; Vogt & Avery 1974). Additional data on the Reykjanes Ridge are from Heirtzler et al. (1966), Godby et al. (1968), Fleischer (1969, 1971), Herron & Talwani (1972) and Johnson & Egloff (1973). In the Labrador Sea and south of Greenland data are from Godby et al. (1966), Mayhew (1969), Mayhew et al. (1970), Le Pichon et al. (1971) and Laughton (1972). On the Iceland–Faeroes Ridge magnetic surveys have been carried out by Fleischer (1971), Bott et al. (1971) and Johnson & Tanner

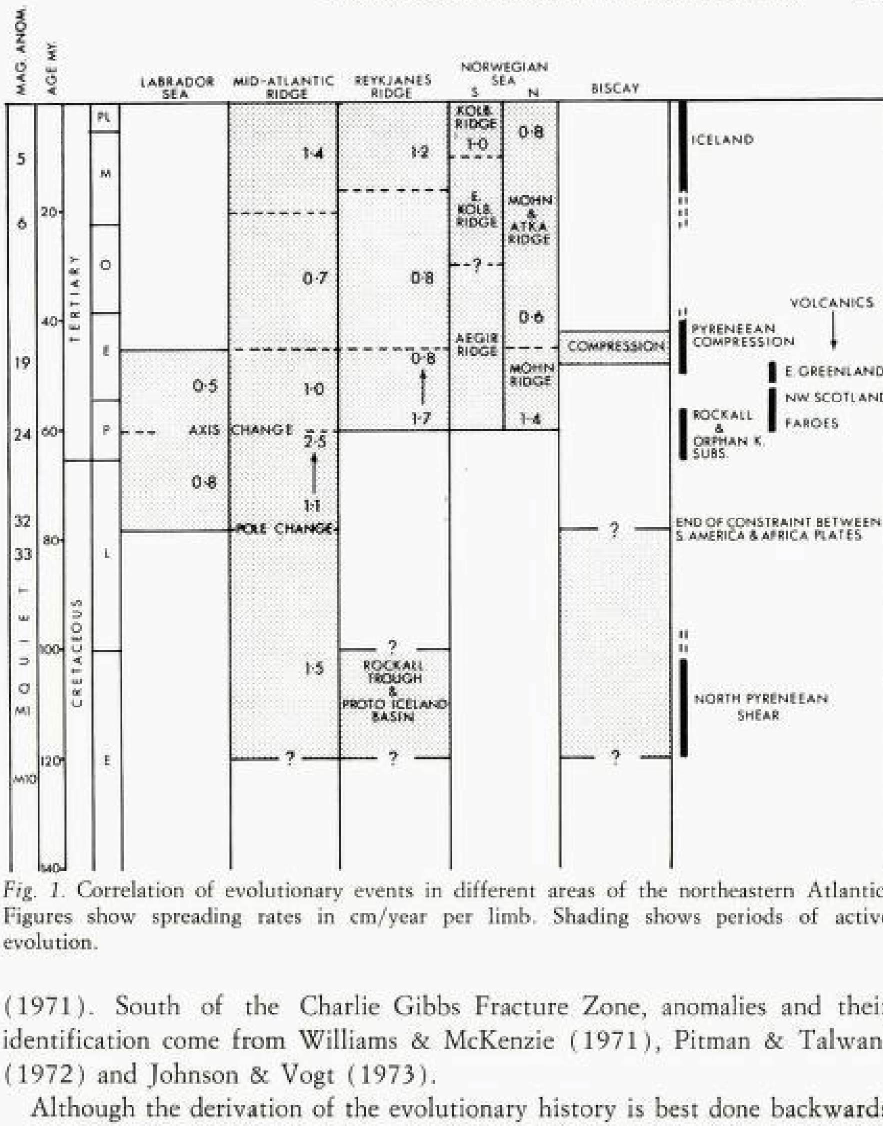

Fig. 1. Correlation of evolutionary events in different areas of the northeastern Atlantic. Figures show spreading rates in cm/year per limb. Shading shows periods of active evolution.

(1971). South of the Charlie Gibbs Fracture Zone, anomalies and their identification come from Williams & McKenzie (1971), Pitman & Talwani (1972) and Johnson & Vogt (1973).

Although the derivation of the evolutionary history is best done backwards in time, the telling of the history is perhaps clearer starting from the pre-split configuration and working towards the present. It is impossible here to elaborate all the data and the arguments that have determined this history. These will be found throughout the literature cited. But I will attempt to relate the stages of evolution as they occur with tectonic events in and around the relevant plates. In Plate 2, the isochron map summarises the results of the palaeogeographic reconstructions shown in Figs. 3 to 8 for the area south of Iceland, using as far as possible the plate boundaries derived from the magnetic surveys, and links them with isochrons in the Norwegian and Greenland Seas and the Arctic derived from the literature. The geometrical consistency between the plate movements south and north of Iceland has not, however, been tested. In Fig. 1, the activities in the various parts of the North Atlantic are related to the geological and magnetic polarity time scales, and to some important geological events.

Pre-split palaeogeography

The reconstruction shown in Fig. 2 represents the closest assemby of continental margins and fragments that can be made by the removal of all known or hypothesised oceanic crust. The closure of Biscay by the rotation of Spain has been made in accordance with the proposal of Le Pichon et al (1971). Rockall Trough and the proto–Iceland Basin have been closed by the translation northwestwards of the European Plate. The position of Africa shown is only approximate and does not take into account the relative movements of the Moroccan and Oranaise plates of northwest Africa described by Dewey et al. (1973).

The reconstruction can be tested by relating major pre-split orogenic features across from one side to the other. In Labrador, the Grenville Front marks the northern boundary of the Grenville orogeny of about 950 my ago. On Rockall Bank, rocks of Grenville (990 my) and Laxfordian (1600 my) age have been sampled (Miller et al. 1973) and it has been suggested that these straddle the eastward extension of the front, which may also be associated with an E–W magnetic lineament and with the E–W scarp bounding the SW edge of Rockall Plateau (Roberts et al. 1973).

The continuity of the Acadian–Caledonian orogenic belt has been extensively discussed (see many papers in Kay 1969), but the rather large separation of the mapped regions of the Acadides and Taconides in Newfoundland, and the Caledonides in Ireland and Scotland make it difficult to use this feature as a test of the correctness of this reconstruction. The continuation to the northeast of the Acadide structure is obscured by a Mesozoic–Tertiary basin of sediments running along the outer edge of the Labrador continental shelf (Grant 1972). The basin is bounded on the land side by a major fault. However, it is cut by NE–SW faults which may result from the reactivation of old Acadian fault lines in the underlying basement. On the European side, uncertainties about the nature of Porcupine Bank prevent any detailed studies of the southwest extension of the Caledonides.

The Hercynian (Variscan) orogeny is found in southern England and Ireland, and in western France. Cogné (1971) and Bard et al. (1971) have discussed the continuity of Hercynian structures into the Iberian peninsula prior to the opening of Biscay. The structures are not, however, clearly seen in the continental margin west of France (Montadert et al. 1971) and may have been obscured by Mesozoic and Tertiary folding and sedimentation. Bard et al. (1971) link the Hercynian folding in Iberia with that of northwest Africa. The Hercynian system can thus be traced as a sinuous zone on the reconstruction of Fig. 2 from France to Africa. The Hercynian Front, which is well mapped in southern Ireland and England but which is not seen in Newfoundland, must therefore run across the Grand Banks and lie subparallel to the continent shelf edge SE of Newfoundland.

Associated with the Front in Wales and southern Ireland, there are high grade anthracites. On Orphan Knoll, Jurassic sandstones of non-marine or

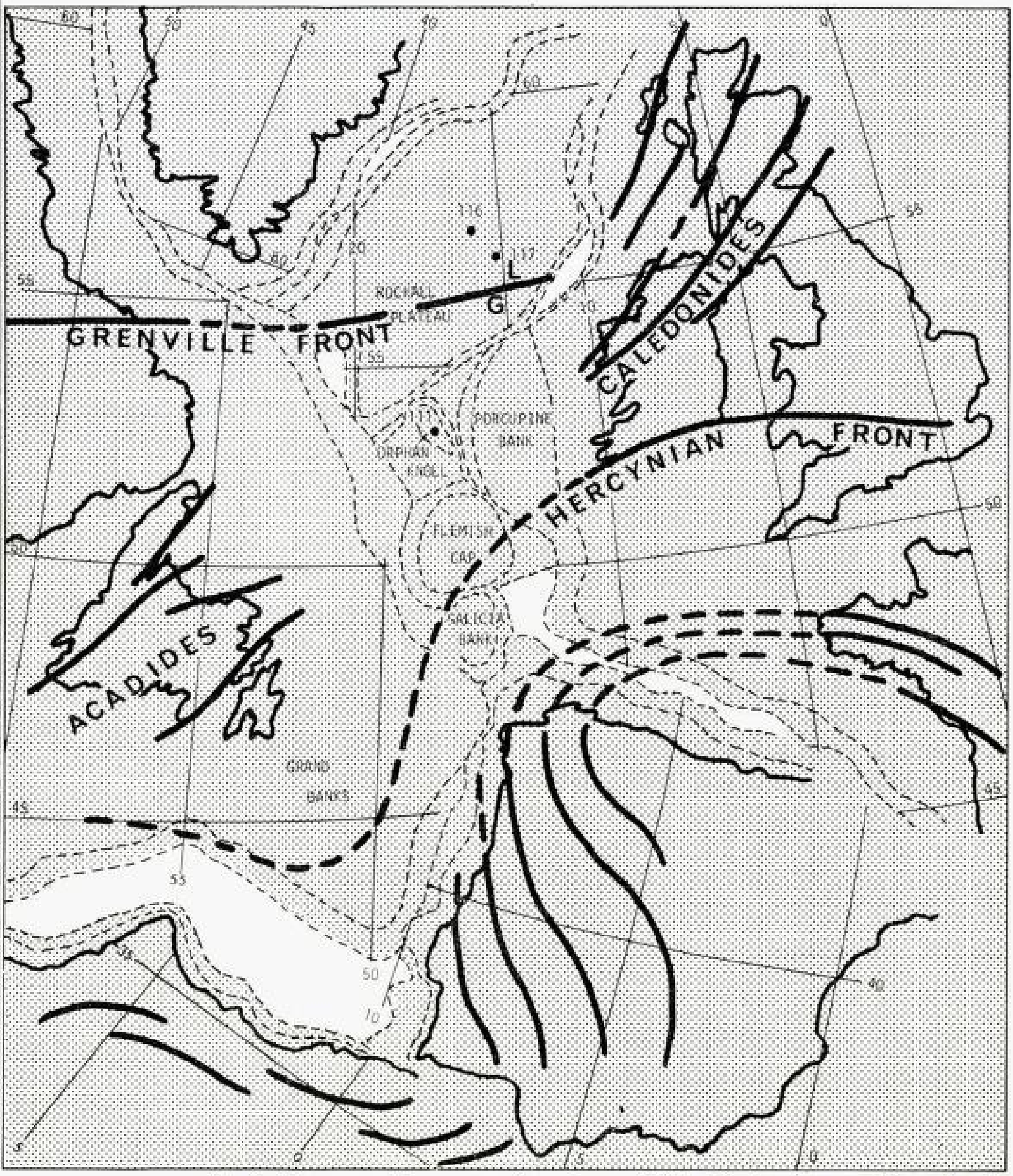

Fig. 2. Palaegeographic reconstruction prior to initial split showing correlation of major orogenic belts (N.B. Orphan Knoll, Flemish Cap and Galicia Bank have been moved arbitrarily to fill the gap between Porcupine Bank and the Labrador Shelf). L and G show positions of Laxfordian and Grenvillian rocks sampled on Rockall Bank.

very shallow coastal environment were sampled which contained detrital anthracite similar to South Wales anthracite (Hole 111, Laughton et al. 1972). This suggests that Orphan Knoll lay south of the Hercynian Front and thus south of the position shown in Fig. 2.

The initial split of Africa away from N. America took place at the beginning of the Early Jurassic at about 180 my based on magnetic anomaly interpretations (Pitman & Talwani 1972) and on the age of Jurassic sediments sampled at hole 105 near the continental margin of N. America (Hollister et al. 1972). During the Early Jurassic, Africa moved SE relative to the America–Eurasia plate along a transform fault running from south of Newfoundland to the south of Iberia. Dewey et al. (1973) suggest that during this movement smaller plates broke off NW Africa, but that north of the transform fault, the America–Eurasia plate remained intact until the middle of the Early Cretaceous.

Early Cretaceous

The sequence of events associated with the first propagation of the Atlantic split northwards is difficult to establish with certainty. The split between Iberia and the Grand Banks could have been continued north of Spain and into the Pyrenees as a transform fault, carving off the Iberian plate, as has been suggested by Dewey et al (1973). In this case the date of the initial split can be related to geological data in Iberia and northwest Europe and has been argued by Dewey et al. to be Hauterivian (120 my). Choukrane et al. (1973) proposed, from similar evidence and from data on the Bay of Biscay, that the Iberian plate started to rotate in the Cretaceous following a subsidence of the margins of Biscay at the beginning of the Early Cretaceous.

However, it is possible that the split also extended northwest in the direction of the Labrador Sea. Grant (1972) speculated that there had been an 'intracratonic' depression in the proto Labrador Sea and running SE to the Grand Banks since the early Palaeozoic, in which Mesozoic sediments accumulated. Seismic profiles clearly delineate the landward side of this depression. This syncline could have been the forerunner of a line of split running from Baffin Bay to the Pyrenees. Mesozoic sediments have been postulated (but not proved) to lie deep in Rockall Trough over oceanic crust (Stride et al. 1969; Scrutton & Roberts 1971) and possibly in the region off SE Greenland. Thus the opening between Iberia and the Grand Banks in Early Cretaceous may have continued northward into Rockall Trough and the proto-Iceland Basin through a transform fault parallel to the shelf edge NE of Newfoundland (Fig. 3). If the pole of rotation implied by such geometry were not too far to the northeast, the amount of opening would be rapidly reduced northwards, where oceanic crust would be hard to recognise or may be absent if a degree of continental crustal thinning took place.

It is possible that both spreading in Rockall Trough and the proto-Iceland Basin, and shear and opening of the Bay of Biscay may have taken place

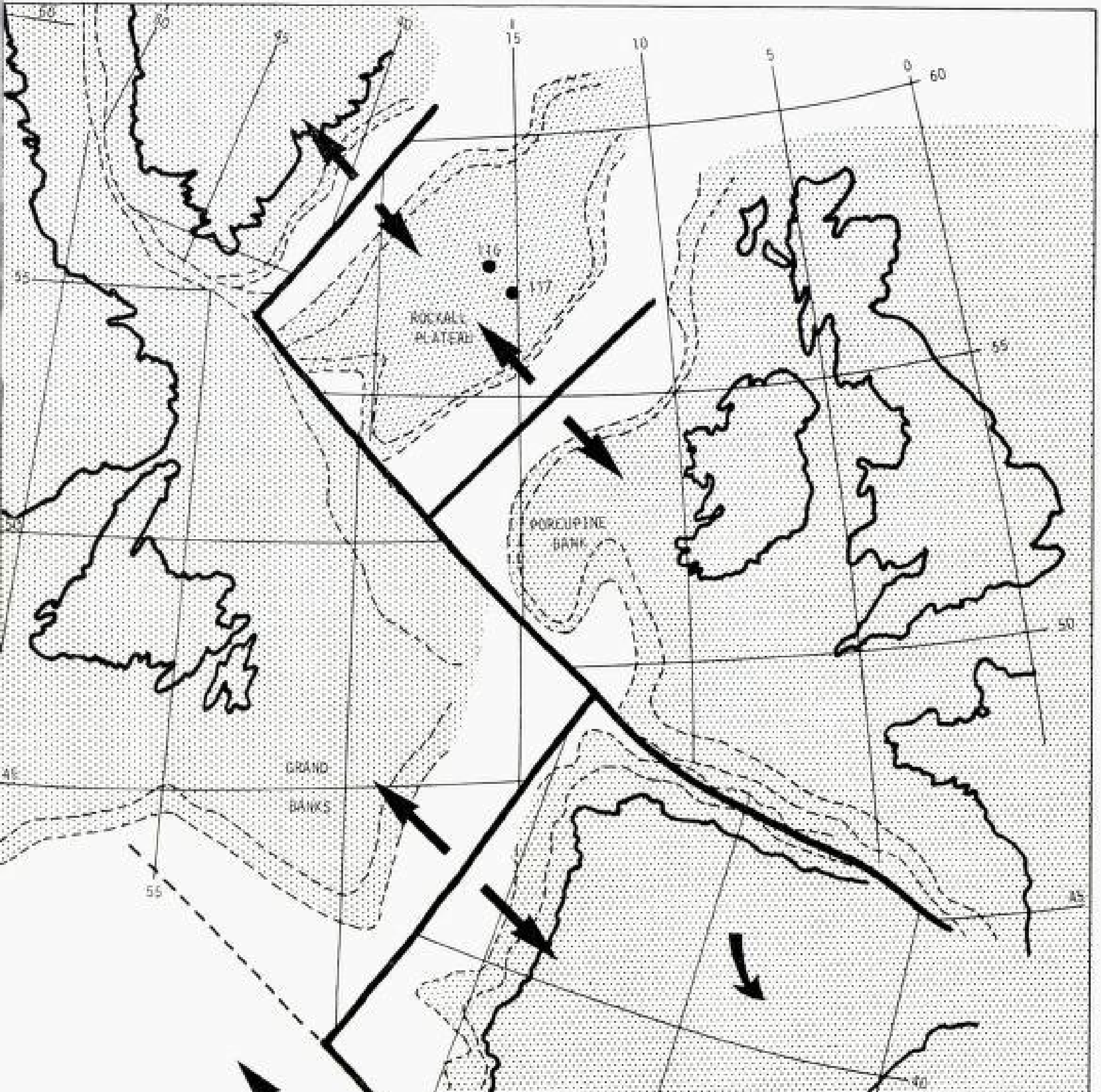

Fig. 3. Palaeogeographic reconstruction in Early Cretaceous (110 + 10 my) showing the end of the evolution of proto-Iceland basin and Rockall Trough and the start of the opening of Biscay. These events may, however, have been simultaneous.

simultaneously in the Early Cretaceous to accommodate the opening between Iberia and the Grand Banks. Between 85 and 110 my, Larson & Pitman (1972) have shown that there were no magnetic reversals and therefore in this period there would be no magnetic anomalies created to be used to date the ocean crust. In hole 120 of DSDP, samples of bathypelagic Barremian (115 my) marl ooze were recovered from the uplifted Gorringe Bank (Ryan et al. 1973), but the position of this hole is somewhat south of the Iberian–Grand Banks split and may in fact belong to ocean crust generated by the separation of Africa and North America.

Two-plate spreading during the Late Cretaceous and Palaeocene (80-60 my)

The Bay of Biscay was fully opened by the Late Senonian (75 my) since anomaly 31–32 can be traced unequivocally more or less parallel to the mid-Atlantic Ridge, cutting across the E–W magnetic lineations of Biscay (Williams & McKenzie 1971). Choukrane et al. (1973) argue that the westward migration of the rotation pole of Iberia in the Late Cretaceous blocked the shear movement along the Pyrenean fault and prevented further rotation of Iberia. However, the opening may alternatively have been stopped as a result of a major change of the stress field between the plates, resulting in an E–W separation of the American and Eurasian plates. The transform fault between the United Kingdom and Labrador became at this time a spreading centre, extending NW between Labrador and Greenland (Fig. 4).

Geophysical data in the Labrador Sea (Le Pichon et al. 1971) followed by deep drilling (Laughton et al. 1972) show that Greenland started to separate from Labrador about 80 my ago, although the separation is relatively small in Baffin Bay (Keen & Barrett 1972). At the same time north of Greenland, the Wegener fault linked the Baffin Bay opening to the Arctic Ocean (Ostenso 1973).

Le Pichon & Fox (1971) attribute the change of stress field at 80 my throughout the North and South Atlantic regions to the fact that the thick continental lithosphere of West Africa ceased to be constrained against the thick continental lithosphere of the northeast part of South America, as they gradually moved apart. This initiated a widespread change of spreading axes and spreading rates. Greenland, Europe and Iberia were now acting as one plate separating from North America. Any earlier spreading in the proto-Iceland Basin and Rockall Trough had ceased and thick sediments accumulated from the nearby continental masses. At this time Rockall Plateau was still above sea level and may have been mountainous along its eastern margin. (See sites 116 and 117 in Laughton et al. 1972.)

Spreading proceeded with this geometry into the Palaeocene (from 80 to 60 my), giving linear magnetic anomalies that are well mapped in the Atlantic as far north as southern Greenland but which are not so easily identified (except for the anomalies 24 and 25) in the Labrador Sea, perhaps on account of the thick sediments, marginal downwarping and slow initial spreading rate. The spreading rate per limb in the Atlantic at about 45°N increased from 1.1 cm/year at 80 my to about 2.5 cm/pear at 60 my (Williams & McKenzie 1971) whereas in the Labrador Sea, the mean rate was only 0.8 cm/year

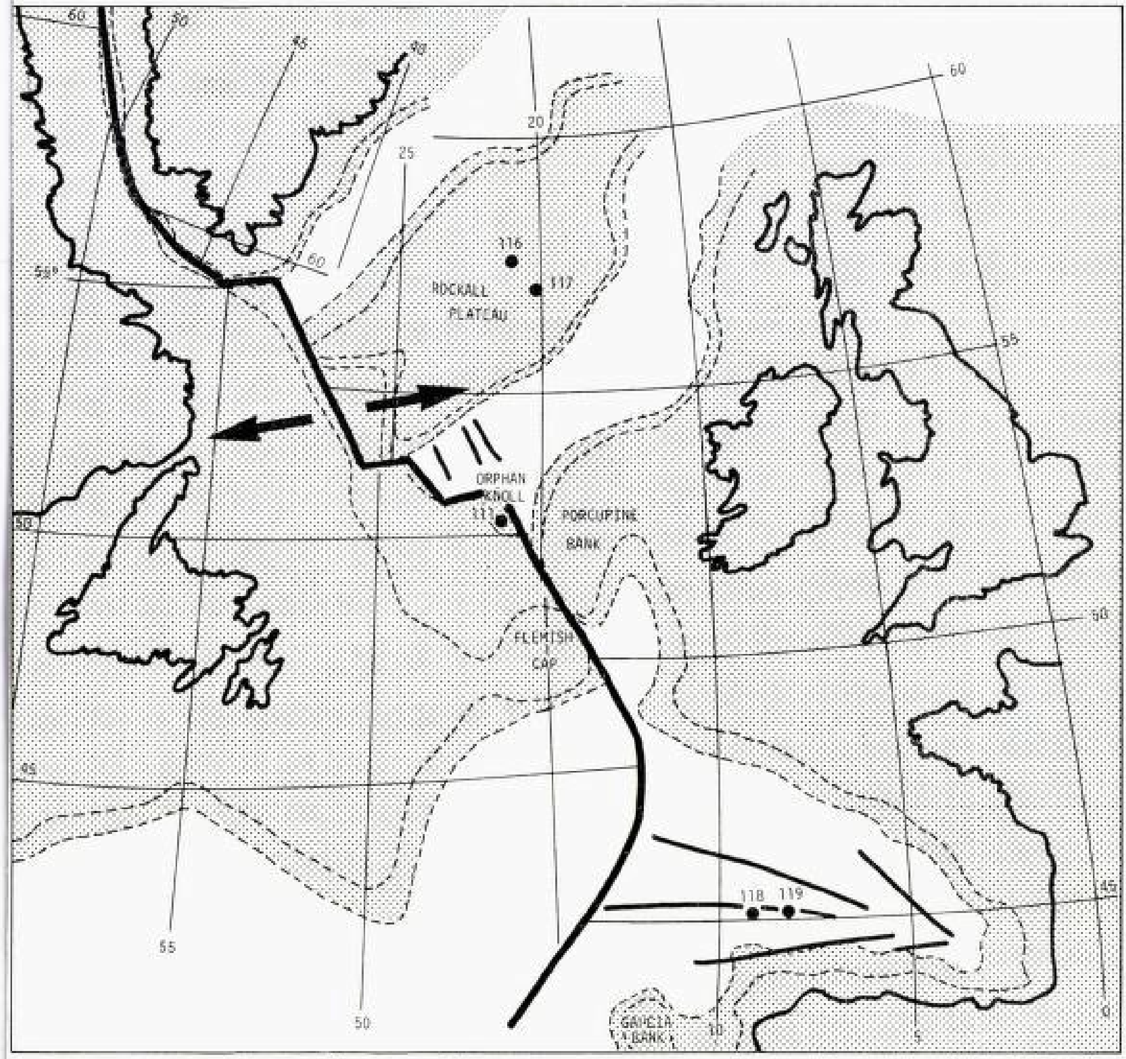

Fig. 4. Palaeogeographic reconstruction at anomaly 32 (78 my–Santonian). Black lines show mapped magnetic anomalies.

(Le Pichon et al. 1971). If both America and Eurasian plates were rigid during this period, then in the Labrador Sea a similar acceleration of spreading occurred which would have been from 0.5 to 1.1 cm/year in order to give the correct mean rate. The initial spreading rate may therefore have been too slow to give identifiable magnetic anomalies.

Fracture zones south of Greenland and between Newfoundland and Rockall Plateau reflect offsets (Fig. 4) in the initial line of opening, which may in turn have been determined by Grenvillian and Acadian fault trends (Grant 1972; Olivet et al. 1974). Spreading continued through the Campanian (Fig. 5) and Maestrichtian until the Palaeocene. The dotted line in Fig. 6 shows the last position of the spreading axis prior to the major change at

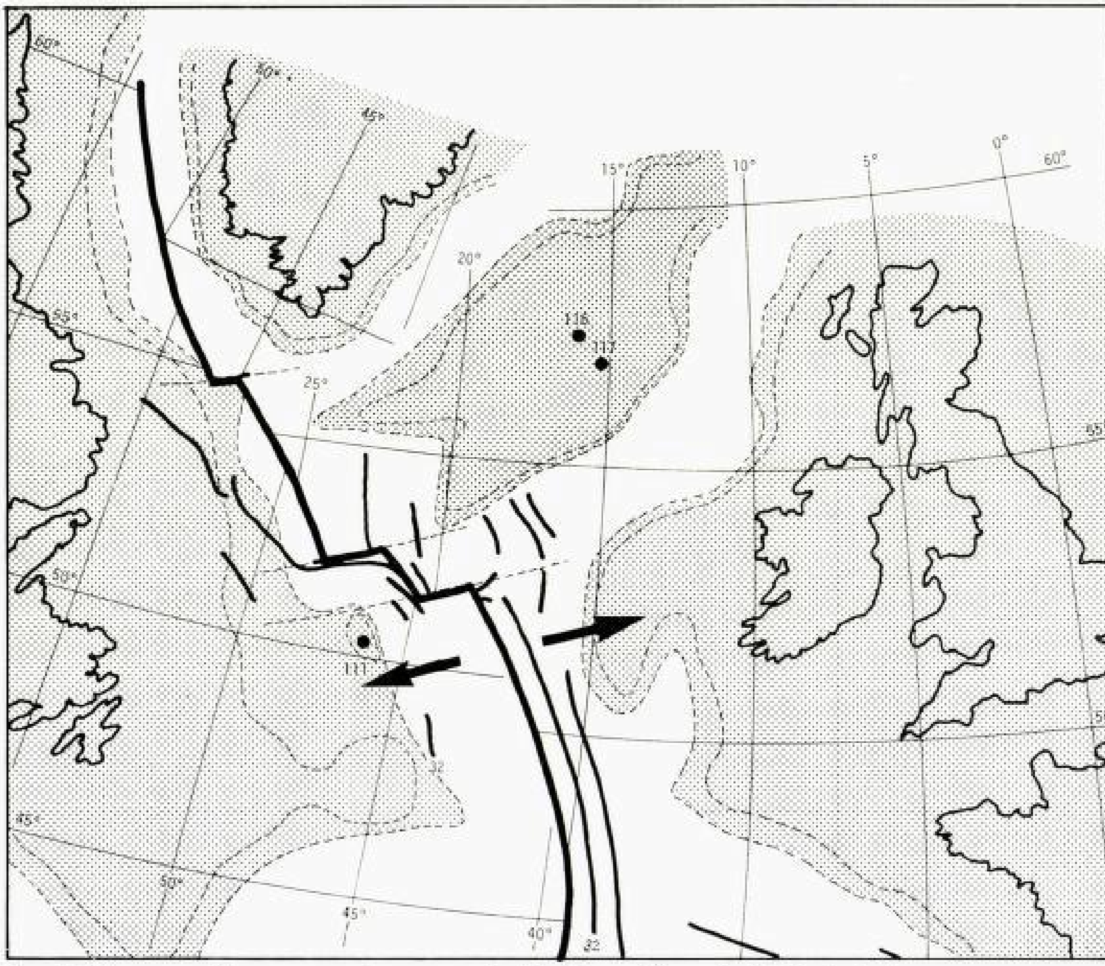

Fig. 5. Palaeogeographic reconstruction at anomaly 31 (72 my–Campanian). The dashed lines show small circles of transform faults.

60 my. Hole 112 sampled basement, just east of this axis, the age of which was estimated at 65 my (Laughton et al. 1972).

Three-plate spreading during Palaeocene and Eocene (60-47 my)

In the Palaeocene at about the time of anomaly 24 (60 my), another major readjustment of spreading geometry occurred caused by a further change in the pattern of forces driving the plates (Fig. 6). The Greenland plate became detached from the Eurasia plate and the spreading centre propagated northeastwards between Greenland to the west, and Rockall Plateau, Faeroe

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Fig. 6. Palaeogeographic reconstruction at anomaly 24 (60 my–Palaeocene). The dotted lines show spreading axes prior to 60 my. Arrows denote relative spreading direction between plate pairs.


IRELAND
116
117
118
119
112
111
31
32

Islands, Vøring Plateau, the northwest Norwegian continental shelf and the Barents shelf to the east. Northeast of Greenland, Spitsbergen split away along a transform fault coupling the Norwegian Sea spreading axis to the Arctic spreading axis. The date of initiation of this split is well determined from the mapped anomaly 24 which lies close to the west margin of Rockall Plateau (Vogt & Avery 1974; Ruddiman 1973), and which can be identified near the margins of the Norwegian Sea (Avery et al. 1968; Phillips 1973). The evolutionary story of the Norwegian Sea has been developed through the last five years by many workers using the data both from the sea floor and from the geology of the neighbouring continents (Johnson & Heezen 1967; Avery et al. 1968; Vogt et al. 1970; Johnson et al. 1971, 1972; Bott 1973), and their data and arguments will not be repeated here. Only their main results will be used in this chronological account of the north Atlantic evolution.

Between Rockall Plateau and Greenland, the new split lay to the east of the axis of the proto-Iceland Basin, against the foot of Hatton Bank. The Reykjanes Ridge began to develop leaving the older sea floor attached to Greenland. At about the latitude of the Iceland–Faeroes Ridge the axis was offset to the east creating the Norwegian Seamount chain (also called the Aegir Ridge). The spreading centre continued northward along the flanks of the Mohns Ridge which was linked to the Nansen Ridge in the Arctic by the Spitsbergen Fracture Zone (de Geer shear zone).

The effect of the new split east of Greenland was to change the geometry from two to three plates, since the Labrador Sea continued to open (Laughton 1971, 1972). The stresses on the three plates, and hence their relative motions, changed, resulting in an alteration of spreading axes. South of Greenland a triple junction appeared in relatively old ocean crust 150 miles from the previous spreading axis (Fig. 6). In the southern Labrador Sea new fracture zones appeared at a different orientation, but further north the axis stayed constant (Le Pichon et al. 1971). South of the triple junction the spreading axes became N–S not NW–SE, and E–W fracture zones developed starting in the position of the older ones (Olivet et al. 1970). Oblique spreading gave way to orthogonal spreading and the spreading rate was substantially reduced from 2.5 to 1.0 cm/year (Williams & McKenzie 1971; Pitman & Talwani 1972).

Johnson & Vogt (1973) associate a topographic ridge in the eastern N. Atlantic with the 60 my isochron and suggest that this was caused by increased mantle plume activity related to the change of stress pattern. They also recognise a change in spreading style from oblique to a series of en échelon orthogonal spreading axes. Considerable volcanic activity occurred at this time in NW Scotland (50–60 my), on Rockall (60 my), in the Faeroe Islands (50 my) and in East Greenland (50–60 my) (Bott 1973) and apparently continued to the present day from the position of Iceland, throughout the period of separation of Europe and Greenland to create the Iceland–Faeroes Ridge (Bott et al. 1971).

After splitting from Greenland, Rockall Plateau began to subside. Evidence

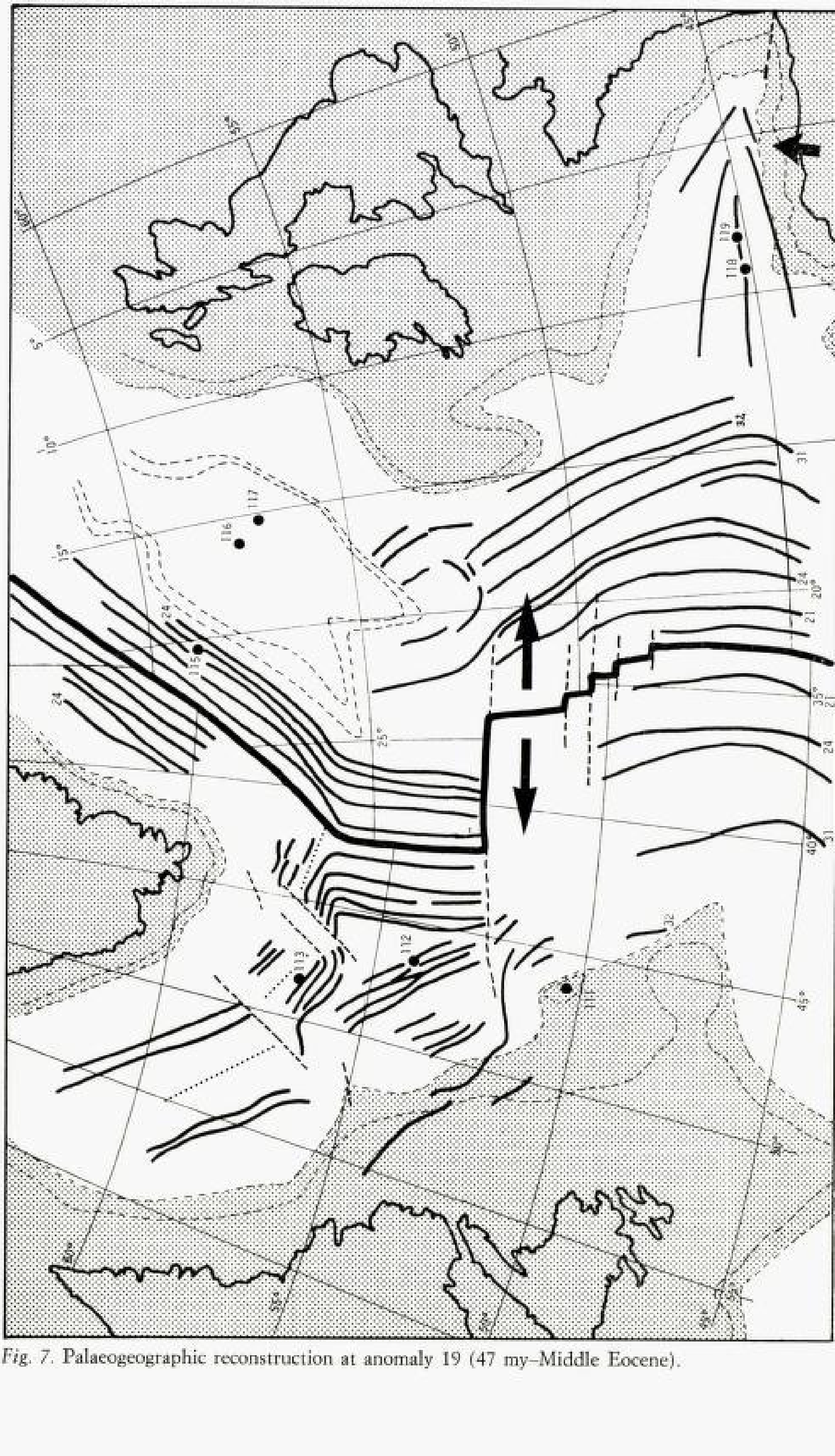

Fig. 7. Palaeogeographic reconstruction at anomaly 19 (47 my–Middle Eocene).

of the palaeoenvironment of sedimentary deposition in samples from drill holes 116 and 117 has provided a subsidence curve showing that between 55 and 50 my, Rockall Bank sank 1200 metres (Laughton et al. 1972). At the same time, Orphan Knoll subsided by 1800 m (Hole 111) even though it was not close to the new split. Evidence of subsidence of the continental sediment basin on the Vøring Plateau and of the region between the Faeroes and the Shetland Islands after the start of the formation of the Norwegian Sea, has been presented by Talwani & Eldholm (1972). These subsidences may be associated with the downwarping and block faulting along continental margins subsequent to split as discussed by Bott (1973) and Sleep (1973).

Two-plate spreading from Eocene to Early Miocene (47-20 my)

The youngest anomalies in the centre of the Labrador Sea are about 47 my, indicating that spreading virtually ceased at that time (Le Pichon et al. 1971; Laughton 1972). Sedimentation from the neighbouring continents and from sediment laden ocean currents from the east and northeast have covered the basement relief associated with the spreading centre (cf. Hole 113 in Laughton et al. 1972). The triple junction ceased to exist and subsequent spreading was between the two plates of America–Greenland and Eurasia (Fig. 7). The spreading rate progressively slowed until it was at a minimum value of about 0.7 cm/year per limb on the Reykjanes Ridge at 30 my (Johnson & Vogt 1973). The Charlie Gibbs Fracture Zone, with an offset of the N. Atlantic spreading axes of 350 km at 60 my, continued to indicate the direction of spreading between the two plates (Olivet et al. 1970).

However, a change occurred in the style of spreading at about this time. On the Reykjanes Ridge, oblique spreading, which had produced strikingly linear anomalies between 60 and 45 my (Fig. 7), gave way to nearly orthogonal spreading in a series of en échelon sections separated by fracture zones, the change being attributed to a small change in the spreading direction (Vogt et al. 1969; Ruddiman 1972). This tendency for oblique spreading axes to change to orthogonal axes was noted by Menard & Atwater (1968) in the Pacific.

Further south the northward movement of Africa (Dewey et al. 1973) during the Eocene resulted in compression along the northern edge of the Iberian plate and along the Pyrenees. This caused a limited subduction of the Biscay oceanic lithosphere beneath the Spanish continental lithosphere (Choukrane et al. 1973) and uplift of a central ridge in the Biscay sea floor (Holes 118 and 119 in Laughton et al. 1972). Le Pichon & Sibuet (1971) associate this Eocene compression phase with the evolution of King's Trough, a 370 km long tectonic feature cutting diagonally across the eastern flanks of the mid-Atlantic Ridge (Matthews et al. 1969). However, the age of crust cut by the western end of King's Trough is Oligocene (29 my) and this would imply that the compression lasted well beyond the Eocene. Certainly some western continuity of the subduction zone is necessary on geometrical grounds but this could be acommodated by a shear zone parallel of the western edge of the Iberian plate.

North of Iceland, the spreading axis jumped westwards from the Aegir Ridge to the Greenland continental margin possibly cutting off a piece of the outer shelf which may be preserved today as the northern part of the Jan Mayen Ridge (Johnson & Heezen 1967), although there are some doubts about the continental nature of all this Ridge (Hinz, this volume). Johnson et al. (1971) suggested an alternative model where the Ridge is part of the early product of the relocated spreading centre. The age of the jump in spreading axis has been variously estimated as 42 my (Vogt et al. 1970), before 30 my (Johnson et al. 1971), 30 my (Johnson et al. 1972) and 18.6 my (Eldholm & Talwani 1973). The new spreading axis is thought to have lain somewhat east of the present axis of the Iceland–Jan Mayen Ridge (also called the Kolbeinsey Ridge).

North of the Jan Mayen Fracture Zone the spreading axis has apparently remained along Mohns Ridge. The older aeromagnetic data in the southern Norwegian Sea of Avery et al. (1968) have now been extended northwards from Jan Mayen Island to the north of Greenland (Phillips 1973). The data reveal that the Mohns Ridge, which in its flank regions comprises linear anomalies arising from oblique spreading, consists on its crest of a series of en échelon sections with right lateral offsets.

Further north still, the fracture zone splitting Spitsbergen from Greenland, changed at about the time of anomaly 20 (49 my) into a spreading axis as the result of a change in the position of the pole of rotation between the Norwegian and Greenland plates (Phillips 1973) and gave rise to the Atka Ridge.

Early Miocene to Present (20-0 my)

From the Early Miocene to the Present (Fig. 8) the plate boundaries did not move much, although the style of spreading changed. South of the Charlie Gibbs Fracture Zone, the en échelon sections of orthogonal spreading that had formed since 60 my now changed gradually into sections of orthogonal spreading linked by oblique spreading, transform faults being nearly absent in crust younger than 10–20 my (Johnson & Vogt 1973). On the Reykjanes Ridge, after 18 my the transform faults disappeared and spreading once more became oblique and parallel to the present axis of the Ridge (Vogt et al. 1970; Ruddiman 1972). At about the same time Iceland began to emerge as an island. The oldest rocks sampled in Iceland are 16 my (Piper 1973) although erosion may have cut back some of the older margins. Ward (1971) has interpreted the geology of Iceland in terms of plate tectonics and identifies two rather diffuse fracture zones on the northern and southern limits of the island which displace the present active spreading centre (the neo-volcanic zone) about 70 miles east of the line joining the Reykjanes Ridge to the Kolbeinsey Ridge. An older spreading centre active about 10 my ago lies in

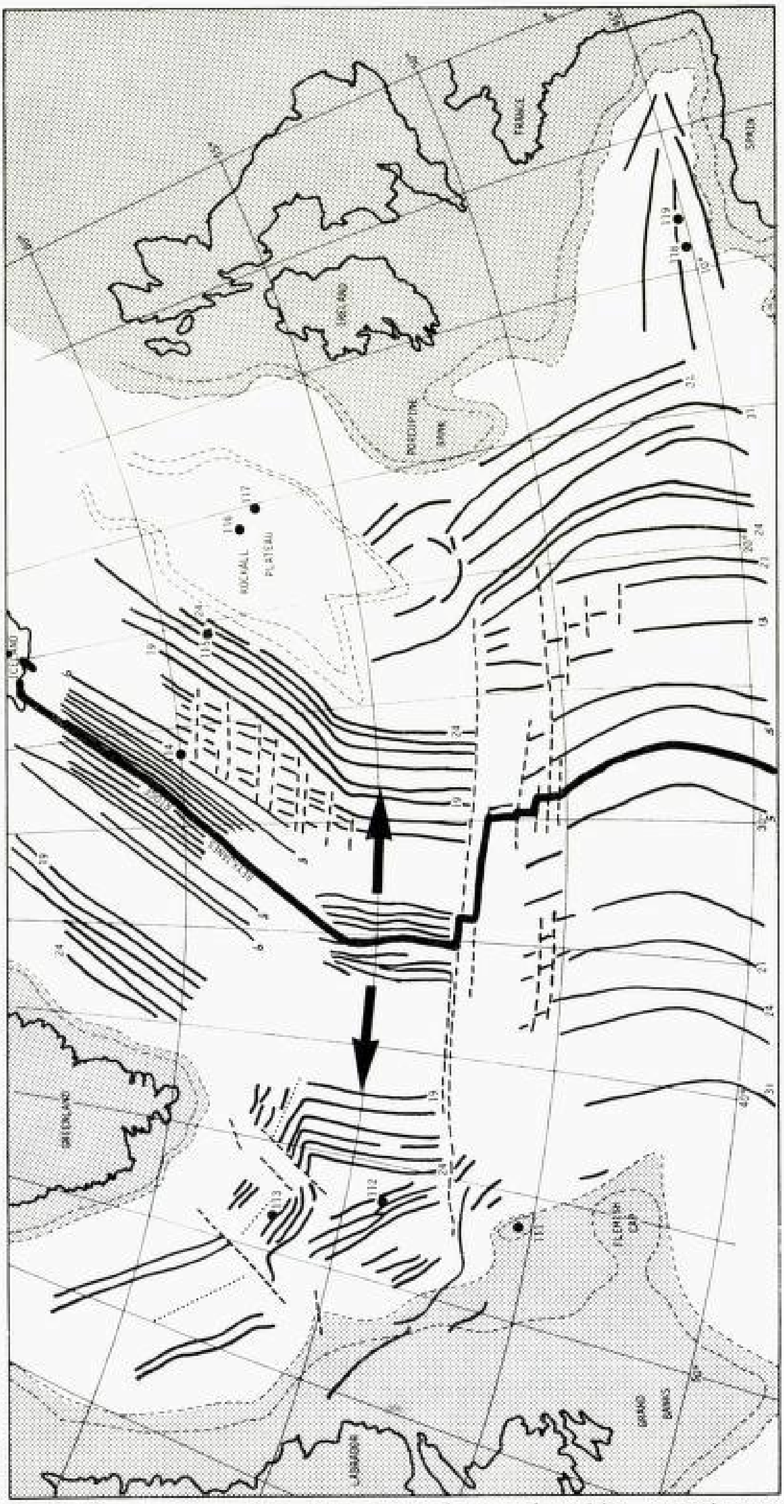

Fig. 8. Present. Heavy lines show principal mapped positive magnetic anomalies with identifications. Heavy dashed lines are fracture zones. Dotted lines are extinct spreading axes. Thin dashed lines are top and bottom of continental slope. Black dots are drill sites.

the western part of the island. Piper (1973) believes that the western zone ultimately became extinct about 5 my ago. Iceland is clearly the site of excessive lava production from the asthenosphere, several times that of the mid-Atlantic Ridge, and may be the surface expression of a mantle plume or hot spot, that has given rise also to the anomalous Iceland–Faeroes Ridge (Bott 1973). The mode of crustal spreading and dyke injection is not, however, the same as that in an oceanic ridge, being spread over a wider area and thus the magnetic anomaly pattern is not so linear (Gibson & Piper 1972).

Between Iceland and Jan Mayen Island, a second jump of the spreading axis to the west may have occurred 10 my ago (Johnson et al. 1972) to bring it to its present position. A detailed survey reported by Meyer et al. (1972) detected that 3 my ago a minor perturbation in the spreading axis created the Spar Fracture Zone.

Throughout the Neogene, the spreading rates have generally increased. At 50°N, the rate increased from 0.7 to 1.4 cm/year (Johnson & Vogt 1973). Similar increases of rate are found on the Reykjanes Ridge (Vogt et al. 1970) and in the Norwegian Sea (Phillips 1973).

Seismic activity continues today along the axis of the mid-ocean ridge system and along the transform faults. The mechanisms of emplacement of the new oceanic lithosphere and of the tectonic forces which in places uplift the ridge either side of the injection axis leaving a median valley and in other places building only a ridge (e.g. Reykjanes Ridge) are not yet fully understood. Nor are the forces understood which are driving the plates and which periodically change to give rise to a different pattern of spreading. However, the existence of sea floor spreading and the related kinematics are now established beyond doubt and must be the basis for understanding oceanic evolution.

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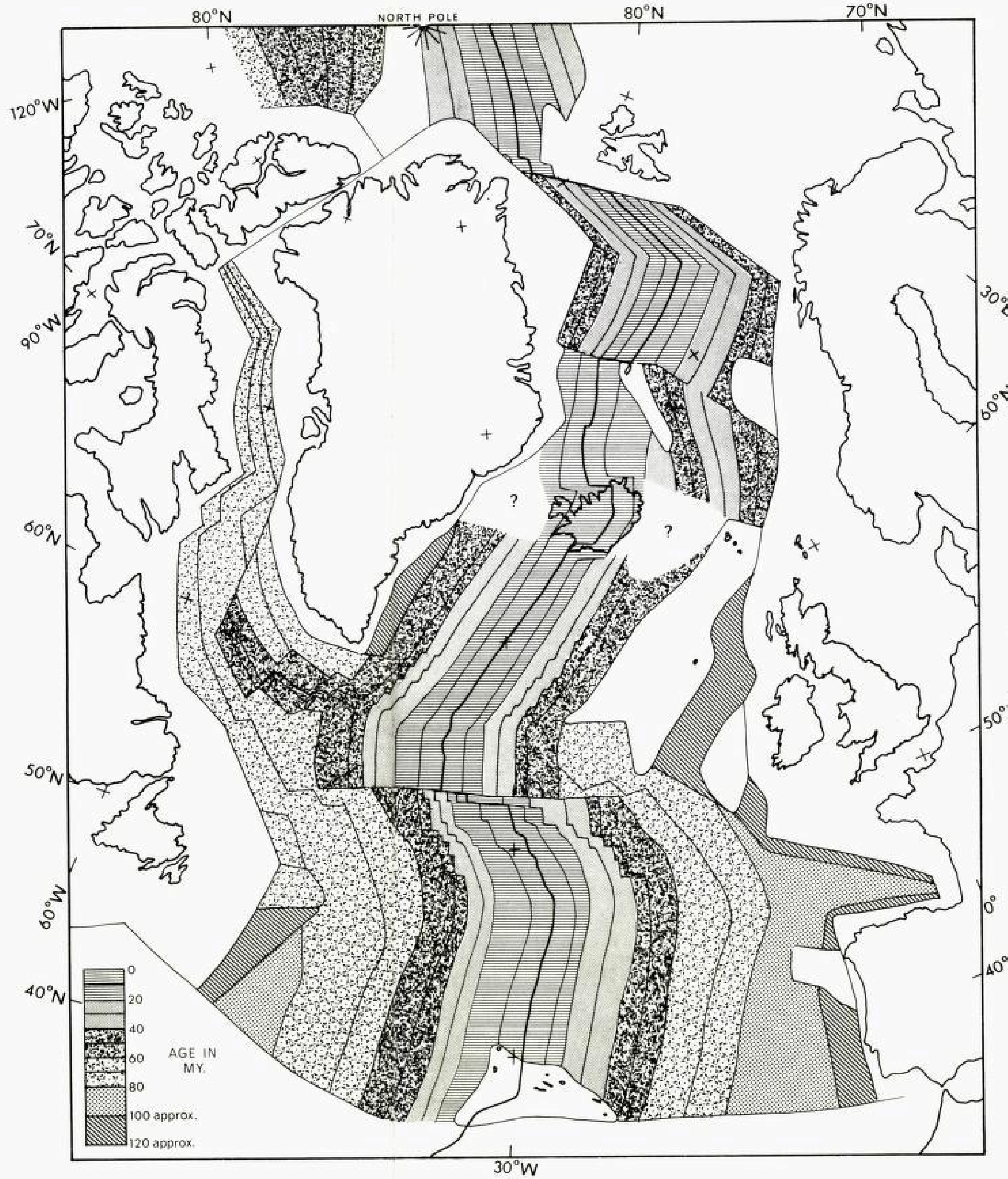

Plate 2. Isochrons of the age of oceanic lithosphere derived from magnetic anomaly surveys, and plate tectonic reconstructions.

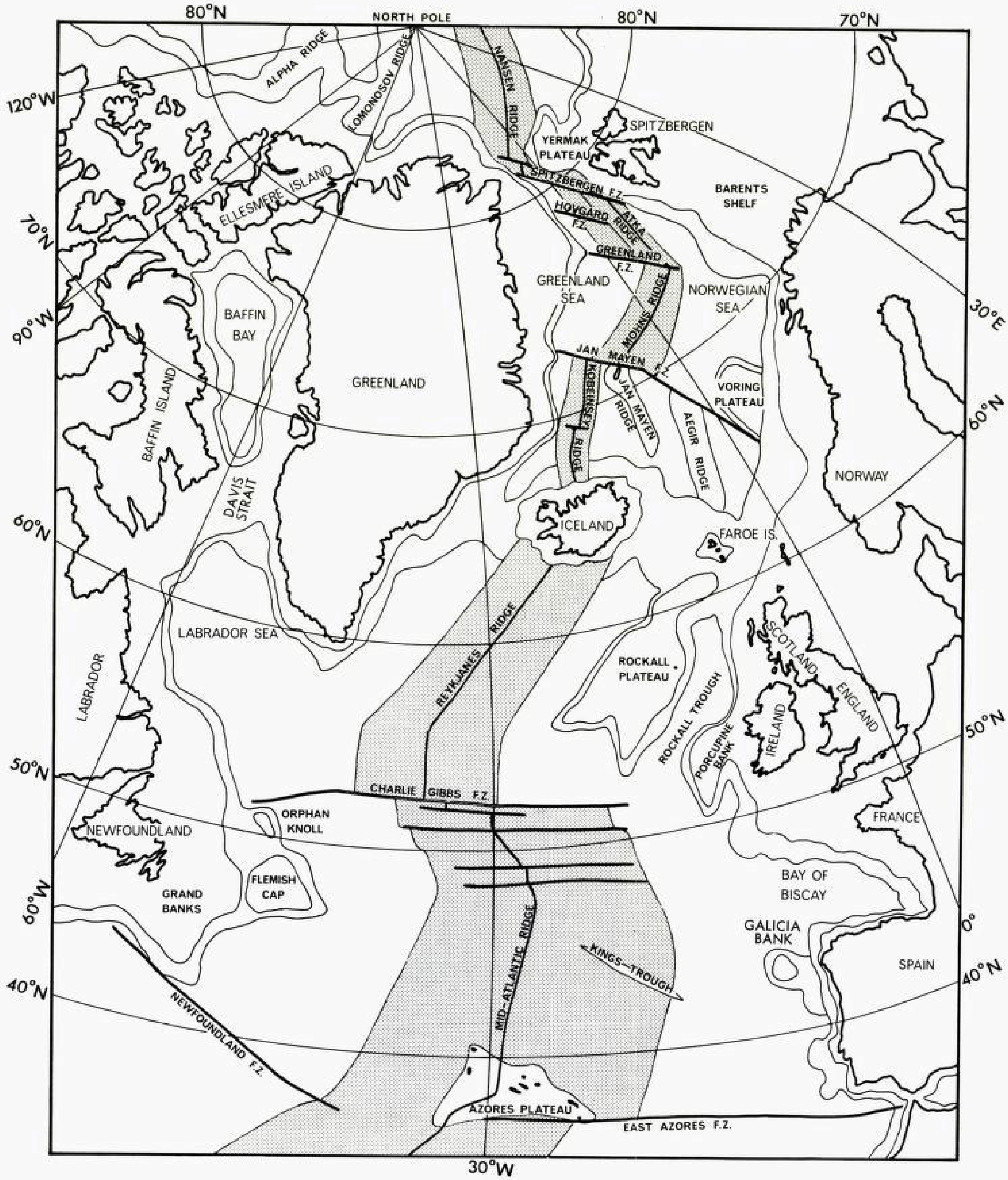

Plate 1. Principle physiographic features of the sea floor in the northeast Atlantic. The two lines bordering the continents represent the top and bottom of the continental slope. The mid-ocean ridge system is shaded. (The projection, which is approximately equal area, has been taken from Dietrich & Ulrich 1968).

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Structure and Evolution of the Atlantic Floor between Northern Scotland and Iceland

M. H. P. BOTT

Bott, M. H. P. 1975: Structure and evolution of the Atlantic floor between northern Scotland and Iceland. *Norges geol. Unders. 316*, 195–199.

The continental shelf west of the Shetland and Orkney Islands is a region of varied geology traversed by the Caledonian front, including some areas of shallow basement and some partially fault-bounded Mesozoic basins such as the west Shetland basin; Tertiary strata appear to be absent except on the slope and near it. At the other extremity of the region, the Iceland–Faeroe Ridge is formed by anomalously thick 'Icelandic-type' oceanic crust probably originating between 60 and 45 m.y. ago. Recent evidence confirms that the shelf region around the Faeroe Islands is underlain by continental crust, but the origin of the Faeroe–Shetland Channel is problematical.

M. H. P. Bott, Department of Geological Sciences, University of Durham, England

Introduction

This paper briefly summarises the deep structure of the region between the North Scottish continental shelf and the Iceland–Faeroe Ridge (Fig. 1). The region forms the southeastern end of the highly anomalous strip of the North Atlantic extending from the Denmark Strait through Iceland to North Scotland, and can be subdivided into the following four main structural units: the North Scottish continental shelf, the Faeroe–Shetland Channel, the Faeroe Block and the Iceland–Faeroe Ridge.

Continental shelf west of the Shetland and Orkney Islands

The aeromagnetic map of the Institute of Geological Sciences, London, was the first geophysical evidence on the deep structure of this shelf region. A qualitative interpretation of the depth and character of the basement using this map was made by Flinn (1969). This was soon followed by a regional marine geophysical investigation of the area by Durham University, using gravity, magnetic, shallow seismic reflection and seismic refraction methods (Bott & Watts 1970, 1971; Watts 1971; Browitt 1972; Bott & Browitt 1975). These early investigations led to a broad understanding of the deep geological structure of the region and outlined the regions of thick sediments.

The basement rocks of this shelf are cut by the Caledonian front, which would be expected to separate the foreland Lewisian and Torridonian basement rocks to the west from the Moinian and Dalradian Caledonian rocks to the east. The Great Glen fault may traverse the eastern extremity of the region, possibly occurring in the Shetland Islands along the line of the Walls boundary fault (Flinn 1961). Old Red Sandstone rocks probably occur

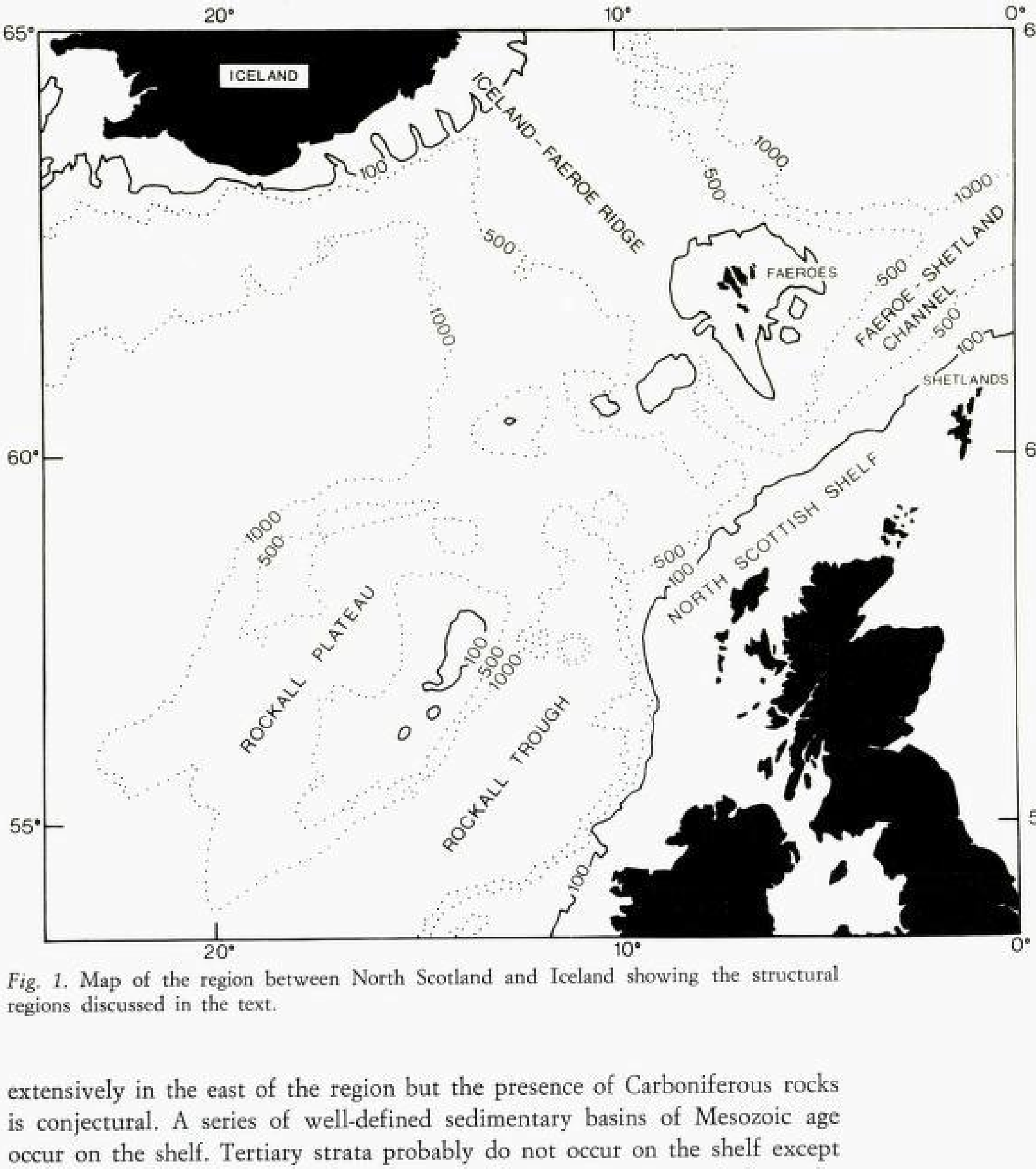

Fig. 1. Map of the region between North Scotland and Iceland showing the structural regions discussed in the text.

extensively in the east of the region but the presence of Carboniferous rocks is conjectural. A series of well-defined sedimentary basins of Mesozoic age occur on the shelf. Tertiary strata probably do not occur on the shelf except near the slope and on it, where a prograding sequence extends into the Shetland–Faeroe trough.

A conspicuous Bouguer anomaly high reaching 94 mgal occurs west of the Shetland Islands. It is about 250 km long and has a NNE strike which gently transgresses the margin. With support from sparker and magnetic observations, this has been interpreted as a region of shallow or outcropping Lewisian rocks of high density, probably of granulites. Other regions of shallow or outcropping basement rocks further south are recognised by their high local gravity and short-wavelength magnetic anomalies.

The Shetland–Orkney shelf is notable for a series of fairly deep and local sedimentary basins which are probably mainly filled by Mesozoic sediments. These are clearly visible on the Bouguer anomaly map, and characteristically steep gravity gradients indicate that they are partially fault-bounded, particularly along their eastern margins. The largest and most spectacular of these is the west Shetland basin, which runs along the west side of the abovementioned gravity high, being separated from the shallow basement rocks by a normal fault or fault zone. This basin causes Bouguer anomalies of about 70 mgal below those of the adjacent high, and the intervening gradient attributed to the fault is particularly steep. A ridge of basement uplift occurs along the west side of the west Shetland basin. According to gravity and seismic refraction observations, the basement is about 6 km deep beneath the basin, but the deeper sedimentary rocks within it are possibly of pre-Mesozoic age.

The Iceland—Faeroe Ridge

The Iceland–Faeroe Ridge occurs at the other extremity of our region. It is an aseismic ridge of relatively smooth bathymetry and north-westerly trend which joins Iceland and the shelf region around the Faeroe Islands. The depth to the crest is about 400 m along the length, and the ridge is separated from the Icelandic and Faeroe shelf regions by small but steep bathymetric scarps. The deep structure and affinity of this problematical quasi-continental feature have been clarified in recent years by geophysical investigations (Bott et al. 1971; Fleischer 1971; Johnson & Tanner 1972).

It is widely believed that the north-eastern branch of the North Atlantic formed over the last 65 m.y. by ocean-floor spreading as Greenland separated from North Europe. If correct, this implies that the Iceland–Faeroe Ridge also formed by the ocean-floor spreading mechanism, probably over the period 60–45 m.y. ago, despite its shallow bathymetry. Seismic investigations indicate that the Ridge is underlain by a type of crust similar to that beneath Iceland, although of greater thickness. Such thick crust is probably produced as a result of intense differentiation from the underlying mantle because of its local high temperature (hot spot theory).

Sediments are generally thin or absent from the crestal region of the Iceland–Faeroe Ridge, except for local pockets, some of which are probably volcanic in origin. The upper crust extends to a depth of about 8 km and it is formed of highly magnetic igneous rocks. It is unlikely that significant hydrocarbon deposits will be found here.

The Faeroe Block

The region of relatively shallow shelf bathymetry surrounding the Faeroe Islands is referred to as the Faeroe Block. The Faeroe Islands themselves are formed by nearly horizontal basaltic lavas of early Tertiary age. The nature of the underlying crustal structure has been a subject of controversy, but has recently been clarified.

Early seismic refraction lines reaching the basement beneath the lavas (Pálmason 1965) suggested a similar upper crustal structure to that of Iceland, indicating oceanic rather than continental affinities. More recently, this conclusion has been reversed. The observed gravity gradient between the Iceland–Faeroe Ridge and the Faeroe Block indicated that these regions are underlain by different types of crust (Bott et al. 1971) and a pre-Tertiary continental reconstruction including the Rockall Plateau indicates that the Faeroe Block itself is probably continental. The most definite evidence comes from a recent crustal seismic project which indicates absence of the characteristics 6.4–6.7 km s^{-1} layer of Iceland and the Ridge, and a crustal thickness of over about 30 km (Bott et al. 1974). Thus continental crust is now believed to underlie the Faeroe Block.

Although some pre-Tertiary sediments may be sandwiched between the Faeroe lavas and the basement, the geophysical evidence indicates that they are unlikely to be more than about 1 km thick. Some thicker sediments may occur on the Faeroes shelf, particularly in the east.

The Faeroe—Shetland Channel

The Faeroe–Shetland Channel forms a north-easterly extension of the Rockall Trough, being separated from it by the transverse Wyville–Thomson Ridge. It is about 1200 m deep and opens at its north-eastern end into the Norwegian Sea. Flat-lying sediments overlie the magnetic basement and these generally thicken towards the north-east. Gravity investigations (Bott & Watts 1971) indicate that the Channel is underlain by slightly thinner crust than that beneath the adjacent continental regions of the North Scottish shelf and Faeroe Block.

The origin of the Channel is problematical. One view, taken for example by Talwani & Eldholm (1972), is that it was formed by subsidence of continental crust. The alternative view, which the author favours, is that the Channel originated as oceanic crust during a preliminary stage of opening of the Atlantic in the Mesozoic when the Rockall Trough also formed. If such a preliminary opening occurred, then one would expect it to extend further north for a considerable distance; a suggestion is that this extension may form the trough of thick sediments beneath the eastern part of the Vøring Plateau, which has been mapped by Talwani & Eldholm (1972).

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Results of Geophysical Surveys in the Area of the Aegir Ridge, the Iceland Plateau and the Kolbeinsey Ridge

K. HINZ

Hinz, K. 1975: Results of geophysical surveys in the area of the Aegir Ridge, the Iceland Plateau and the Kolbeinsey ridge. *Norges geol. Unders. 316*, 201–203.

K. Hinz, Bundesanstalt für Bodenforschung, 3 Hannover 23, Postfach 23 01 53, W. Germany

It is today accepted that after break-up and separation of the European and Greenland blocks, expansion caused by production of oceanic lithosphere at the Mid-Atlantic ridge proceeded, developing the present Norwegian and Greenland Seas. The asymmetrical location of the present active Mid-Atlantic ridge is explained by a westward relocation of the ridge producing oceanic lithosphere through Tertiary time.

In 1969, 1971 and 1972 geophysical investigations of the German Hydrographic Institute and the Bundesanstalt für Bodenforschung (Fig. 1) were carried out by *R. V. Planet*, *Komet* and *Meteor* partly together with the Seismological Institute of Bergen and the Norwegian R.V. *Nordskap.* Some of these results will be presented in this report.

From refraction seismic measurements it can be assumed that the crust within the 200 km wide *Aegir ridge* consists of blocks of oceanic crust as encountered near the Jan Mayen ridge (Hinz & Moe 1971; Hinz 1972) interspersed with ultrabasic sections. If this is true it would mean that the zone of the Aegir ridge represents an important rift zone in addition to the Mid-Atlantic ridge.

The *Iceland plateau* is divided by a north-south trending escarpment into an upper and lower plateau. Within the upper Iceland plateau up to 300 m of sediments overlie the completely flat, presumably basaltic basement. The thickness of the unconsolidated sediments increases to the south. Magnetically the zone west of the escarpment is characterized by anomalies with amplitudes of about 1000 n.T. and more. At 16°W the character of the anomalies changes abruptly. East of the escarpment follows a zone with a flat-lying basement and a thin sedimentary cover. Further to the east the acoustic basement steepens eastwards and suddenly has a considerable relief. The sediments thicken to the east, reaching a thickness of up to 2000 m in the area of the southern part of the Jan Mayen ridge. The area east of 16°W has lower magnetic amplitude anomalies (100 n.T.–400 n.T.). The anomalies show no preferred trend.

Within the investigated part of the Jan Mayen ridge the upper sediments have nearly horizontal layering and seem to fill in an older relief. This stratigraphic relationship points to an important regression and erosion phase

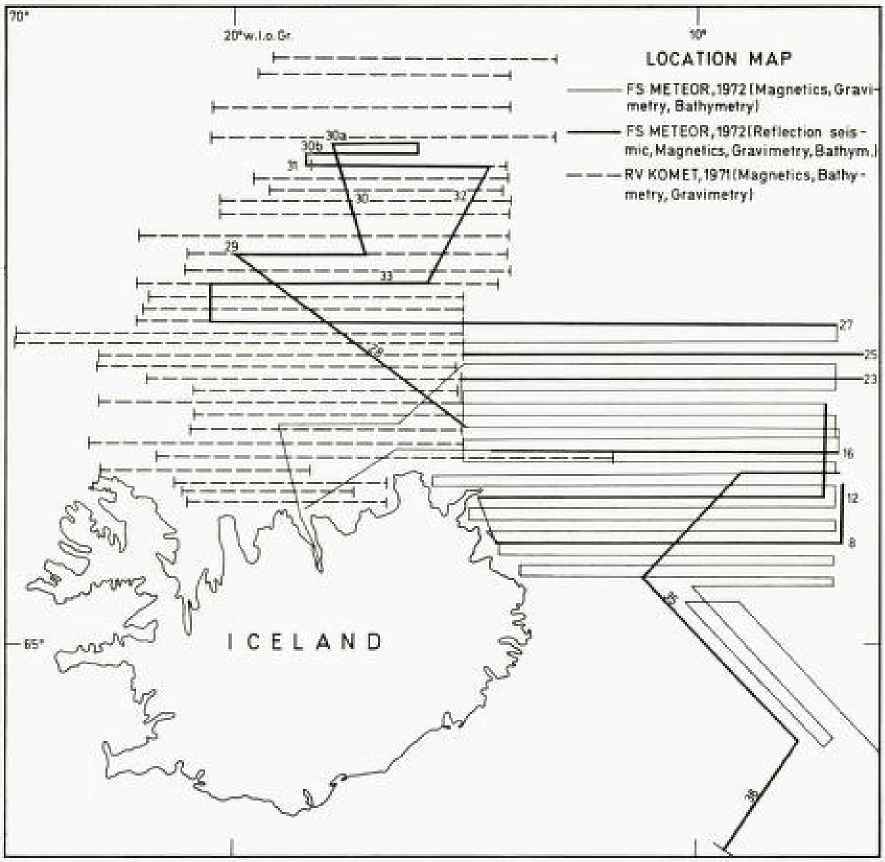

Fig. 1. Location map showing the traverses of the FS. Meteor and RV. Komet.

caused by eustatic changes of sea level and/or rise of the Jan Mayen ridge. The seismic records indicate that after a period of erosion/regression and flow basalt activity the present lower Iceland plateau foundered.

Directly west of the Iceland plateau lies the *Kolbeinsey ridge*. In general, the Kolbeinsey ridge crest is almost devoid of sediments. The sediment thickness generally increases towards the flanks and is thicker on the western flank than on the eastern flank. Sediments more than 200 m thick were found mainly within the area of the Spar fracture zone and locally in the rift valley. The results of the magnetic-gravimetric survey have been published by Meyer et al. (1972).

From the geophysical data it seems that spreading and production of oceanic lithosphere have not occurred only along a single ridge at one particular geologic time. The bifurcation of the Kolbeinsey ridge (Meyer et al. 1972), the refraction seismic results from the Aegir ridge (Hinz & Moe 1971; Hinz 1972), and the thinned crust within the Vøring plateau (Hinz 1972) favour the idea that in the Norwegian Sea spreading and production of oceanic lithosphere sometimes occurred along several tectonic zones.

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Gravel Fraction on the Spitsbergen Bank, NW Barents Shelf*

MARC B. EDWARDS

Edwards, M. B. 1975: Gravel fraction on the Spitsbergen Bank, NW Barents Shelf. *Norges geol. Unders. 316*, 205–217.

The gravel fraction of the bottom sediment on the Spitsbergen Bank, located south of Svalbard in the NW Barents Sea, is composed predominantly of clastic sedimentary rocks, especially sandstone and shale. Between sampling stations the proportion of eight lithologic types is markedly different, ruling out the possibility of large-scale transport of the gravel by ice-rafting. Striated pebbles occur in small numbers: a few are exotic in composition, but most are similar to a non-striated rock-type at a given station. This suggests that the gravel was formed by reworking of previously deposited glacial material, which tends to be locally derived.

The sandstone pebbles in the gravel include a variety of petrographic types, most of which are identical or similar to sandstones of known stratigraphic position on Svalbard. Observations on Mesozoic rocks on Bjørnøya, Hopen and southern Spitsbergen, and on the distribution of pebbles as described herein, suggest that the Spitsbergen Bank is underlain by nearly flat-lying Mesozoic sedimentary rocks similar to those known on Svalbard. The overall structure is a gentle syncline; a southward continuation of the dominant regional structure of Spitsbergen.

Marc B. Edwards, Continental Shelf Division, Royal Norwegian Council for Scientific and Industrial Research; Norsk Polarinstitutt, N–1330 Oslo Lufthavn, Norway

Introduction

The Spitsbergen Bank is approximately 200 by 350 km, elongated NE–SW, lying south of Svalbard in the NW part of the Barents Sea (Fig. 1). It includes Bjørnøya (Bear Island) at its SW end, and Hopen at its NE end. To the north it is separated from Spitsbergen by the Storfjord Trough, and to the south is the Bjørnøya Trough. The surface of the bank varies from rough to smooth, and generally lies 40–100 m below sea-level, though highs of 17 m occur in one area. The bank is covered by gravel, sand and shell material, with some large stones (cobbles and boulders) and clay. Bedrock crops out near Hopen. The Storfjord Trough reaches depths of over 300 m, and is covered by clay, locally with sand and gravel.

The present knowledge of the geology and superficial deposits of the Barents Sea is based mainly on work carried out by Soviet scientists, summarised in English by Klenova (1966) and Emelyanov et al. (1971). Dibner et al. (1970) described Lower Carboniferous to Late Cretaceous/Paleocene fossiliferous limestones dredged from 21 stations in the Barents Sea. As Soviet workers (op. cit.) maintain that the bottom sediment is locally derived, as opposed

* Publication No. 46 in NTNF's Continental Shelf Project

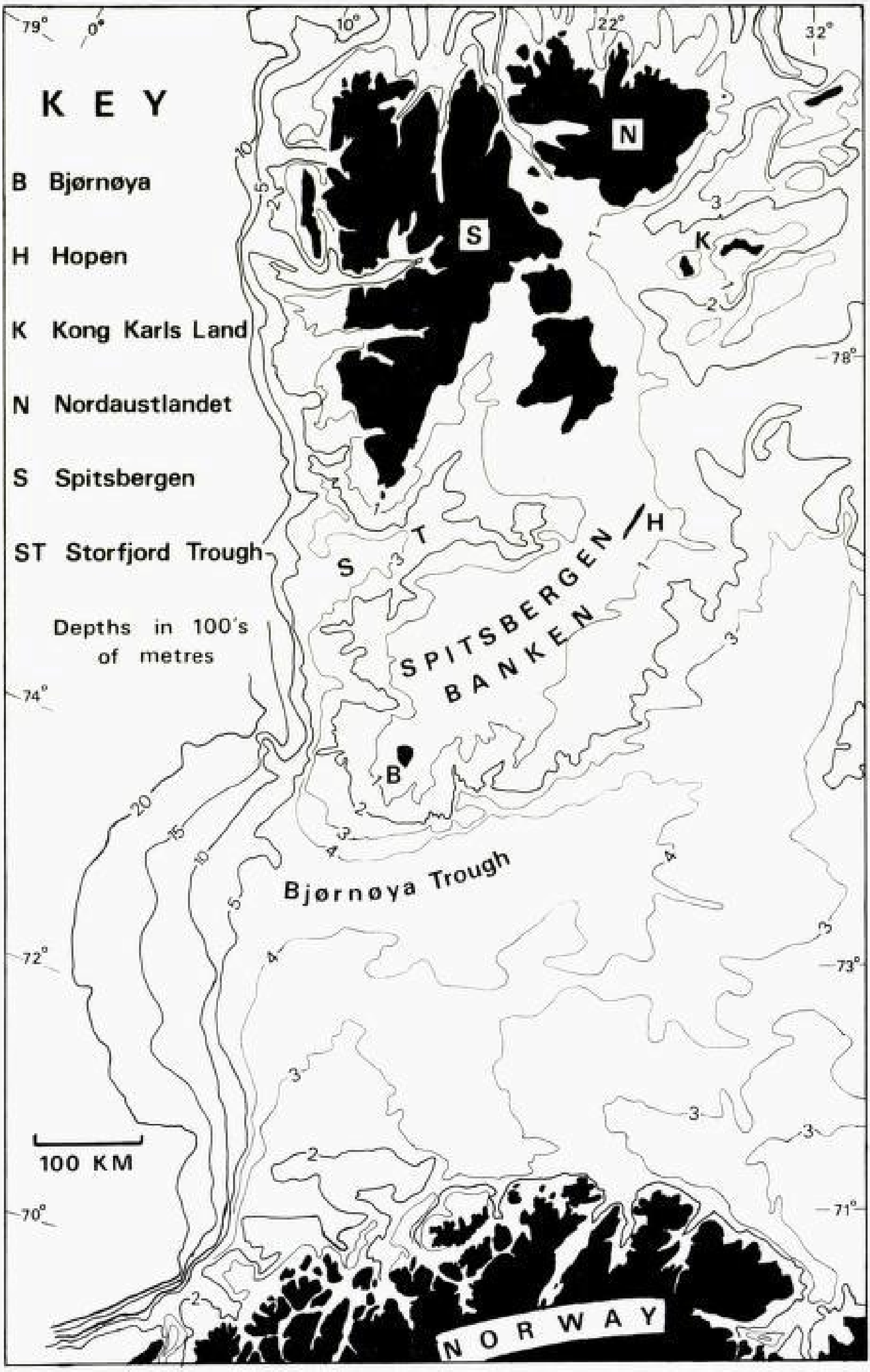

Fig. 1. Spitsbergen Bank and surrounding area.

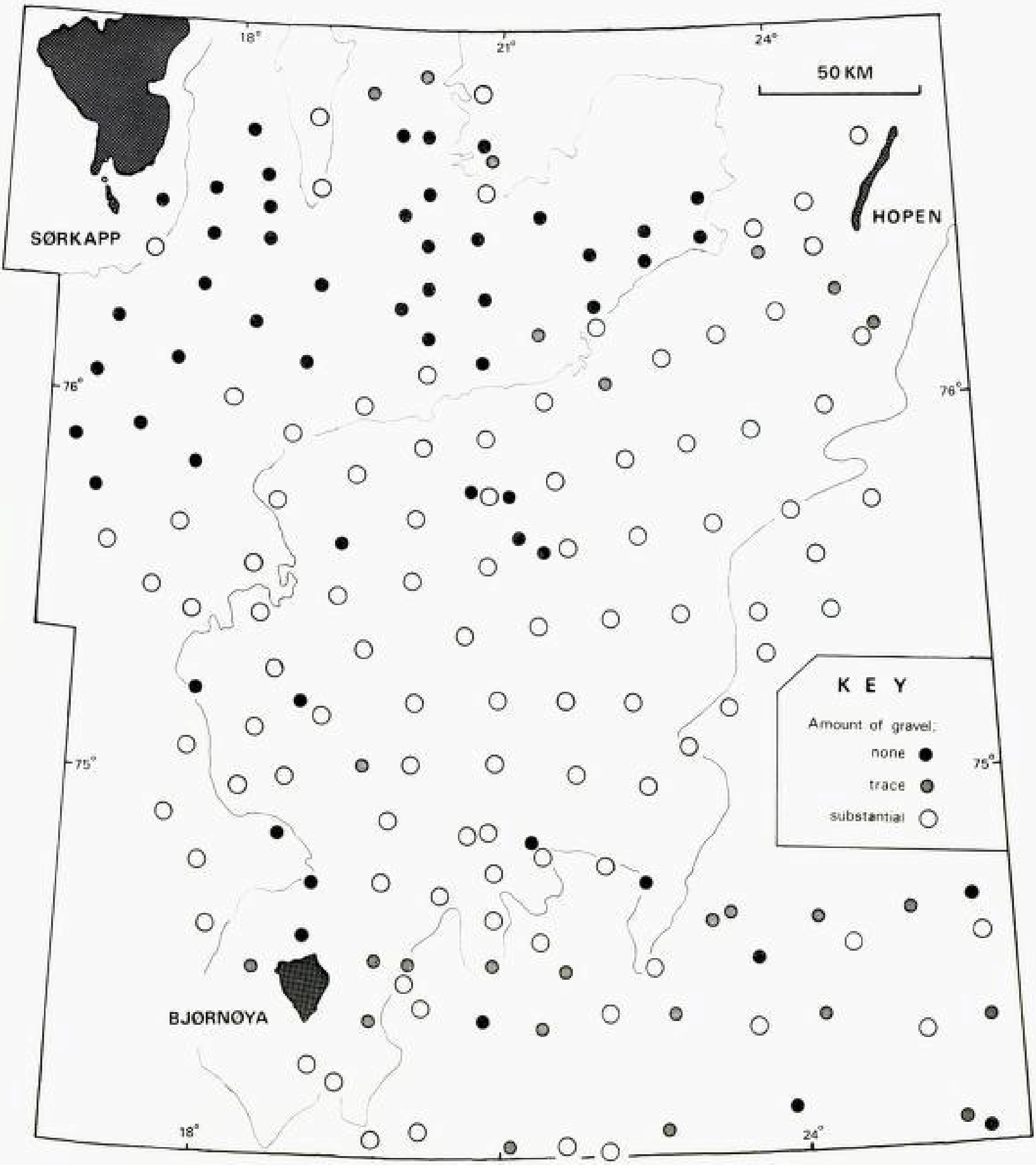

Fig. 2. Location of sampling stations, with relative abundance of gravel. Hundred metre contour shown.

to ice-rafted, glacial material, Dibner et al. concluded that the Spitsbergen Bank is underlain mostly by Carboniferous to Permian bedrock, with Mesozioc rock in the E and SE part of the Bank.

Spjeldnæs (1971), in a short note, described a siltstone slab collected from the central part of the Barents Sea, which he interpreted as Triassic and of local origin.

In response to the growing need for scientific information with regard to locating areas of oil potential, the Continental Shelf Division of the Royal Norwegian Council for Scientific and Industrial Research (NTNFK) initiated the Barents Sea Project, based on a proposal by the Norwegian Polar Institute in 1969.

In 1971, bottom samples were recovered by the use of large and small grabs, dredges, gravity cores, and diving. Bottom photographs were taken and continuous seismic profiling was carried out. The gravel fraction was picked out of the sample, and described lithologically within arbitrary size classes. In a few cases, only some of the pebbles were retained from the sample. About 90 of the 177 sampling stations contain some gravel (Fig. 2).

Methods

In a preliminary study, 10 sandstone pebbles were selected from each of eight stations over a wide area to determine the degree of variation in sandstones both within and between stations. Striated pebbles were also picked out. For the detailed study, 22 stations covering most of the area were selected (one subsequently discarded), each with relatively large quantities of gravel. The samples were treated in the following way: material smaller than 2 cm was removed and set aside. The remainder was divided into eight lithological classes, or 'types', of which all but one are sedimentary rocks. Each type was further subdivided into striated and non-striated components. Groups of pebbles were then weighed and converted to per cent of the type at a particular station. Taking all the stations together, sandstones make up 56% of the gravel fraction, shales 17%, siltstone and shale 11%, siltstone 9%, and the remaining four types 7%. Other lithologies, such as flint, compose less than one per cent at any given station, and so were not included in the study.

The characteristics of the eight lithological types are:

1) Sandstone: usually very fine-grained to coarse-grained, mostly well indurated, grey to greenish grey, with occasional brownish weathering rims. Green sandstones with abundant glauconite occur at some stations. Calcite cement is occasionally present; the amount of matrix is generally low.

2) Shale: dark grey to black, somewhat fissile mudstones with high clay content and occasionally fossiliferous. Small mica flakes may be present.

3) Siltstone and Shale: pebbles with thin alternations of dark grey shale and grey siltstone (sometimes grading into very fine sandstone) with marked contrast in grain size; bioturbation is abundant.

4) Siltstone: mainly grey to dark grey massive or faintly laminated siltstone, occasionally with a high clay content.

5) Clay Ironstone: dense claystones with a characteristic dark purple surface colour, and often with a brown-yellow weathering rim several millimetres thick. Unoxidized claystone is medium to dark grey.

6) Limestones: light coloured, massive, fine-grained limestones, oolitic limestones, and dark, coarse-grained fossiliferous limestones.

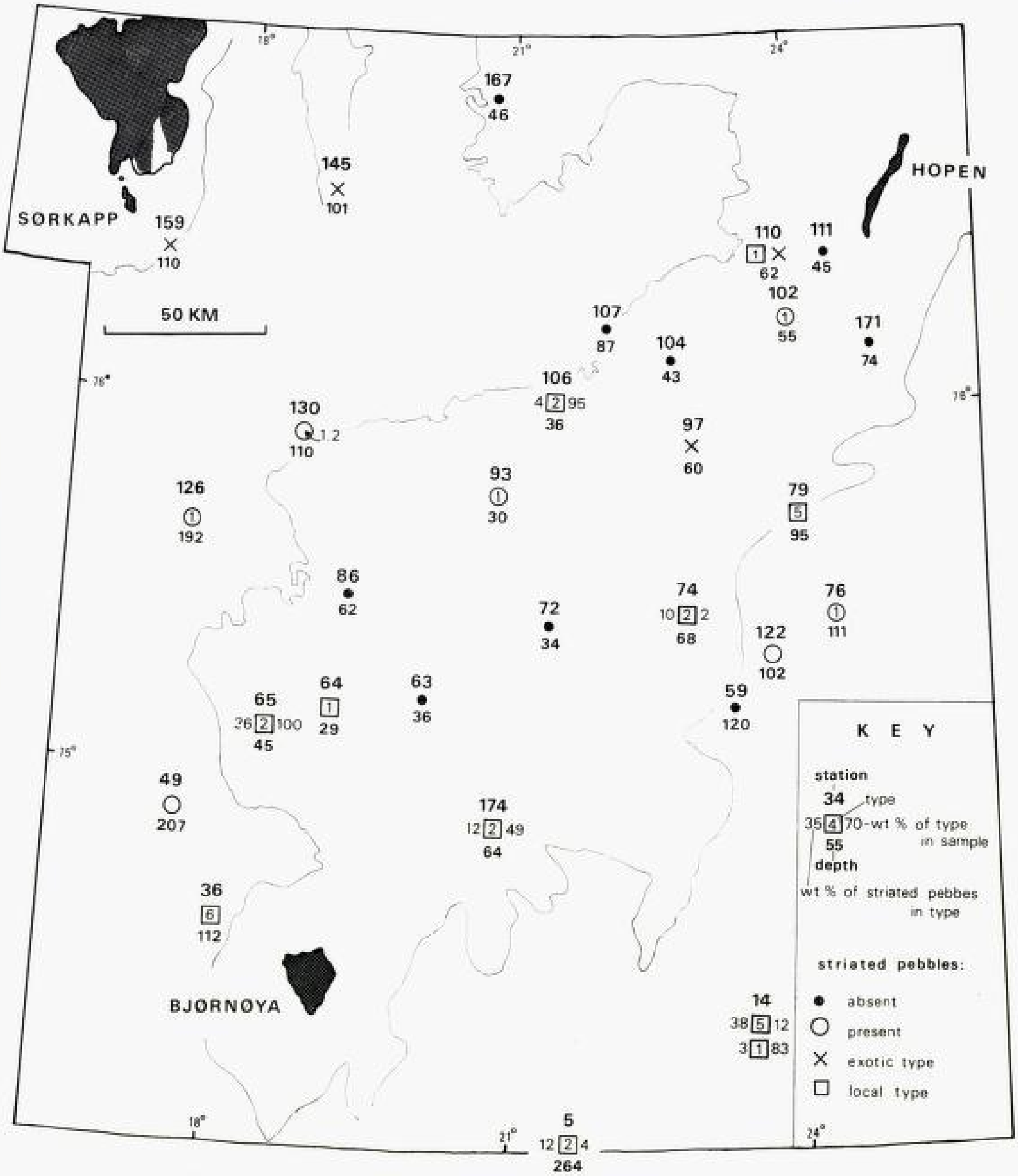

Fig. 3. Occurrence of striated pebbles. The lithologic type numbers are explained in the text. Hundred metre contour shown.

7) Crystalline Rocks: dark diabase, and coarse, reddish and grey granite and slightly foliated gneiss.

8) Marl: light grey, lime-rich claystone.

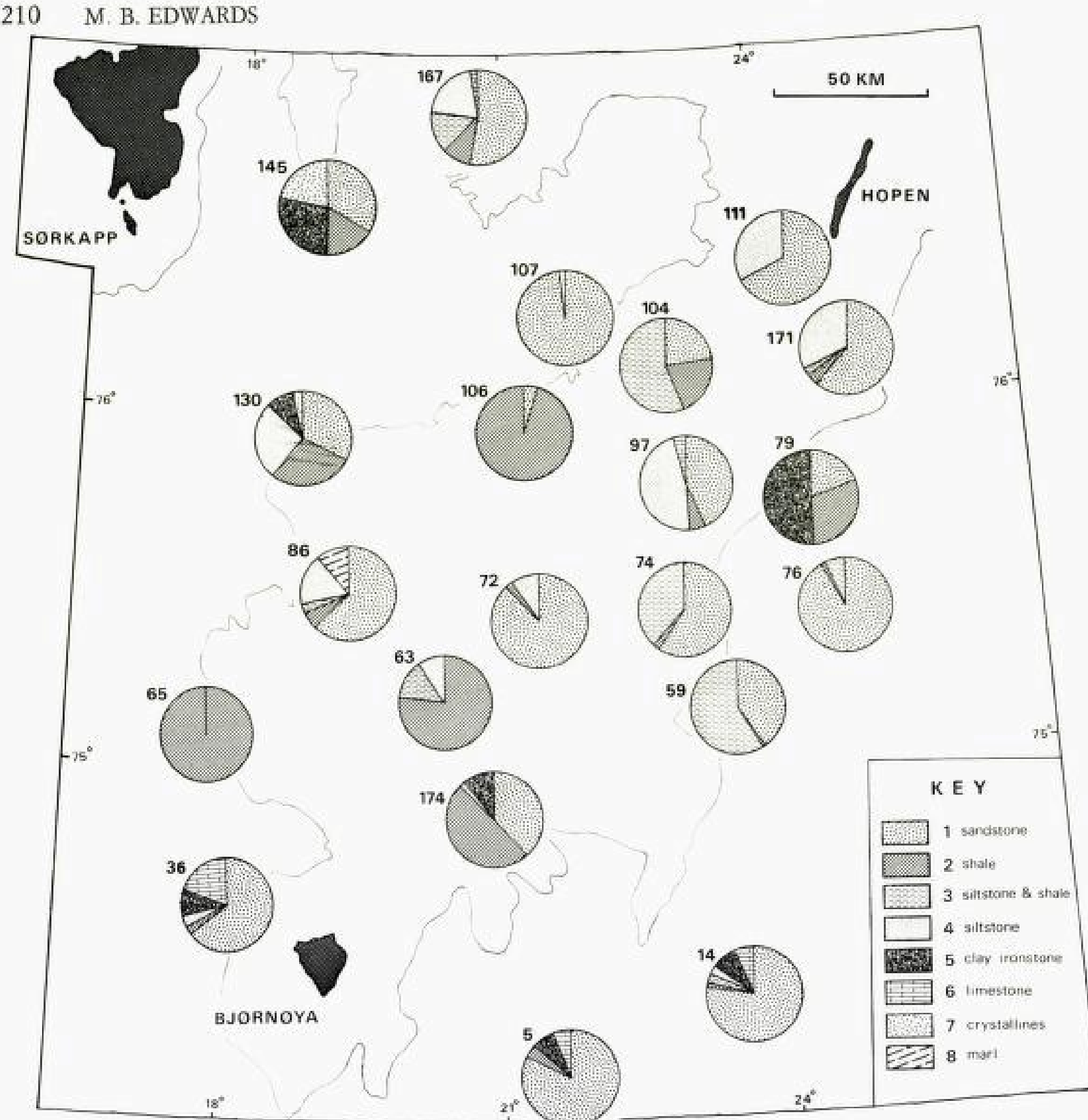

Fig. 4. Distribution and abundance of lithologic types. Pie diagrams show the weight per cent by lithologic type of non-striated pebbles at 21 stations (station number is given with each sample), as discussed in the text. Hundred metre contour shown.

Striated pebbles

Striated pebbles, found in 18 of 28 stations, are broadly distributed (Fig. 3). In 13 of the 22 stations in the detailed study, the striated pebbles are similar in composition to a substantial part of the non-striated pebbles in a given sample.

Lithological type

The distribution of the eight lithological types is shown in Fig. 4. The large differences in lithology between adjacent stations is striking. In general, either sandstone or shale is the predominant lithology at a given station, but siltstone and shale, and clay ironstone are locally abundant.

Sandstone petrography

Sandstones in the gravel material vary greatly in appearance, both in hand specimen and in thin-section. While some stations contain a homogeneous assemblage of sandstone pebbles, other stations may have several petrographic types mixed together. Rather than studying exclusively the gravel material, it was felt that more significant findings would result from a comparison between the sandstones in the gravel and those on the neighbouring land areas. A reconnaissance of the petrography of Upper Paleozoic and Mesozoic sandstones of Svalbard suggested that the sandstones in the gravel are similar to Mesozoic types. Accordingly, during field work in south Spitsbergen in 1973, close sampling of Mesozoic sandstones was carried out. Fig. 5 shows results of petrographic observations on these sandstones, which range from the base of the Sassendalen Group (Triassic) to the middle part of the Helvetiafjellet Formation (L. Cretaceous), and which are supplemented by material made available by other workers from other parts of Svalbard. These observations, discussed below, form the basis for study of the sandstones in the gravel.

The most easily distinguished sandstones in the sequence (Fig. 5) are those in the Kapp Toscana Group, particularly the upper, sandstone-rich De Geerdalen Formation. These sandstones always contain large quantities of feldspar and rock fragments, and the quartz content may be less than 50%. Volcanic rock fragments (an important indicator type), chert and polycrystalline quartz are abundant. The fine grain-size of the underlying Sassendalen Group (in the material collected from Sørkappland) prohibits the identification of rock fragments, but the feldspar content is similar to that of the Kapp Toscana Group. The overlying Wilhelmøya Formation (Worsley 1973) contains less feldspar and rock fragments than the Kapp Toscana Group, but the grain-size is comparable.

Following a thick sequence of shales, sandstones are again developed, locally, in the upper part of the Janusfjellet Sub-group and in the overlying sediments. The lower part of the Helvetiafjellet Formation, the Festningen Sandstone Member, consists predominantly of medium to coarse, sometimes very coarse and conglomeratic sandstone. Similar sandstones also occur locally, intercalated with other lithologies, in the upper part of the Janusfjellet Sub-group, and in the upper part of the Helvetiafjellet Formation. The somewhat coarser grain-size and lower feldspar content generally distinguished these sandstones from those of the Wilhelmøya Formation.

Fine-grained, feldspathic sandstones appear in the upper part of the Helvetiafjellet Formation, and are present throughout the Carolinefjellet Formation.

The observations in the stratigraphic column (Fig. 5) show that 1) heterogeneous sandstone assemblages in the gravel can be explained by the

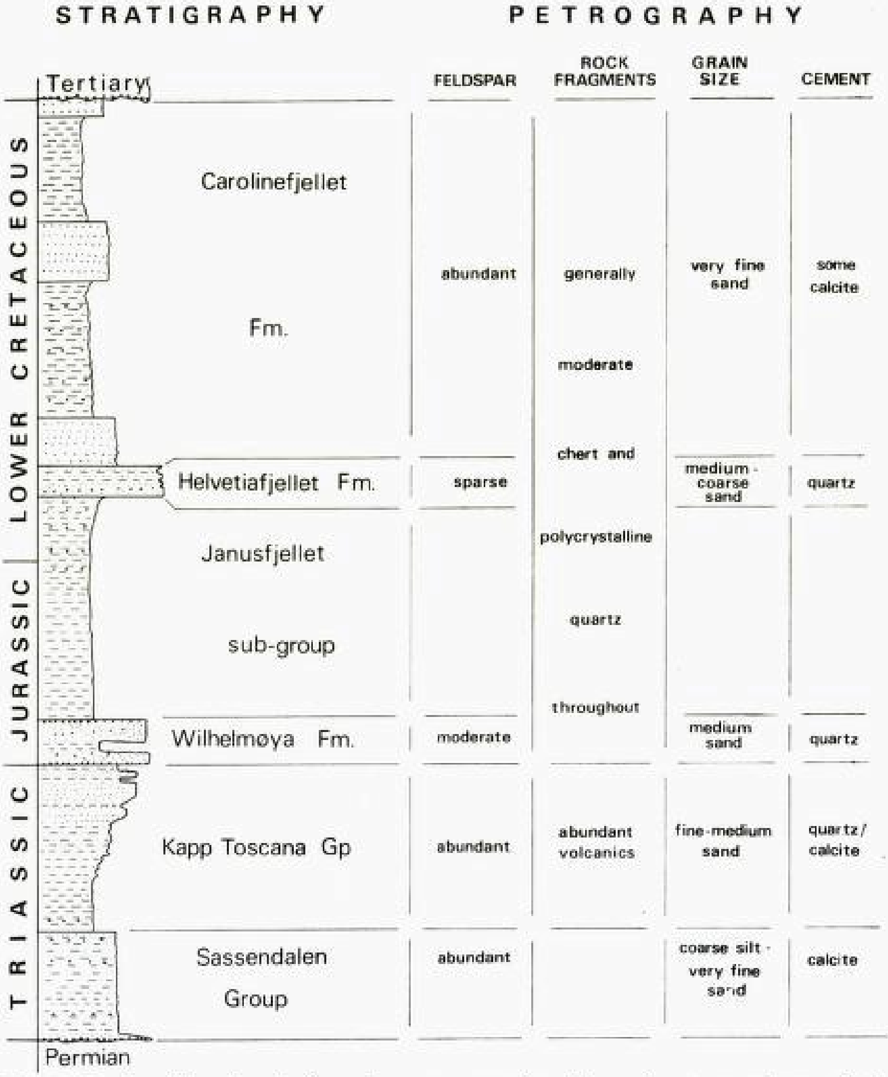

Fig. 5. Stratigraphic résumé of sandstone petrography. The sedimentary column, about 2,000 m thick, is generalised from parts of the Mesozoic in different areas of Svalbard. Descriptive terms indicate characteristic or typical features, and are not necessarily inclusive of all the sandstones in a given unit. Stratigraphic nomenclature is taken from Buchan et al. (1965), Flood et al. (1971a), and Parker (1967).

occurrence of contrasting types of sandstone in adjacent parts of the succession, and 2) the stratigraphic position of a large portion of the sandstone pebbles in the gravel can be determined. The two key indicator types are the sandstones in the Kapp Toscana Group and those in the Helvetiafjellet Formation. The occasionally low feldspar content of the sandstones in the Wilhelmøya Formation may allow confusion with those in the Helvetiafjellet Formation. Similarly, the very fine sandstones in the Sassendalen Group and the Carolinefjellet Formation may lead to confusion between these two units. More data, especially quantitative observations, could reveal significant differences

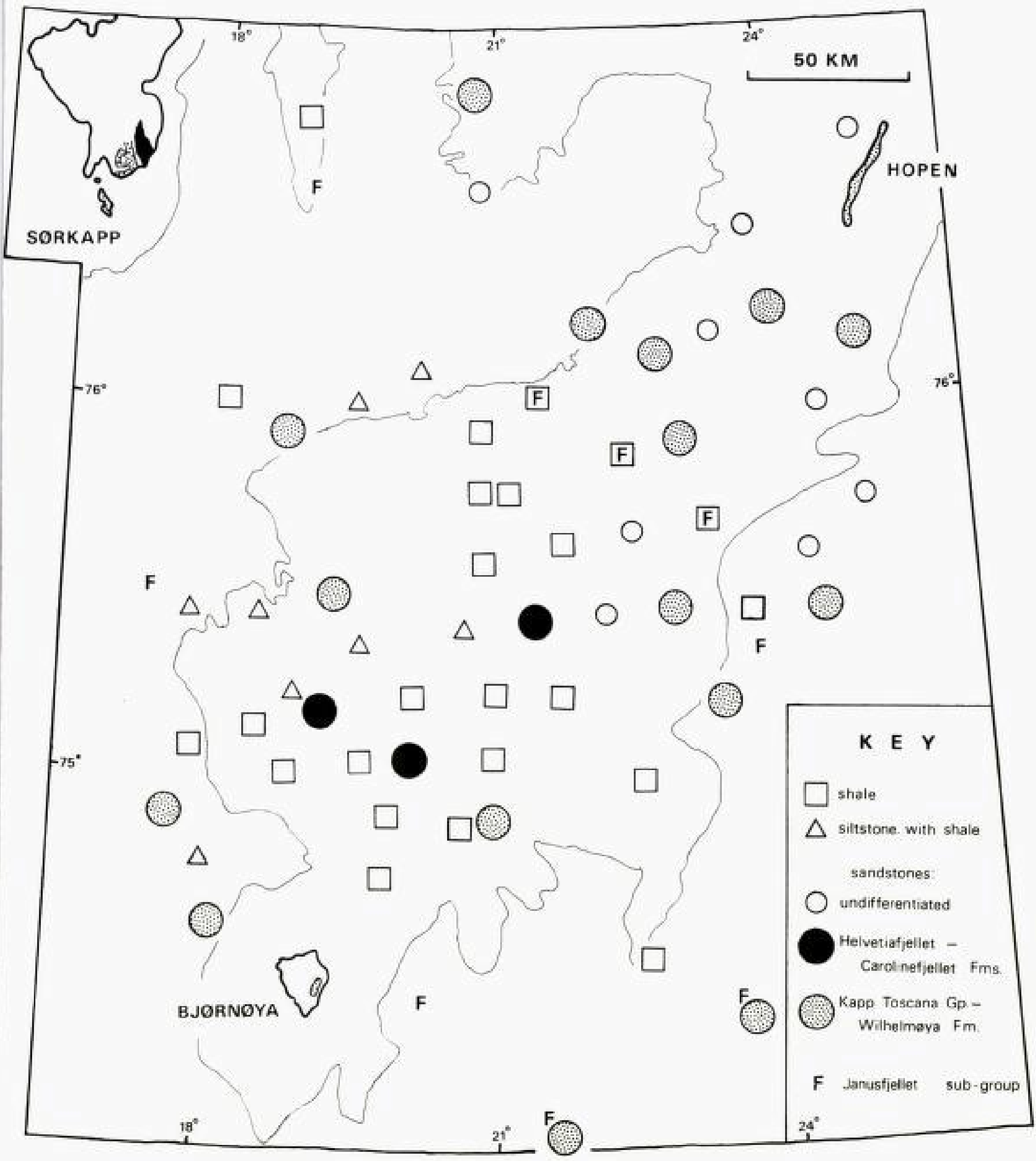

Fig. 6. Distribution of significant lithologies. Compiled from Fig. 4, data on sandstone petrography, and from an early account of the gravel lithology prepared by NTNFK. Geology of adjacent land areas based on Flood et al. (1971b) and Pčelina (1972). Hundred metre contour shown.

between these major stratigraphic units. Because of these uncertainties, the sandstone units are divided stratigraphically into two groups: those underlying, and those overlying the shales of the Janusfjellet Sub-group.

Stations containing a large proportion of pebbles which can be attributed to either of the two groups of sandstones are designated in Fig. 6.

Discussion

The differences in gravel lithology between various stations (Fig. 4) indicate that the material has not been rafted into place by icebergs. This has also been noted by Soviet scientists (Klenova, cited in Dibner et al. 1970), and is supported by the rarity of pebbles of definitely exotic lithologies, such as Caledonian granites and gneisses. Icebergs containing debris could originate from glaciers on Franz Joseph Land (400 km east of Svalbard) and on eastern Svalbard. Drift ice movements are controlled by both currents and wind. The main currents affecting the Spitsbergen Bank are the East Spitsbergen Current and the Bear Island Current (U.S. Hydrographic Office 1958), both of which are cold streams from the NE, and the North Cape Current, an east-flowing warm stream situated along the southern margin of the Bank. Winds are highly variable, and may temporarily disturb the effects of the currents (Lunde 1965). A considerable portion of the drift ice is carried southwestwards along the east coast of Spitsbergen, driven by the East Spitsbergen Current. It can be expected that only a small proportion of the ice finds its way southwards over the Spitsbergen Bank, which is consistent with the meagre quantities of apparently drifted material in the gravel.

The fact that striated pebbles are generally of the same lithology as non-striated pebbles in a sample suggests that both may be parts of the same glacial deposit. The distribution of this feature (Fig. 3) suggests that glacial deposition was widespread on the Spitsbergen Bank at an earlier time. Evidence supporting the glaciation of a large area of the Barents Shelf, probably during the Würm, has recently been put forward. A study of shorelines, with associated C^{14} datings, on Spitsbergen, Nordaustlandet, Kong Karls Land, Hopen and other areas in Svalbard indicates that the whole area of the NW Barents Shelf, possibly extending to the edge of the continental shelf, was covered by ice, with the greatest uplift (within the last 6500 years) in the southeast (Hoppe et al. 1970). These data, in conjunction with the inferred direction of ice movement from the NE on Hopen (Hoppe 1969), suggest that the ice sheet had its centre to the east of southern Svalbard, in the Barents Sea.

Although detailed work was not done on the roundness and sorting of the gravel, several observations are of interest. Most pebbles are sub-angular to sub-rounded, with other types relatively scarce. At some stations, rounded and well-rounded pebbles are dominant, suggesting extensive reworking. The gravel component appears rather poorly sorted. If the primary deposit was a subglacial till, then the presence of the gravel attests to some reworking. However, the preservation of striations on relatively soft lithologies, and the roundness of the pebbles suggests that this was not extended in time. Such reworking may have been due to glaciofluvial or marine currents. The effectiveness of the latter in winnowing the bottom at the present time is suggested by the current speeds of up to 2 knots (U.S. Hydrographic Office, 1958) and by observations made by divers during the sampling.

Thus, of the several mechanisms involved in the formation of the gravel, glaciers and icebergs, particularly the former, were probably responsible for transport, while glaciofluvial or marine currents resulted in reworking and some abrasion, with little significant lateral transport.

Applying the widely observed fact that subglacial tills tend to be composed mainly of local material (e.g. Flint 1971, p. 174), the composition of the gravel can be used to reconstruct the bedrock geology of the Spitsbergen Bank. The overall Mesozoic character of the bedrock is indicated by the predominance of sandstones with strong Mesozoic affinities, as well as by the occurrence of fossiliferous dark shales which have been assigned to the Janusfjellet Sub-group (Nagy, in prep). Omitted is material which, although datable, is a minor lithological constituent at a station. This applies, for example, to the limestones at stations 5 and 14 (and scattered pebbles at other stations) some of which closely resemble limestones on Bjørnøya, particularly the Permian Spirifer Limestone, but which are overshadowed by the large quantities of Mesozoic pebbles at those stations. The exotic pebbles found on the southern part of the Spitsbergen Bank may have been rafted into place by icebergs derived from calving glaciers on Bjørnøya during the last glaciation.

The resulting distribution of stratigraphic units does not form a simple pattern (Fig. 6). However, the occurrence of the older sandstone group on the NE and SW parts of the Bank, and the restriction of the younger sandstone group to a central area suggests a gentle synclinal structure to the Spitsbergen Bank, consistent with the regional structural pattern of southern Spitsbergen (Fig. 7). The concentration of limestone at station 36 may indicate the proximity of the station to the Permian–Triassic boundary. The anomalous central occurrences of the lower sandstone group in the central area could represent remnants of an end moraine complex or, on the other hand, reflect structural complications in the bedrock.

The inferred broad distribution of the Mesozoic formations, and the flat-lying aspect of the Triassic strata exposed on Bjørnøya and Hopen suggest that the strata underlying the greater part of the Spitsbergen Bank are, similarly, nearly flat-lying and largely undeformed. Additional evidence supporting these conclusions is provided by unpublished reports of NTNFK based on the continuous seismic profiling carried out in 1971.

Conclusions

1) Gravel material on the Spitsbergen Bank is essentially locally derived, transported into place largely by glaciers.
2) Petrographic observations can distinguish some of the sandsone units in the Mesozoic of Svalbard, and can be used to stratigraphically locate sandstones in the gravel from the Spitsbergen Bank.
3) Sandstones of the Kapp Toscana Group and the Wilhelmøya Formation have a wide distribution on the bank. Sandstones from the Helvetiafjellet and Carolinefjellet Formations have a limited distribution. The occurrences

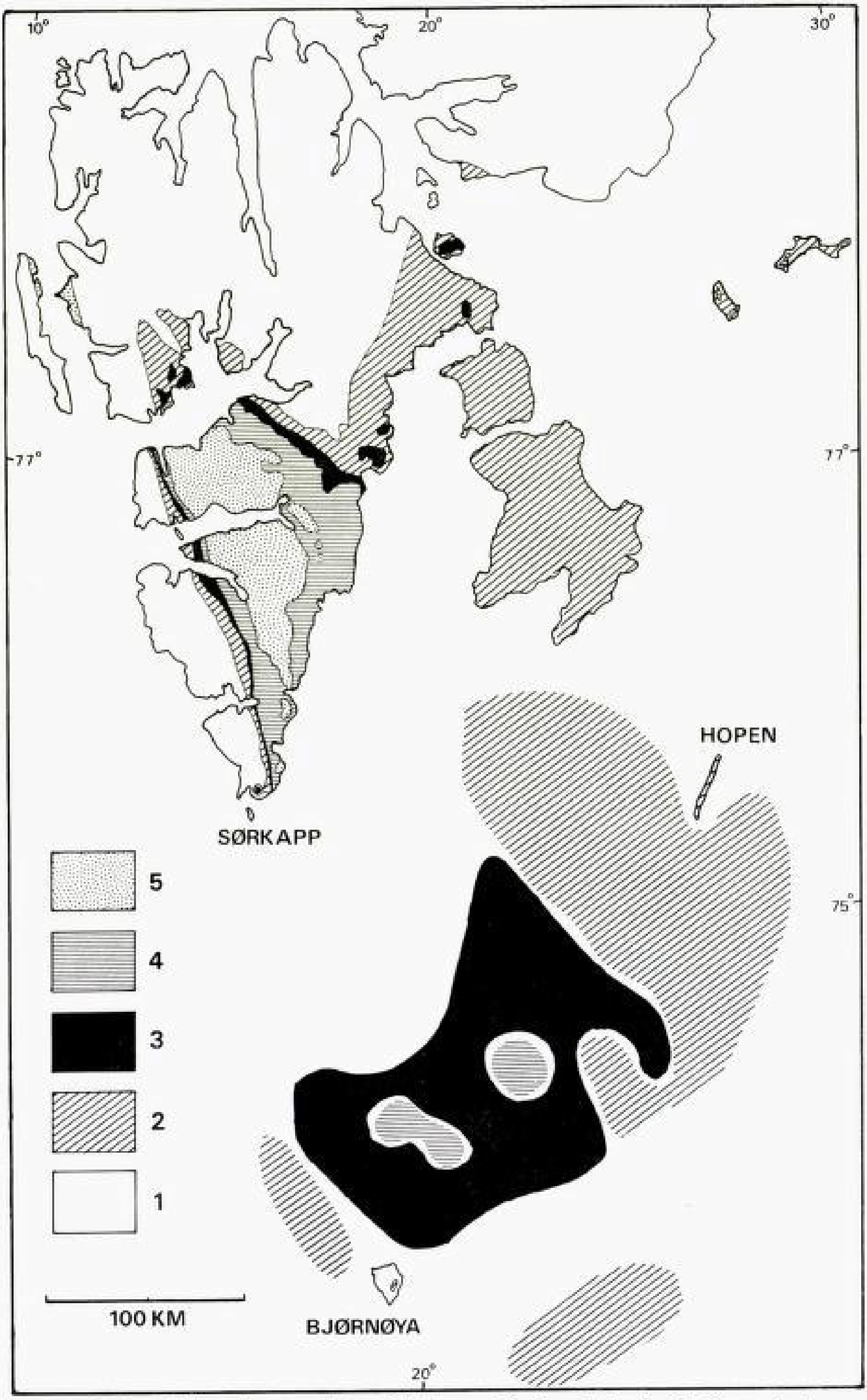

Fig. 7. Geological sketch map of the Spitsbergen Bank and Svalbard: 1) on land, pre-Triassic; on the Spitsbergen Bank, uncertain, 2) Triassic Sassendalen and Kapp Toscana Groups, and Liassic Wilhelmøya Formation, 3) Janusfjellet sub-group, 4) Lower Cretaceous (Barremian to Albian) Helvetiafjellet and Carolinefjellet Formations, 5) Tertiary. Based on Buchan et al. 1965); Flood et al. (1971b) and Orvin (1940).

show that these formations may continue for a considerable distance (at least 200 km) to the south of Spitsbergen.

4) The basic structure of Spitsbergen Bank appears to be a gentle syncline, a southward continuation of the dominant structure of Spitsbergen.
5) Shallow drilling at selected sites could test the hypothesis of the local nature of the gravel. Support for this hypothesis would render gravel a valuable aid in the investigation of the bedrock of the Barents Sea.

Acknowledgements. – Sampling of bottom material on the Spitsbergen Bank in 1971 was coordinated and financed by NTNFK, and carried out by NTNFK and the Norwegian Polar Institute.

The field work at Sørkappland was carried out while the author was a member of the 1973 Norwegian Polar Institute expedition to Svalbard. Colleagues at the Polar Institute kindly lent the author specimens and thin-sections of sandstones from Svalbard.

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Seismic Refraction Measurements and Continuous Seismic Profiling on the Continental Margin off Norway between 60°N and 69°N

MARKVARD A. SELLEVOLL

Sellevoll, M. A. 1975: Seismic refraction measurements and continuous seismic profiling on the continental margin off Norway between 60°N and 69°N. *Norges geol. Unders. 316*, 219–235.

The present study is mainly based on seismic measurements on the continental margin off Norway between 60°N and 69°N carried out by the Seismological Observatory, University of Bergen. Available information from other studies has also been utilized.

The sedimentary basin off Norway wedges out near the Norwegian coastline. The ocean side of the sedimentary basin is in contact with oceanic crust along escarpments and the Jan Mayen Fracture Zone. The present study reveals that the contact zone between the crust of oceanic and continental origin on the continental margin may be very complex and difficult to identify confidently on the basis of work carried out to date.

The strike beneath the base of Pleistocene deposits is generally parallel to the coastline. The dip just beneath the base of the Pleistocene deposits reflects the uplift of the landmasses from the coastline across the continental margin. The steep dips observed between the coastline and the Storegga region together with major submarine slides suggest that the shelf region between the coastline and Storegga may have undergone greater uplift relative to the rest of the shelf area (except for the Lofoten and Vesterålen shelf area).

Seismic velocity-information and the velocity-age relation suggest that the continental margin consists of extensive areas of sedimentary rocks of Cenozoic, Mesozoic and Paleozoic age.

Isopach maps show the calculated distribution of assumed Cenozoic and Mesozoic sedimentary rocks on the continental margin. Cenozoic deposits cover a great part of the continental margin and reach the greatest thickness between 62°N and 67°N. A deltaic structure which is probably of lower Tertiary age is especially well developed between Frøyabanken and Sklinnabanken. A large diapiric structure is observed in the Cenozoic sediments at 64½°N, 01°E, in a water-depth of about 2500 m. Extensive areas on the continental margin are covered by Mesozoic sediments, reaching a maximum thickness of about 2.7 km.

M. A. Sellevoll, Seismological Observatory, University of Bergen, Villavei 9, N–5014 Bergen, Norway

Introduction

The continental margin north of 60°N is of interest because of recent oil discoveries to the south, and the geology and structures of the outer and deeper marginal areas are of great interest in view of the latest ideas on the plate-tectonic evolution of the Norwegian–Greenland Sea. Thus, it is also of special interest to locate the boundary between the oceanic and continental crust.

A long-term study of the continental margin off Norway has been carried out by the Seismological Observatory, University of Bergen, as a part of the

'Continental Shelf Project', sponsored by the Royal Norwegian Council for Scientific and Industrial Research (NTNF) from 1967 up to 1972, and from 1973 onwards by Statens Oljedirektorat. The main objective of this long range project has been to study the distribution and structure of the sedimentary rocks underlying the continental margin off Norway north of 62°N.

The principal objective of the present study was to study the distribution, thickness and structures of Pre-Quaternary sedimentary rocks on the continental margin off Norway between 60°N and 69°N mainly on the basis of seismic data. Results from earlier seismic reflection and refraction data as well as from magnetic data (Ewing & Ewing 1959; Johnson & Heezen 1967; Eldholm & Nysæther 1969; Eldholm 1970; &m 1970; Sundvor & Sellevoll 1971; Sundvor 1972; Talwani & Eldholm 1972; Hinz 1972; Sellevoll & Sundvor 1974; Sellevoll 1974) are also included to give the most complete geological picture.

Several writers have previously discussed the distribution of the Pleistocene sediments and the development of the Pre-Pleistocene rocks and structures on the continental margin off Norway (O. Holtedahl 1940, 1960; H. Holtedahl 1955; Nysæther et al. 1969; Holtedahl & Sellevoll 1971, 1972; Talwani & Eldholm 1972).

Seismic refraction studies and continuous profiling

Continuous seismic profiling has been done along the track locations indicated in Fig. 1. Heavy lines indicate where deep seismic reflection measurements have been carried out recently using a DFS–10,000 instrumentation and a 12-section streamer. The processing of the multicovered reflection lines is in progress at the Observatory.

The seismic refraction data (Fig. 2 and Table 1) used in the present study are partly from two-ship refraction measurements and partly from expandable sonobuoy experiments using an air gun as energy source. The apparent velocities are calculated by visually fitting straight lines to segments of travel time curves and measuring their slopes relative to the slope of the direct water wave. These refraction data have been gathered by scientists from the Seismological Observatory, University of Bergen, and from the Lamont Doherty Geological Observatory of Columbia University. Profile locations, P-wave velocities and thicknesses of the sedimentary layers are shown in Fig. 2 and Table 1. All of the profiles shown on Fig .2 are short, and thus not amenable for acquisition of information on the deep crust or the mantle.

Seismic crustal studies have also been performed along a profile line on the Vøring Plateau during a cooperative geophysical programme between Bundesanstalt für Bodenforschung, Hannover, and Seismological Observatory, University of Bergen. These data have been interpreted by Hinz (1972).

The histogram in Fig. 3 shows the seismic P-wave velocities obtained from the refraction measurements presented in Table 1. It is well known that velocities of seismic waves in sedimentary rocks are a function of thickness,

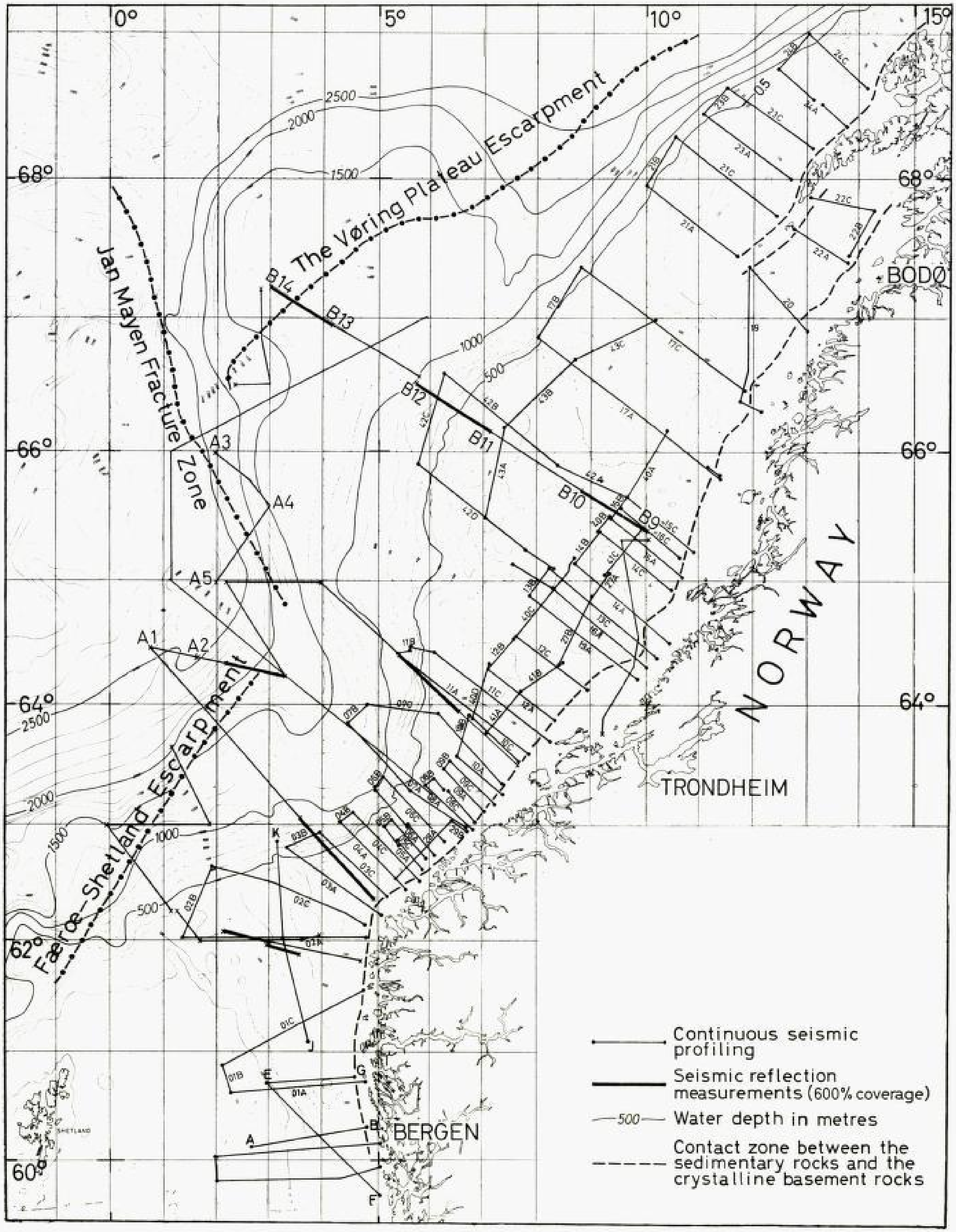


Fig. 1. Tracks on the continental margin off Norway. Measured by the Seismological Observatory, University of Bergen. (Escarpments and fracture zone after Talwani & Eldholm 1972).

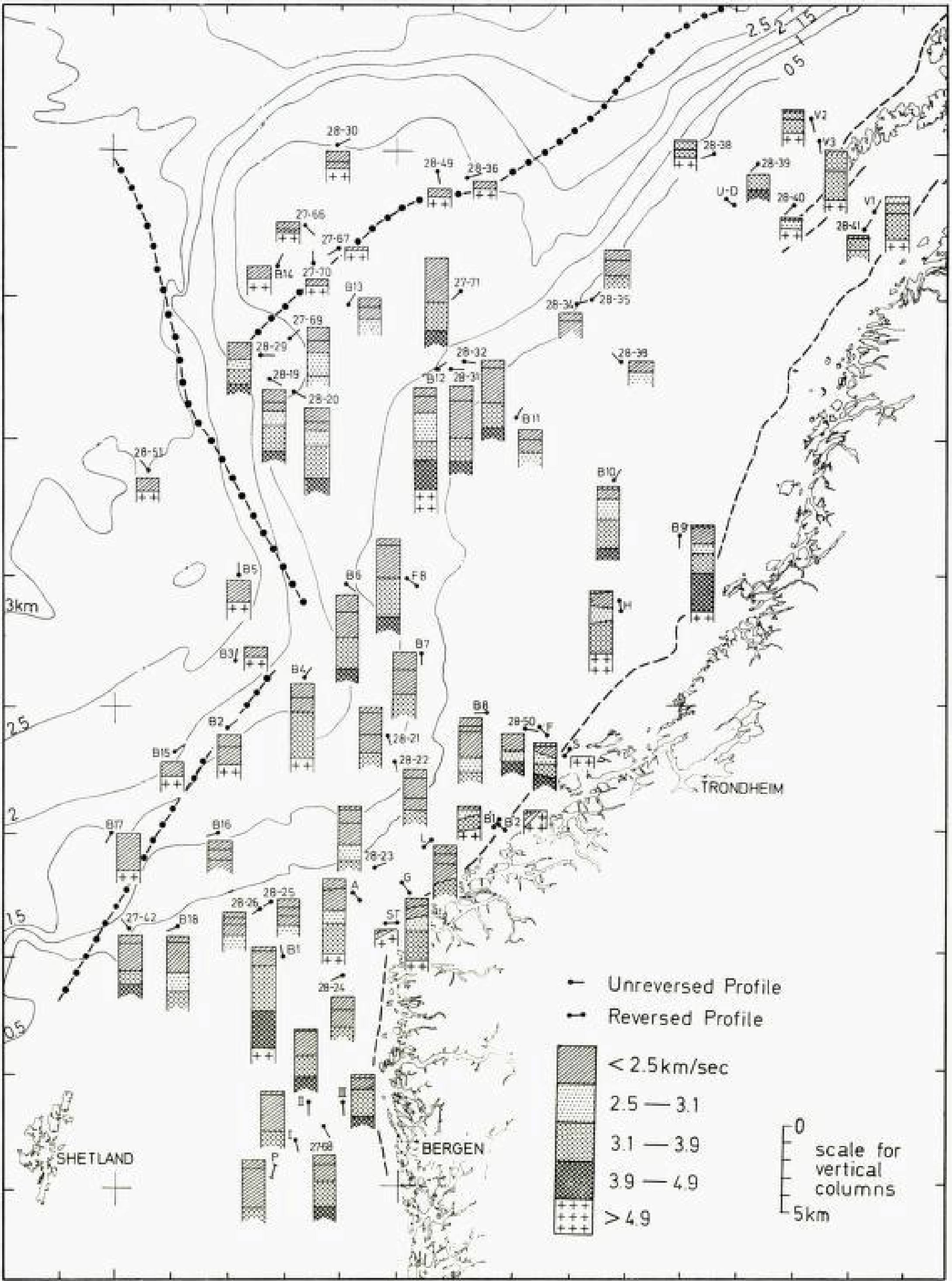


Fig. 2. Map showing seismic refraction data (see Table 1 and text).

Table 1. Listing of seismic refraction results. All units are in km and km/s. Parentheses indicate assumed veloc

Profile	Lat. (N)	Long. (– = W)	Water depth	V1	H1	V2	H2	V3	H3	V4	H4	V5	H5	V6	Data sourc
B1	61°58.0′	2°57.0′	0.40	(1.80)	0.25	2.40	0.64	3.35	2.70	4.30	2.02	5.20			Seism
B2	63°50.8′	2°00.0′	1.66	1.75	0.59	2.10	1.00	5.00							Obsv
B3	64°09.0′	2°09.0′	1.50	2.00	0.56	5.05									1973
B4	64°13.5′	3°22.0′	1.85	1.82	0.80	2.20	0.76	3.50	2.57	5.20					"
B5	65°00.0′	2°08.0′	2.50	2.15	1.08	5.10	1.00	6.45							"
B6	64°57.0′	4°04.0′	1.05	1.85	0.87	2.20	1.39	3.55	1.68	4.00					"
B7	64°26.5′	5°23.5′	0.93	1.80	0.97	2.10	1.40	3.30							"
B8	63°56.5′	6°33.0′	0.25	1.85	0.39	2.10	0.27	2.35	1.40	2.70	1.40	3.20			"
B9	65°17.8′	10°00.5′	0.16	(1.80)	0.23	2.10	0.61	2.60	0.56	3.70	1.17	4.40	2.10	5.50	"
B10	65°42.8′	8°52.0′	0.43	(1.80)	0.16	2.35	0.51	2.90	1.02	3.55	1.65	4.45			"
B11	66°10.2′	7°09.5′	0.35	1.85	0.54	2.30	0.68	2.70							"
B12	66°32.0′	5°46.5′	0.90	1.85	0.47	2.10	0.89	2.55	1.57	3.90	1.02	4.45	1.61	4.90	"
B13	66°58.0′	4°10.2′	1.33	1.75	0.49	2.10	0.74	2.80	0.80	3.45					"
B14	67°14′8	2°50.0′	1.30	2.30	0.77	5.25									"
B15	63°40.0′	1°11.0′	1.90	2.30	0.88	5.10									"
B16	63°00.0′	1°36.0′	1.05	(1.80)	0.37	2.05	0.51	2.30							"
B17	63°00.0′	-0°10.0′	1.44	2.25	2.06	5.30									"
B18	62°15.8′	1°06.5′	0.50	(1.80)	0.32	2.10	1.68	2.55	1.02	3.95					"
27–42	62°13.0′	0°16.0′	0.67	(1.85)	0.43	2.02	1.51	3.32	0.70	4.65					Talwan
27–66	67°29.0′	3°21.0′	1.28	1.56	0.41	2.26	0.15	5.20							Eldhol
27–67	67°21.0′	3°59.0′	1.21	2.12	0.14										1972
27–68	60°32.0′	3°42.0′	0.30	1.85	0.49	2.25	0.94	3.45	1.44	4.50					"
27–69	66°42.0′	3°05.0′	1.44	1.78	0.73	2.02	0.69	2.95	1.22	4.15					"
27–70	67°15.0′	3°30.0′	1.19	1.71	0.34										"
27–71	67°03.0′	6°05.0′	1.31	(2.30)	2.30	3.70	1.55	4.70							"
28–19	66°25.0′	2°42.0′	1.59	1.86	0.69	2.24	0.49	2.47	0.86	3.59	1.41	4.09			"
28–20	66°20.0′	3°10.0′	1.48	1.87	0.81	2.24	0.39	2.54	0.86	3.64	1.82	4.43			"
28–21	63°45.0′	4°48.0′	1.37	(1.85)	0.23	2.11	1.24	2.49	1.01	3.48					"
28–22	63°33.0′	4°56.0′	1.19	(1.85)	0.49	2.09	1.15	2.44	0.71	3.43					"
28–23	62°42.0′	4°35.0′	0.19	1.85	0.22	2.17	0.73	2.45	1.19	3.10	0.92	3.85			"
28–24	61°50.0′	4°01.0′	0.20	(1.85)	0.69	1.98	0.99	3.76							"
28–25	62°26.0′	2°45.0′	0.40	(1.85)	0.47	1.94	0.43	2.28	0.51	2.85					"
28–26	62°23.0′	2°35.0′	0.40	(1.85)	0.34	1.97	0.32	2.38	0.58	2.88					"
28–29	66°35.0′	2°34.0′	1.60	1.76	0.95	2.46	0.54	3.17	0.80	4.26					"
28–30	68°07.0′	3°57.0′	1.55	1.74	0.62	2.23	0.37	5.35							"
28–31	66°30.0′	5°58.0′	0.85	1.94	0.81	2.48	2.11	3.70	1.24	4.65					"
28–32	66°33.0′	6°09.0′	1.03	(1.85)	0.40	2.23	1.99	3.62	1.38	4.32					"
28–33	66°33.0′	8°56.0′	0.30	2.00	0.65	2.63									"
28–34	66°57.0′	8°11.0′	0.46	1.94	0.46	2.38									"
28–35	67°00.0′	8°27.0′	0.31	1.93	0.47	2.35	0.92	3.25							"
28–36	67°49.0′	6′15.0°	1.23	1.75	0.34										"
28–38	68°00.0′	10°37.0′	0.23	3.75	0.54	4.65	0.48	5.24							"
28–39	67°56.0′	11°24.0′	0.18	3.57	0.91	4.23									"
28–40	67°39.0′	12°03.0′	0.16	(2.00)	0.10	2.75	0.48	5.35							"
28–41	67°29.0′	13°19.0′	0.27	(2.00)	0.14	3.07	0.62	3.70							"
28–49	67°52.0′	5°42.0′	1.38	1.77	0.43	5.10									"
28–50	63°49.0′	7°13.0′	0.18	1.97	0.98	2.75	0.56	4.10							"
28–51	65°46.0′	0°37.0′	3.15	2.20	0.77										
A	62°27.1′	4°19.6′	0.17	1.82	0.38	2.28	1.10	2.79	0.75	3.33	1.70	5.00			Eldholt
	62°30.5′	4°10.0′			0.59		1.10		0.75		1.70				1970
G	62°35.0′	5°05.6′	0.10	1.95	0.30	2.17	0.56	2.62	0.92	3.57	1.61	5.25			"
	62°40.1′	4°58.9′			0.56		0.59		0.64		1.57				
L	62°51.9′	5°29.4′	0.10	1.94	0.46	2.09	0.52	2.45	0.78	3.60					"
	62°49.0′	5°19.7′			0.46		0.54		1.05						

ble 1. (Continue)

file	Lat. (N)	Long. (– = W)	Water depth	V1	H1	V2	H2	V3	H3	V4	H4	V5	H5	V6	Data source
	62°15.5′	4°57.6′	0.18	2.01	0.00	2.51	0.00	5.60							"
	62°16.3′	4°47.9′			0.23		0.22								
I	63°03.0′	6°41.6′	0.08	2.05	0.11	2.56	0.59	3.79	0.56	5.24					"
	63°08.4′	6°49.0′			0.19		0.23		0.79						
II	63°02.0′	6°55.5′	0.12	2.08	0.00	2.76	0.00	3.50	0.00	5.20					"
	63°04.6′	6°45.8′			0.13		0.34		0.44						
	63°35.5′	7°57.0′	0.20	5.02											"
	63°38.5′	8°00.5′													
	63°45.3′	7°37.8′	0.17	1.80	0.19	1.96	0.23	2.60	0.70	3.28	0.54	4.38			"
	63°47.7′	7°29.0′			0.19		0.63		0.23		0.82				
	64°40.2′	8°57.0′	0.12	2.00	0.12	2.16	0.70	2.58	1.02	3.54	1.64	5.50			"
	64°46.7′	8°55.2′			0.00		0.91		0.60						
	67°36.0′	13°28.8′	0.25	2.69	0.32	3.22	0.51	3.76	1.48	5.15					Sundvor &
	68°13.8′	12°21.0′	0.24	(1.90)	0.05	3.29	0.41	3.76	0.83	5.18					Sellevoll,
	68°04.8′	12°28.2′	0.19	(1.90)	0.13	3.20	0.92	3.81	1.76	5.20					1971
D	67°38.0′	10°59.0′	0.18	1.90	0.25	2.40	1.10	3.50							Eldholm &
	67°40.0′	10°53.0′			0.25		1.10								Nysæther, 1969
	60°24.5′	3°09.3′	0.13	(1.85)	0.14	2.15	2.06	3.35							Sellevoll &
	60°44.3′	3°24.0′	0.30	(1.85)	0.11	2.20	1.34	3.25	1.18	4.50					Sundvor,
	60°44.4′	4°06.5′	0.30	(1.85)	0.23	2.10	0.58	3.05	1.49	4.70					1974
	60°11.6′	2°41.7′	0.09	1.82	0.37	2.08	2.23	2.97							Sundvor, 1972
	60°04.7′	2°28.3′													
	64°55.0′	5°19.0′	0.64	1.66	0.36	2.08	1.78	3.44	2.12	4.01					Ewing &
	65°14.0′	4°35.0′			0.36		1.78		2.12						Ewing, 1959

geological age and lithology of the rocks. As a consequence it is impossible to get an exact correlation of the seismic velocities with geological age alone. Seismic velocity profiles at different parts of the continental margin off Norway are fairly similar to one another concerning the velocity so it should be possible at least to make correlation within the velocity-structure from one profile to another, although the stratigraphic ages suggested in Fig. 3 may be somewhat incorrect.

P-wave velocities in Tertiary rocks seldom exceed 2.25 km/sec in the North Sea (Hornabrock 1962; Wyvobeck 1969), and seismic velocities in Mesozoic rocks are generally less than 4.0 km/sec. Thus, on Andøya, the island in Northern Norway, P-wave velocities are about 2.5 km/sec in Cretaceous rocks and 3.1 km/sec in sedimentary rocks of Upper Jurassic age (Sellevoll & Sundvor 1972). The velocity structure of the upper sedimentary sequences on the westernmost part of the Barents Shelf is similar to the Norwegian shelf area farther south. Eldholm & Ewing (1971) and Sundvor (1971) interpreted the 2.2 km/sec layer to be of Tertiary age. The velocity-age relation used in the present study (Fig. 3) is the same that Talwani & Eldholm (1972) have used for almost the same region.

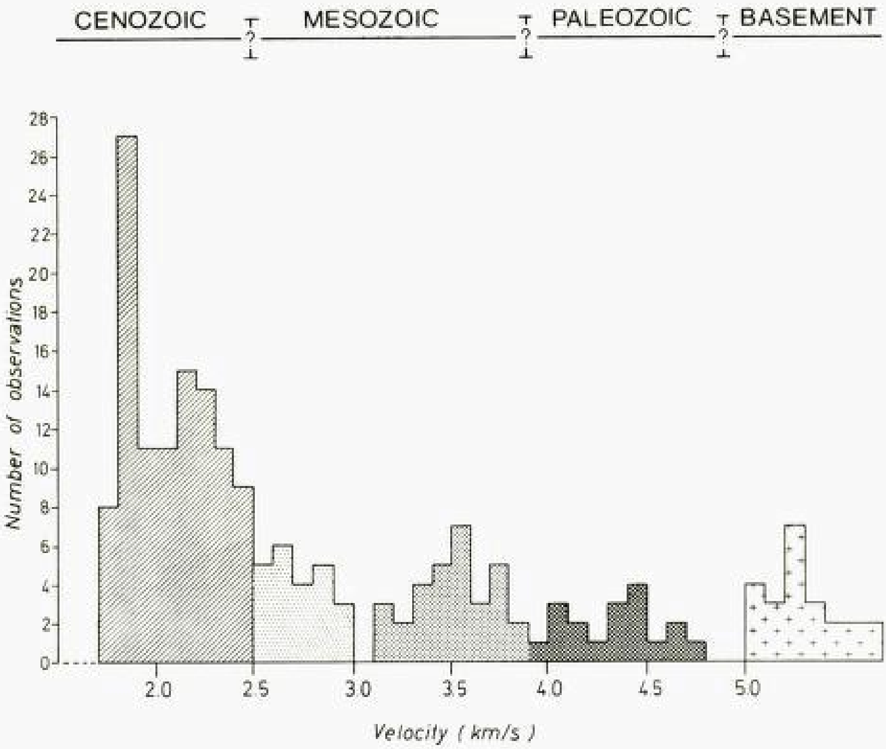
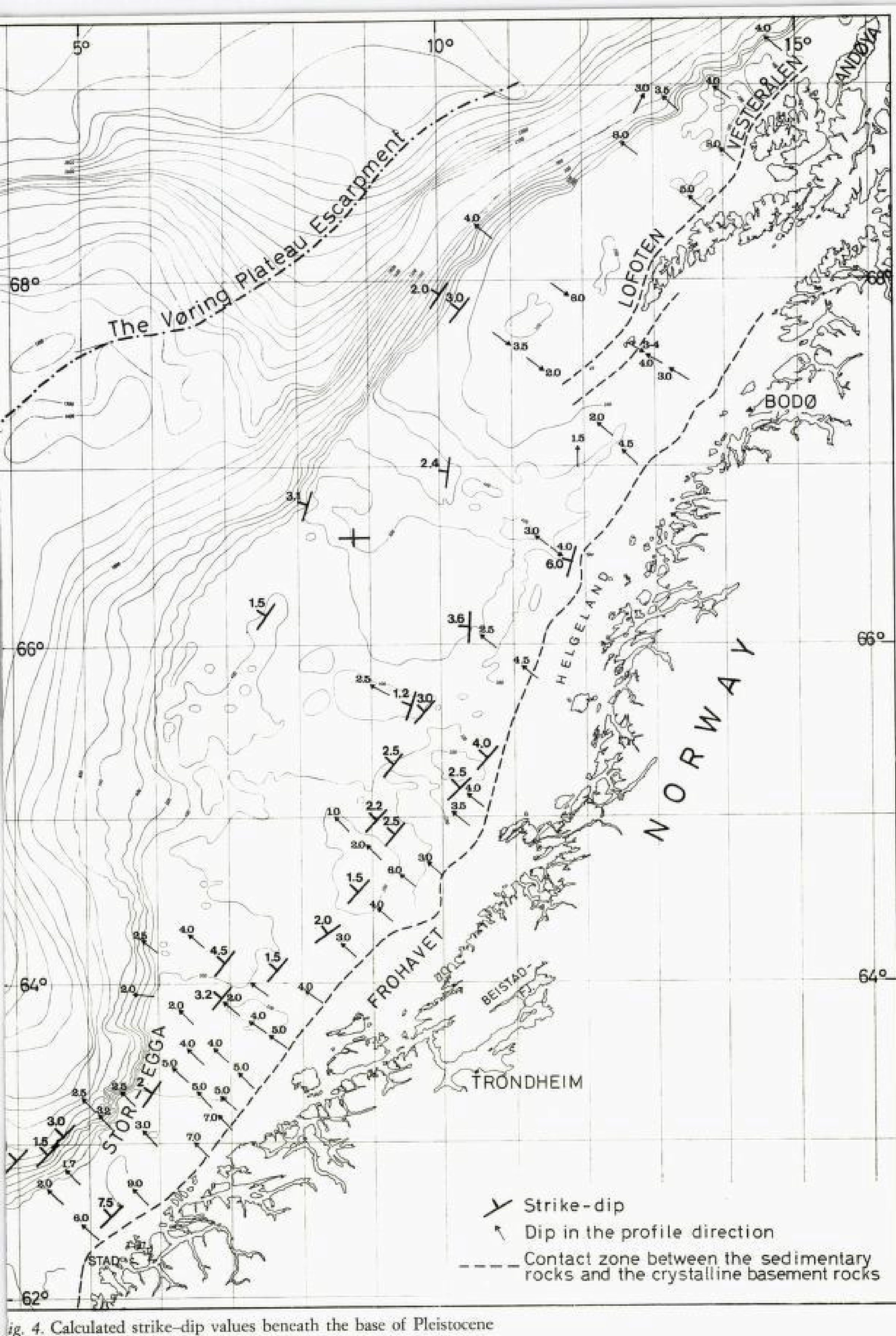


Fig. 3. Histogram of velocities from seismic refraction measurements with proposed velocity-age relation. (See text for discussion of these relations.)

The sedimentary basin between 60° and 69°N

The continental shelf varies in width from about 60 km seawards off Stad to 200 km off Helgeland to less than 60 km off Lofoten–Vesterålen (Fig. 1). Some of the boundaries of the main sedimentary basin on the continental margin off Norway are shown on Fig. 4. The contact between Phanerozoic sedimentary rocks and crystalline rocks on the Norwegian mainland is subparallel to the coastline as shown by continuous seismic profiling (Nysæther et al. 1969). The rate of change of dip measured just beneath the base of Pleistocene decreases seawards and across the shelf. The dip near the contact varies generally between 3° and 9° (predominantly to the northwest), as shown in Fig. 4 by some calculated strike-dip values and a number of apparent dip-indications along the continuous seismic profile lines. These apparent dips are equal to or less than the real dips. The dips are almost all northwesterly on the continental margin with two exceptions: one near the cost between 66°N and 67°N (Fig. 4) and another in the Lofoten region. The steep dips observed along the Norwegian coast are probably too steep to be primary dips. Nysæther et al. (1969) concluded that primary dips were steepened as a result of the uplift of the Scandinavian landmasses. Thus, according to Nysæther et al. (1969), the eastern part of the Vestfjorden area must have been a basin



ig. 4. Calculated strike–dip values beneath the base of Pleistocene

of Pre-Tertiary age, which was centred where the Lofoten Islands are situated today. The high grade metamorphic basement rocks were uplifted through the basin sediments and formed a horst. Smaller — probably downfaulted — sedimentary basins are also observed on the east side of the main contact zone near Frohavet and Andøya. Oftedahl (pers. comm., 1972) has reported one smaller basin beneath Beitstadfjorden.

Structure and development of the Vøring Plateau

An extensive study of aero-magnetic data between the Norwegian Coast and 15°W and from 60°N to 73°N has been completed by Avery et al. (1968). They found that a magnetic quiet zone covers a great part of the Norwegian continental margin, but his feature changes to linear magnetic anomalies on the seaward side of the Norwegian continental margin. This change is especially well marked on the Vøring Plateau.

The interpretation of the linear magnetic anomalies in terms of sea-floor spreading by Avery et al. (1968) indicates that the separation of Greenland and Norway started about 60–70 m.y. ago. Prior to this, Greenland and Norway were a part of the same crustal plate. According to Talwani & Eldholm (1972) the Vøring Plateau escarpment and the Faeroe–Shetland escarpment are remnants of the original rift zone between Norway and Greenland.

From geophysical studies of the Vøring Plateau, Johnson et al. (1968) pointed out that the magnetic anomalies over the outer edge of the plateau indicated a zone of volcanic activity, and that smooth anomalies over the crest of the plateau are characteristic of a thick sequence of overlying sedimentary rocks. Åm (1970) made a depth estimate to magnetic basement and found a marked change in the magnetic patterns on the western part of the Norwegian continental margin which corresponds to an abrupt shallowing of the magnetic basement. On basis of seismic data Hinz (1972) found that the abrupt change in the magnetic pattern on the Vøring Plateau corresponds to a fault which divides the Vøring Plateau into an eastern and a western part. Extensive seismic, magnetic and gravimetric studies by Talwani & Eldholm (1972) show that a prominent feature on the continental slope off Norway is an abrupt shallowing of the crystalline basement seawards along escarpments (Fig. 2 and 4).

On the basis of our crossings of the escarpments and Jan Mayen Fracture Zone we have noticed that the Vøring Plateau Escarpment is a well-defined contact between two contrasting lithologic units. Our measurements on the southern part of the Vøring Plateau and along Faeroe–Shetland Escarpment indicate, however, that a much more complicated and less distinct configuration of the contact-zone exists between oceanic and continental crust in this region. To some degree this can be demonstrated in Fig. 5, which show continuous seismic sections from A3 to A4 and from A4 to A5. Several 'escarpments' seem to exist indicating a rather complicated rifting of the crust combined

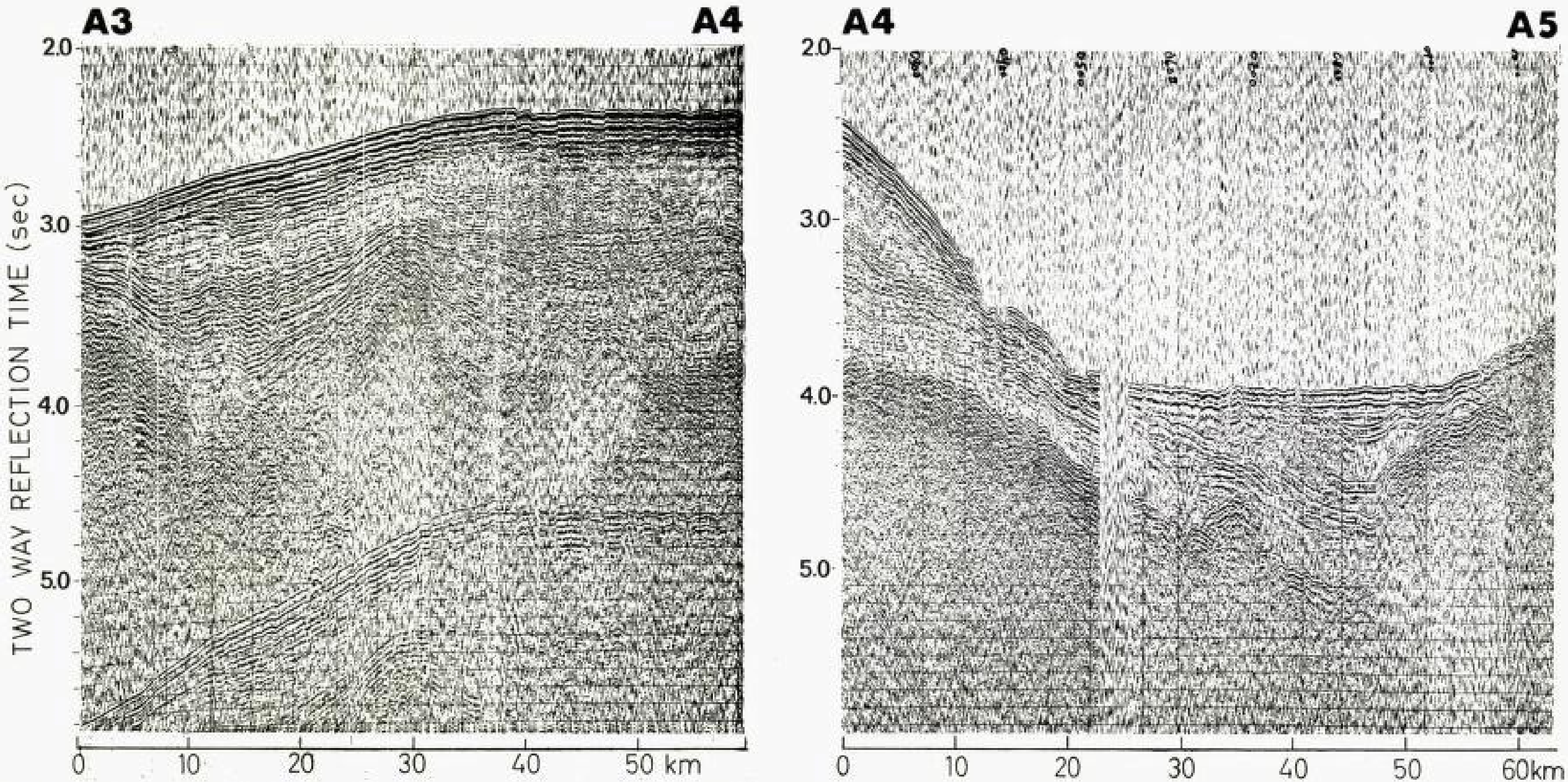


Fig. 5. Continuous seismic profiling between A3–A4 and A4–A5. (Profile locations shown in Fig. 1.)

with vertical block movements. Sediments have filled in between the blocks and as a consequence it is rather difficult to trace thicknesses of the sediments within the area between the Vøring Plateau Escarpment and the Faeroe–Shetland escarpment. The present study indicates that the Jan Mayen Fracture Zone may not be a single continuous fracture zone on the south-western side of the Vøring Plateau.

It is reasonable to assume that the initial opening of the Norwegian Sea started at the marginal escarpments, as proposed by Talwani & Eldholm (1972), but the configuration of these escarpments may vary considerably from one location to another according to the processes associated with the initial rifting and later spreading of the sea-floor.

Prior to the opening of the Norwegian Sea it is reasonable to assume that there was a very large sedimentary basin between Greenland and Norway, as suggested by Talwani & Eldholm (1972). This basin has probably developed in the same way as a normal basin with subsidence and sedimentation as the main processes. It is also reasonable to assume that the sedimentary basin between Greenland and Norway rested on a crust which could be classified as continental.

The opening of the Norwegian Sea was associated with intrusion of basaltic magma through the rifted crust and into the rifted sedimentary basin, and a new basaltic crust was created along rifts and the sea floor spreading processes separated Greenland from Norway during Tertiary time. Isostasy requires the newly formed oceanic crust and the continental crust to be in balance with each other, thus different crustal thicknesses across the transition zone between the oceanic and continental crust also require adjustments of the sea bottom elevation, resulting in development of the continental slope, rise and ocean floor.

The schematic model (Fig. 6) (Hinz 1972) presented to show the crustal structure beneath the Vøring Plateau and adjacent regions, indicates that the crustal and upper mantle processes may have been very complex in the transition zone between the continental and the oceanic crusts. The Moho is less than 20 km beneath the central part of the Vøring Plateau which is a crustal thickness intermediate between average oceanic and continental crustal thickness. Thus, Hinz (1972) has assumed that the Vøring Plateau to the east of the escarpment developed from a continental crust by subsidence and rising of mantle material beneath the Vøring Plateau as a result of the sea floor spreading.

That the outer part west of the escarpment on the Vøring Plateau also consists of a 'continental crust' overlain by basaltic lava cannot yet be completely ruled out. It is not unlikely that a 'contact' exists between the oceanic and continental crust in this area, and it is also reasonable to assume that the escarpments represent an important indication about where the contact is located; but there may exist several escarpments within the complex transition zone between the two crustal types and available data do not permit exact delineation of the contact.

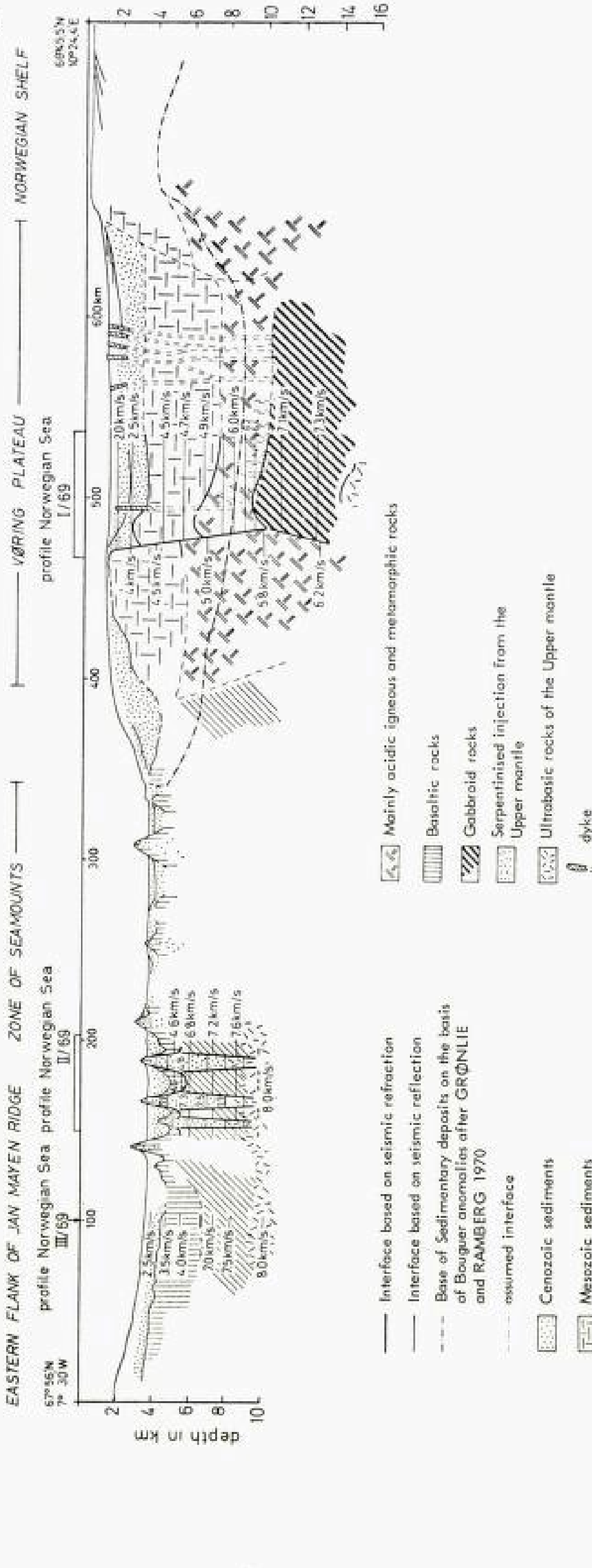


Fig. 6. Schematic crustal section (after Hinz 1972).

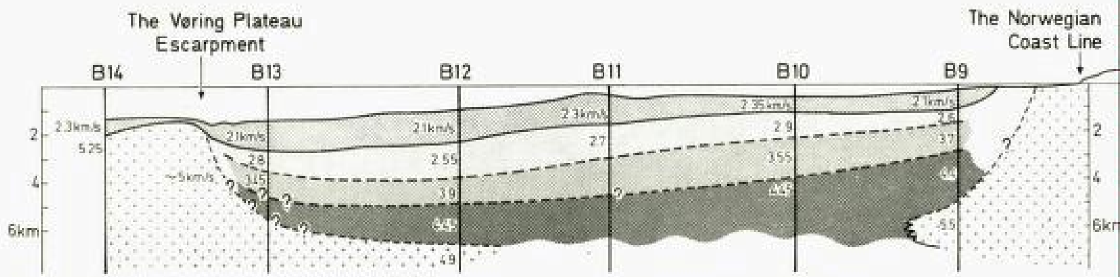


Fig. 7. Seismic structure section across the Norwegian continental margin. (Profile location shown in Fig. 1.)

A schematic cross-section of the sedimentary basin from the Norwegian coast to the western side of the Vøring Plateau is shown in Fig. 7. The assumed surface of the crystalline basement slopes rather steeply away from the mainland underneath the sedimentary rocks. Typical crystalline basement-velocities have not been observed at B10 and B11. This means that the deep crystalline basement is unusually deep along that part of the cross-section, as suggested previously on the basis of magnetic measurements (Åm 1970). Crystalline basements is observed a little deeper than 6.0 km below sea level at B12 on the eastern part of the Vøring Plateau (Fig. 7). The crystalline basement is not observed at B13 and little can be said about the structure of the deeper sedimentary layers in the vicinity of the escarpment (Fig. 7) on the basis of the available data.

According to the velocity-age relation (Fig. 3) Cenozoic, Mesozoic and Paleozoic rocks should exist along the profile from the Norwegian mainland to the escarpment on the Vøring Plateau. The cross-section (Fig. 7) indicates no major structural change at the shelf edge. Thus the present difference in the elevation of the shelf area and the Vøring Plateau area may be the results of differential subsidence along the shelf edge and continental slope.

TERTIARY SEDIMENTS

The calculated thicknesses of the Tertiary sediments compiled from seismic refraction and seismic continuous profiling data are shown on Fig. 8. These thicknesses include a layer of Quaternary sediments (from a few metres to 400 m, but normally less than 200 m), except along the contact between the Tertiary sediments and the Mesozoic rocks along the Norwegian coast. The Cenozoic sediments wedge out landwards, whereas they thicken rapidly seawards, especially in the southern part of the sedimentary basin, where they reach a maximum thickness of about 2.5 km at 64°N, 5°E. Farther west the thickness decreases at the marginal escarpments. On the shelf area between 66°N and 69°N the Tertiary sediments are rather thin or absent.

Correlations of the sonobuoy results obtained at B-9, B-10, B-11, B-12 and B-13 indicate a refractor with the respective velocities 2.6 km/s (B-9), 2.9 km/s (B-10), 2.7 km/s (B-11), 2.55 km/s (B-13). We consider this refractor to be the base of the Tertiary.

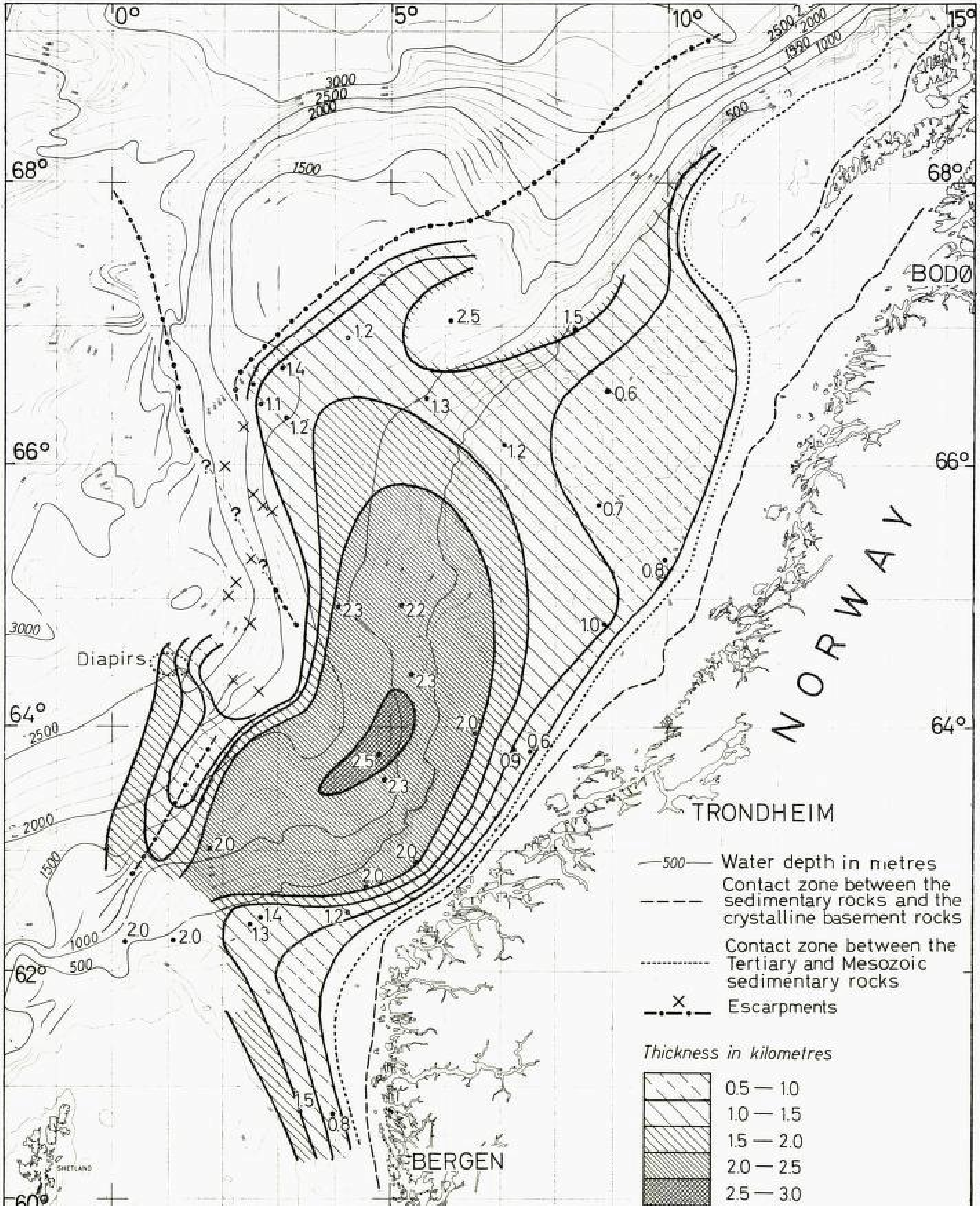


Fig. 8. Map showing thickness of the Tertiary sedimentary sequence (a thin Quaternary sequence is included), basement and Tertiary outcrops beneath the base of Pleistocene.

The sedimentary layer on the western part of the Vøring Plateau is relatively thin, suggesting that the escarpment has been an effective barrier to transportation of Cenozoic sediments. The area between the Vøring Plateau and the Færoe–Shetland Escarpment appears to be elevated in general, but the fracturing in this area reduced the effectiveness of the barrier and as a consequence of this a great amount of Cenozoic sediments have flowed over the 'barrier' into the ocean basin.

The northern part of the Færoe–Shetland Escarpment is rather easy to observe beneath a moderately thick sedimentary cover. Farther south the escarpment almost disappears beneath a constantly increasing sedimentary cover, and a great amount of Cenozoic sediments have flowed over the escarpment and accumulated on the western side of the Færoe–Shetland Escarpment. A large diapiric structure was observed within these sediments, as is shown on a continuous seismic section from A1 to A2 across the diapiric structure (Fig. 9). Unfortunately, the magnetometer broke down during profiling here, and as a consequence we have no observation of the magnetic field across the diapir. The seismic cross-section indicates, however, that the diapir has developed only within the sedimentary sequence.

The seismic refraction data obtained at profile 27–71 (see Figs. 2 and 8) on the Vøring Plateau indicate another Cenozoic depression, but the seismic data available from this region are too limited to permit a more detailed interpretation.

Several good reflectors have been observed within the Tertiary sequence. Some refractors can be correlated from one profile to another, but the profile spacing is in general too great to do so except for a few horizons, such as a deltaic structure which is well developed between Frøyabanken and Sklinnabanken. The seismic sections show that the deltaic sequence consists of smaller continuous deltas where the average transport direction has been from ESE to WNW. It has previously been assumed that the deltaic sequence marks the base of Tertiary, but seismic velocities indicate that the base of Tertiary reaches closer to the Norwegian coast-line than previously thought. This means that the deltaic structure is a younger sequence within the Tertiary succession than previously assumed.

PRE-TERTIARY SEDIMENTS

The thickness of the Mesozoic sedimentary sequence was compiled from seismic refraction and reflection data as shown in Fig. 10; one must keep in mind, however, the uncertainties of the velocity determination and velocity-age relationship.

It seems from Table 1 and Fig. 3 as if the main area of sedimentation during the Upper Mesozoic era is limited in the north-south direction by the parallels of 65°N and 67°N — in the east-west direction by 4°E and 10°E. The main area of sedimentation during the Lower Mesozoic era seems to have been between the parallels 61° and 66° in the north-south direction and on the western part of the continental margin between the Færoe–Shetland Escarpment and the Norwegian mainland. On both sides of the Lofoten Islands

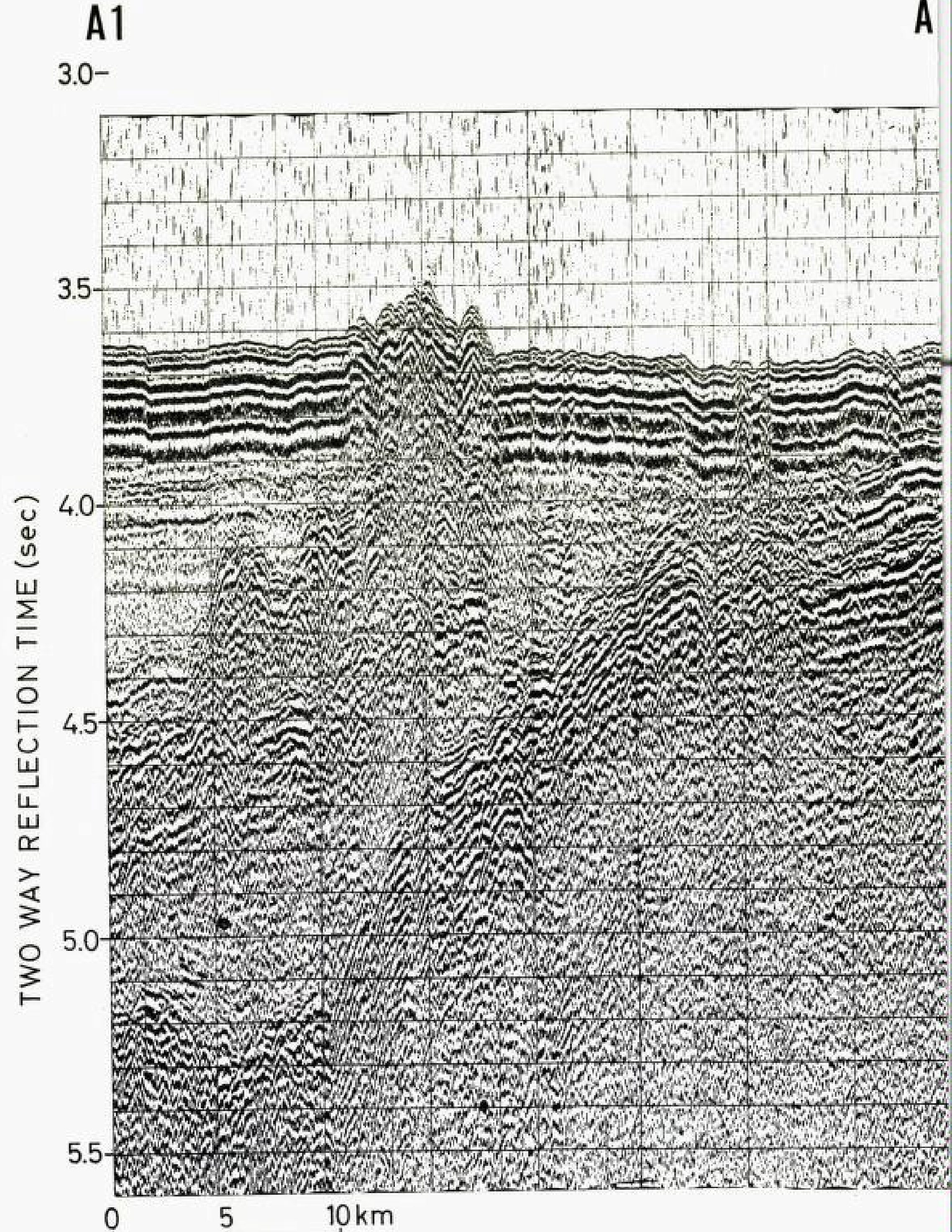


Fig. 9. Continuous seismic profile west of the Faeroe–Shetland Escarpment. (Profile location shown in Fig. 1.)

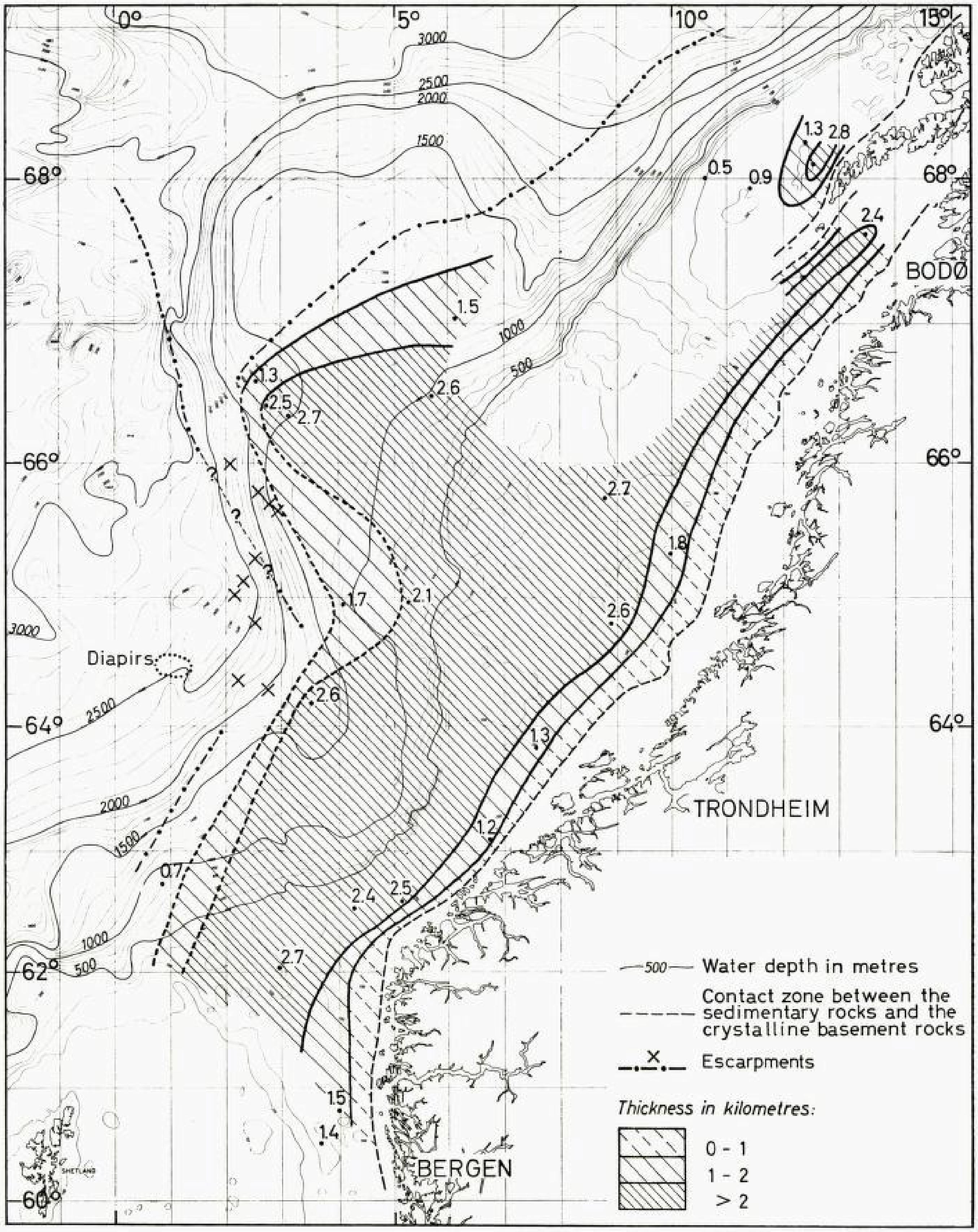


Fig. 10. Map showing thickness of Mesozoic sedimentary sequence.

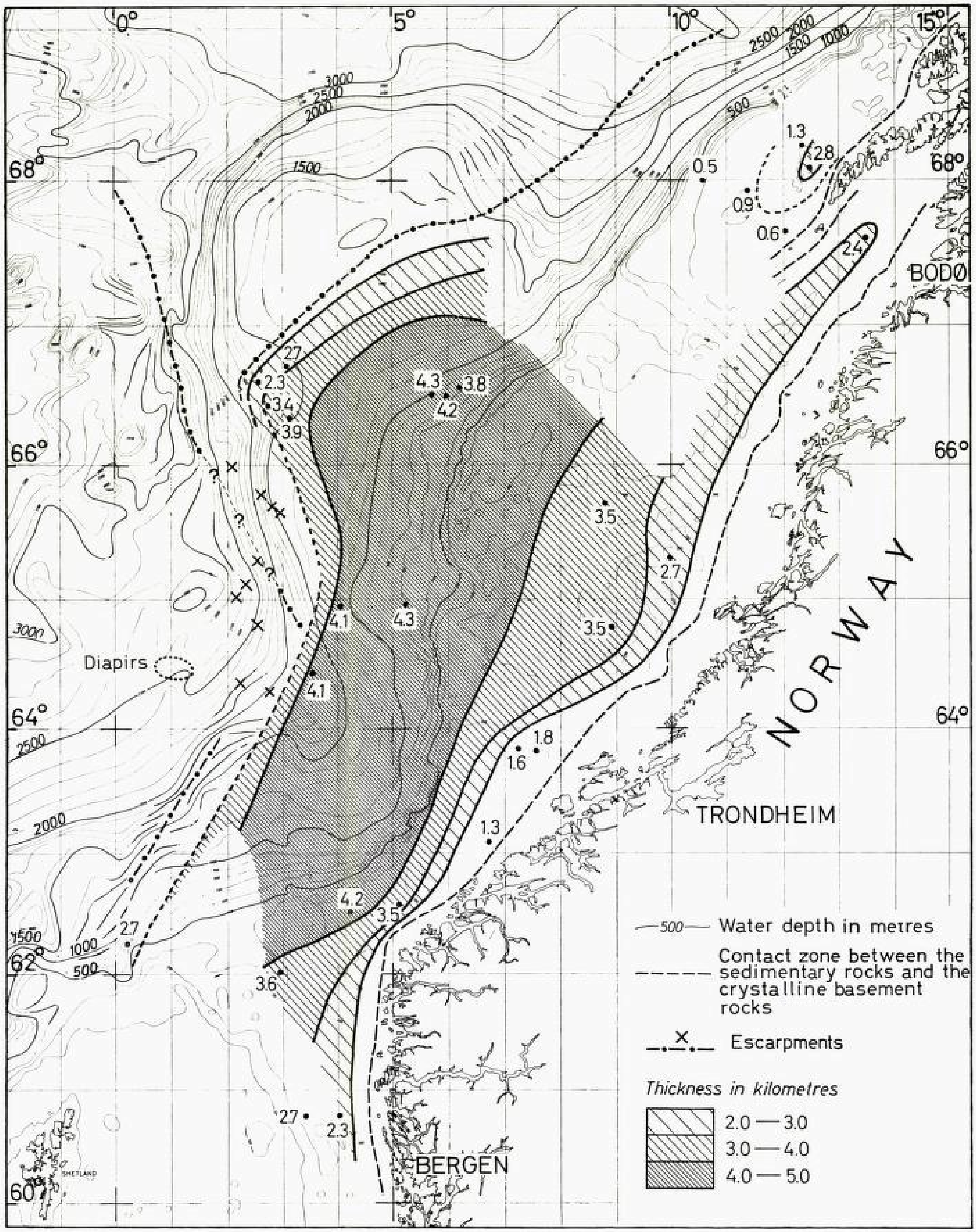


Fig. 11. Map showing total thickness of the Cenozoic and Mesozoic sequences.

smaller sedimentary basins are observed. As already mentioned, these two basins may once have been one single basin which has been divided into two basins by a horst which is now the Lofoten Islands.

The total thickness of the Cenozoic and Mesozoic sequences based upon the assumed relation between seismic velocity and geological age are shown in Fig. 11. This Figure shows an elongated sedimentary basin on the continental margin bounded to the east by the crystalline rocks along the Norwegian coast and to the west by the escarpments and the Jan Mayen Fracture Zone. The maximum thickness of the Cenozoic and Mesozoic sequences which has been observed is approximately 4.3 km, and a great area is covered by a sequence of more than 4 km in thickness.

Comments and remarks

Our seismic profiling has shown a well marked escarpment on the Vøring Plateau which coincides with the escarpment found by Talwani & Eldholm (1972) at all locations where we have crossed this escarpment. The Færoe-Shetland Escarpment has not been observed in all our crossings. Figs. 4, 8 and 11 show where we have observed the Færoe–Shetland Escarpments in the seismic sections.

The region between the southern end of the Vøring Plateau Escarpment seems to be strongly fractured and faulted, and we are not able to follow a single fracture zone along the southern part of the Vøring Plateau. More investigations are needed in this region to document and explain the complicated structure here.

There are many arguments that the crust west of Vøring Plateau Escarpment is oceanic. These arguments are supported by the basement velocity observed at shallow depth and by the typical, oceanic, magnetic features over the outer part of the Vøring Plateau. On the other hand it seems rather unlikely that an oceanic and continental crust should balance so well isostatically on the Vøring Plateau without any elevation difference on either side of the escarpment. The contrasting character of the magnetic and seismic velocities on both halves of the Vøring Plateau may be explained if we assume the western half to be covered by layers of flow basalt. This must be carefully investigated before the final conclusion can be drawn concerning the crustal structure beneath the Vøring Plateau. One more fact that should be taken into account is that the results obtained from seismic, magnetic and gravimetric measurements made by Bundesanstalt für Bodenforschung, Hannover, and the Seismological Observatory, Bergen University, Bergen, indicated that the western part of the Vøring Plateau is also underlain by continental crust (Hinz 1972).

The author fully realizes the limitation of determining the velocity-depth function by using the sonobuoy method as well as the limitation of the velocity-age correlation, but he considers that the seismic data used and the interpretation of these data reflect the main Cenozoic and Mesozoic structural trends on the continental margin off Norway.

The different isopach maps of the continental margin in the present paper show that subsidence and sedimentation have shifted from one part of the continental margin to another during the Cenozoic and Mesozoic time period.

The Pleistocene uplift of Fennoscandia is well 'recorded' by the dips (Fig. 4) which generally decrease from the mainland across the continental margin. Torske (1972) has discussed the relation between the uplift of Fennoscandia and the opening and spreading of the Norwegian Sea. His conclusion was that these two events were associated with each other. Talwani & Eldholm (1972) are also of the same opinion. It seems in general as if the dips are greatest in the two shelf areas between 62°N–64°N and 68°N–69°N (see Fig. 4). This should indicate a greater uplift in these two areas in relation to the shelf area in between. It is interesting to notice that these two areas are the most narrow shelf areas on the continental margin off Norway. It may be suggested that the Tertiary uplift may to some extent have caused the narrowing of these shelf areas by sliding, erosion and transportation of material from the outer part of the shelf. This may especially have been the case in the Storegga region where a great transportation of sedimentary 'material' has been observed downslope from the shelf edge (Sellevoll 1974). It is reasonable to assume that the shelf edge was previously located west of the Storegga region where it is located today, and that the uplift caused great mass-transport of sedimentary deposits downslope from the Storegga region. Thus the width of the shelf between the mainland and the Storegga region has been greatly reduced in the past and continues to be so reduced at present.

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Thickness and Distribution of Sedimentary Rocks in the Southern Barents Sea

EIRIK SUNDVOR

Sundvor, E. 1975: Thickness and distribution of sedimentary rocks in the southern Barents Sea. *Norges geol. Unders. 316*, 237–240.

The southern Barents Sea is underlain by sediments of varying thickness and physical properties. The thick sequence of Tertiary sedimentary rocks, which is representative of the southern provinces of the Norwegian continental margin, is found only as a wedge near the shelf edge of the Barents Sea. Structures that can be associated with salt diapirs are recorded in a limited area on Tromsøflaket.

Eirik Sundvor, Seismological Observatory, University of Bergen, N–5014 Bergen, Norway

The data presented in this short article, which is a summary of a longer paper (Sundvor 1974), show that the Barents Sea is covered by layered strata. A sediment isopach map has been constructed from all available data from the area (Plate 1). The structural information revealed by the seismic results indicates that the investigated part of the Barents Sea does not represent a uniform area. The results from various locations are therefore summarized in Table 1, in terms of average velocities.

The velocity distribution along the coast of Troms and Finnmark corresponds very well with the velocities measured by Eldholm (1970) on the southern part of the Norwegian shelf. Farther north in the Barents Sea the velocities are not significantly different from those reported by Eldholm & Ewing (1971). However, by using explosives as energy source it has been possible to penetrate the considerable thickness of sediments and record the basement velocity in the areas near the shelf edge.

In the main part of the Barents Sea, high seismic velocities (2.8–3.3 km/s) occur near the sea floor, while a low velocity (1.8–2.2 km/s) sedimentary wedge was recorded both in the shelf edge areas of the Barents Sea and along the coast of Troms and Finnmark. The seismic data confirm the suggestions made by Eldholm & Ewing (1971) and the regional geological model proposed by Frebold (1951) that the main part of the Barents shelf was emergent during the Tertiary. Eroded material from the uplifted landmass to the east was deposited in this period on the present shelf edge areas to form an upper section of prograded Cenozoic sedimentary rocks. The Paleozoic–Mesozoic succession in the main part of the Barents shelf was probably deposited during a changing pattern of transgressions and regressions in a shallow sea environ-

Table 1. Velocity structure of various areas of the continental shelf between Andøya and Bjørnøya (69°N–75°N)

Area	Average velocities (km/s)						
Coast of Troms and Finnmark	1.8	2.2	2.55	3.2	3.85		5.2
Southern Barents Sea (main shelf)			2.8	3.3	3.8	4.4	5.1
Southern Barents Sea (shelf edge)	1.75	2.15	2.65	3.4		4.35	5.05

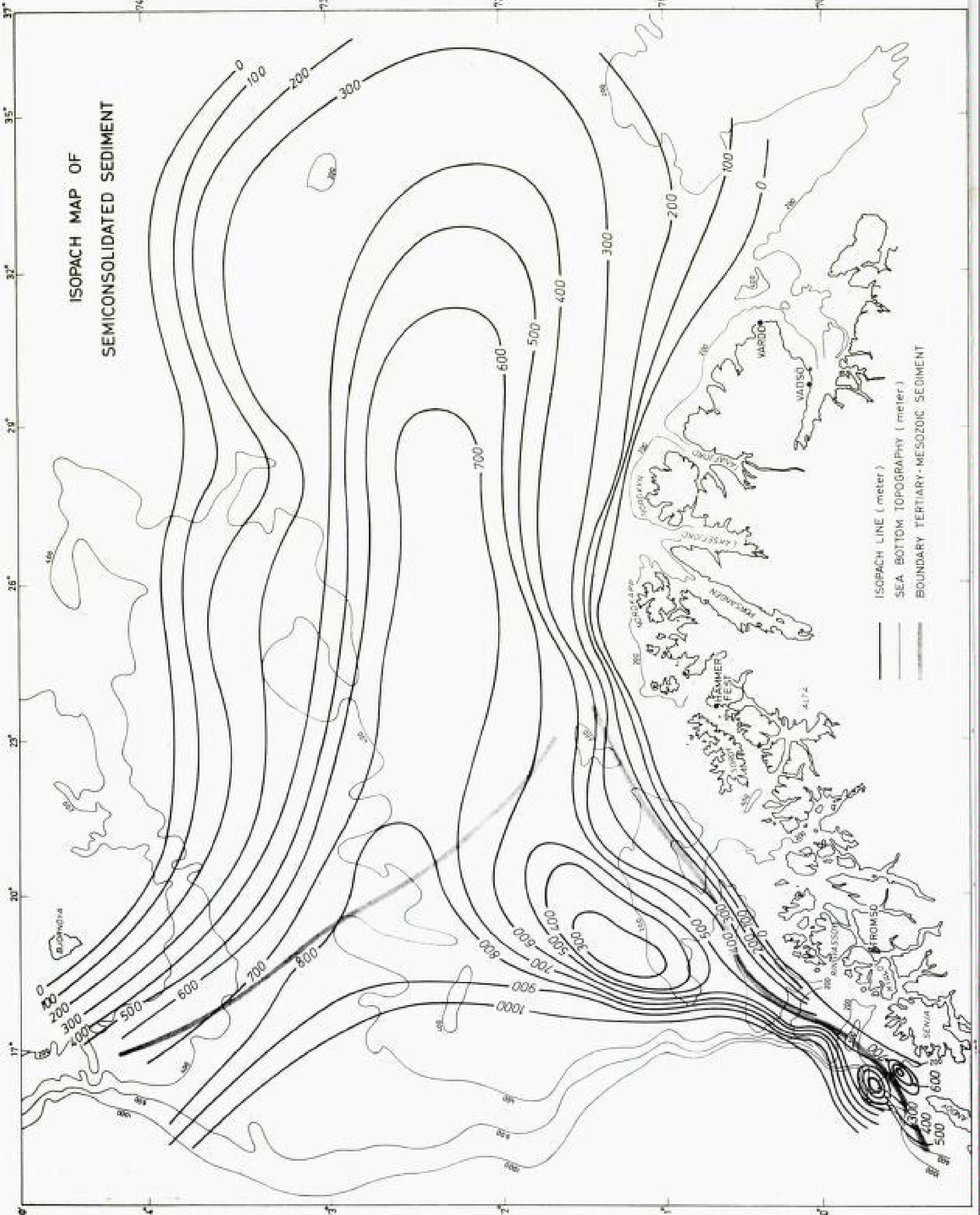

Fig. 1. Isopach map of semi-consolidated, probably Cretaceous sediment and outline of the base of Tertiary outcro

ment. A gentle westward-dipping basement surface is noted in the main part of the Barents shelf and there are no special structural highs in the basement topography which may reflect a continuation of the Caledonides through the western portions of the investigated area.

Previous studies in the North Sea have shown difficulties in correlating seismic velocities with the ages of Mesozoic rocks. The semi-consolidated layer in the southern Barents Sea with velocities from 2.65 to 2.8 km/s lies within the range of reported Cretaceous velocities. The continuous profiler data from the same area show that a prominent reflecting interface is present in many of the records. This layer corresponds very well with the refractor showing an increase in seismic velocities to values greater than 3.0 km/s. Fig. 1 was compiled on the basis of refraction and reflection data, and shows contours of the semi-consolidated, probably Cretaceous strata. Between the Norwegian coast and the northern parts of Bjørnøyrenna, these strata either crop out or, in some locations, are covered by relatively thin layers of Quaternary deposits, forming an apparent basin with a maximum thickness of 800 metres. Fig. 1 also outlines the areas of the base of Tertiary outcrop, dividing the southern Barents shelf into two different sedimentary provinces. West of the limit of Tertiary outcrop, the thickness of the semi-consolidated layer increases rapidly, except for the local areas north of Andøya and on Tromsøflaket where areas of a thinner sedimentary sequence are found.

In contrast to the southern provinces of the Norwegian continental shelf, some of the profiler records north of Andøya show the occurrence of faulting and tectonism. Locations of such structural features are shown in Fig. 2. The most prominent area of faulting is in the trough south of Bjørnøya. The records show that the areas of faulting and vertical movements are located in the southern parts of the trough, and that the sedimentary layers wedge out towards the north. All of these traverses lie east of the Tertiary boundary and the fault structures occur in the semi-consolidated, probably Cretaceous section. This may indicate a late Cretaceous or Tertiary age of the faulting.

The continuous reflection records show structures that appear to be diapiric in a limited area on Tromsøflaket (Fig. 2). These apparent intrusive structures occur at shallow depths and do not show internal reflecting horizons. A seismic velocity of 4.9 km/s is recorded in these structures in the refraction profile S13. The magnetometer records show no significant anomalies which can be associated with these structures, and the fact that the structures are located in an area of thick sedimentary deposits (Plate 1) strongly supports an assumption that they are sedimentary diapirs rather than uplifted crystalline basement. It has not been possible to identify the layer from which the diapiric structures have originated, but Fig. 1 shows that the semi-consolidated, probably Cretaceous sediments show a local decrease in thickness in the Tromsøflaket area. The thinned strata may very well be associated with the observed intrusive structures indicating that the structures are pre-Cretaceous in age.

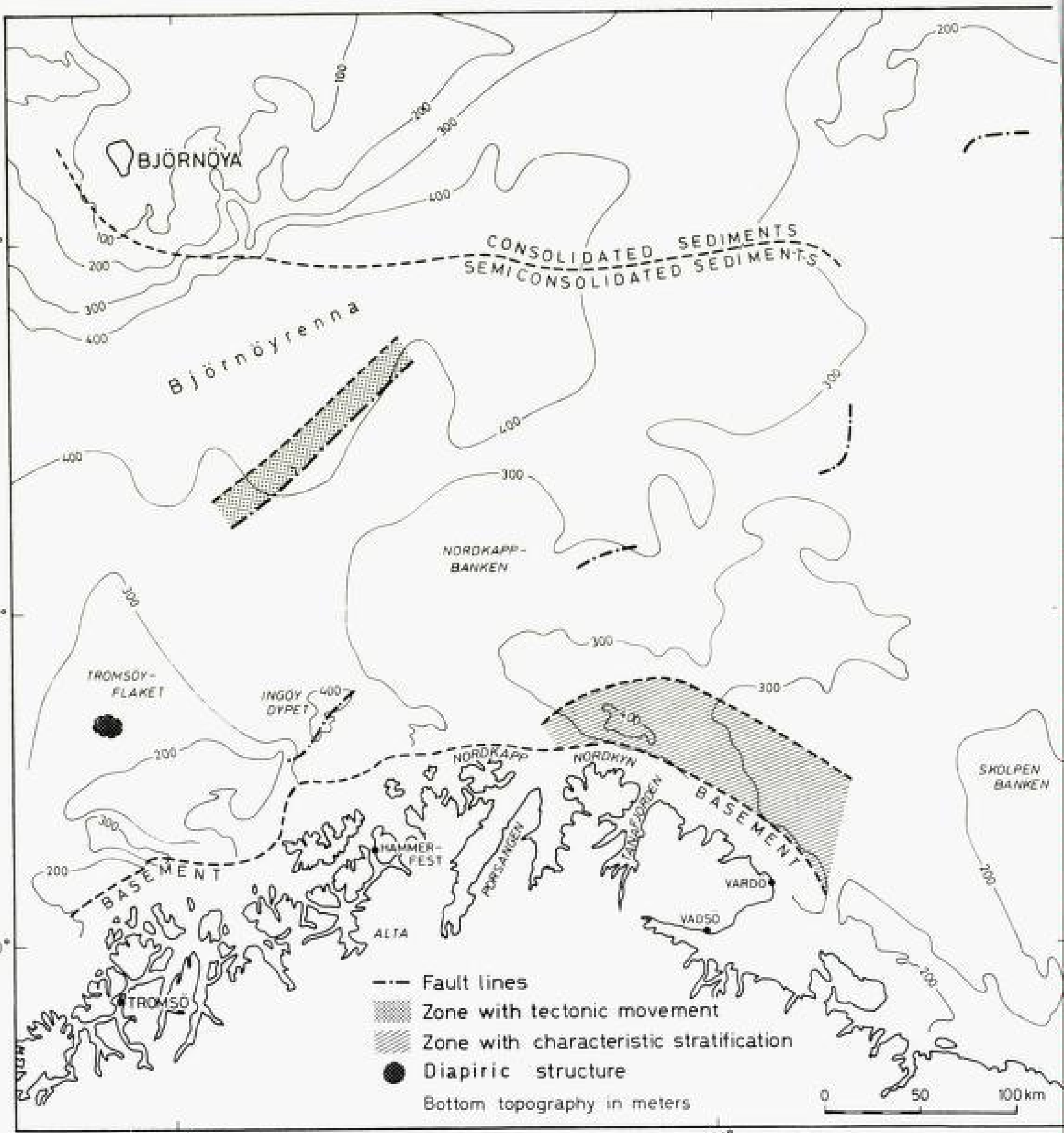

Fig. 2. Locations of structural features discussed in the text.

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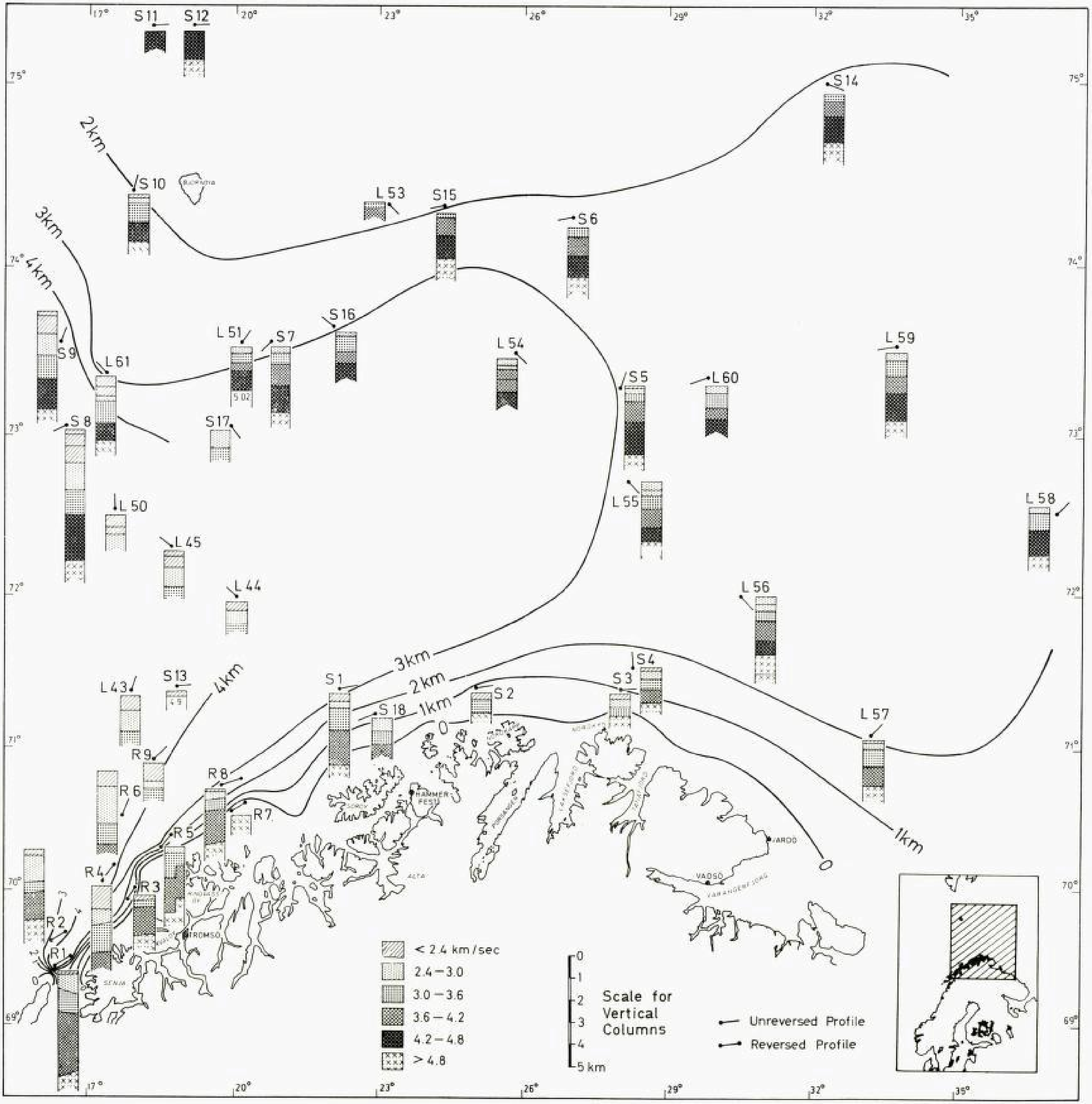


Plate 1. Map showing seismic refraction data and depths to acoustic basement.

Pleistocene and Recent Sediments of the Norwegian Continental Shelf (62°N-71°N), and the Norwegian Channel Area

H. HOLTEDAHL & K. BJERKLI

Holtedahl, H. & Bjerkli, K. 1975: Pleistocene and recent sediments of the Norwegian continental shelf (62°N–71°N) and the Norwegian Channel area. *Norges geol. Unders. 316*, 241–252.

The surface sediments covering the area under consideration can be divided into 4 groups: 1. Till, mainly composed of material derived from the adjacent mainland. 2. Till, mainly containing material derived from Mesozoic–Tertiary bedrock on the continental shelf. 3. Lag deposits from 1 and 2, mainly in shallow areas. 4. Secondarily transported sand and finer material winnowed out from 1 and 2, probably partly mixed with meltwater deposits, and deposited at intermediate and greater depths.

There is a clear relationship between depth and sediment texture. The thickness of post-Tertiary sediments varies between zero and about 400 m. In areas with thin cover, the glacial sediments have a lithology greatly influenced by the subsurface rocks.

H. Holtedahl and K. Bjerkli, Geologisk Institutt, avd. B, Universitetet i Bergen, N–5014 Bergen Univ., Norway

Introduction

Looking at the bathymetrical features of the Norwegian continental shelf, as well as the post-Tertiary sediments covering it, one has to consider the very special events which took place during the Pleistocene and Holocene epochs. We know that ice sheets developed in northern Europe several times during the Pleistocene, and that the mountainous areas of Norway and Sweden were glacial growth-centres for these ice sheets.

While the outer limits of the ice-covered areas during the various glaciations in the continental parts of northern Europe are fairly well established, they are far less clear in the areas now covered by the sea. Considering the Würm glaciation, the extension of the ice sheet in the present sea areas has been much disputed. Later papers and cartographical compilations often show separate ice sheets over Scandinavia, the British Isles and the Shetlands during the Würm, while these areas are assumed to have been covered by one continuous ice sheet during previous gleaciations. Some authors, however, have held the view that coalescence between the Scandinavian and British ice sheets did occur during the Last Glaciation.

From field studies in the Shetland islands, Hoppe (1970, 1972) and coworkers have given strong support to the view that this area was overridden by the Scandinavian ice during the Würm. From field work in the Svalbard area, especially from studies on isobases of uplift, they have also assumed the existence of an ice-centre situated in the central parts of the Barents Sea during the same period, and suggested the possibility that this ice-sheet was connected with the ice-sheet of northern Norway.

With regard to the western limit of the Scandinavian ice, various authors have expressed divergent opinions, and strong views, especially from biologists, have been put forward in favour of ice-free areas along the Norwegian west coast during the Würm glaciation, and possibly also during previous glaciations.

Most geologists, however, have had difficulties in accepting these views, and one of the main arguments has been the special bathymetrical features of the Norwegian continental shelf, which has a clear relation to glacial erosion and accumulation, a fact which was stressed by F. Nansen seventy years ago (Nansen 1904; also O. Holtedahl 1929, 1940; Shepard 1931, 1973; H. Holtedahl 1955; H. Holtedahl & Sellevoll 1971, 1972).

Some information on the types of sediments which are present on the Norwegian continental margin was obtained around the turn of the century by various oceanographic expeditions, which, however, spent most of their time exploring the deep sea (see H. Holtedahl 1955, p. 139). The information was extremely sparse, and one may say that our knowledge of the sediments on the Norwegian continental shelf before 1940 mainly came from the notations on the nautical charts obtained by the sounding work with plumb-lines by the Norwegian Hydrographic Office, as well as from information from fishermen.

In 1950, the first systematic investigations of the sediments on the Norwegian continental shelf, as well as geomorphological studies, were started at the University of Bergen, but it was not until 1968 that one was able to continue these studies as a result of growing interest in the Norwegian continental shelf, primarily due to oil and gas prospects in the North Sea area. In this work financial backing has mainly come from the Continental Shelf Division under the NTNF (Norwegian Research Council for Science and Technology). It may be mentioned that, recently, Soviet researchers have published some information on the sediment distribution on the Norwegian continental shelf (Litvin 1970).

Even though more systematic investigation of the surficial deposits on the Norwegian continental shelf has been carried out in later years, with the collection of a great number of sediment samples especially in the shelf-area off the coast of Møre–Trøndelag and off the coast of Troms, and to some extent in the Skagerrak area, there are still large areas which lack detailed information. This is especially true of the relatively deep parts of the continental shelf off the coast of Nordland, the shelf off Lofoten, and the Norwegian part of the North Sea.

An attempt to draw a sediment-distribution map of the Norwegian continental shelf from latitude 62°N up to latitude 71°N, and of the parts of the Norwegian continental shelf which include the Norwegian Channel, must therefore, to a great extent, still be carried out using, old chart notations and geological assumptions. Plate 1 shows a map which has been drawn from all the available information.

The inner continental shelf ('skjærgård') region

Along the entire Norwegian coast, with a few exceptions, is a submarine continuation of the coastal platform – the '*strandflat*' with the '*skjærgård*' region. This submarine area is extremely irregular, mainly rocky, and with a topography caused by glacial erosion in crystalline rocks with marked fracture patterns (O. Holtedahl 1940; H. Holtedahl 1960a, 1960b). Sediments are generally found in depressions, and in certain areas ice marginal accumulations are present, such as the Younger Dryas terminal moraine on the Skagerrak coast and a somewhat older terminal moraine along the south-west coast in the Lista–Jæren area, which can be traced as ridges or longitudinal accumulations over great distances (Andersen 1954, 1960; Klemsdal 1969). This latter coastal area is also one of the few where thick glacial deposits have been formed on land, covering large regions and reaching thicknesses of more than 100 m.

Some detailed work has been done in the submarine terminal moraine areas off the south coast by the senior author, and the distribution of sediments is shown in Fig. 1. The Younger Dryas moraine, which occurs as a very marked submarine ridge and which also appears above sea level as an island (Jomfruland), can be traced for very long distances. It is assumed to consist of boulder clay (as can be seen from subsurface excavations on the island Jomfruland), and has a covering lag-deposit of course material, mainly boulders, stones and gravel, down to a water depth of 50 m. Sand, washed out from the moraine, covers the bottom further out in a zone of varying width and depth. Further out, boulder clay appears in certain areas, and below a depth of about 100 m sandy and silty clay occurs.

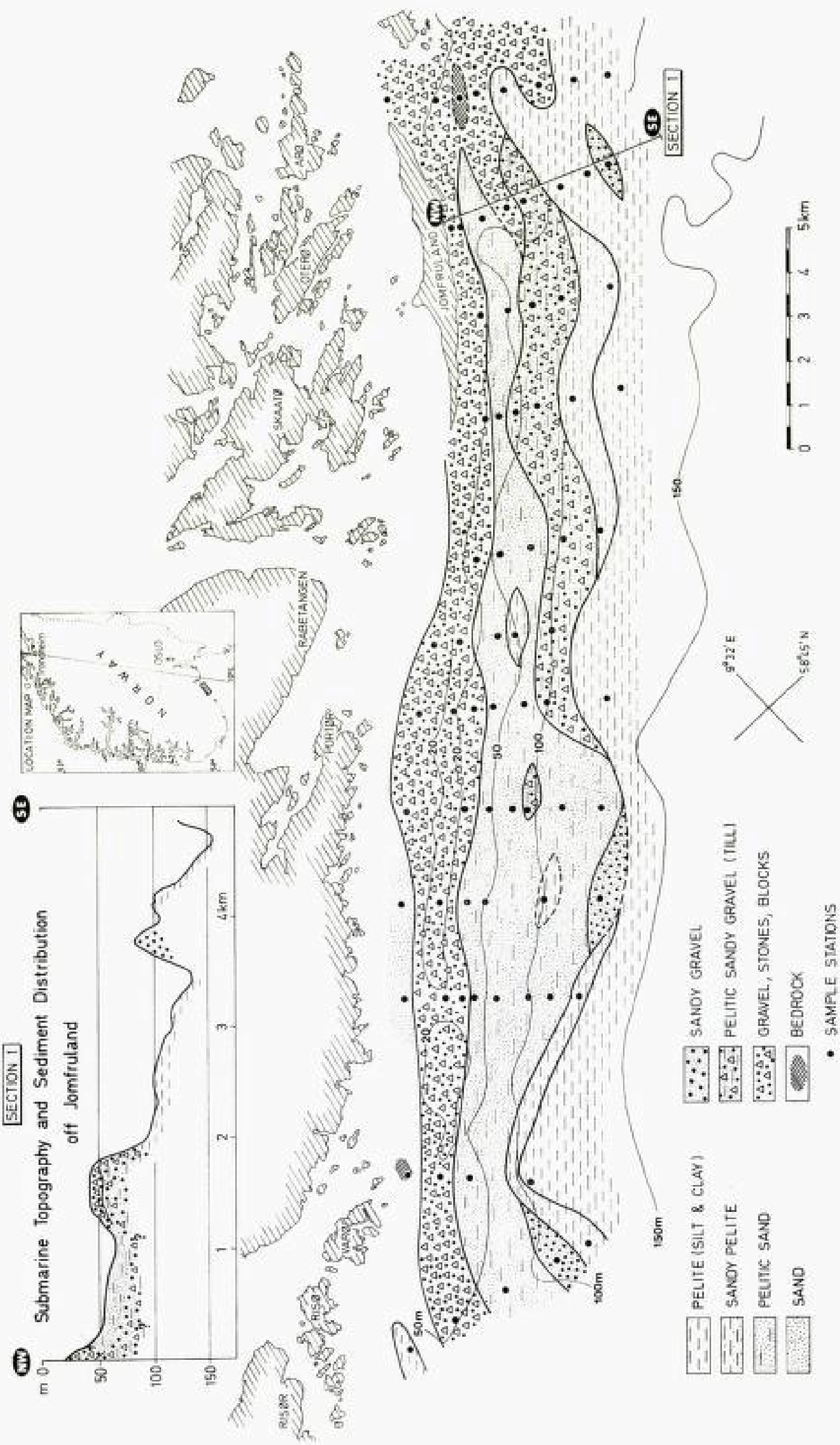
The Norwegian Channel area

With regard to the Norwegian Channel area, Quaternary deposits in the northern part of the Skagerrak have been shown by Sellevoll & Aalstad (1971) to attain a maximum thickness of about 380 m. As will be seen from the line-drawing from a seismic profile (Fig. 2a) the deepest part of the channel is covered by relatively thin deposits, while the thickest accumulations are present on the southern side towards Jutland. A maximum value of 380 m is not unreasonable, bearing in mind the thickness of some 220 m at Skagen in northern Jutland.

A recent acoustic reflection study by van Weering et al. (1973) in the Skagerrak and Norwegian Channel area further north, reveals several sedimentary units in the Quaternary accumulations (Fig. 2b).

The upper, most recent unit, which is acoustically transparent, is present in the deeper parts of the Channel, and has a maximum thickness of about 30 m in the Skagerrak, and 15–20 m off Egersund further west. Towards the sides it wedges out.

The upper unit rests on a well-stratified deposit which has about the same


Submarine Topography and Sediment Distribution off Jomfruland
SECTION 1
NW
SE
JOMFRULAND
RABETANGEN
LOCATION MAP
NORWAY
OSLO
PELITE (SILT & CLAY)
SANDY PELITE
PELITIC SAND
SAND
SANDY GRAVEL
PELITIC SANDY GRAVEL (TILL)
GRAVEL, STONES, BLOCKS
BEDROCK
SAMPLE STATIONS
5km
4km
50m
100m
150m
9°32'E
58°45'N

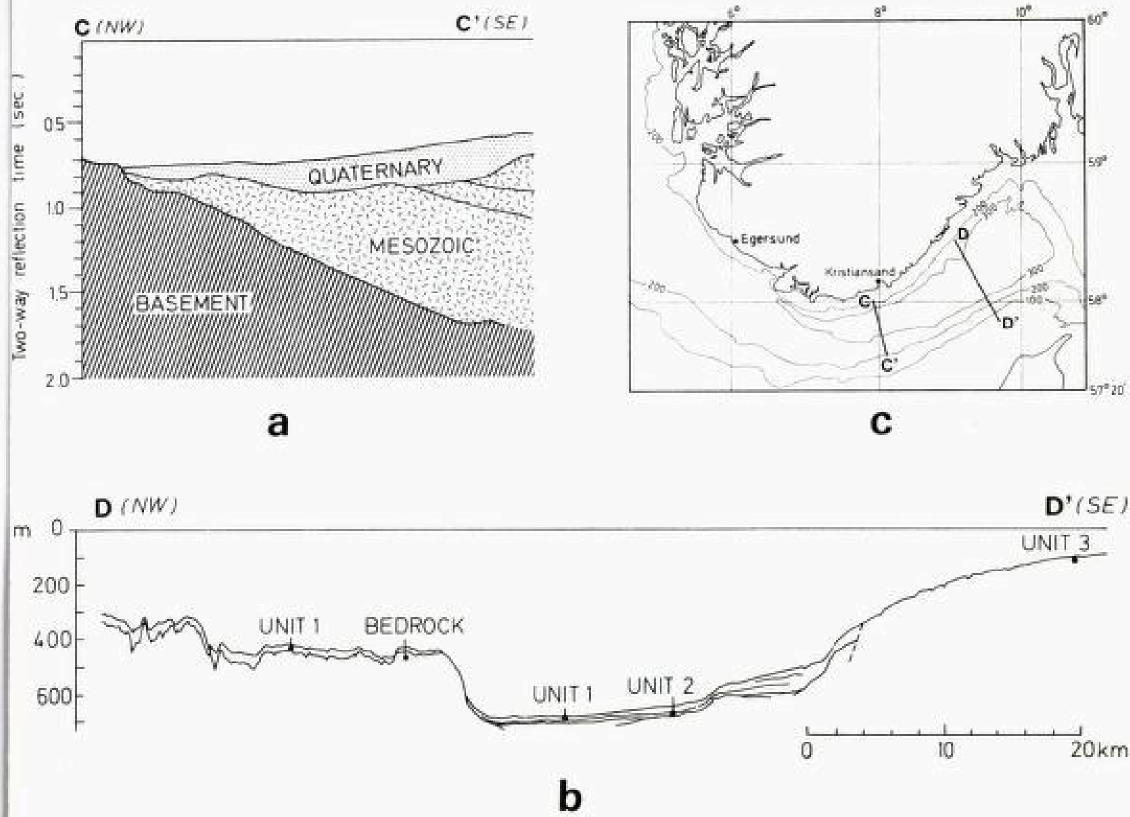


Fig. 2a. Line-drawing of seismic profile in Skagerrak south of Kristiansand. From Sellevoll & Aalstad (1971).
b. Line-drawing of acoustic profile in Skagerrak, south of Arendal. From van Weering et al. (1973).
c. Location map.

areal distribution, but which extends more to the east from Egersund and southwards. This lower unit has a greater thickness than the upper, with a maximum thickness of 90 m in the Skagerrak and more than 60 m off Egersund.

These two sedimentary units rest, with a few exceptions, on a thick deposit which has an irregular undulating surface giving a strong reflection and limited acoustic penetration. The thickness of this deposit, which is interpreted as glacial drift, could only be determined locally where a fourth sedimentary unit, or bedrock, could be traced underneath.

The most likely conclusion, according to van Weering et al. (1973), is that the supposed glacial drift found below the two upper units is of Würmian age, and deposited by a 'Skagerrak glacier' moving along the deepest part of the

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Fig. 1. Sediment map and profile of the submarine Younger Dryas terminal moraine and distal slope on the Skagerrak coast. Sediment classification based mainly on field studies of samples.

Channel, and that the two upper units represent deposition during the Holocene.

The view of the present authors, however, is that the glacial drift was not deposited by a 'Skagerrak glacier' moving *along* the Channel, but by glaciers moving *across* the Channel. This is in agreement with the views of Andersen (1964), Feyling-Hanssen (1964) and Sellevoll & Sundvor (1974). According to Andersen and Feyling-Hanssen, the deposits at Jæren, present up to 200 m above sea level, which were previously believed to have accumulated by a glacier moving along the south-west coast, are probably marine deposits containing coarse material dropped from icebergs.

A series of bottom samples was collected in the Skagerrak by the senior author in 1965 from the Naval Research vessel *H. U. Sverdrup*. The sediment map of Plate 1 is partly based on analysed surface samples from this cruise. A study of the foraminifera fauna in some of the cores was carried out by Kihle (1971). Lange (1956) found in the Skagerrak that at least the upper 10 m of the clays were deposited after 7,000 B.P., i.e. during the Holocene.

A recent seismic survey by Sellevoll & Sundvor (1974) in the Norwegian Channel west of Bergen and northwards (see also Flodén & Sellevoll 1972), shows Quaternary deposits with a thickness varying between 245 m and 45 m (1.85 m/sec.). A seismic profile along the central part of the northern part of the Channel shows a thickness of 220–240 m increasing to about 500 m approaching the shelf edge. It is interesting to note that the morphology of this part of the Norwegian Channel is due to varying thicknesses of Quaternary sediments, and not to erosional features of the underlying pre-Quaternary basement. The profiles show a number of continuous reflectors in the Quaternary deposits, but it is yet too early to interpret these features. Interesting is the observation by Sellevoll & Sundvor (1974) that the stratification in the northern part of the Channel indicates two main directions of transport, one a glacial transport across the trench, and a later mass transport of material towards the north, possibly by fluvial or glacifluvial action. If the stratified deposits, resting on the supposed foreset beds, are fluvial or glacifluvial, it requires a relative sea level depression of 500–600 m, an amount which is far greater than that concluded from the known facts about sea-level oscillations due to eustatic and isostatic movements during the Quaternary. This point obviously needs clarification, and it is hoped that drilling projects in connection with oil and gas exploration in this part of the North Sea may be able to give us some information.

Data concerning the surface sediments in the Norwegian Channel off the west coast of Norway are very sparse. A few core-samples off the Sognefjord at the bottom of the trench have been described by the senior author (H. Holtedahl 1955). The typical sediments were silty clays.

The sediments between 62°N and 71°N

The special morphological features which characterize the Norwegian continental shelf north of latitude 62°N are great variations of depth, irregularity of surface, submarine continuations of fjords, large trough-like depressions cutting across the shelf, and more or less narrow channels running almost parallel to the coast. The most shallow areas are also the narrowest ones; one is the bank-area off the coast of Møre, with depths mainly less than 200 m and a few banks even less than 100 m; and the other is the area off the Lofoten islands and the coast of Troms with similar depth conditions.

The submarine regions between these two relatively shallow areas have depths mainly greater than 200 m, and depressions down to depths of 500 m are known. It is obvious that depth variations of this order will have an effect on the sediments which have been deposited, and which are being deposited.

In the areas between latitude 62°N and 71°N (Plate 1) where information has been collected from echo-soundings, continuous seismic profiles, and from analysed sediment samples (grab, dredge, corer) from about 600 sampling stations, the surface sediments may be divided into several categories:

1. Till, mainly composed of material derived from the adjacent mainland.
2. Till, mainly containing material derived from Mesozoic–Tertiary bedrock on the continental shelf.
3. Lag deposits from 1 and 2, mainly in shallow areas.
4. Secondarily transported sand and finer material winnowed out from 1 and 2, and deposited at intermediate and greater depths. This material is probably partly mixed with meltwater deposits.

The distribution of the various categories of sediments reveals a clear relationship to depth. In the comparatively shallow areas off the coast of Møre–Romsdal and Lofoten–Troms, with depths mainly less than 200 m, the surface consists to a large extent of coarse lag deposits, mostly boulders, stones and gravel, but also sand which is secondarily transported by the action of waves and currents. Sand is also found at greater depths, and is also here supposedly secondarily transported.

On the deeper parts of the shelf, as well as on the continental slope, boulder clay occurs, showing little or no winnowing at the surface. This type of deposit is found to cover large areas of the bottom. In other deep parts silty clays are present, which to a great extent have been deposited during the later part of Würm and during the Holocene. Probably large proportions of this clay have been washed out from tills in more shallow water, but during the withdrawal of the ice-front from the continental shelf area, a large amount of fine material from meltwater must have settled. It is believed, however, that very little of the fine particles brought to the sea from land during the major part of the Holocene actually settled on the continental shelf. Most of this material presumably settled in the fjords, and the very fine fraction bypassed the shelf (H. Holtedahl 1965).

The thickness of the Quaternary deposits of the continental shelf north of latitude 62°N outside the '*strandflat*' shows great variation, according to the seismic investigations carried out by the Seismological Observatory at the University of Bergen (Eldholm & Nysæther 1969; Nysæther et al. 1969; H. Holtedahl & Sellevoll 1971, 1972). Off the coast of Trøndelag and Nordland the sediment cover varies in thickness from zero to about 400 m. The greatest thicknesses are found in the peripheral areas close to the continental slope, while areas with sparse accumulations are especially noticeable between Haltenbanken and Frøyabanken.

The present submarine topography, with great variations in depth, can be ascribed partly to an unevenly eroded sub-Quaternary surface, and partly to the varying thickness of the Quaternary sediments.

Glacier movement along the entire coastline has been more or less normal to the coast and the main ice-movement on the continental shelf between latitudes 62°N and 71°N during the maximum extent of the glaciation was directed towards the west and north-west.

LITHOLOGY

The lithological composition of the Quaternary surface sediments which have been sampled on the continental shelf between 62°N and 71°N is clearly related to the total thickness of the Quaternary deposits. In areas of great thickness, the material is primarily derived from the crystalline rocks on the mainland and consists of rocks and mineral grains, including clay minerals, typical of glacial deposits on land. An admixture of types, obviously dropped from icebergs, and with a long transport to a large extent from the Oslo area, has been shown to occur (H. Holtedahl 1955).

In areas of thin Quaternary cover, the deposits contain a large percentage of sedimentary rocks of Tertiary and Jurassic–Cretaceous age (H. Holtedahl & Sellevoll 1971), to a large extent unmetamorphosed sandstones, claystones and limestones. Of macro-fossils belemnite rostra were commonly found in the dredge samples. This material has presumably been picked up by the glaciers from the underlying bedrocks, and transported in the direction of ice-flow over a distance which may in certain cases have been very short.

A sediment core 4.5 m long, sampled on the northwest slope of a submarine ridge between Haltenbanken and Frøyabanken at a depth of 310 m, consisted of an upper, mainly postglacial clay, 150 cm thick, and a clayey deposit with more or less disintegrated sedimentary rock fragments and a high organic content, supposedly representing a glacial till (Fig. 3). The fossil assemblage of the rock fragments, as well as of the matrix, was of Middle and Upper Jurassic, Cretaceous and Lower Tertiary age. Furthermore, the clay mineral assemblage differed markedly from the assemblage in the upper part, in having a high content of kaolinite (30–40%) and montmorillonite (30–40%). This assemblage was clearly related to clay minerals in the nearby pre-Quaternary rocks. The major part of the material of till was considered to have had a fairly short transport, thus giving an indication of the bedrock lithology to the east of the sampling station (H. Holtedahl et al. 1974).

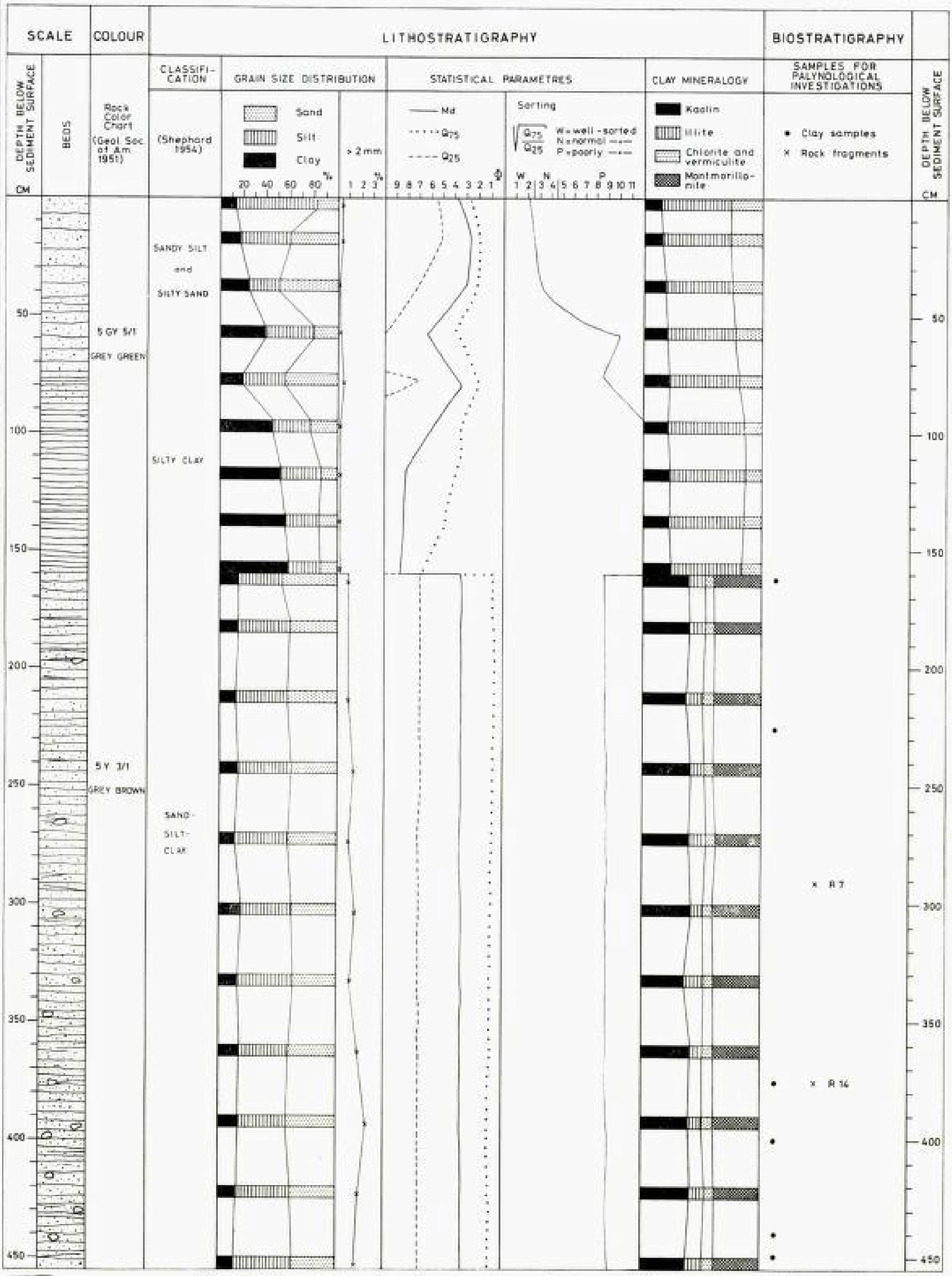

Fig. 3. Lithostratigraphy of sediment core between Haltenbanken and Frøyabanken. From Holtedahl et al. 1974).

Another relationship between components of Quaternary deposits and the underlying sedimentary rocks in thinly covered areas has been demonstrated by the junior author (Bjerkli 1972) in the distribution of certain minerals of the sand fraction. Bjerkli has found a concentration of glauconite in an area between Haltenbanken and Frøyabanken, which points to a source in the underlying Mesozoic and Tertiary rocks.

A lithological analysis of the components of the Quaternary sediment cover is, according to these investigations, a good indicator of the thickness of this cover, as well as of the composition of the underlying pre-Quaternary bedrock surface.

Special textural investigations on quartz grains from Quaternary sediments off the coast of Møre have been carried out by Strass (1973). By using scanning electron microscopy, she was able to distinguish grains with a different origin and transport history. According to Strass, the very well-rounded and frosted quartz grains, which occur as a minor part of the glacial sediments on the continental shelf off Møre, have not acquired their eolian texture on dry bank areas during the Pleistocene, as has been suggested by the senior author (H. Holtedahl 1955), but are probably derived from eolian sandstones of the underlying pre-Quaternary bedrock surface.

General consideration of Quaternary history

The Norwegian continental shelf, including the Channel area in the North Sea and the Skagerrak, has been greatly influenced by glacial activities during the Pleistocene. It is believed that the inland ice, which during its maximum had its culmination in the central parts of the Fennoscandian area, covered the entire continental shelf. During other stages, the high mountains of especially the western parts of Norway were ice accumulation centres, producing glaciers which extended more locally over the continental shelf to varying distances from the coast. The great depths, present especially near the coast in the Norwegian Channel area as well as off the west coast, are probably due to heavy glacial erosion in rocks with limited resistance. The increased thickness of Quaternary sediments towards the peripheral areas is natural when these regions are considered as the peripheral parts of the continental ice sheet.

Sea level oscillations on the continental shelf have no doubt been extensive during various phases of the Pleistocene, and have greatly effected sedimentation. At the outer coast, especially the west coast, caves formed by sea abrasion, and other sea-abrasional features, at levels high above the maximum sea level during the final disappearance of the ice in Late Würm times, are commonly found. Along the coast of Møre, signs of sea abrasion are distinct up to levels of 80 m above present sea level, while the Late Würm marine limit is about 20 m (H. Holtedahl 1960a, 1960b). This indicates a considerable isostatic depression of the crust in these coastal areas during the maximum of glaciation, which again indicates a considerable ice thickness. A great isostatic depression of large parts of the continental shelf must therefore be assumed, interfering with the general eustatic lowering of sea level.

High former sea levels, far above the Late Würm marine limit, are also known from the Jæren–Sandnes area on the south-west coast, where supposed glacial-marine clays are found up to a level of 200 m. In this area Late Würm marine limits are only 10–20 m, and as mentioned above the clays have previously been reckoned as deposits from a 'Skagerrak glacier'.

As to indications of lower sea levels on the continental shelf, the senior author has proviously pointed to a very marked increase in rounded stones and gravel at a depth of about 100 m, suggesting that banks less than this depth have been dry during the later parts of the Würm glaciation (H. Holtedahl 1955). The study of textures on quartz grains by scanning electron microscopy, carried out by Strass (1973), has supported this assumption.

A lowering of sea level to the extent which was thought necessary to explain the assumed foreset–topset structures in the northern part of the Norwegian Channel has not been verified by studies on the continental shelf north of the Channel. In any case, the relative sea level has shown great oscillations and the crustal movements have been extensive, due primarily to the glacial isostatic effect. To what extent, if any, these verical movements of the crust have influenced the migration of oil, has yet to be seen.

Acknowledgements. – The present contribution is to a great extent based on surveys carried out by the Geologisk Institutt, Avd. B, and the Seismological Observatory, University of Bergen, on various cruises. Financial support has been given by NAVF, NTNFK, NATO and Elkem A/S. Furthermore, field surveys have been carried out in co-operation with the Institute of Oceanography, University of Bergen, and the Institute of Ocean Research, Directory of Fisheries, Bergen. The support of all these institutions is greatly appreciated. A number of people, among them persons attached to the Geologisk Institutt, Avd. B, have taken part in the collection of data at sea, and in the laboratory studies. To all these we extend our thanks.

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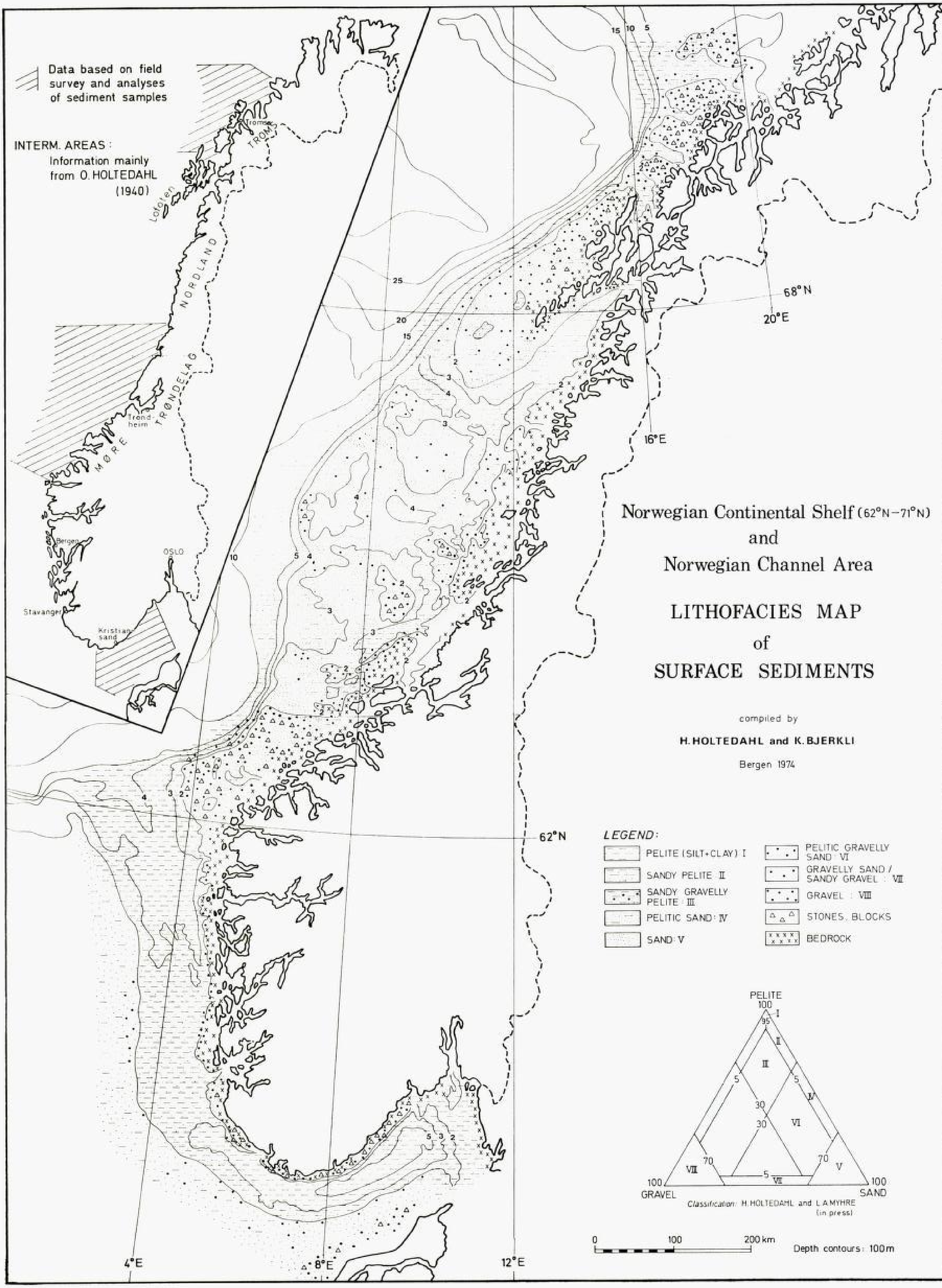


Plate 1. Sediment map showing lithofacies of surface sediments. Areas where information is based on field surveys and detailed analyses of collected sediment samples (grab, dredge, core) are marked on the inset map. Information from other areas mainly from O. Holtedahl (1940).

Geological Investigation of a Lower Tertiary/Quaternary Core, Offshore Trøndelag, Norway*

T. BUGGE, M. LØFALDLI, G. H. MAISEY, K. ROKOENGEN, F. E. SKAAR & B. THUSU

Bugge, T., Løfaldli, M., Maisey, G. H., Rokoengen, K., Skaar, F. E. & Thusu, B. 1975: Geological investigation of a Lower Tertiary/Quaternary core, offshore Trøndelag, Norway. *Norges geol. Unders. 316*, 253–269.

Seismic profiling carried out offshore Trøndelag, Norway, has provided a basis for selecting localities for core sampling of the bedrock.

A continuous, metre-long core revealing new information was recorded. Sedimentological, mineralogical, geotechnical and micropaleontological investigations indicate that the core consists of near-shore, silty claystone of Upper Cretaceous/Lower Tertiary age at the base, of Quaternary till in the middle, and of a Late Quaternary cover sand at the top.

T. Bugge, M. Løfaldli, G. H. Maisey, K. Rokoengen, F. E. Skaar & B. Thusu, Continental Shelf Division of the Royal Norwegian Council for Scientific and Industrial Research (NTNFK), Hoffsveien 13, Oslo 2, Norway

Previous work

The Continental Shelf adjacent to Trøndelag has been subject of several scientific investigations. Holtedahl (1940) has described the bathymetry of the Norwegian coastal zone. Holtedahl (1955) discussed the glacial history of the shelf area off Møre, and concluded that during the last glaciation, the ice-sheet must have extended at least 40 km from the present coast. Later investigations showed that the Quaternary sediments are thin over large areas (Eldholm & Nysæter 1969). Holtedahl & Sellevoll (1971) pointed out that in some sediment samples from the area there is a high content of local, short-transported, Mesozoic material. Bjerkli (1972) has made mineralogical analyses of the surface sediments, and Haldorsen (1974) has investigated the mineralogy of sediment cores. Bjerkli & Østmo-Sæther (1973) have found authigenic glauconite inside shells of foraminifera west of Frøya, Trøndelag. Studies of marine geology, mineralogy and palynology have been carried out by Holtedahl et al. (1974).

Several geophysical investigations including estimations of the thickness of sediments have also been carried out (Grønlie & Ramberg 1970; Åm 1970; Talwani & Eldholm 1972).

* Publication No. 48 in the NTNF Continental Shelf Project.

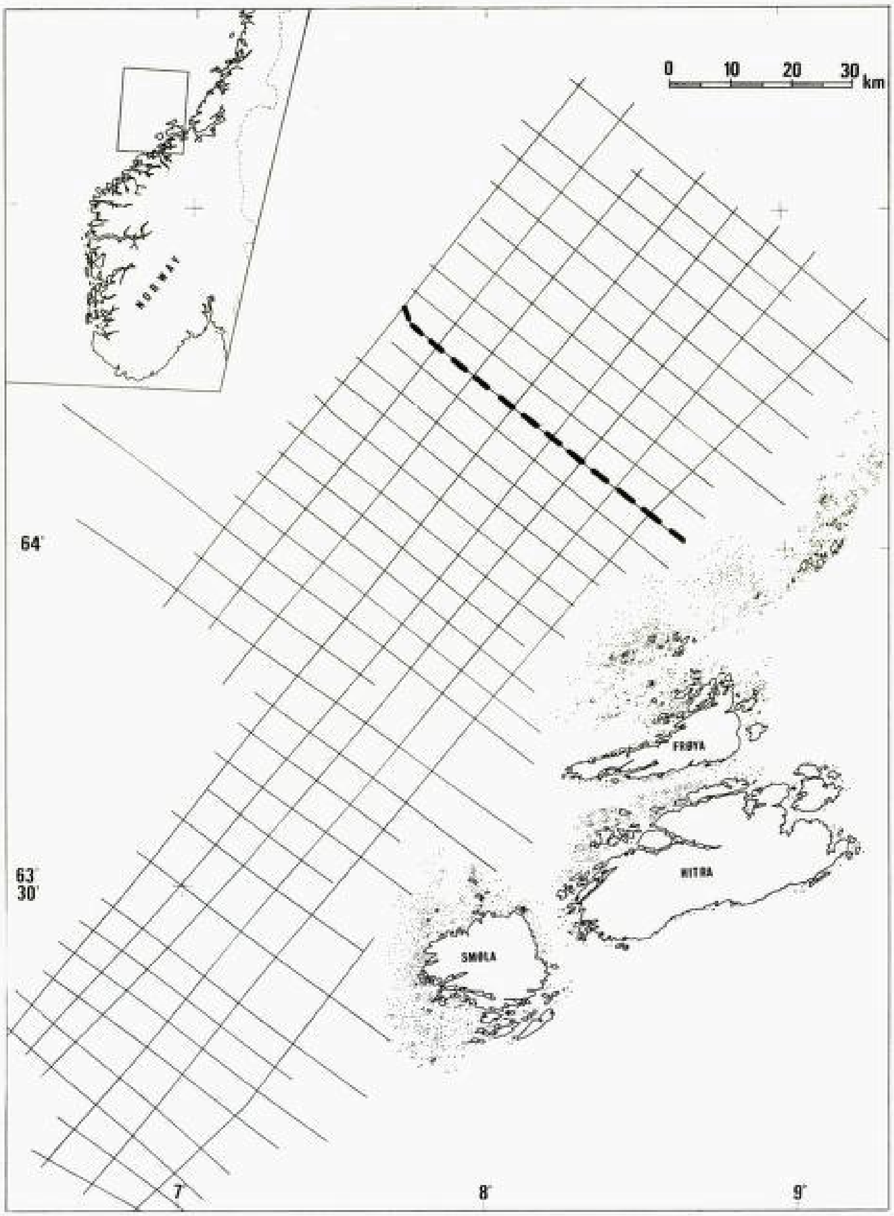


Fig. 1. Sparker and boomer profiling carried out by NTNFK 1973, offshore Trøndelag, Norway. Dashed line is the profile shown in Fig. 2.

Geological setting

NTNFK has carried out geological and geophysical studies on the continental shelf north of 62° since 1969 and has also supported and coordinated investigations carried out by other institutions.

In June 1973 a grid of about 3,500 km of sparker and boomer profiles was run offshore Trøndelag (Fig. 1). The distance between the profiles

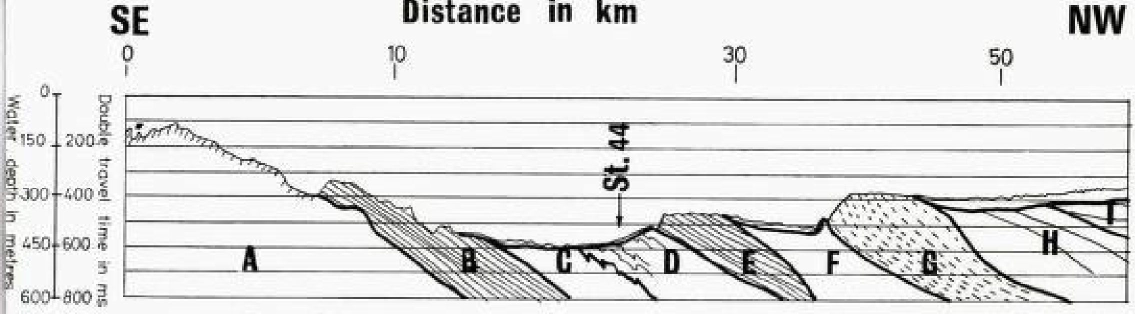


Fig. 2. Schematic interpretation of sparker profile. Letters indicate the various identifiable stratigraphical units. The thin top layer is Quaternary deposits.

is about 5 km. The profiles showed a number of identifiable rock units. A schematic interpretation of a sparker profile is shown in Fig. 2. Formation A, basement, is easy to recognize from its uneven top surface and typical parabolical reflection patterns. Formation B has thin and even layers that are only slightly disturbed. Formation C has no visible layering. Formation D has undergone block-faulting with offset in the order of about 30 m vertically. Formation E is layered and forms an elevation that can be traced topographically along strike for some distance. Formation F has no clear layering. Formation G has distinct foreset bedding and forms a topographical elevation in parts of the area. A more complete sequence to the west shows a topset, foreset and bottomset element. This formation has previously been mapped and was tentatively called a 'delta' (Eldholm & Nysæther 1969). Formation H has layers partly parallel to the delta. Formation I is discordant to H, and the layers are nearly horizontal.

These formations could be recognized in adjacent profiles and a preliminary geological map was constructed (Fig. 3). The Quaternary deposits are not yet mapped in detail, but some possible windows through the Quaternary cover have been located. It was decided to piston core these windows to obtain samples of the bedrock on a subsequent cruise in September.

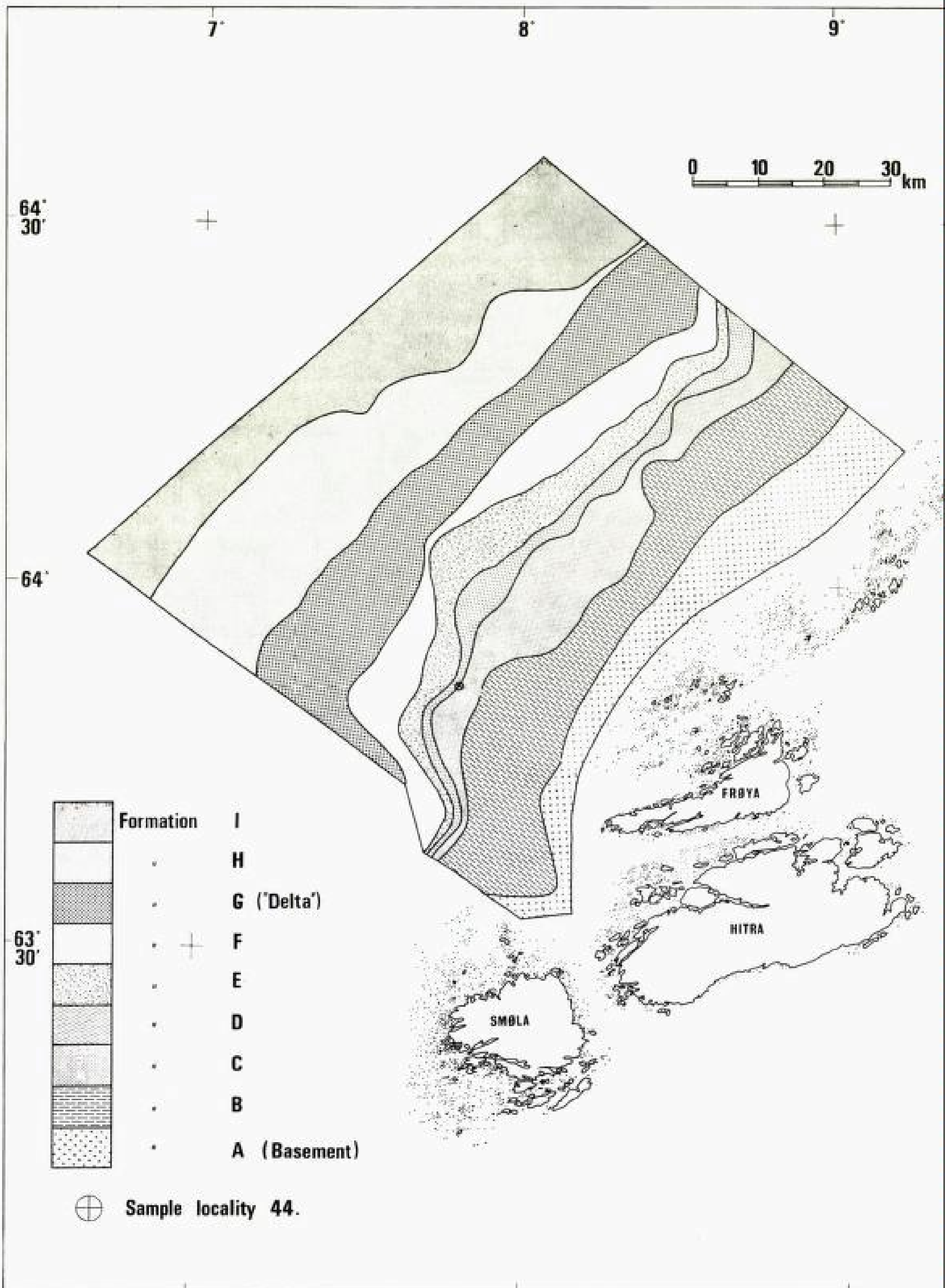
Sampling

One of the selected localities was No. 44 on cruise 5 (station 44/7305).

Bottom photographs of the selected site showed a bioturbated sandy bottom with scattered pebbles (Fig. 5). A dredge sample yielded about 15 litres of sandy mud which contained a handful of pebbles.

Table 1. Composition of pebbles in dredge sample

Rock type	Number	No. %	Weight grams	Wt. %
Magmatic	69	37	150	40
Gneiss	75	40	124	33
Sediments	34	18	89	24
Chert	9	5	14	4
Undetermined	5		21	



7°
8°
9°
64° 30'
64°
63° 30'
0 10 20 30 km
Formation I
H
G ('Delta')
F
E
D
C
B
A (Basement)
Sample locality 44.
FRØYA
HITRA
SMØLA

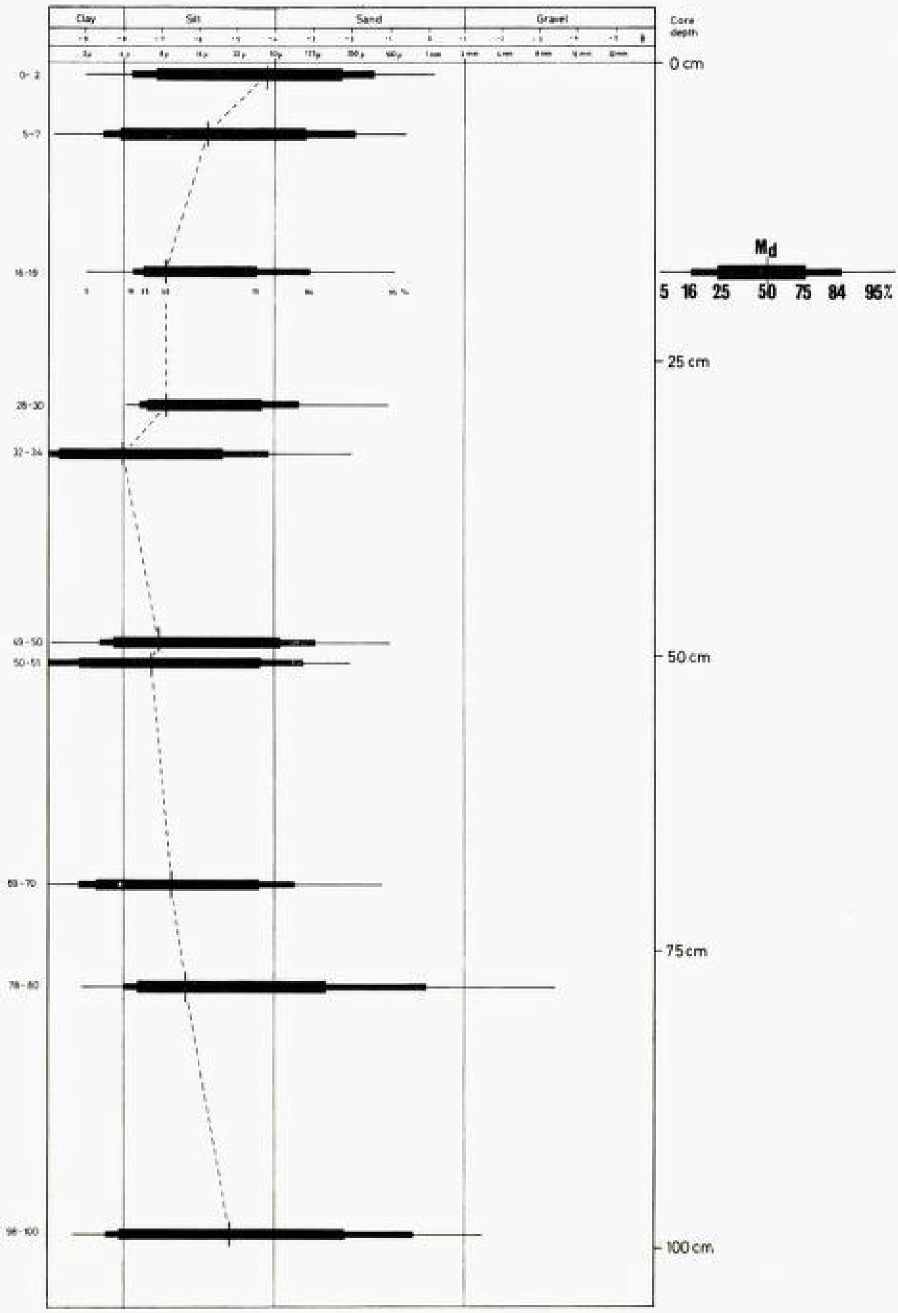


Fig. 4. Grain-size distribution, Station 44.

Fig. 3. Preliminary geological map of stratigraphic units, offshore Trøndelag, Norway.

Some pebbles consist of chert and sandstones foreign to the mainland rock types (Table 1). The degree of roundness varies from angular to well-rounded with about 80% subangular to subrounded.

At 63°52.0'N, 7°49.0'E a piston corer was tried at a depth of 230 m. The corer came up severely bent with about 5 cm of hard clay protruding from the nose. Unfortunately, bad weather prevented further sampling of the selected localities.

The core consists of a top layer of silty sand (c. 10 cm). Between c. 10 and c. 65 cm is silty clay with alternating greenish and brownish coloration. The lowermost part (c. 40 cm) is a hard clay with green glauconite grains. Scattered throughout the core are pebbles of brittle claystone up to about 1 cm in diameter with a dark, sometimes chocolate-brown colour.

Laboratory investigations

The following properties of the core have been investigated: grain-size, mineralogy, geotechnical properties, micropaleontology, chemistry of interstitial water and carbon content. A partial account of the results is given below. It is planned to present more detailed information in a subsequent NTNFK report.

Grain-size distribution

Ten levels in the core were selected for grain-size and mineralogical analyses. Claystone pebbles from three levels (0–2 cm, 50–51 cm and 98–100 cm) were analysed separately. The grain-size distribution (Fig. 4) was determined by wet sieve analysis and by pipette analysis of the fraction between 63 μm and 2 μm. The claystone pebbles were treated as grains.

The analyses show that the sediment is a poorly sorted sandy silt in the top and the bottom layers, and a poorly sorted silty clay in the middle part of the core. The median diameter for all examined samples is within the silt-size range.

Mineralogy

BULK SAMPLES

One thin-section was made from the 98 cm level by impregnating the claystone with Epofix resin. Quartz, plagioclase, amphibole, glauconite and K–feldspar are the principal minerals; in addition, fine-grained clay-minerals are present.

The quartz grains are angular and fairly uniform in size, about 50–200 μm. Some of them show secondary growth along the rims.

The claystone pebbles have distinct boundaries and consist of a uniform mixture of fine-grained clay-minerals and quartz. The quartz grains are very small and evenly distributed, and show a preferred optical orientation in each individual pebble.

Differential thermal analyses, X-ray diffractometer analyses and staining tests (Mielenz & King 1951) on bulk samples at the 25 cm and 85 cm levels show

Fig. 5. Photograph of the bottom at Station 44. Trigger weight and release wire are seen in the upper right.

small differences between the two levels. The samples consist mainly of feldspar and mica/hydromica, with about 10% quartz and less than 5% montmorillonite. The upper sample contains some calcite.

FRACTION FINER THAN 63 μm

This fraction has been examined by means of X-ray diffraction of randomly orientated specimens (Fig. 6). The samples were treated with ethyleneglycol and heated to 450°C. Montmorillonite was identified by expansion of the 14Å reflection to 17Å in glycolated samples and a collapse of the same reflection to 10Å in the heated sample.

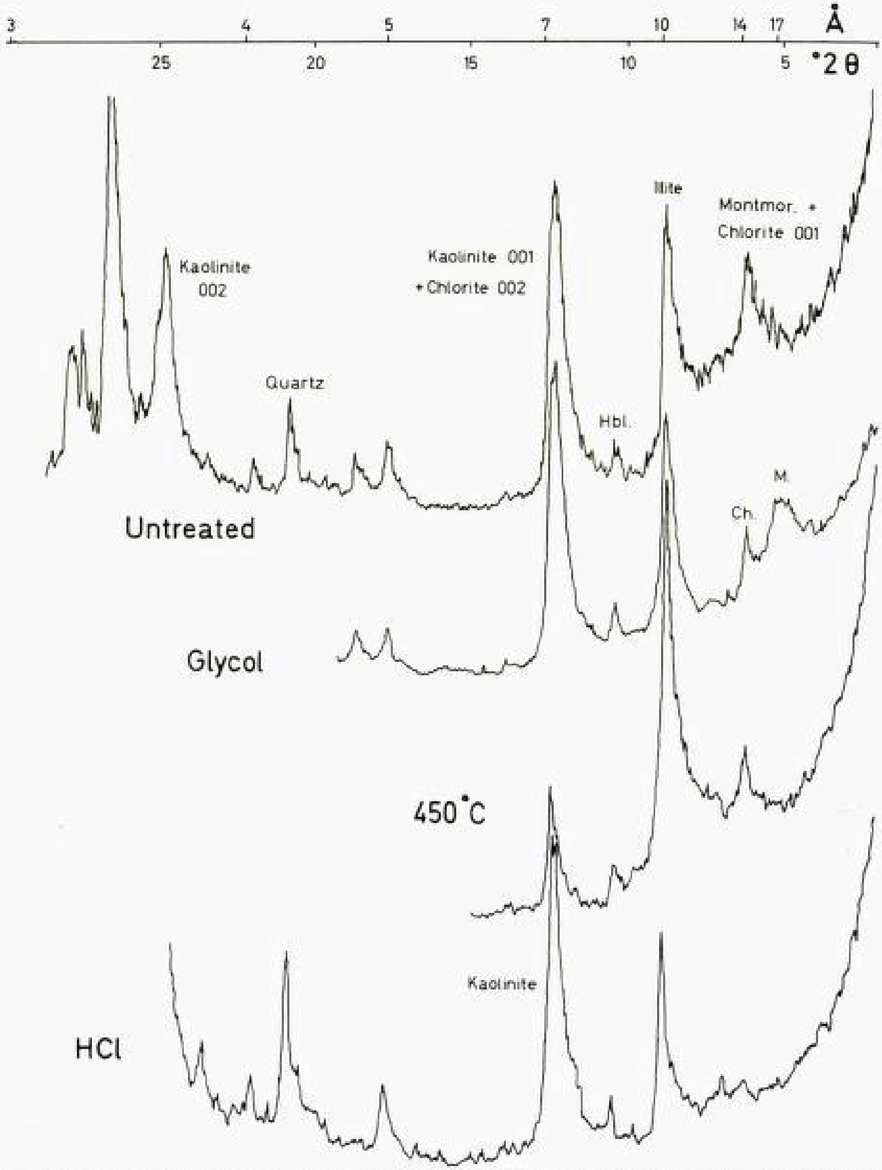


Fig. 6. X-ray diffractograms of unoriented samples of the fraction < 63 μm, 78–80 cm core depth, $CuK\alpha_1$-radiation.

The sediment consists of illite, quartz, plagioclase, K-feldspar, montmorillonite, kaolinite, chlorite and amphiboles, in decreasing order of abundance. The content of chlorite increases upwards in the core. Towards the top the amount of calcite, in the form of animal tests, increases.

The claystone pebbles consist mostly of montmorillonite. The degree of crystallinity decreases in the uppermost pebbles where the montmorillonite is replaced by an illite–montmorillonite mixed-layer mineral. In addition to the

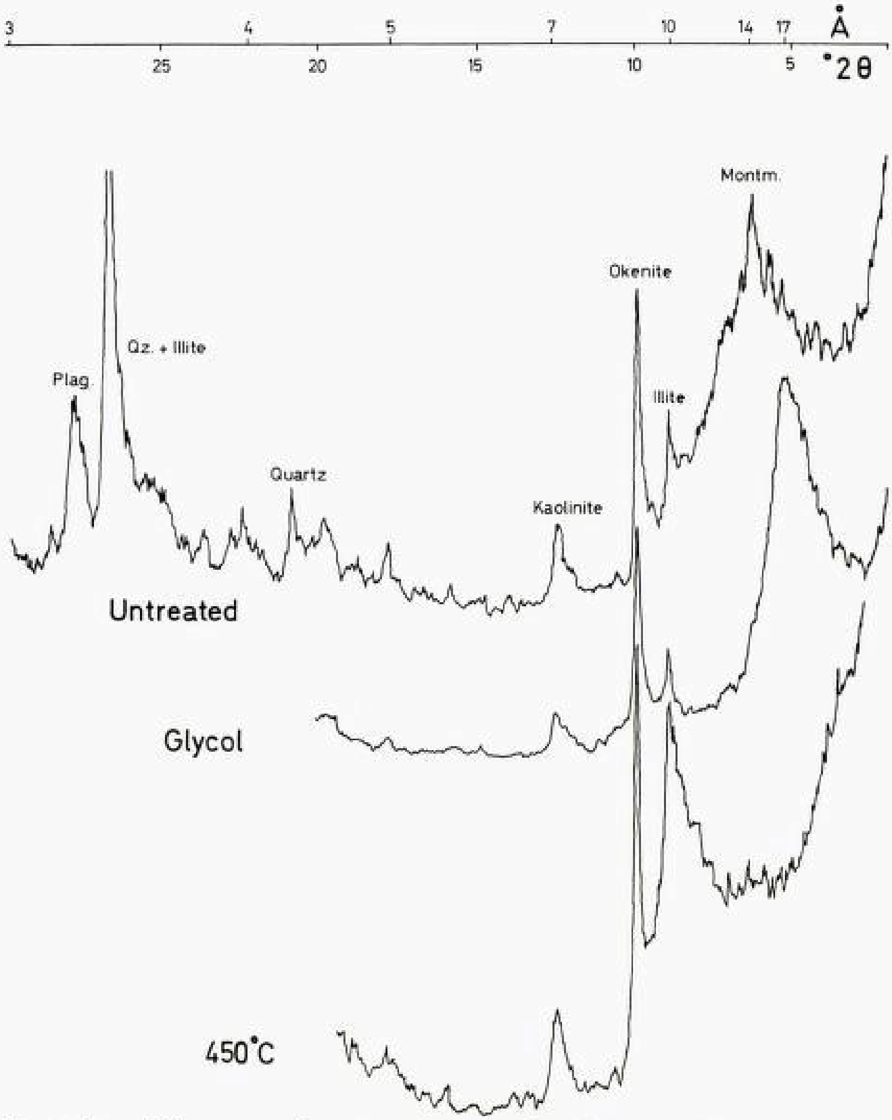


Fig. 7. X-ray diffractograms of unoriented samples of a claystone pebble at 99 cm core depth, CuKα_1-radiation.

montmorillonite there is a zeolite mineral, okenite, with a distinct 8.91Å reflection. Quartz, illite and kaolinite are also present in minor amounts (Fig. 7).

DISCUSSION

The mineral composition of the upper part of the core is somewhat different from that of typical marine clays of Quaternary age in Norway (Roaldset 1972). The fairly uniform composition of the montmorillonite pebbles throughout the core indicates the reworking of older material: the kaolinite

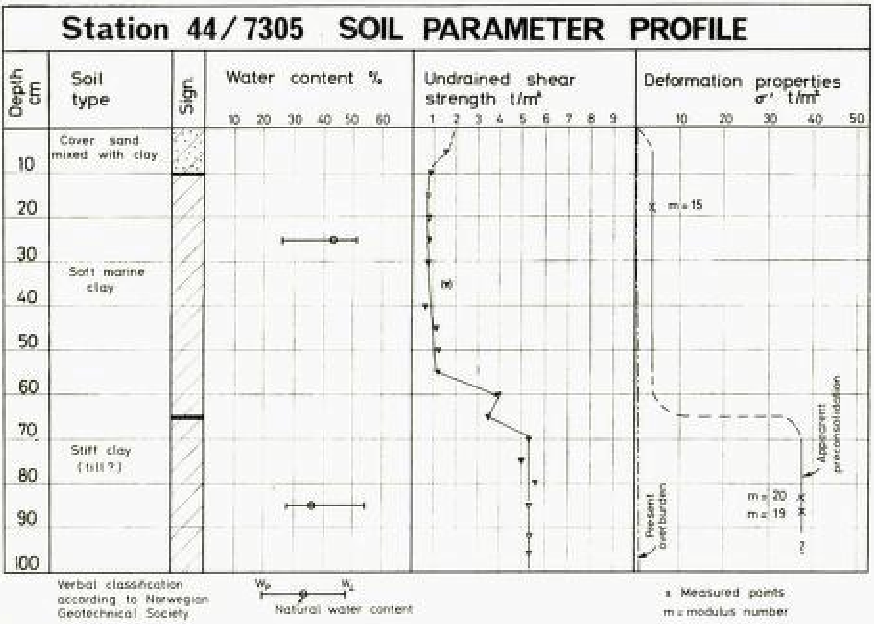


Fig. 8. Station 44, soil parameter profile.

also suggests this. The parallel greenish and brownish layers could indicate a migration and a segregation of ferrous ions, but the chemical analyses do not provide any evidence for this.

The preferred optical orientation of the quartz grains seen in the thin-section indicates diagenetic precipitation of silica. The high content of montmorillonite could indicate that the claystone pebbles originally came from a bentonite bed. During the alteration of volcanic ash to montmorillonite, silica is liberated and may precipitate within the sediment and zeolite minerals may be formed (Grim 1968).

Geotechnical investigations

The main geotechnical properties are given in the soil parameter profile (Fig. 8).

The undrained shear strength along the core was measured by the falling cone method. The profile shows several different layers. In the top layer, 0–10 cm, the higher shear strength is probably due to sand grains. From 10 to 55 cm, the shear strength, s_u, about 1 t/m², seems to increase slightly downwards. Between 55 and 70 cm, s_u is 3.5–4 t/m². From 70 to 95 cm, s_u is fairly constant between 5 and 6 t/m². In the lowermost part of the core, below about 95 cm, the water content may have increased after sampling, so that the natural shear strength here may have been higher than the one measured.

Two representative samples, 17–30 cm and 80–90 cm, were chosen for a more extensive geotechnical investigation of layers 10–55 cm and 70–95 cm, respectively. Natural water content and index properties, liquid limit, W_L , and plastic limit, W_P , were measured. Prior to measuring the index properties, the sample was remoulded and clay fragments crushed smaller than 0.5 mm. Compared with Quaternary clays from Norway, the liquid limit, W_L , and the plasticity index, $I_P = W_L - W_P$, are high. Both values are slightly higher in the 80–90 cm sample than in the 17–30 cm sample. A comparison of natural water content with index properties (Fig. 8) reveals that the 80–90 cm sample is more highly consolidated.

The deformation properties of three samples (17–19 cm, 82–84 cm and 85–87 cm) were measured by an oedometer and the results were interpreted by means of the resistance concepts of Janbu (1967). From the load-deformation curve one can find the tangent modulus $M = \frac{d\sigma'}{d\varepsilon}$ were σ' = effective stress (load) and ε = relative deformation.

The load-modulus curves show that the modulus is nearly proportional to the pressure $M = m \cdot \sigma'$ for a stress higher than the pre-consolidation pressure, where m = the modulus number. The pre-consolidation pressure represents a break in the load-modulus curve, showing where the behaviour of the sample changes from elastic to plastic (Janbu 1967). The modulus values and pre-consolidation pressures are shown in the soil parameter profile. The difference in modulus numbers shows that sample 17–19 cm is more easily compressed than the lower one. The pre-consolidation pressure, 35–40 t/m^2 for the samples between 80 and 90 cm, is considerably higher than the present overburden.

DISCUSSION

A pre-consolidation effect may have many explanations, such as overlying ice or soil, chemical weathering, cementation between grains, effect from the sampling, etc. The apparent pre-consolidation pressure of 3–4 t/m^2 from 10–65 cm may, with effective unit weight 0.5 t/m^2, have been caused by 6–8 m of overburden which was later removed. The load-deformation curve of sample 17–19 cm, however, indicates the breakdown of a potentially unstable structure. Cementation between the grains may therefore be a preferable explanation.

The layer below 65 cm has a pre-consolidation pressure of 35–40 t/m^2, which could have been caused by ice load.

Palynology

Ten channel samples representing the total length of the core were examined. On the basis of the palynology, the core can be divided into two parts: 0–98 cm and 98–105 cm.

UPPER PART, 0–98 CM

A pure sample of the clay could not be prepared, as all the claystone pebbles could not be separated out. The claystone pebbles contain a well-preserved and diversified assemblage of spores, pollen, dinoflagellates, cuticles and woody matter. The clay contains a similar, but much reduced assemblage to that in the claystone.

LOWER PART, 98–105 CM

This part of the core contains a rich assemblage of palynomorphs similar to that present in the upper section of the core. The taxa identified include:

Spores and pollen:

Lygodium, *Extratriporopollenites*, *Triatriopollenites*, *Taxadiaceaepollenites*, *Sequoiapollenites*, *Tricolpites*, *Caryapollenites*, *Alnus*, *Picea* and *Pinus*.

Dinoflagellates:

Deflandrea, *Hystricosphosphaera*, *Baltisphaeridium* and *Micrystridium*.

DISCUSSION

The assemblage is assigned an Upper Cretaceous/Lower Tertiary age. It has undergone little thermal alteration. Such an assemblage is characteristic of a near-shore marine environment. The pebbles in the upper part of the section appear to have been reworked from deposits of the same age as the lower part. The palynomorphs in the clay in the upper part of the core may have been derived from the breakdown of claystone pebbles. However, the reduced assemblage in the clay suggests that the sediment was formed by the mixing together of clay lacking palynomorphs with the described claystone pebbles.

Foraminifera

Three different foraminiferal faunal assemblages can be recognized (Fig. 9). The upper part of the core (0–10 cm) is very rich in foraminifera (Fig. 10), and contains 20–25% of planktonic specimens most of which are of Pre-Quaternary age. The benthonic fauna is characterized by the boreal–lusitanian deep-water species *Uvigerina peregrina* Cushman, *Trifarina angulosa* (Williamson), with the cosmopolitan species *Cibicides lobatulus* (Walker & Jacob), *Cassidulina laevigata* d'Orbigny, and *Nonion barleeanum* (Williamson) as common forms. Other common boreal–lusitanian species are *Hyalinea baltica* (Scroeter) and *Bulimina marginata* d'Orbigny. The majority of the species mentioned above are common in the deeper parts of the North Sea area today (Høglund 1947; Lange 1956; Jarke 1961; Kihle 1971; Murray 1971). In addition, these samples also have a strong arctic or arctic-boreal element (*Elphidium clavatum* Cushman, *Cassidulina crassa* d'Orbigny). Among the rarer arctic-boreal species are *Islandiella teretis* (Tappan), *Stainforthia loeblichi* (Feyling-Hanssen) and *Nonion labradoricum* (Dawson). The benthonic calcareous fauna indicates a Holocene age for the sediment in the upper portion of the core. A few arenaceous Pre-Quaternary specimens were also found, of much the same composition as the fauna in the core section between 98 and 105 cm.

The middle part of the core (10–98 cm) contains quite poor populations of foraminifera, and is characterized by the arctic or arctic-boreal species

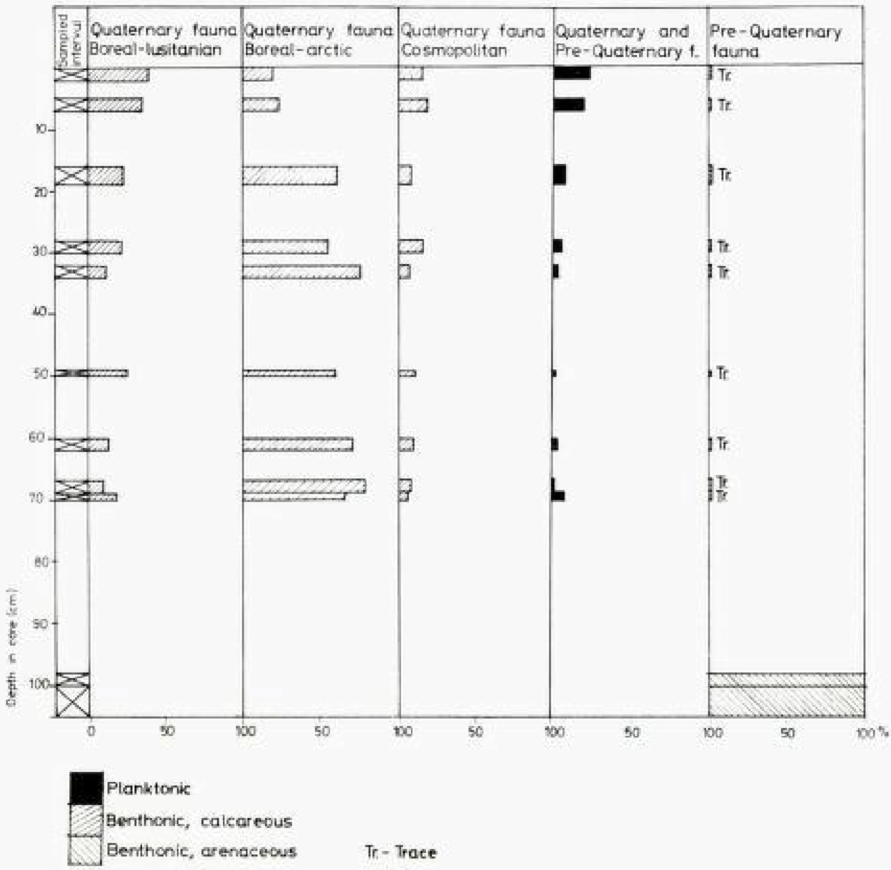


Fig. 9. Distribution of foraminiferal groups in samples selected from the core.

Elphidium clavatum Cushman and *Cassidulina crassa* d'Orbigny. In addition, a relatively strong boreal–lusitanian element occurs (*Uvigerina peregrina* Cushman, *Bulimina marginata* d'Orbigny, *Trifarina angulosa* (Williamson)). Among the rarer species are the cold-water forms *Trifarina fluens* (Todd), *Nonion labradoricum* (Dawson), *Bolivina pseudopunctata* (Høglund), *Elphidium groenlandicum* Cushman, *Quinqueloculina stalkeri* Loeblich and Tappan, *Stainforthia loeblichi* (Feyling-Hanssen) and *Islandiella teretis* (Tappan). This foraminiferal fauna shows affinities to the Late Glacial faunas from the Oslofjord area, but has a much stronger boreal–lusitanian element (Feyling-Hanssen 1964). These foraminifera indicate a Pleistocene age for the sediments in the middle part of the core. A few Pre-Quaternary arenaceous forms were found. The samples in this part of the core also contain 5–10% of Pre-Quaternary planktonic foraminifera.

The lowest portion of the core (98–105 cm) contains quite rich populations of foraminifera (Fig. 11) and is characterized by the arenaceous genera *Ammodiscus*, *Ammolagena*, *Bathysiphon*, *Cyclammina*, *Dorothia*, *Glomospira*, *Haplophragmium*, *Haplophragmoides*, *Psammosphaera*, *Saccammina*, *Spiro-*

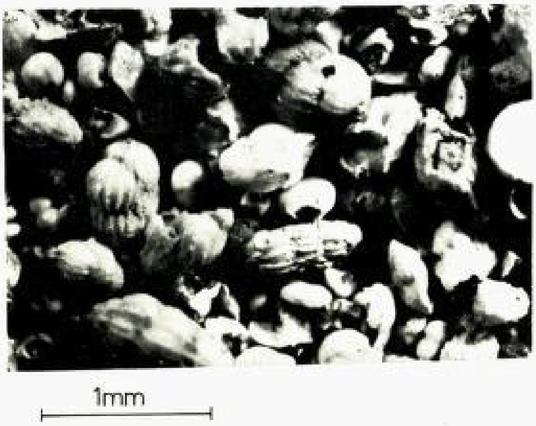


Fig. 10. Foraminifera from the interval 0–2 cm.

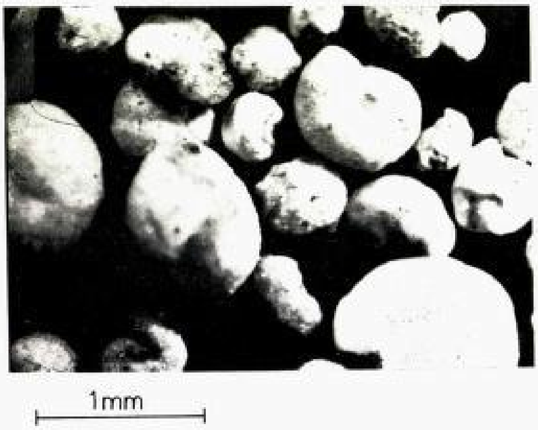


Fig. 11. Foraminifera from the interval 100–105 cm.

plectammina, *Textularia*, *Thalmannammina* and *Trochammina*. Some of the foraminifera found here are worn specimens and cannot be determined. The most common species are:

Ammodiscus cf. *incertus* (d'Orbigny)
Ammolagena clavata (Jones & Parker)
Cyclammina challinori Haynes
Cyclammina incisa (Stache)
Dorothia eocenica Cushman
Glomospira charoides (Jones & Parker)
Haplophragmium bulloides Beissel
Haplophragmoides eggeri Cushman
Haplophragmoides kirki Wickenden
Spiroplectammina spectabilis (Grzybowski)
Textularia smithvillensis Cushman & Ellisor
Thalmannammina recurvoidiformis Neagu & Tocorjescu

This fauna has forms in common with Upper Cretaceous and Lower Tertiary arenaceous faunas from various parts of Europe (Grzybowski 1901; Haynes 1958; Neagu 1970; Săndulescu 1972). The fauna also shows affinities to upper Cretaceous and Paleocene arenaceous faunas from America (Cushman & Waters 1927; Cushman & Ellisor 1933; Mallory 1959). The majority of the represented specimens are of Lower Tertiary age. No calcareous specimens were found in the lowest portion of the core.

DISCUSSION

The boreal–lusitanian element in the upper 10 cm of the core constitute 35–40% of the total foraminiferal fauna, whereas the arctic–boreal species make up 20–25% of the total fauna. The first group of foraminifera is more frequent than the second group; one should therefore expect a Holocene age for the sediments in this portion of the core. The relatively strong arctic–boreal element may be reworked from Pleistocene deposits, whereas the Pre Pleistocene element of the planktonic foraminifera and the few arenaceous specimens are probably reworked from Upper Cretaceous or Lower Tertiary deposits in the area.

The middle part of the core contains 55–80% of arctic or arctic-boreal foraminifera and the sediments are thought to be of Pleistocene age. The content of boreal–lusitanian species is relatively high (10–25%), and these sediments may represent a deposit from a warmer part of the Pleistocene, perhaps from an interglacial or interstadial period or perhaps from the very end of Late Glacial time. The sediments also may represent glacial deposits with reworked material from an interglacial period. The Pre-Pleistocene specimens are probably reworked from Upper Cretaceous or Lower Tertiary deposits in the area.

The lower part of the core contains only arenaceous foraminifera solely of Upper Cretaceous and Lower Tertiary age. The age of this part of the core is discussed below.

Discussion and conclusions

Evidence from the foraminiferal assemblages suggests that the core can be divided into three parts: 0–10 cm, 10–98 cm, and 98–105 cm. The geotechnical properties show that a subdivision of the middle part can be drawn at 65 cm.

The loose, sandy sediment in the upper part of the core (0–10 cm) is similar to the top portion of other cores collected from this area. This cover sand is believed to be the result of winnowing and retransportation on the shallower bank areas during the Weichselian/Late Weichselian low sea-level period. Coarse material was probably contributed by ice rafting. The sand is thus a relict sediment. The bioturbation, shown on bottom photographs from this area, appears to extend down into the underlying sediments. The nature of the cover sand therefore reflects late glacial sedimentation as well as Holocene bioturbation.

The poor sorting and mixed faunas of different ages suggest that the middle part of the core (10–98 cm) is a till. The pre-consolidation stress as well as the banded structure suggest a subglacial origin for the interval 65–98 cm, while a glacial-marine origin is likely for the interval 10–65 cm. The mixed foraminiferal assemblages indicate a Quaternary age.

In contrast to the overlying portion of the core, the palynological and foraminiferal assemblages of the lower part of the core (98–105 cm) together indicate one age, Upper Cretaceous/Lower Tertiary, which may suggest that this part of the core represents in-situ bedrock. This is supported by the very thin nature of the Quaternary sediments as seen on the sparker profile. However, the presence of claystone pebbles and banding may indicate that the sediment is a very locally derived basal till, representative of the immediately underlying bedrock. In this case, the sediment would be of Quaternary age, but nevertheless representative of the age and composition of the local bedrock.

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The Mesozoic Rocks of Andøy, Northern Norway

A. DALLAND

Dalland, A. 1975: The Mesozoic rocks of Andøy, northern Norway. *Norges geol. Unders. 316*, 271–287.

The Jurassic–Lower Cretaceous sediments of Andøy are found in a small, down-faulted area on the north-east coast of the island. The Mesozoic sequence is more than 650 m thick, and rests non-conformably on weathered basement rocks. The lower part of the sequence (about 85 m thick) consists of coarse-grained sandstones interbedded with some siltstones, bituminous shales and a few coal layers. The age of this part is Middle Jurassic or earliest Upper Jurassic. On top of this are micaceous sandstones and siltstones of Upper Jurassic age, with a total thickness of nearly 300 m. The next formation consists of calcareous sandstones and siltstones (about 80 m), in part Valanginian in age, and this is followed by a formation consisting of fine-grained siltstones and shales, more than 200 m thick and probably of Lower Cretaceous age.

The Mesozoic sequence is largely marine, except for a few layers of freshwater and brackish-water deposits in the lowest part. The sediments are part of an onlap sequence which rapidly thins towards the NW. The main basin of deposition probably lay somewhere to the east, the sediments coming in mainly from the north-west. Sedimentation started as a consequence of faulting in Middle Jurassic time. Another period of faulting occurred in early Cretaceous time.

The down-faulted structure which we see to-day, with its tight pattern of NNE–SSW-trending vertical faults, probably owes its origin to the Tertiary fault movements which occurred along much of the western part of Norway as a result of continental margin adjustments to sea-floor spreading in the Norwegian Sea.

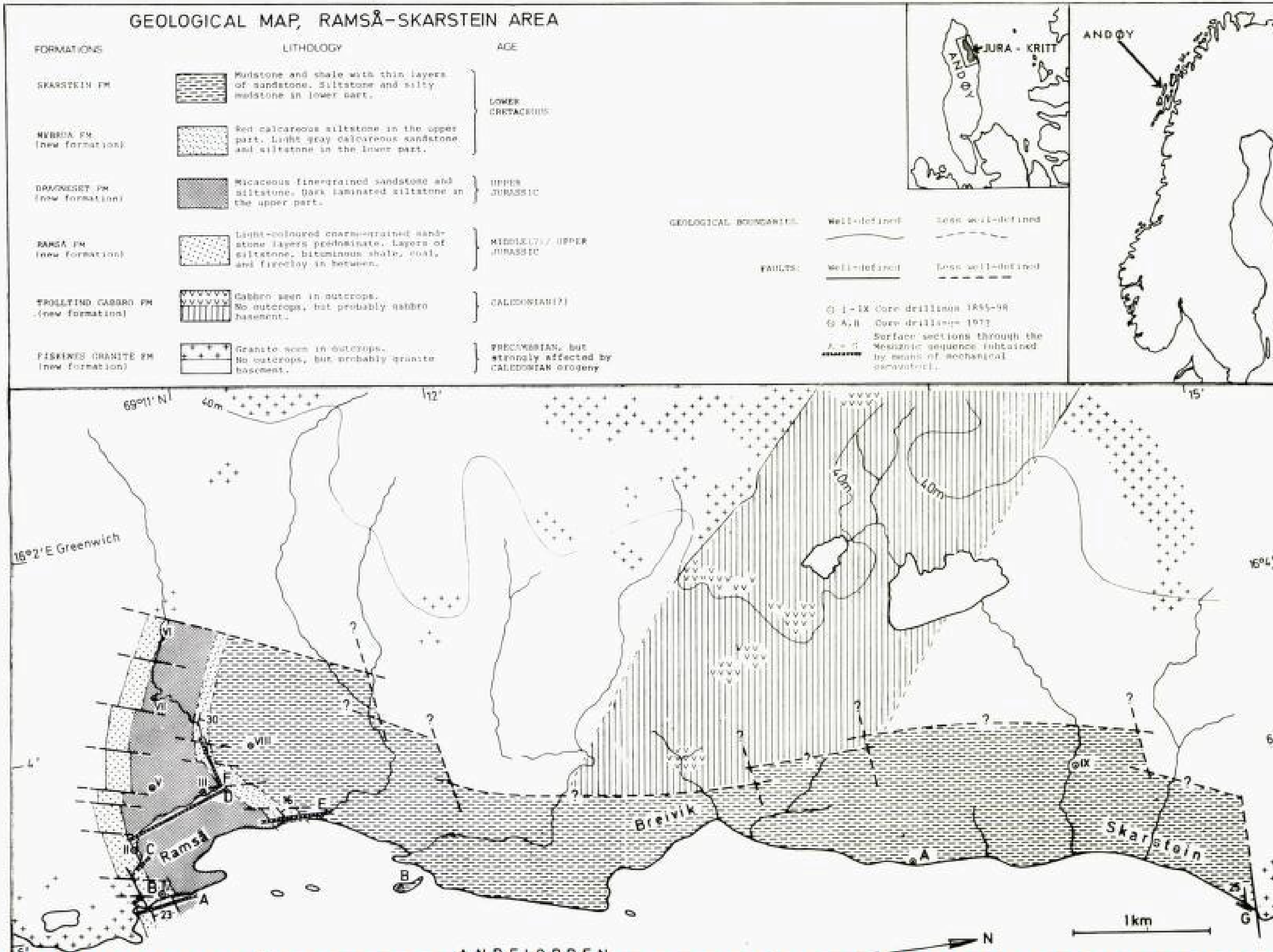
The rocks and fossils from Andøy show close affinities to the Jurassic–Lower Cretaceous deposits of East Greenland.

A. Dalland, Geologisk Institutt, Avd. A, Universitetet i Bergen, N–5014 Bergen, Bergen

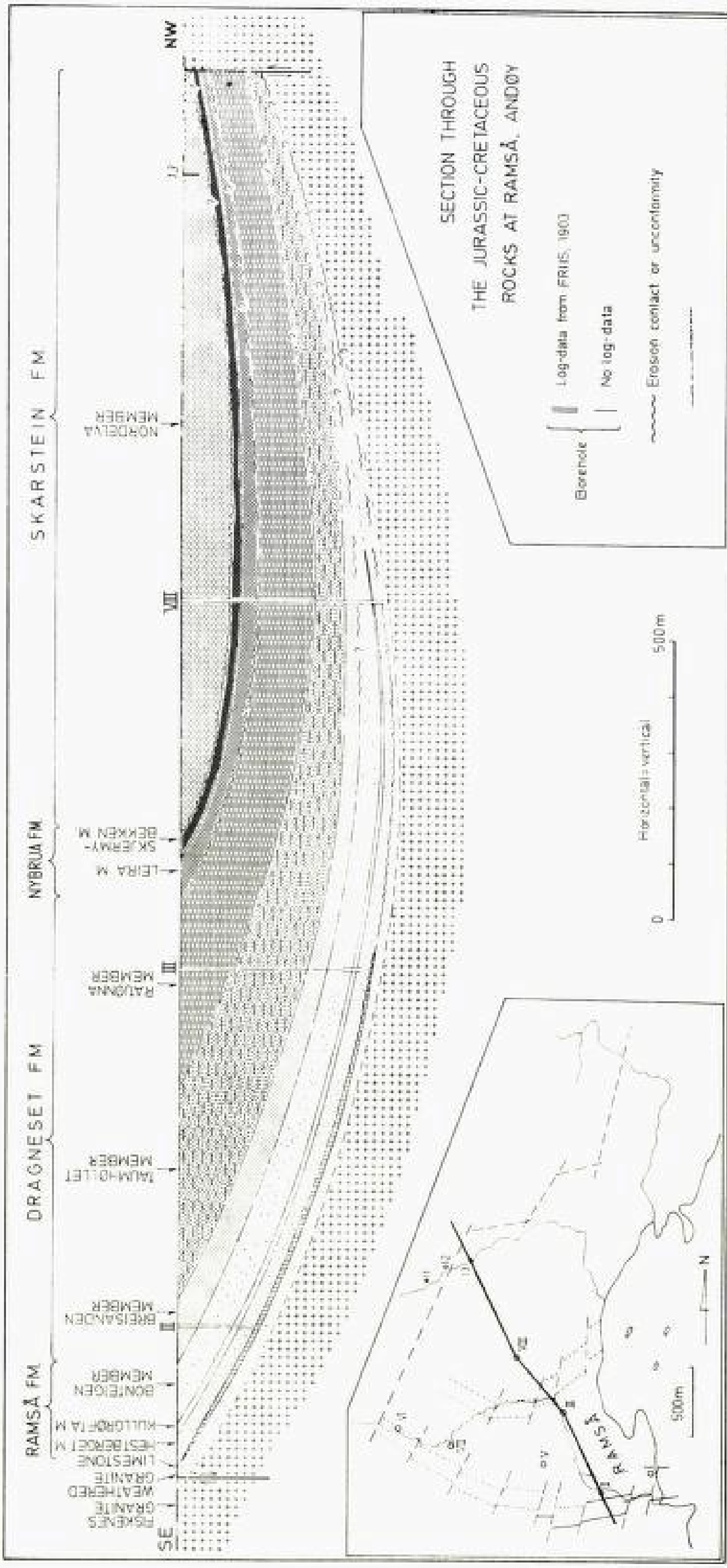
Introduction

Sediments of Jurassic–Lower Cretaceous age occupy an area of about 8 km^2 along the north-east coast of Andøy, an island situated just to the north of the Lofoten group of islands (Fig. 1). Total thickness of the sequence is more than 650 metres. The Mesozoic sediments rest unconformably on weathered basement rocks. The outcrop area is bounded by faults except to the south. To the east the sediments extends under the Andfjord and are probably continuous with the thick pile of sediments (about 5 km) proven by seismic studies in the outer part of the fjord (Sundvor & Sellevoll 1969).

The sediments have escaped erosion because of down-faulting and a tight pattern of post-Lower Cretaceous faults has created two small troughs — one in the southern and the other in the northern part of the outcrop area. The depth to basement in the middle of each of the troughs is at least 450 metres. The southern trough contains mainly rocks of Upper Jurassic age (Fig. 2), whereas rocks of Lower Cretaceous age make up the bulk of the sediments


GEOLOGICAL MAP, RAMSÅ-SKARSTEIN AREA
FORMATIONS
LITHOLOGY
AGE
SKARSTEIN FM
Mudstone and shale with thin layers of sandstone. Siltstone and silty mudstone in lower part.
NYBRUA FM (new formation)
Red calcareous siltstone in the upper part. Light gray calcareous sandstone and siltstone in the lower part.
LOWER CRETACEOUS
DRAGNESET FM (new formation)
Micaceous finegrained sandstone and siltstone. Dark laminated siltstone in the upper part.
UPPER JURASSIC
RAMSÅ FM (new formation)
Light-coloured coarse-grained sandstone layers predominate. Layers of siltstone, bituminous shale, coal, and fireclay in between.
MIDDLE(?)/ UPPER JURASSIC
TROLLTIND GABBRO FM (new formation)
Gabbro seen in outcrops.
No outcrops, but probably gabbro basement.
CALEDONIAN(?)
FISKENES GRANITE FM (new formation)
Granite seen in outcrops.
No outcrops, but probably granite basement.
PRECAMBRIAN, but strongly affected by CALEDONIAN orogeny
GEOLOGICAL BOUNDARIES:
Well-defined
Less well-defined
FAULTS:
Well-defined
Less well-defined
I - IX Core drillings
Surface sections through the Mesozoic sequence (obtained by means of mechanical
JURA - KRITT
ANDØY
ANDØY
69°11' N
40m
12'
15'
16°2' E Greenwich
16°4'
6'
4'
VI
VII
30
VIII
V
III
F
D
E
II
C
Ramså
B
A
23
B
Breivik
IX
A
Skarstein
G
1km
N
ANDFJORDEN

. 2. Section through the Jurassic–
wer Cretaceous rocks at Ramså,
døy.



. 1. Geological map of the Mesozoic
diments on Andøy, northern Norway.

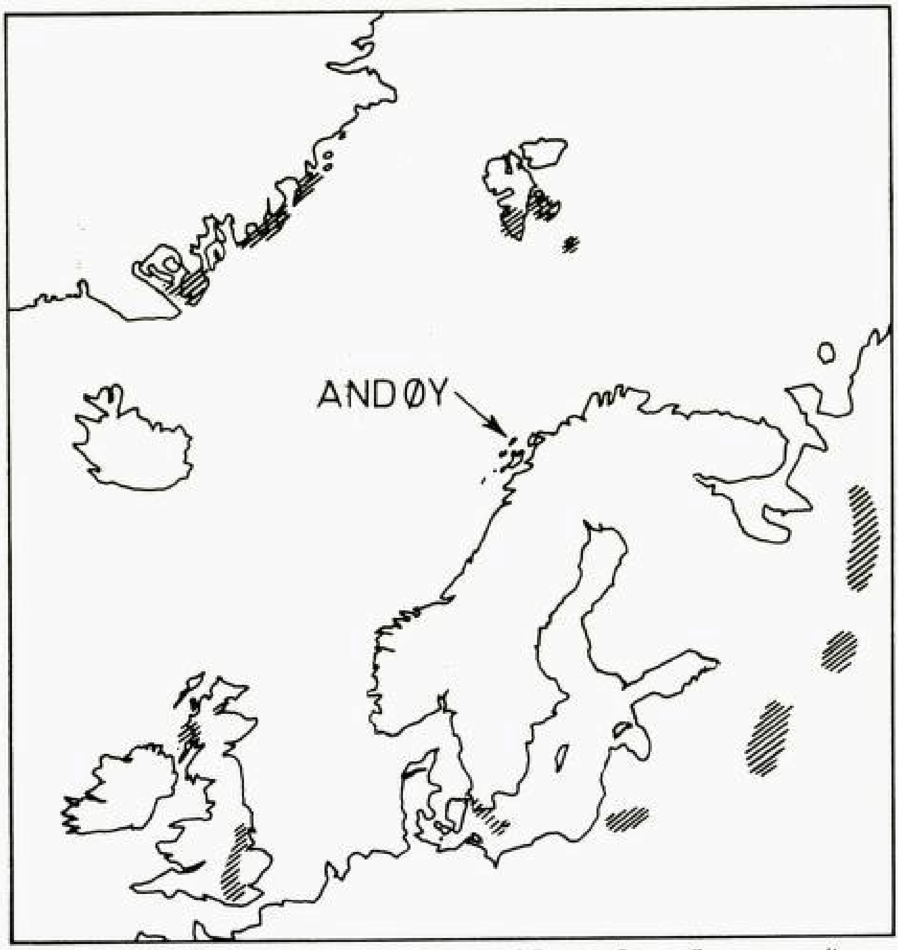


Fig. 3. Andøy in relation to the nearest outcrops of Jurassic–Lower Cretaceous sediments.

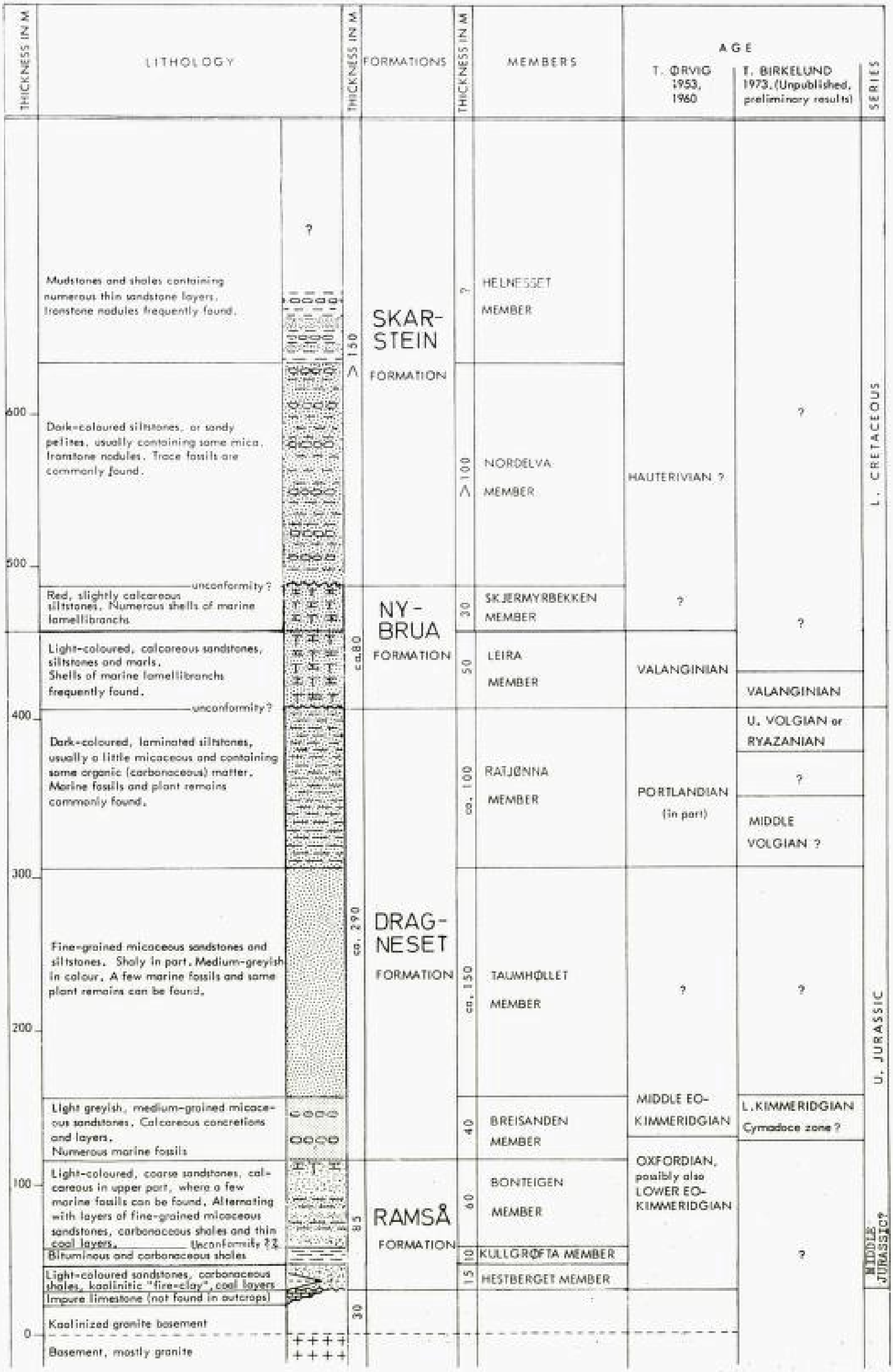
in the northern one. The two troughs are connected by an area with a thinner sedimentary cover extending across a basement high at Breivik (Fig. 1).

Andøy is the only place in Norway where outcrops of Mesozoic rocks are found. The distance to the nearest area of outcrop of sediments of the same age (Fig. 3) makes Andøy a key area for paleogeographic reconstructions, and in recent years the area has developed additional interest because of the possibility of petroleum occurring on the nearby continental shelf.

Exposures are few, and most of them are very small. Because of this, the geological picture of the area has been based mainly on the core drillings made during the late 1890's (Friis 1903; Vogt 1905). The bore-holes were sunk tc investigate the occurrence of coal in the lowest part of the Mesozoic sequence.

Some palaeontological work has been done, but only on samples from restricted parts of the sequence (Heer 1877; Mayer 1877; Lundgren 1894;

Fig. 4. Lithostratigraphy of the Mesozoic rocks of Andøy.


THICKNESS IN M
LITHOLOGY
THICKNESS IN M
FORMATIONS
THICKNESS IN M
MEMBERS
AGE
T. ØRVIG 1953, 1960
T. BIRKELUND 1973, (Unpublished, preliminary results)
SERIES
?
Mudstones and shales containing numerous thin sandstone layers. Ironstone nodules frequently found.
SKAR-STEIN FORMATION
> 150
?
HELNESSET MEMBER
600
Dark-coloured siltstones, or sandy pelites, usually containing some mica. Ironstone nodules. Trace fossils are commonly found.
> 100
NORDELVA MEMBER
HAUTERIVIAN ?
?
L. CRETACEOUS
500
unconformity ?
Red, slightly calcareous siltstones. Numerous shells of marine lamellibranchs
NY-BRUA FORMATION
ca.80
30
SKJERMYRBEKKEN MEMBER
?
?
Light-coloured, calcareous sandstones, siltstones and marls. Shells of marine lamellibranchs frequently found.
50
LEIRA MEMBER
VALANGINIAN
VALANGINIAN
unconformity ?
400
Dark-coloured, laminated siltstones, usually a little micaceous and containing some organic (carbonaceous) matter. Marine fossils and plant remains commonly found.
ca. 100
RATJØNNA MEMBER
PORTLANDIAN (in part)
U. VOLGIAN or RYAZANIAN
?
MIDDLE VOLGIAN ?
300
Fine-grained micaceous sandstones and siltstones. Shaly in part. Medium-greyish in colour. A few marine fossils and some plant remains can be found.
DRAG-NESET FORMATION
ca. 290
ca. 150
TAUMHØLLET MEMBER
?
?
U. JURASSIC
200
Light greyish, medium-grained micaceous sandstones. Calcareous concretions and layers. Numerous marine fossils
40
BREISANDEN MEMBER
MIDDLE EO-KIMMERIDGIAN
L. KIMMERIDGIAN Cymodoce zone ?
100
Light-coloured, coarse sandstones, calcareous in upper part, where a few marine fossils can be found. Alternating with layers of fine-grained micaceous sandstones, carbonaceous shales and thin coal layers.
Unconformity ??
RAMSÅ FORMATION
85
60
BONTEIGEN MEMBER
OXFORDIAN, possibly also LOWER EO-KIMMERIDGIAN
?
MIDDLE JURASSIC?
Bituminous and carbonaceous shales
10
KULLGRØFTA MEMBER
Light-coloured sandstones, carbonaceous shales, kaolinitic "fire-clay", coal layers
15
HESTBERGET MEMBER
Impure limestone (not found in outcrops)
30
Kaolinized granite basement
0
Basement, mostly granite

Sokolov 1912; Johansson 1920). Ørvig (1953, 1960) gave a summary of the earlier work in the area.

During the present investigation a mechanical excavator was used in the field work. This enabled us to obtain a nearly continuous surface through the greater part of the sequence, and it also made it easier to correlate the old bore-logs and erect formal lithostratigraphic units (Dalland 1974). The stratigraphic units are shown in Fig. 4, and a short description of each unit is given below.

Basement complex

As Fig. 1 shows, basement in the area consists largely of granite with gabbro cutting through it. The granite is probably of Precambrian age, and has been severely affected by the Caledonian deformation. The gabbro could be of Caledonian age.

Mesozoic system

WEATHERED BASEMENT COMPLEX

The basement was deeply weathered before the sedimentation started, but the weathered granite is preserved only in the southern part of the area. Here it makes up an *in situ* layer, up to 30 metres thick, consisting almost exclusively of kaolin (dickite) and quartz. The weathering profile must have taken a considerable time to form, and stable tectonic conditions and a humid climate must have prevailed. The weathering process could well have started in Early Jurassic time, or even earlier if the tectonic and climatic conditions were suitable, but probably the climate in Triassic time in the area was too dry for weathering of this kind to develop.

BASAL LIMESTONE

This limestone, up to 6 metres thick, rests on the weathered basement in a small part of the area. It is reported only from a few of the old bore-holes, and no samples exist. The age cannot be younger than Middle Jurassic. The limestone unit is poorly described in the old bore-logs. Because of its quite different lithology it is not a natural part of the overlying formation, and it ought to be better known before a formal stratigraphic designation can be given to it.

Ramså Formation (a new formation name; see Dalland 1974)

The deposition of the Ramså Formation was initiated by faulting in the area. A small, down-faulted block served as a local depocentre, and it is only close to this centre that the underlying limestone and weathered basement have escaped erosion. The first sediments to accumulate came from the nearby weathered basement, mainly from the north-west.

Some details of the various members of the formation, which were introduced and described fully in the author's thesis (Dalland 1974), are presented below.

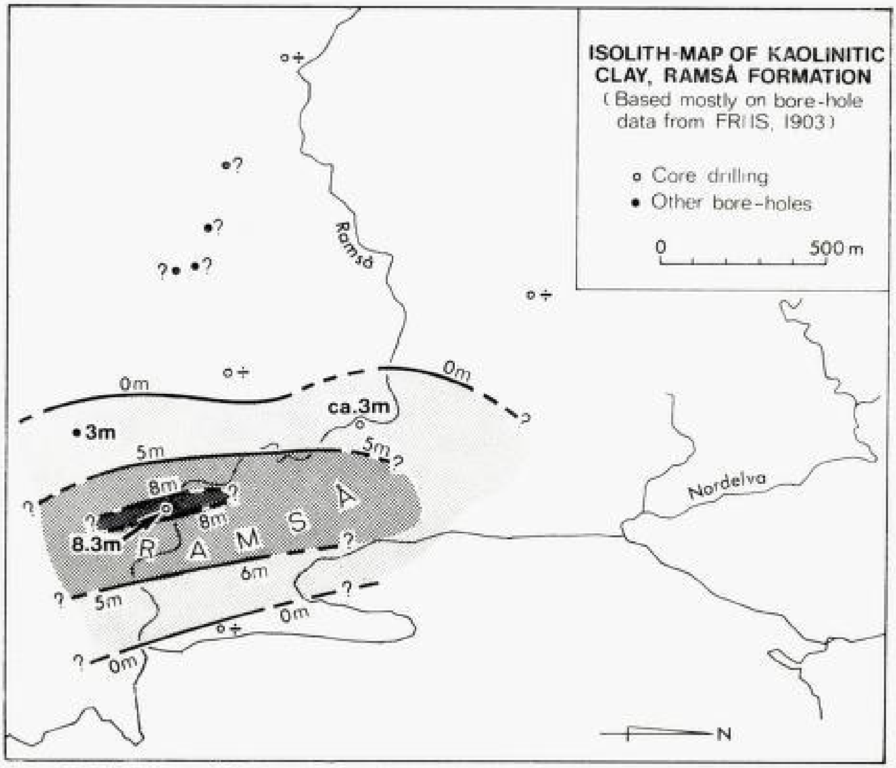


Fig. 5. Isolith map of kaolinitic clay, Ramså Formation.

Hestberget Member

The Hestberget Member consists of coarse-grained sandstone interbedded with layers of kaolin-rich shale ('fireclay'), bituminous shale, micaceous siltstone and a few layers of cannel-coal. The kaolin shales are concentrated in a very small area in the middle of the basin, and must represent outwash from the kaolin-rich weathered basement (Fig. 5). The lowest sandstone layers consist of irregular quartz grains; they contain surprisingly little feldspar, and must also be derived from the nearly feldspar-free weathered layer. There is a tendency for roundness to increase upwards throughout the member. Feldspar and garnet content also increases upwards. Garnets are common in the basement, but have not escaped the intense weathering. There is also a marked upward decrease in fireclay. This points to a gradual erosion through the weathered crust in the source area, and suggests that the source area itself moved a little further away from the depocentre as the basin became filled up. Deposition of this member could have occurred over a relatively short period of time, and there might also have been some syn-sedimentary faulting activity.

The thickness of the Hestberget Member is about 30 m in the centre of the basin, but this decreases rapidly out in all directions, especially towards the north-west. Deposition probably took place in coastal swamps with the down-faulted area acting as a lagoon for part of the time. Palynological investigations indicate a Middle Jurassic age (J. Os Vigran, pers. comm. 1973).

Kullgrøfta Member

This is a sequence consisting mainly of brownish-black bituminous shales. It also contains a few layers of micaceous siltstones and fine-grained sandstones. Except for the higher clay content, the shale layers have about the same composition as the layers of cannel-coal. The maximum thickness of the Kullgrøfta Member is about 12 m, and it thins out away from the depocentre in the same way as the underlying Hestberget Member. During the time of deposition the area was probably a shallow fresh- or brackish-water lagoon with a high accumulation of sapropel-rich mud. The age of the member is probably Middle Jurassic.

Bonteigen Member

The Bonteigen Member consists of medium- to coarse-grained sandstones, interbedded with shaly micaceous siltstones. A few layers of bituminous shale and thin coal layers are found in the middle and lower part. The sandstones in the upper part are calcareous and contain marine fossils (bivalves and belemnites). Near the top of the member a few glauconite-rich layers are present. The local depocentre can still be recognized, but the reduction in thickness outwards from this centre is not as marked as in the lower parts of the formation. Even so, the Bonteigen Member seems to thin out in all directions, and as with the lower members the thinning is especially rapid towards the north-west. The member is about 55 m thick at the depocentre, while the thickness 1 km to the north-west is only 25 m. According to Ørvig (1953, 1960) the Bonteigen Member is probably of Oxfordian age.

Conditions of deposition for the lower part of the Bonteigen Member must have been fairly similar to those suggested for the Hestberget Member. The upper part seems to represent more open marine conditions — this was probably in part high-energy shore-line sedimentation, and in part sedimentation in somewhat deeper sea-water.

An isopach map of the Ramså Formation is shown in Fig. 6, together with other distribution maps of the same sequence. Fig. 7 presents an interpretation of the local paleogeography during the deposition of the formation. The main basin probably lay to the east, in the area to-day occupied by the Andfjord. The small, down-faulted block which became the local depocentre must have been subsiding continually during deposition. The source area for much of the sediment was located at a short distance to the north-west of this small basin, and the entire outcrop area seems to be part of an onlap sequence marginal to the larger basin to the east.

Dragneset Formation

The lowest part of Dragneset Formation (Dalland 1974) consists of medium- to fine-grained sandstones. The middle part is a fairly monotonous sequence

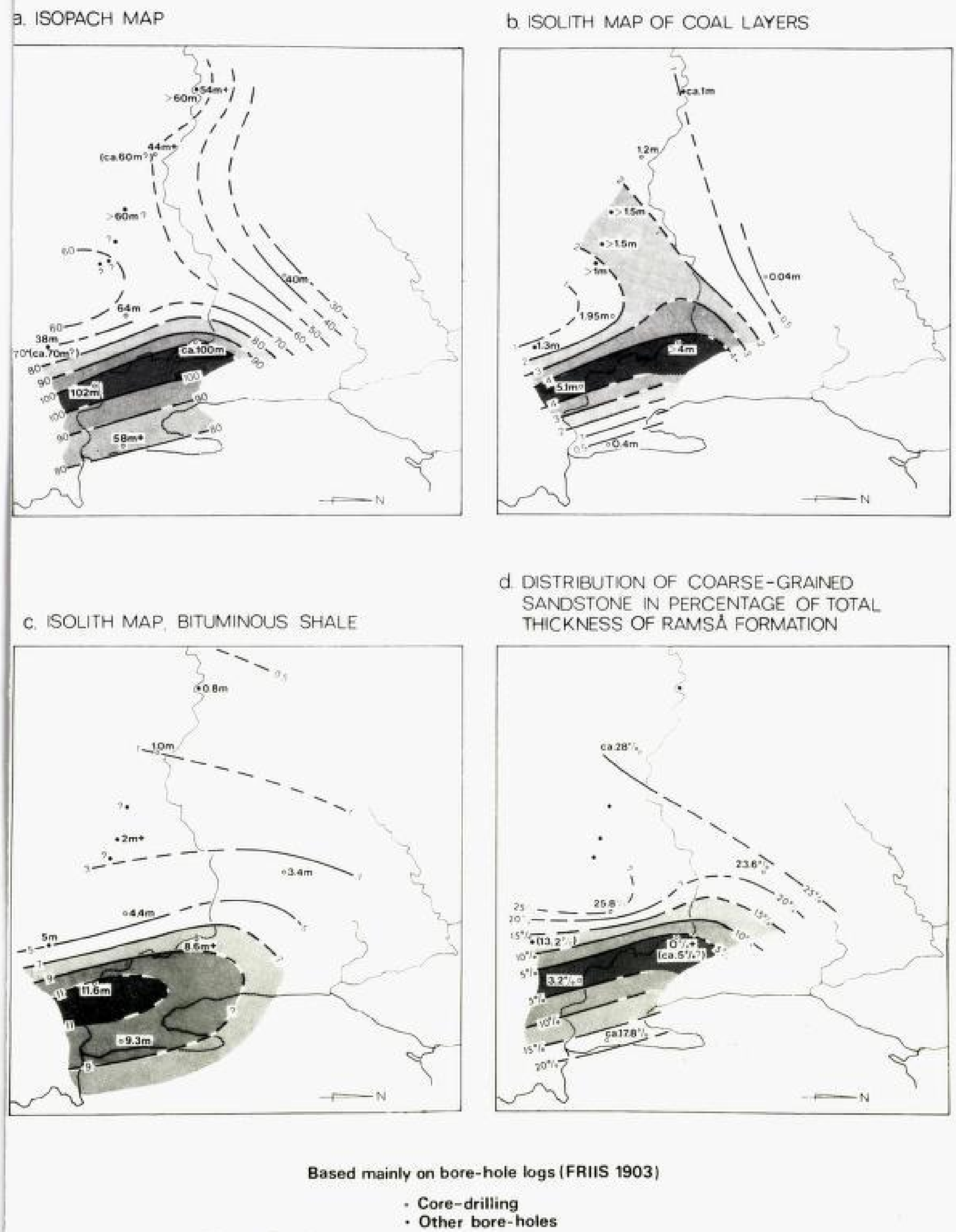


Fig. 6. Distribution maps, Ramså Formation.

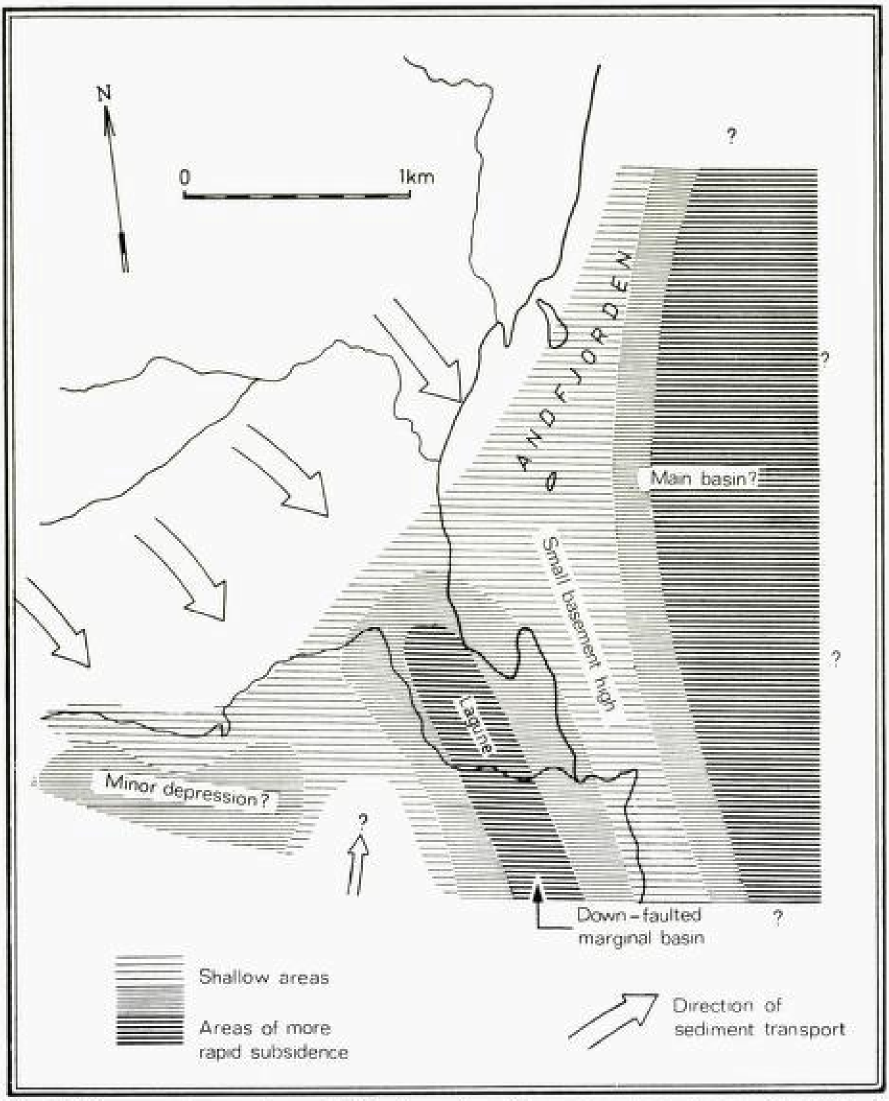


Fig. 7. Basement topography and directions of sediment transport during deposition of the Ramså Formation.

of siltstone and fine-grained sandstone layers, while the upper part is dominated by layers of shaly siltstones. A high content of muscovite is typical for the whole formation. No coarse-grained sandstones are found. The thickness of the formation is up to 290 m. The Dragneset Formation is divided into 3 members.

Breisanden Member

The Breisanden Member consists of layers of medium- to fine-grained sandstone. Some of the layers are a little calcareous and large concretions are common. Many beds contain abundant marine fossils — bivalves, ammonites and belemnites are the most common. A few vertebrate remains have been found. Glauconite-rich layers occur in the lower part. The thickness is up to 40 m, the member thinning out north-westwards as with the underlying members.

Deposition of the lowest part of the member probably took place in an open marine environment, a little below the limit of the littoral zone. The upper part seems to have been deposited in a bay or lagoon with a connection to the sea.

The Breisanden Member may be correlated with the *Rasenia cymodoze* zone of the Lower Kimmeridgian (T. Birkelund, pers. comm., 1973).

Taumhølet Member

The Taumhølet Member consists of micaceous siltstones and fine-grained sandstones. Some of the layers are shaly, but lamination is not common. The colour is dark grey, due to the presence of finely dispersed plant material. A single thin layer, which seems to contain volcanic material, is observed in the lowest part. Marine fossils are found throughout the member, but there is a very limited number of different species. The maximum thickness of the Taumhølet Member is about 150 m in the outcrop area.

Conditions of deposition are not quite certain. The fossil fauna suggests a somewhat restricted marine environment, possibly brackish water. It could perhaps be a mud-flat deposit, or alternatively the member may have been deposited in a shallow, brackish-water lagoon.

The age is also a little uncertain; the whole sequence could have been deposited within Kimmeridgian time, but it is not unlikely that at least the upper part is younger than that (Lower Tithonian or Lower Volgian). It has not been possible to detect any major sedimentological break or unconformity either within or at the bottom or top of the Taumhølet Member, but the very bad exposure in the area makes observation difficult.

Ratjønna Member

Dark, partly laminated, siltstone layers predominate in this member. The dark colour is mainly due to the presence of dispersed plant material. The Ratjønna Member contains abundant marine fossils. Ammonites, bivalves (especially *Aucella* forms) and belemnites are common. The thickness of the sequence in the Ramså area is about 100 metres. Layers of hard, light sandstone occur in the upper part. These layers are slightly calcareous. Small-scale cross-bedding and concentrations of transported shell material are common within the layers. This, together with the lense-shaped cross-section of most of the layers, suggests that these sandstones represent infill deposits of small tidal channels. The dark, shaly siltstone that characterizes most of the sequence was probably deposited in somewhat deeper water. The laminated parts could indicate a deficiency in oxygen at the bottom of the basin during deposition.

According to T. Birkelund (pers. comm. 1973) the age of the lower part of the Ratjønna Member could be Middle Volgian, while the upper part is of

Upper Volgian age; the uppermost layers of the member may also contain sediments of Berriasian age, but this is still uncertain.

The boundary between the Dragneset Formation and the overlying Nybrua Formation seems to represent an unconformity or at least a period of non-deposition.

Nybrua Formation

The Nybrua Formation (Dalland 1974) consists of calcareous sandstones and siltstones. The thickness of the formation is a little less than 80 m; it is divided into two members.

Leira Member

Layers of hard, calcareous sandstone with softer intercalations of siltstone predominate in this member. The siltstone layers are also a little calcareous and commonly contain concretions. Fossils of marine pelecypods are common (*Aucella* types dominate). The shell material is often fragmentary and shows signs of transportation. Deep vertical burrows are common in many of the sandstone layers. Very little plant material has been found.

The age of the Leira Member is Valanginian. Many of the sandstone layers seem to have been deposited on a shore, exposed to the open sea, while most of the siltstone layers may have been deposited just below the littoral zone.

Skjermyrbekken Member

The Skjermyrbekken Member consists of brownish-red siltstone. In addition to the colour, slump structures are the most characteristic feature, and these are found throughout the sequence in the outcrop area. Because of the presence of these slump structures, original layering is difficult to find. A few, often fragmented belemnites and badly preserved lamellibranch shells are present. Plant material has not been observed. The thickness of the sequence is up to 30 metres.

The age of the Skjermyrbekken Member is uncertain, but as there seems to be a gradual downward transition at the lower boundary, it is probably fairly close to that of the Leira Member (?Valanginian–Hauterivian).

Deposition must have taken place in a marine environment, not very different from that in which the siltstones of Leira Member were deposited. Small relics of greenish-grey coloration can be found within the rock, and these parts very much resemble the siltstones in the underlying Leira Member. The red colour is almost certainly secondary; this could probably have originated during a very short period of uplift and weathering. Oxidizing conditions within the bottom sediment shortly after deposition could also have caused the change in colour (by oxidizing Fe^{++} minerals), and the process of slumping could perhaps have played a role here. The slump structures are indicative of faulting activity in the area shortly after deposition.

Skarstein Formation (Ørvig 1960)

The Skarstein Formation, which lies probably unconformably upon the Skjermyrbekken Member, consists of dark, fine-grained rocks — siltstones and silty shales in the lower part, mudstones, shales and thin beds of sandstones in the upper part. The total thickness is unknown, although it must exceed 200 m. The formation is divided into two members.

Nordelva Member

Dark siltstones and silty shales and mudstones predominate in the Nordelva Member (Dalland 1974). Most of the beds contain a few per cent of white mica (muscovite/sericite). Sideritic concretions, most of which display an internal system of carbonate-filled cracks, are common. Besides finely dispersed plant remains, a few imprints of small leaves have been observed. Trace fossils are common, and belemnites, ammonites and bivalve fossils have been found. The thickness of the Nordelva Member is at least 70 metres.

The sequence was deposited in a marine environment; perhaps in somewhat deeper water than for most of the underlying formations. The plant remains indicate that the distance to the nearest shore-line was short. The precise age of the Nordelva Member is not known (?Hauterivian–Aptian).

Helnesset Member

The Helnesset Member (Dalland 1974) consists of dark shale and mudstone with a few interlayers of sandstone. The sandstone layers show graded bedding and other structures commonly found in turbidity current deposits. Concretions of the same type as those found in the Nordelva Member are common. The fine-grained rocks contain a fairly large amount of dispersed organic material. A few marine fossils are found, and trace fossils are common. The thickness of the member is at least 125 metres.

Microfossils indicate an Aptian age, at least for the upper part of the member (Bergsaker 1973, information given at the Bergen Oil Conference). The sediments were probably deposited in deep water, in part by turbidity currents.

Structure

BASEMENT COMPLEX

Brecciated zones, most of them approximately vertical and trending N–S, have been observed in the basement close to the Mesozoic outcrops. The age of the deformation which produced these zones is unknown.

MESOZOIC SYSTEM

Sedimentation during this time was tectonically controlled. Faulting probably began in late (?) Middle Jurassic time, and lasted perhaps until early Late Jurassic. Another period of faulting occurred in the area in Early Cretaceous time (?Hauterivian–Aptian).

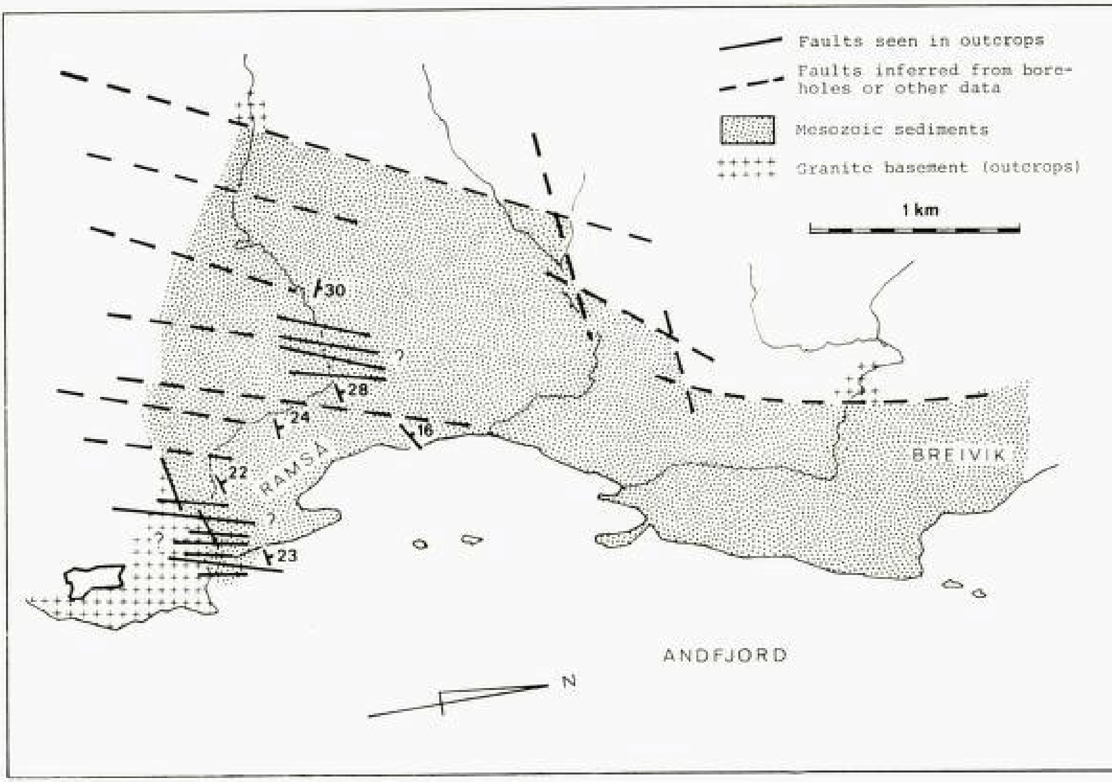


Fig. 8. Fault pattern, Ramså area, Andøy.

Most of the faults shown in Fig. 8, however, must be of post-Early Cretaceous age. These faults are probably part of the Tertiary fault system which occurs along much of the western part of Norway, and which is thought to have developed as a consequence of continental margin adjustments following sea floor spreading in the Norwegian Sea area. Nearly all the faults are of small displacement and are nearly vertical; their main direction is NNE–SSW, but E–W trending faults are also observed. Limited exposure makes it difficult to trace the faults, but judging from their concentration in the Ramså area (Fig. 8) where they are easiest to detect, the pattern must be extremely dense.

The faults, of course, are not restricted to the Mesozoic sediments, but they are much more difficult to detect in the basement rocks. The tight fault pattern suggests that the outcrop area at Andøy is situated within a NNE–SSW-trending zone of very intense Tertiary faulting.

The Tertiary faults to some extent are likely to follow earlier fault lines, as there are indications that the approximate N–S direction was also dominant for older fault systems.

As mentioned, the faulting has produced two small graben or trough structures in the outcrop area. The numerous fault blocks make the structure complicated, but in general the sediments seem to be dipping with angles of 15°–30° towards the centres of the troughs.

A section through the southern trough is shown in Fig. 3. This section also shows the rapid north-westward pinching out of the lower part of the sequence. Almost all the faults are omitted in this diagram.

Relationship to other areas

The Mesozoic rocks and fossils of Andøy show closer affinities to the Jurassic–Lower Cretaceous deposits and fossils of East Greenland (Aldinger 1935; Spath 1935, 1936; Donovan 1957; Surlyk et al. 1973) than to any other of the outcrop areas shown in Fig. 3, but as Norway and Greenland were juxtaposed in the pre-drift situation, this is really no surprise. Most of the outcrop areas of East Greenland were originally situated just a little to the south-west of Andøy. As on Andøy, most of the faults which affect the Mesozoic rocks of East Greenland trend in about N–S direction.

Fig. 9 shows the possible relationships of the outcrop area on Andøy to the nearby shelf sediments. Sediments are about 5 km thick in the outer part of Andfjord and probably continue into the fjord (Nysæther et al. 1969; Sundvor & Sellevoll 1969, 1971; Sellevoll 1972, Fig. 4). Little is known about the structures and sedimentary thicknesses along the submerged part of the section, and the structural interpretation presented in Fig. 9 must therefore be considered as tentative.

Petroleum prospects

Potential hydrocarbon source rocks on Andøy are the bituminous shales near the base of the sequence (Fig. 6c). The much thicker Lower Cretaceous shales and mudstones are also of interest as possible source rocks.

Porosity measurements of Jurassic sandstones give values between 12 and 30%, but the porosity of the samples has probably been affected by weathering.

The fine-grained Cretaceous rocks in the upper part of the sequence could act as sealing horizons.

Signs of gas have been reported from bore-holes (core-drillings, 1972–73, by the local company Norminal — at least one of the holes was drilled down to below 500 metres), but the fault pattern and the small size of the outcrop area do not favour the occurrence of hydrocarbons on land.

Nevertheless the area is important, as it provides some idea of the rock-types and structures that one can expect to find on the nearby shelf. Especially in Andfjord and in the area just to the north of the fjord, the depositional and structural history of Jurassic–Lower Cretaceous times could be very similar to that known from Andøy. Sedimentation was probably controlled by the same pattern of mainly N–S trending fault lines. The faults probably dissected the area into long, troughlike basins, with considerable

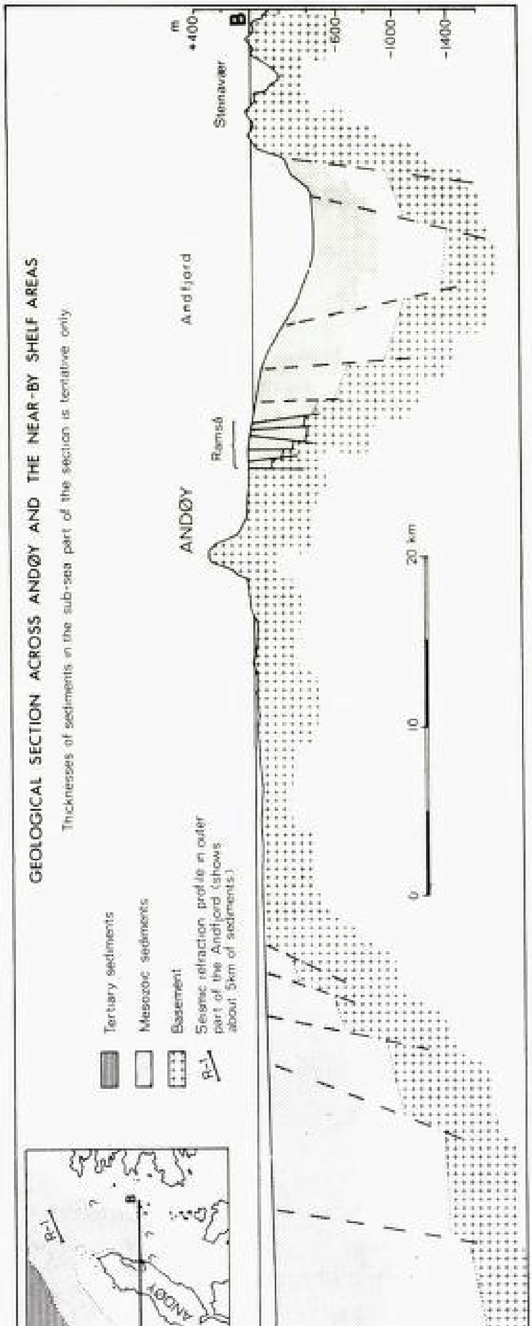


Fig. 9. Geological section across Andøy and the nearby shelf areas.

thicknesses of Mesozoic and perhaps also older sediments occurring just to the north of Andfjord.

If the Jurassic–Cretaceous sediments in this area of the shelf are of the same types as those occurring on Andøy, and if the structural conditions are favourable, than the area could be highly prospective.

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Palaeogeography and Facies Distribution in the Tertiary of Denmark and Surrounding Areas

NILS SPJELDNÆS

Spjeldnæs, N. 1975: Palaegeography and facies distribution in the Tertiary of Denmark and surrounding areas. *Norges geol. Unders. 316*, 289–311.

A short review of the stratigraphical terminology and nomenclature of the Danish Tertiary is presented. The Lower Paleocene (Danian) is lithologically similar to the Upper Cretaceous but has a more variable facies distribution, and is mainly restricted to the area north of the Fyn–Ringkøbing ridge. The Upper Paleocene (Selandian) is also variable in facies and consists mostly of glauconitic clays and marls. In the Eocene, volcanic ash, diatomites and extremely fine-grained clays were deposited. One of the still incompletely understood palaeogeographical peculiarities in the area during the Cretaceous to Oligocene is the predominance of sediments and faunas of open oceanic, and partly deep water type close to the Fennoscandian (Baltic) Shield. Possible explanations for this anomaly are discussed.

In the Upper Oligocene terrigenous clastic sand begins to appear in quantity, and the Miocene is dominated by an alternation between terrestrial/limnic sands with lignite, and marine, glauconitic and micaceous clays and sands. A Miocene drainage system from the east and north-east can be discerned, and it appears to be shifted towards the south during the Pliocene, possibly reflecting the uplift of the Atlantic margin of the Fennoscandian Shield. The mechanism and and timing of this tectonic event is not completely understood, but it must have been of major importance for the palaeogeographical development of the region, and for the formation of the sedimentary basins.

It is concluded that sediment supply and facies were influenced mainly by biological and tectonic parameters, whereas climate played a less important role. The climate appears to have been tropical to sub-tropical in the region up to, and including, the Upper Miocene, with a rapid cooling during the Pliocene, culminating in the Pleistocene glaciations.

Except for minor fluctuations and Southern Sweden, the coastlines were roughly parallel and close to the present coasts of Norway and Sweden until the Miocene. Since then the area has been low, fluctuating between flats and shallow sea.

N. Spjeldnæs, Palæokologisk Afdeling, Aarhus Universitet, DK–8000 Aarhus C, Denmark

Introduction

A glance at a map of the distribution of Tertiary strata around the North Sea (Fig. 1) shows that Denmark has a key position in this respect. Here the exposed Tertiary is closest to the productive areas in the northern part of the North Sea, and because of the structural trends, the development in Denmark can be extrapolated out into the off-shore region.

The Tertiary of Denmark, and surrounding regions, has been studied for a long time, and one would immediately suppose that these areas were known in detail. There are, however, restrictions which must be considered when evaluating the data from Denmark. Besides the imprecision inherent in the drilling methods used, there are two additional sources of error and inaccuracy.

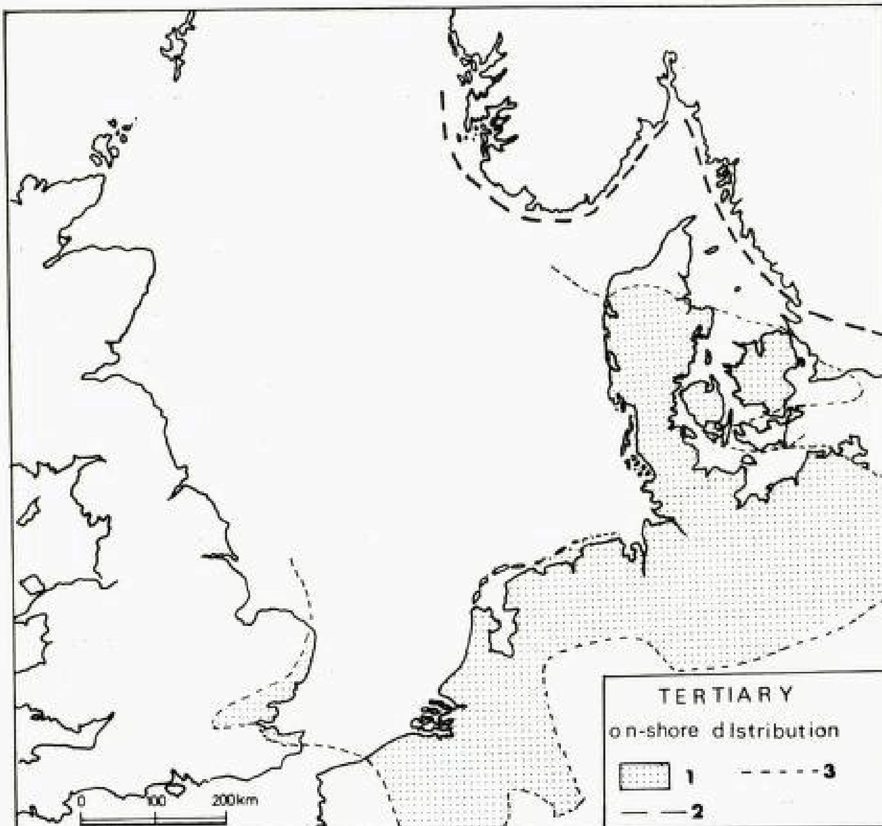

Fig. 1. The distribution of the Danish Tertiary in relation to the North Sea basins.
1. Areas where Tertiary strata are found (partly covered with Pleistocene and Holocene).
2. The generalized coast-line, from the end of the Lower Cretaceous to the Oligocene.
3. Approximate outcrop boundary of the Tertiary.

During the Pleistocene glaciations, thick ice-sheets invaded Danish territory, and to some extent ploughed up the soft Tertiary beds which lay on top of the thick sedimentary sequence. This has produced some spectacular tectonics, with intense folding and thrusting, which obscures contacts and make precise measurements of thickness rather difficult. The result is that the thicknesses given must be corrected, and the best ones are those which result from educated estimates made by experienced geologists. It would be easy to construct complicated patterns of distribution, but most of the information should not be taken at face value because of these sources of error.

Salt tectonics have deformed the soft Tertiary sediments, and on and around these structures the thickness may vary considerably (Madirazza 1968a, 1968b; Lind et al. 1972). This is partly due to original changes in thickness, and even in facies, because the salt structures were topographic features at the time of deposition. In other cases the changes in thickness may be due to differential compaction, to slumping or faulting due to post-depositional movements of the salt, and to solution features.

Most of the palaeontological and sedimentological work on the Danish Tertiary has traditionally been made on material from exposures. As mentioned above, the quality of the exposures — which at the surface have been overridden by the Pleistocene ice — is generally not good and an increasing amount of work is now carried out on material from drillings.

The bulk of the borings are short ones, for water or engineering studies (the most important ones in connection with construction of bridges). Even though most of these borings are short ones, and the information — especially in the older ones — is not always complete, they give an excellent coverage over almost all Denmark. They are all registered in the drilling archive of the Danish Geological Survey, where they form a wealth of useful information. The first maps with this kind of drilling data have been published (Andersen 1973), and will be an indispensable tool for the study of the geology of Denmark when they have been completed.

The deep drillings have been made in the search for hydrocarbons, and especially those made in the first phase of exploration (cf. Sorgenfrei & Buch 1964; Sorgenfrei 1966) must be used with considerable caution, since most of them were made on or near salt-structures where the Tertiary strata may have been deformed.

The new borings, made in the second phase of exploration, are better in this respect (Rasmussen 1973b, 1974), and are the key points in the description of the Danish Tertiary. Seismic studies, done in connection with the exploration, have not and cannot be used extensively in working out the detailed stratigraphy of the Tertiary, because of the low velocity contrast between the different formations. The only useful information obtained is the depth to the top of the carbonates (Danian, or where it is missing, the Maastrichtian).

Much of the work on the Danish Tertiary is done by the Geological Survey of Denmark, and most of the information used in this paper is based on the work of the Survey staff. It should also be noted that an enormous amount of information exists — mostly unpublished, and partly classified — in the Survey archives. The presentation given here is therefore incomplete, as part of the information has not been available. It should be regarded as a review of the present status of publicly available information.

Danian

Lithologically the Danian consists of carbonates, resembling those of the Upper Cretaceous (Ødum 1926; Brotzen 1948). Three main types are found: coccolite limestone, Bryozoan limestone and coral limestone. The coccolite limestone is an ooze, where planctonic foraminifera and coccolites dominate among the biotic constituents. It is the most common rock-type, especially in the central part of the basin.

The Bryozoan limestone is characterized by small bryozoan fragments. It grades into the coccolite facies through limestones with fewer bryozoans, floating in a coccolite ooze matrix. The most striking development is in the bryozoan banks, elongate mound-like structures often occurring in large numbers with more or less parallel orientation. This rock typically has a high porosity, part of which is residing in the cavities within the bryozoan and other fossils, which are normally not filled diagenetically. The permeability varies widely, depending on the fracturing which seems to be related to the specific tectonic setting in the individual localities.

The formation of the bryozoan bank has been much debated, and has recently been studied by Cheetham (1971), Thomsen (1973a, 1973b) and Nielsen (1973). Thomsen (1973b) concludes that the banks have been formed in an environment with only weak water movement (currents below wave-base), and that they are authigenic and not the result of long-distance transportation of the bryozoan, which have apparently been fragmented biogenetically and not mechanically.

Such bryozoan carbonates are not uncommon on continental shelves in regions of slight supply of terrigenous sediments, and at depths of 100–400 m.

The coral limestone is found especially at Fakse, which, together with Stevns Klint nearby, is the type locality for the Danian. The limestone is no bioherm, but rather a type of coral thicket with a rich and varied fauna, but without algae. There are few good depth indicators, and none of wave activity. H. W. Rasmussen (1973) has indicated that some of the burrows found may belong to nocturnal animals seeking darkness during the day. This would imply that the coral thicket was in the photic zone, but more evidence is needed before a good estimate of the depth of deposition can be given. The spatial relationship to the bryozoan limestone indicates a similar, or perhaps slightly shallower depth, for the coral limestone.

At one locality on the Fyn–Ringkøbing High, Klintholm, the bryozoan limestone is somewhat differently developed. It is more coarse-grained and well-washed, and the grain-size distribution both of the bryozoan fragments and of other fossils is different from that of the bryozoan banks. This may imply shallower depths, perhaps even above wave-base, and is consistent with the position of this locality (and Fakse) on the Fyn–Ringkøbing High.

The Danian is resting on the Maastrichtian with a slight disconformity. Detailed studies indicate that the disconformity was a double, or even multiple one, which stops in the sedimentation, with some erosion, or at least the formation of intensely burrowed 'hard-grounds'. It is not known if this represents a total emergence above sea level, or only a period of submarine non-sedimentation, but anyhow it indicates a shallower water depth.

Thomsen (1974) has recently shown that the presumed thick, Danian, bryozoan limestone on the northern flank of the Fyn–Ringkøbing Ridge really is of Maastrichtian age. Such bryozoan limestones of Upper Maastrichtian age have been known, but only as thin and local developments. Our present knowledge indicates that the Maastrichtian is developed in bryozoan limestone facies on the ridge, which was therefore a topographic high, exposed to slight currents but probably not to wave activity.

The Danian seems to be missing on the ridge itself, but occurs as a thin

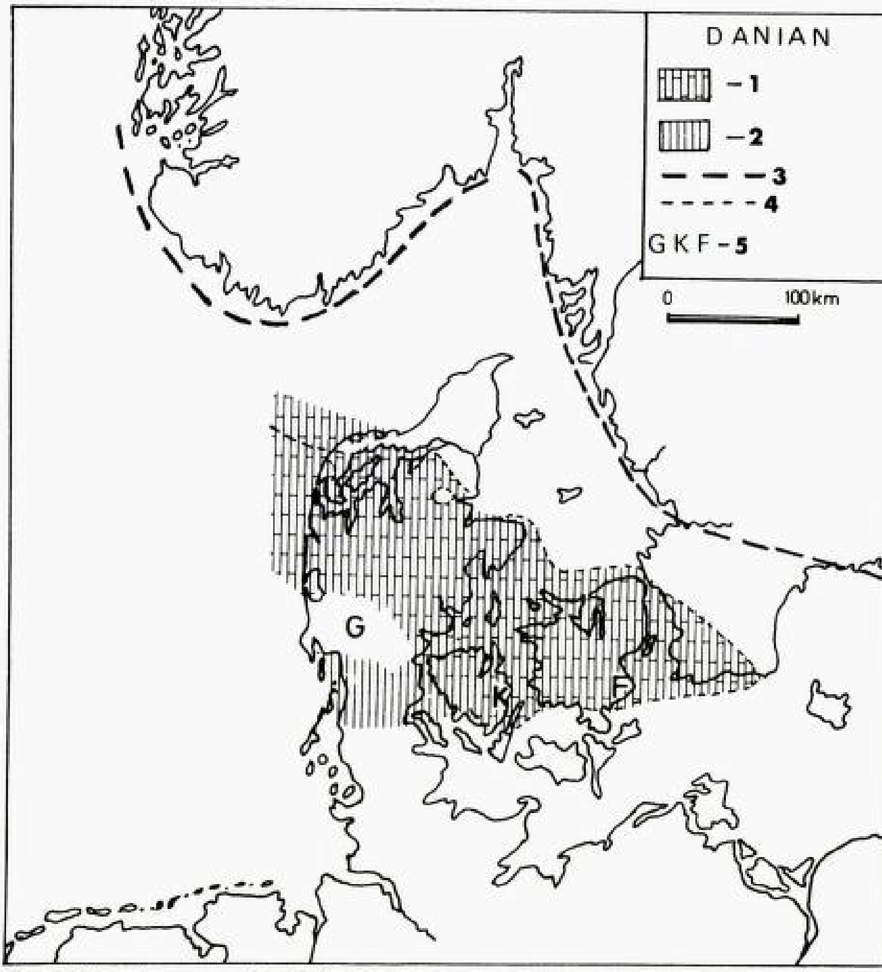

Fig. 2. The palaeogeography of the Danian.

1. Normal Danian represented.
2. Only thin Danian IV represented.
3. Generalized coast-line.
4. Outcrop boundary.
5. Important localities (G – Grindsted, F – Fakse, K – Klintholm).

sliver on the southern flank or in a special, shallower facies (at Klintholm) which differs from the ordinary bryozoan limestone, both in faunal composition and granulometry.

The Danian is subdivided either by planctonic foraminifera (cf. Berggren 1962, 1971; Bang 1969, 1973; Hansen 1970; Bang, in Rasmussen 1972), by coccoliths (Perch-Nielsen 1971), or by echinid spines (cf. Brotzen 1959). The methods are in general agreement, even though there are some discrepancies. Of the four biostratigraphical zones, only the uppermost one (Danian IV) is found on the southern flank of the Fyn–Ringkøbing Ridge (Fig. 2), and this period represents the greatest extent of the Danian transgression.

Selandian

In contrast to the Danian, the Selandian consists of clays with minor quantities of glauconite sand and marls (Gry 1935; Müller 1937). The detailed distribution of facies is not known, but the greatest variety of facies is found in the east, on Sealand. Here the basal beds are glauconite sand (Lellinge Greensand) followed by marl (Kerteminde Marl). Upwards the clays are poorer in carbonates, and grade imperceptibly into the unfossiliferous carbonate-free clays forming the transition to the Eocene.

Further west, the basal beds are thinner and less varied, and the bulk of the sequence consists of marls and clays with very little coarse-grained terrigenous components, and a decreasing carbonate content towards the top. The detailed stratigraphy of this interval is not known, but it is not altogether unfossiliferous. Foraminifera preserved as pseudomorphs of clinoptilolite (Hansen & Andersen 1969) in one of the localities, indicate that at least part of this sequence really belongs in the Palaeocene.

The Selandian sedimentation seems almost unaffected by movements of the Fyn–Ringkøbing High, but in the absence of detailed studies it is not definitely known if the basal beds on the high are younger than those in the basin. The age difference seems to be small.

The mollusc fauna of the Selandian is fairly well known from a series of older papers, e.g. by Ravn, but the modern biostratigraphy is based mostly on the foraminifera, and the nannofossils (Hansen 1968, Bang 1973).

The boundary towards the Danian is not only a sharp lithological one, but the basal Selandian is often resting on an eroded Danian, where the upper part is missing. In the basal beds of the Selandian there are reworked fossils of Maastrichtian age, indicating that the erosion cut down right through the Danian. This can be explained by assuming a Danian/Selandian faulting, creating fault escarpments on which rapid erosion took place. The vertical components of the faulting exceeded the thickness of the Danian ($\approx$ 250 m.). The direction of the faults is not known, but it was probably parallel to the basin and high (NW–SE). Most of the erosion probably took place below sea level. There are also indications of a certain topography in the pre-Selandian carbonate surface, probably a result of the same erosion.

The Selandian has been studied in less detail than the Danian, and much remains to be done on the detailed stratigraphy. The description given by Gry (1935) is the main source for the sedimentology and for the older literature.

Eocene

The Selandian grades imperceptibly into the Eocene, and because of the lack of fossil evidence it may be difficult to define the boundary precisely.

The practical boundary is therefore taken at the base of the volcanic ash series in the Moler Formation and its lateral equivalents. This formation is restricted to the basin in the northernmost outcrop of the Eocene, but the lower Eocene Røsnes and Lillebelt Formations are distributed all over that part of Denmark were the Eocene is found; very little in the way of systematic changes in facies has been observed.

There are some doubts as to where the boundary between the Palaeocene (Selandian) and Eocene should be placed in the North Sea area. Particularly in the subsurface geology, there is a tendency to place the boundary above the volcanic ash beds. The boundary used here is the traditional Danish one, which is convenient lithologically and supported by micropalaeontological evidence (Dinesen 1973).

The well known volcanic ash beds were the subject of the classical monograph by Bøggild (1918), and have later been studied by e.g., S. A. Andersen (1937) and Norin (1940). Most of the layers are of basaltic composition, but a few acid ones are also found. Studies of the thickness of the individual beds (which may, however, change due to compaction and tectonic deformation (cf. Madirazza & Fregerslev 1969)) and the granulometry indicate a northerly or north-westerly source for the ash and a distance to this source of 100–200 km, or perhaps less. The ash beds are known throughout the North Sea but are thickest and best preserved in the Moler Basin in North Jutland (Fig. 3). This has probably been substantiated by the finding of a magnetic structure, interpreted as a basaltic volcanic cone, off Kristiansand in the Skagerrak (Sharma 1970). Basaltic rocks, presumably in situ and intrusive into Jurassic strata, have been dredged further to the east (Noe-Nygaard 1967).

The 'Skagerrak Volcano' has not yet been positively identified by seismic studies, and it is not known with certainty if the magnetic anomaly is really a basaltic volcano, or whether it was one or more centres of volcanic activity in the Skagerrak in the early Eocene.

The Diatomites of the Moler Basin are richly fossiliferous, with a varied fish fauna dominated by oceanic, open water forms (Bonde 1966), plant fragments and wind-blown insects indicating forest conditions (Henriksen 1934; J. P. Andersen 1947; Heie 1970) on a neighbouring land, presumably situated to the north. Because of the distribution pattern of the volcanic ash, the prevailing winds must have come from the north and norh-west.

The microfossils are dominated by diatoms (Stolley 1899), spores and pollen, with radiolarians and archeomonadids (Deflandre 1932) as minor constituents. Calcareous forms are normally absent, and benthonic fauna (ophiurids and a few molluscs) are restricted to a few bioturbated bedding planes.

The Moler diatomite biota was therefore dominantly planctonic with no bottom fauna, except in some few special cases. This is so striking that it cannot be due solely to lack of nutrients, but must be attributed to lack of oxygen in the bottom waters. The rather high content of organic matter, and the usually completely preserved fossils (fishes and insects) indicating absence of scavengers also support this.

The Moler diatomite contains some carbonate concretions and bands

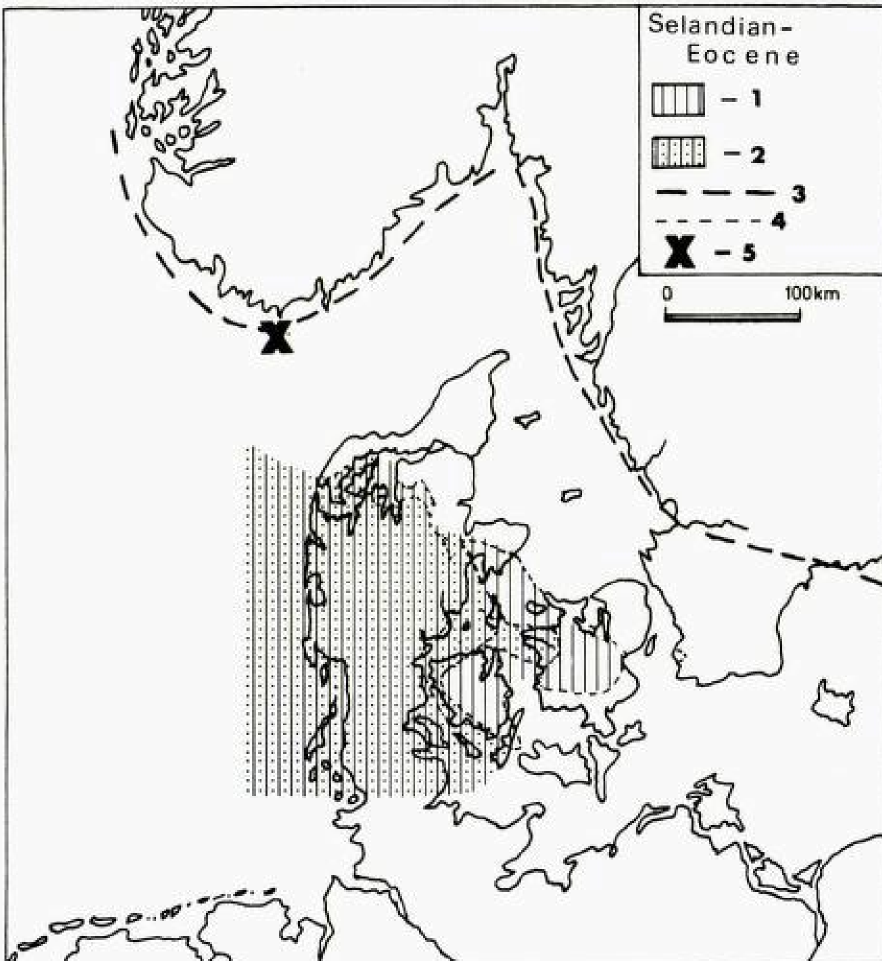

Fig. 3. The palaeogeography of the Selandian and Eocene.
1. Areas where only Selandian is represented.
2. Areas where both Selandian and Eocene are represented.
3. Generalized coast-line.
4. Outcrop boundary.
5. Presumed location of the Skagerrak volcano (geophysical anomaly).

(cement-stone) which are finely laminated like the rest of the diatomite, with the exception of the few bioturbated beds mentioned above.

There are two opposite views about the sedimentation rate of the diatomite. Bonde (1973) suggests that the lamination represents annual varves, indicating a high production rate of organic material, and consequently high net sedimentation rate (ca. 1000 mm/1000 years) in a short period (ca. 60,000 years). Sharma's studies (1969) of the magnetic reversals of the volcanic ashbeds indicate a much longer time for the duration of the formation, approximately 3 mill. years, given an average net sedimentation rate of only 20 mm/1000 years. Both models are compatible with the fact that there are over 130 well-defined ash beds in the sequence, and with the observations on sedimentation rates of similar recent sediments.

Sharma's model is most extreme, requiring sedimentation rates of oceanic type, indicating low nutrient supply and restricted organic production. No breaks have been detected in the sequence to explain the low sedimentation rate, and it would be surprising if the Eocene 1 (of which the Moler formation is a part) had a duration of over 3 mill. years.

Bonde's model requires a rich nutrient supply and high organic productivity, and is compatible with sedimentation rates found in such environments today. The anaerobic bottom conditions also indicate a high organic productivity in the surface water, rather than a low one. In the clays, which are the lateral equivalent to the Moler with the same volcanic ash beds, Bonde's model would give a net average sedimentation rate of about 200 mm/1000 year, which seems to be realistic. If this sedimentation rate were typical of the similar sediments in the rest of the Eocene in Denmark, the sequence would have been several times thicker, and a consequence of Bonde's model is that less than 1/6 of the Eocene is represented as continuous sediments in Denmark. The lower sedimentation rate implicit in Sharma's model would — under the same assumptions — give an unrealistically long time interval for the Eocene.

The Eocene above the volcanic ash beds (and the lateral equivalents of the Moler) has a rather poor macrofauna — some fishes, molluscs and trace fossils. The microfossils are much better represented, and the foraminifera have been studied by Dinesen (1972, 1973) and the coccoliths by Perch-Nielsen (1971). There are also numerous hystricospaerids, diatoms and other siliceous microfossils.

In contrast to the Lower Eocene, which is uniformly distributed, the Upper Eocene is most completely developed in the North Jutland basin (Fig. 4), whereas only the lower part is found in the Fyn–Ringkøbing ridge (cf. Dinesen 1973).

The development of the Eocene, and partly also the Selandian (Upper Paleocene), poses some interesting paleogeographical questions which are important to the understanding of the development of the Danish Tertiary.

The sediments are normally extremely fine-grained, with more than 70–80% less than 2 μ, and often more than half below 0.2 μ. Coarse clastic material is extremely sparse, the small coarse fraction consisting almost exclusively of organic remains or products of diagenetic processes. The sedimentation rate (calculated from the number of microfossils, particularly hystricospheres, and the ash beds — see above) is very low, at least for many beds.

The fauna consists mostly of microfossils; three types of foraminifer-faunas occur (Dinesen 1972), one dominated (75–90%) by planctonics, one with almost exclusively agglutinating, benthonic types, and a third dominated by calcareous, benthonic ones, but often with a much higher content of planctonics than in the present North Sea. The details of the successions of these faunal assemblages in the individual sections are not known, but a general review has been given by Dinesen (1972).

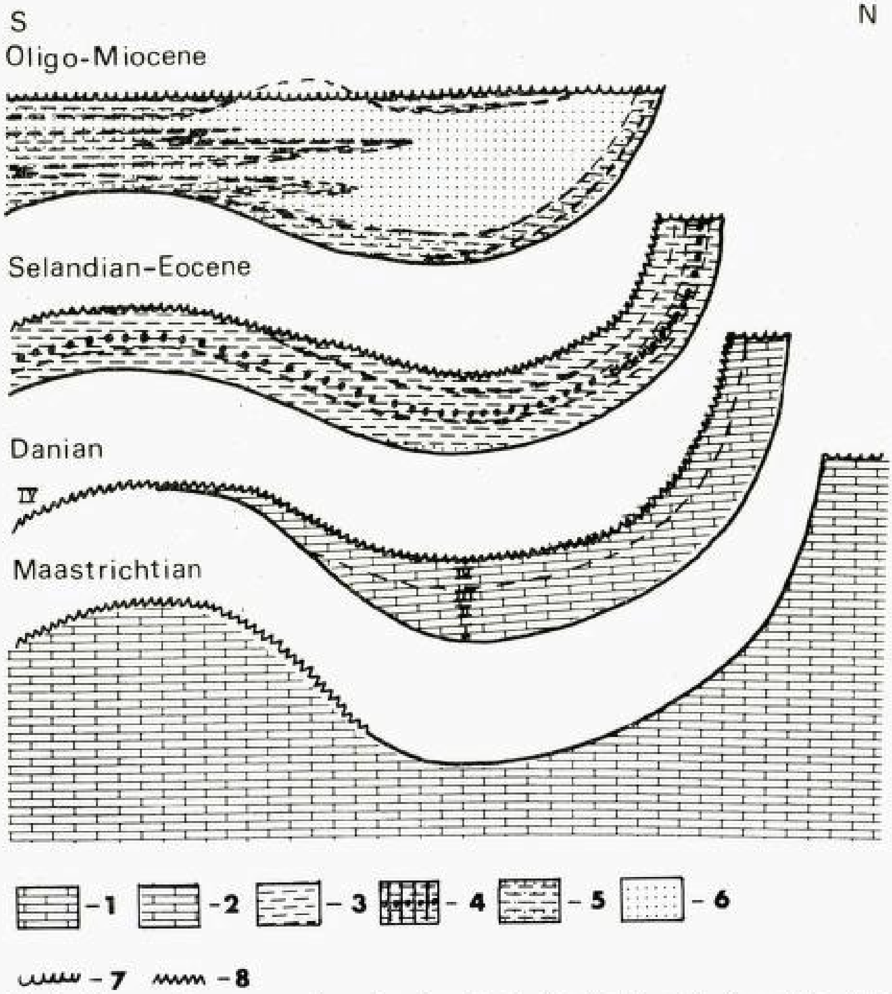

Fig. 4. Diagramatic, exaggerated section through the North Jutland Basin and the Fyn-Ringkøbing High. The thicknesses are only approximate.

1. Chalk (mostly coccolith ooze carbonates).
2. Bryozoan carbonates of chalk type.
3. Fine-grained clays.
4. Diatomite, with volcanic ash beds.
5. Micaceous clays.
6. Sand and silt with lignite.
7. Truncation of beds due to glacial erosion.
8. Major surface of erosion or non-deposition.

Taken at face value, these data would indicate very deep and partly cold water. The carbonate-free sediments with agglutinating faunas recall those deposited below the compensation depth. An unsuspecting student would most likely refer some of the samples from the Danish Eocene to abyssal sediments in an oceanic basin. Since it is geologically improbable that the Danish basin was more than 2000 m deep in the Eocene, one must conclude that the sediments mimic abyssal, oceanic sediments in a way which reflects rather unusual geological and oceanographic conditions.

The two important questions are: (i) why did not the Fennoscandian Shield yield any appreciable amount of terrigenous clastics to the basin (it did up to the lower Cretaceous, and from the Oligocene on) ?; and (ii) what kind of oceanographic model can explain the peculiar faunas and sediments?

(i) The amount of clastics coming from a land area will depend on the relief and the run-off. Most of the Shield probably had a low relief, such as today, or even lower, but some geomorphological studies (Gjessing 1967) indicate that at least parts of Norway had a moderate relief even in the Mesozoic. Under the prevailing subtropical climate, considerable amounts of weathering products would be delivered even from a low-relief area. If the Shield were at least partly covered by shallow sea, currents generated by wind and tides would certainly transport more and coarser material into the basin.

It is difficult to estimate the amount and the seasonality of the run-off. It has been suggested that the area was an arid one, but that would be difficult to explain meteorologically at these (palaeontological) latitudes, even if the climatic belts were displaced and more diffuse than now. Under arid conditions, in the absence of a vegetation cover the weathering products would be more mobile, and arid regions normally are also centres of high pressure, which would give considerable eolian sedimentation into adjoining basins. Sediments which could be identified as eolian have not been observed in the Danish Eocene.

The presence of a rich insect fauna, indicating forest conditions (J. P. Andersen 1947, Bonde 1973), in the diatomite beds in the basin supports an interpretation with a coast-line close by, or nearer than the present one, and the land area covered with a permanent vegetation cover in a subtropical, at least partly, humid climate.

The same lack of terrigenous clastics is found in the Upper Cretaceous and Danian, when carbonate sediments prevailed. In one instance, at Åsen, north of Ivøsjøn in Scania, there are indications of mangrove-swamp-like conditions. Here, at the extreme limit of the maximal Campanian transgression, the thin marine beds resting on kaolinitic clay with plants contain oysters which were attached to roots or branches, such as one finds in mangrove swamps to-day (Kegel Christensen, pers. comm.). If the estuaries during this period were characterized by mangrove swamps (or similar vegetation), the terrigenous sediments would be effectively filtered away and only put into recirculation when the 'filter' broke down for climatic and/or tectonic reasons in the Oligocene.

It is dangerous to generalize from one single locality, but the best model to explain the lack of terrigenous deposits in the period from the end of the Middle Cretaceous to the Lower Oligocene in our area seems to be that the land area had a low relief, a thick and permanent vegetation cover, moderate run-off rather evenly distributed over the year, and with mangrove swamps or similar vegetation operating as filters in the estuaries. The oceanographic model to explain this situation must be one which explains the seemingly deep and cold water, under conditions giving low sedimentation rates close to the continent, and under a subtropical climate.

In the Lower Eocene London Clay, there is also a discrepancy between the plant remains washed in from land, indicating a tropical to subtropical climate, and the marine invertebrate bottom fauna which appears to indicate cooler conditions.

To explain the diatomitic rocks in the centre of the basin, Bonde (1973) suggested a model with winds from the north (suggested by the distribution pattern of the volcanic ash) creating a surface current. Deflected by the Coriolis force, this would lead to upwelling of cold water. A more westerly placed counter-current would deprive the base of nutrients. This hypothesis provides a logical explanation for the conditions observed, even if the upwelling takes place farther from the continental margin than in most modern upwellings. It supposes that conditions along the Norwegian coast were compatible with the model (with a marked N–S current along the coast), and this can certainly be tested by studies of the Eocene beds on the continental shelf of Norway.

It should be noted that this model is different from normal upwellings as it will bring cold water, not from the oceans, but from the deep-water layers of the shelf. These waters are normally less rich in nutrient than upwelling deep-ocean water. The model may therefore be better to explain the normal situation in the Selandian and Eocene, rather than the presumably short episode ($\approx$ 60,000 years) during which time the diatomite was formed, when parts of the basin probably received a rich supply of nutrients.

Oligocene

Comparatively little work has been done on the Danish Oligocene. The most recent work, with references to earlier literature, is found in Larsen & Dinesen (1959), Christensen (1969) and Christensen & Ulleberg (1973, 1974).

The Lower Oligocene appears to be totally missing, and the Middle Oligocene is found mostly in the basin. The sequence consists mostly of clays with the Viborg Formation at the base, sometimes with glauconitic basal beds (the Grunfør Formation) lying with a marked lithological boundary and a considerable and variable hiatus upon the Eocene Søvind Formation. The succeeding Branden Formation is lithologically similar, but the higher Cilleborg and Sofienlund Formations have an increasing content of silt and sand towards the top.

On the Fyn–Ringkøbing Ridge, the whole of the Oligocene is missing; towards the north and west a more complete sequence is found, with more littoral sediments in the south-east.

The biostratigraphy is not completely known, but recent studies by Christensen & Ulleberg (1973, 1974) indicate that the faunas known at present represent only few zones (such as Ruppel 2), whereas others either are missing or are not yet identified.

The Middle Oligocene forms the base of the Upper Tertiary in the Danish stratigraphic terminology. In sharp contrast to the fine-grained clays of typical marine origin in the older Tertiary, the Oligo–Miocene presents a more varied picture, with sands, silts, micaceous clays and lignites, partly of fresh water and partly of marine origin.

In the Oligocene it is possible to discern the contours of a coast-line, even if littoral sediments are missing in the basin. The clay minerals show a zonation similar to that found parallel to recent coast-lines on a continental shelf (Christensen 1969). From this it is possible to reconstruct a N–S trending coast-line through Kattegat turning westwards in N. Jutland. It is difficult to estimate the exact distance to the shore-line, but from sedimentological and ecological data it is reasonable to expect it to have been at or just west and south of the present coast of Sweden and S. Norway.

The position of the coast-line is difficult to define in the Danian to Eocene, since no direct evidence of proximity to a landmass is found in the sediments from this time interval. The sum of the geological evidence indicates that the generalized shore-line was fairly stable from the end of the Lower Cretaceous to the Oligocene, along and close to the present coast of S. Sweden and S. Norway (Fig. 5).

Considering the good exposures and the minute remains of older transgressions (Middle Cambrian and Campanian) which have been discovered along the west coast of Sweden, widespread Tertiary transgressions on to the Shield would certainly have left some traces. Geomorphological studies (Gjessing 1967) also support the idea that the landscape of S. Norway is an old, pre-Pleistocene or even pre-Tertiary one.

The geological mechanism governing this stable boundary between the Precambrian shield, which is now mostly on land, and the sediment area, which is mostly submerged also today (with Denmark as the most notable exception), is still imperfectly understood. Holtedahl (1953, 1960), who first pointed out this important boundary, suggested simple faulting or flexures (Holtedahl 1970). Geophysical studies have failed to show major faults, even if minor ones are common both in the Kattegat and in the Skagerrak. The old structural lines in the Precambrian are often parallel to the 'Holtedahl-line', and at least in S. Norway there is a spectacular fault system paralleling the coast which has been active intermittently from the Precambrian to the Pleistocene The tectonic movements along the Holtedahl-line were probably rather complicated in detail, and in chronology. They may have been comparatively 'plastic' deformations of flexture-type, alternating with sharply defined faulting, both in time and space. In a transitional area like Denmark, there are complicating elements such as the Fyn–Ringkøbing High and the comparatively high Precambrian area in northernmost Jutland, and the tectonic development was certainly not simple.

The development of the North Jutland basin and the structure of the region strongly suggest that this shield-basin boundary is an old and regular, large-scale feature, probably related to movements along neighbouring plate boundaries. The actual coast-line may have fluctuated somewhat, but basically the Shield was an area of erosion and uplift where ephemeral sediments were

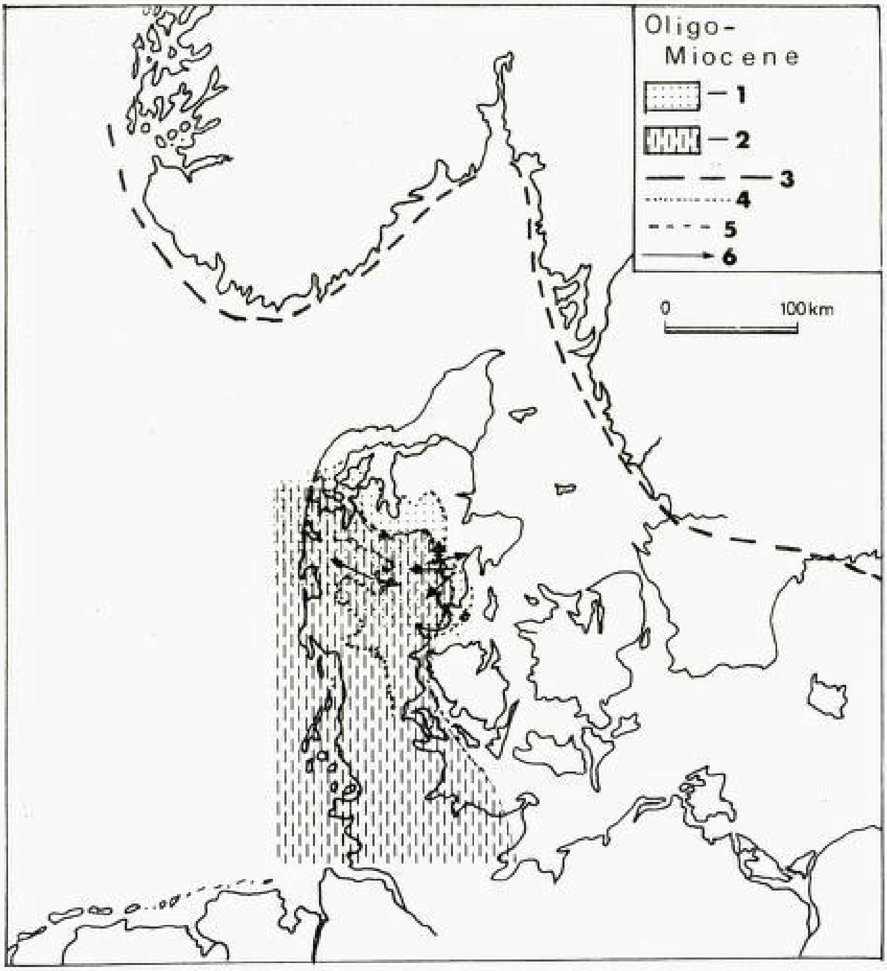

Fig. 5. Palaeogeography of the Oligo–Miocene.

1. Area where Oligocene is represented.
2. Area where Miocene is represented.
3. Generalized coast-line (this moved towards west and south during the Oligo–Miocene).
4. Western limit of the present outcrop of the marine Miocene.
5. Outcrop boundary.
6. Major transport directions (from the observations of Larsen & Friis 1973 and the present author).

rapidly and effectively removed by erosion before consolidation. Similarly, the basins were areas of sedimentation and subsidence. Even if there are hiatuses in the sequence, they are more due to non-deposition than to erosion, except locally on the Fyn–Ringkøbing High and on fault-cliffs formed by small-scale tectonics (small-scale = 200–300 m, compared to the 3000–6000 m along the Holtedahl-line).

Finally, the Pleistocene glaciers have levelled off and eroded both on the Shield and in the basin (especially along the Norwegian Channel), and disturbed the soft sediments. This disturbance is not only due to the direct erosion of the glaciers, but is also associated with isostatic movements resulting from loading and unloading of the ice-sheets. This process resulted in, among other things, reactivation of the salt-domes in the basin, with considerable deformation and even faulting, which could be confused with the tectonic activity of earlier phases (Madirazza 1968a, b).

In the Baltic the palaeogeography is more complex than in the Kattegat and the Skagerrak, and Scania forms a transitional zone between the Shield and the basin over which the transgressions are partly preserved. The Campanian transgression is — as could be expected — the one which reaches farthest in the Shield, but remains of Danian and Eocene transgressions are also found. It is important to note that the Tertiary sediments on the Kattegat side — just as the Cretaceous ones — retain their open oceanic character up to the contact with the Shield, with very little by way of real littoral sediments. They are found in the Cretaceous only where the beds are in virtual contact with the Precambrian rocks. The only exception to this rule, the Upper Cretaceous or Danian Åhus sandstone in E. Scania, is well into the Baltic. This indicates that there were sediment filters (mangrove swamps or the like) in the estuaries, which in connection with low relief on the Shield and run-off evenly distributed over the year, reduced the influx of terrigenous material into the basins to a minimum.

In the Baltic there are indications of Tertiary littoral sediments. Most of them are found as erratic boulders along the north coast of Germany. A review of their age, lithology and distribution is given by Hucke (1967, pp. 106–108). Our present knowledge is insufficient to locate the original source of these rocks, and thereby the Tertiary coast-line in the region, but it is likely that it was situated largely between a N–S line through Bornholm and a NW–SE line through the southern part of Öland, until the Oligo–Miocene regression transferred it towards the west and south.

No littoral sediments of Tertiary age are known from the Kattegat region. The only observation is by Gottsche (1883), who referred the so-called *Cyrena*-boulders, which are found as erratic boulders in part of E. Denmark, to the Oligocene. This observation has been widely repeated in the literature.

Studies in progress by F. Strauch (Cologne), O. Bruun Christensen (Copenhagen) and the author indicate that these boulders are of Wealden age, and in a facies which is well known from the brackish-water Wealden in N. Germany. They probably come from submarine exposures in the Kattegat, close to the Swedish coast.

Miocene

The Danish Miocene is well known through the classical monographs of Sorgenfrei (1940, 1958, 1960) and L. B. Rasmussen (1961, 1966, 1968). Lithologically the Miocene consists of an alternation of micaceous clays and sands. The micaceous clays carry a marine fauna dominated by molluscs (cf.

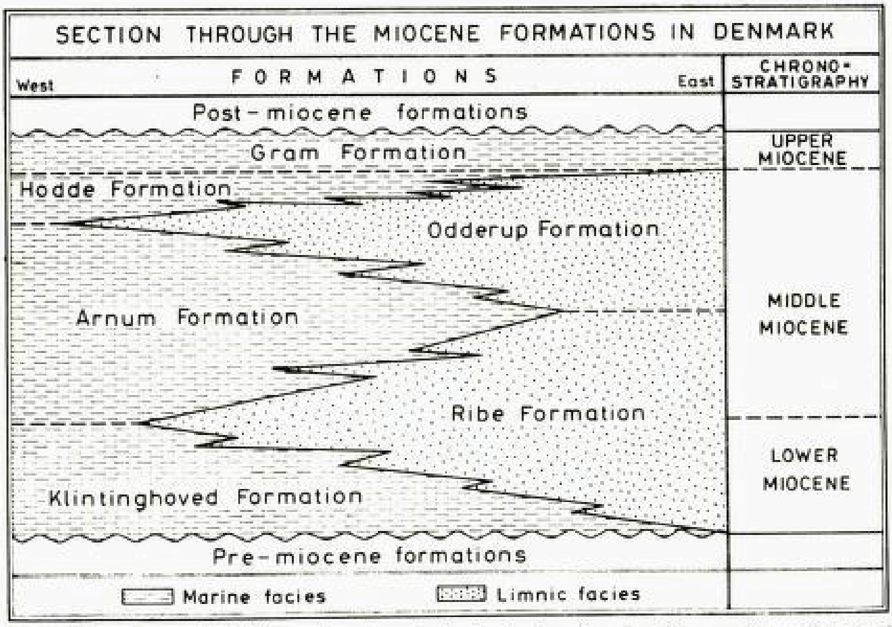

Fig. 6. Diagrammatic E–W section through Jutland, showing the Miocene formations and their inter-relationship (from Rasmussen 1966).

Sorgenfrei 1958; L. B. Rasmussen 1968) and foraminifera (Kristoffersen 1972, 1973 – in L. B. Rasmussen 1973b). Bryozoans, ostracods, fish otoliths and hystricospherids are common, and whales are often met with. Because of the well-studied faunas and the proximity in time to modern faunas, it is easy to evaluate the depth, temperature and other ecological parameters. The environment of the upper part of the Miocene (the Gram Formation) was one of an open shelf with marine water of normal salinity, climatic conditions similar to the west coast of North Africa or the SW Iberian Peninsula, and depths of around 100 m. The organic production was high, as was the sedimentation rate. The pore water of the sediments, but not the bottom water itself, was anaerobic, and pyrite is common in unweathered sediments. When the clay is exposed to weathering, the resulting acidity leads to leaching of the carbonates. Many localities, therefore, do not carry any carbonate-shelled fossils, only corroded ones. The content of fresh mica (dominantly muscovite) in flakes from more than 1 mm in diameter down almost into the clay fraction, indicates rapid transport and sedimentation, and unusually immature weathering considering the climatic situation where the source rock normally would be exposed to much more mature weathering. This indicates that the relief of the Shield had been considerably sharpened, and that this process continued throughout the Miocene, supplying fresh terrigenous material.

The detailed stratigraphy is seen in Fig. 6 (from L. B. Rasmussen 1966).

The lower, typical marine beds are lithologically the same type as the Gram Formation, but restricted faunas indicate reduced salinity or other adverse conditions. The occurrence of high percentages of planctonic forms in some of these restricted foraminiferal faunas indicates that the salinity was not the only restricting factor, and that the hydrographic situation was probably complicated and rapidly changing. Kristoffersen (1972) has described the faunal changes, and discussed this problem, in connection with the definition of the Oligo–Miocene boundary in Denmark.

The Hodde Formation has a characteristic lithology — a black, stiff clay with glauconite where the calcareous fossils are mostly missing, presumably due to leaching. The transitional zone to the Gram Formation is a strongly glauconitiferous clay horizon. This Hodde/Gram sequence is easily identified even without fossils, and forms an excellent marker horizon.

The top of the Miocene is always a surface of glacial erosion (except in the extreme south-west corner of Denmark, cf. L. B. Rasmussen 1958, 1966), and the local distribution and structure of the Miocene is much affected by glacial activity (cf. L. B. Rasmussen 1966). The beds below the base of the Gram Formation increase rapidly in thickness towards the west; at the Danish west coast they reach over 300 m, and in the North Sea they exceed 800 m (L. B. Rasmussen 1973a, 1974). This isopach pattern, and the fact that the sediments apparently do not become finer westwards, indicate that the basin pattern also changed radically in the Oligo–Miocene, and that at least the greater part of the terrigenous sediments in the central North Sea came from the north (Norway) or north-west (Scotland), rather than through the drainage system across Denmark.

The sand sequence in the Miocene (cf. Fig. 6) is partly fluviatile–lacustrine–estuarine, and partly marine–littoral. The sedimentary environment with micaceous clays requires large bodies of sand from which the clay was winnowed, between the area where the clays was deposited and the land. Part of this, low-lying, sand area was probably at least partly above sea level, but was dominated by marine influences. The transport directions found by Larsen & Friis (1973) show that directions from between W and NW and from between NE and SE dominate. Even if the observational material is not really sufficient, the author would like to interpret the NE/SE set as indicating the original direction of fluviatile transport (i.e., towards SW to NW), whereas the W–NW trends are more likely due to tongues of marine transgressions coming in from the W and NW. This is also supported by observations in the lignite pits, where the sand between the lignites shows transport from the east, whereas the coarsely cross-bedded sand above the lignites shows current directions from W and NW. The real picture was certainly a rather complicated one, with rapidly changing environments and transport directions and many separate sand tongues. Fjeldsø Christensen (in Koch et al. 1973) has shown that there are plant-bearing clays between the presumably marine sand and the marine Hodde Clay in one locality. More observations are therefore needed in order to unravel the detailed depositional picture of the lignite basin.

The lignites, and their flora and depositional environment, have been studied by Koch & Friedrich (1970) and Koch et al. (1973). Their preliminary results on partly excellently preserved plant remains show a strong ecological resemblance to the North German lignites, and further studies are likely to give precise datings of the lignite beds.

Pliocene

In Denmark, Pliocene rocks are found only in the extreme south-west corner (L. B. Rasmussen 1958, 1973a). They represent a continuation of the Upper Miocene, and like the Miocene they appear to thicken and become more complete towards the west, beneath the North Sea (L. B. Rasmussen 1974).

LATE TERTIARY MOVEMENTS

There are two types of derived siliceous material which are relevant to the interpretation of the youngest Tertiary in Denmark. The first is the chert conglomerate described by Ødum (1968). This is a siliceous rock, composed of angular or slightly rounded fragments of chert, mostly or wholly of Danian age, cemented with a silica matrix. The type of chert fragments indicates weathering on land in a fluviatile environment, and is entirely different from the well-rounded, often flat, ellipsoidal chert stones found in the beaches below cliffs of Maastrichtian or Danian chert-bearing carbonates. The distribution of the boulders of chert conglomerate indicate (Ødum 1968) a source between north of the present outcrop of the Danian (Fig. 2) and the Oslo Fjord. This indicates that in this area Danian carbonates, without an appreciable cover of younger rocks, were subject to subaerial weathering in Tertiary time. The original bedrock for the chert conglomerate has either been completely removed by glacial erosion (second last glaciation), or is still preserved on the bottom of the Skagerrak.

The only fossil found in the chert conglomerate — a pine-cone of Neogene affinities — also supports the interpretation of the rock being a product of subaerial weathering, and lag deposition.

In the sands intercalated with the lignites there are also silicified fossils ('silicificates'); these have been described by the author (Spjeldnæs, in Koch & Friedrich 1970, pp. 180–181). The fossils are of Early Palaeozoic age, and come from a restricted region in the Baltic where silicified fossils are common in rocks of this age, and where they have also been reported weathered free from submarine exposures (Veltheim 1962). Such silicificates have been known for a long time, but it must be mentioned that the term is used here in a restricted sense, including only the determinable fossils of well-known age (Middle Ordovician to Lower Silurian) and provenance (the area shaded in Fig. 7). As shown by Voigt (1970), silicified fossils also of Danian (and unknown) age can be found in similar situations, and uncritical use of the silicificates has led to erroneous stratigraphic conclusions (cf. Friis 1972). The author interprets the occurrence of the silicificates in the sands at Fasterholt (the main locality in the North Jutland lignite basin), as being derived from

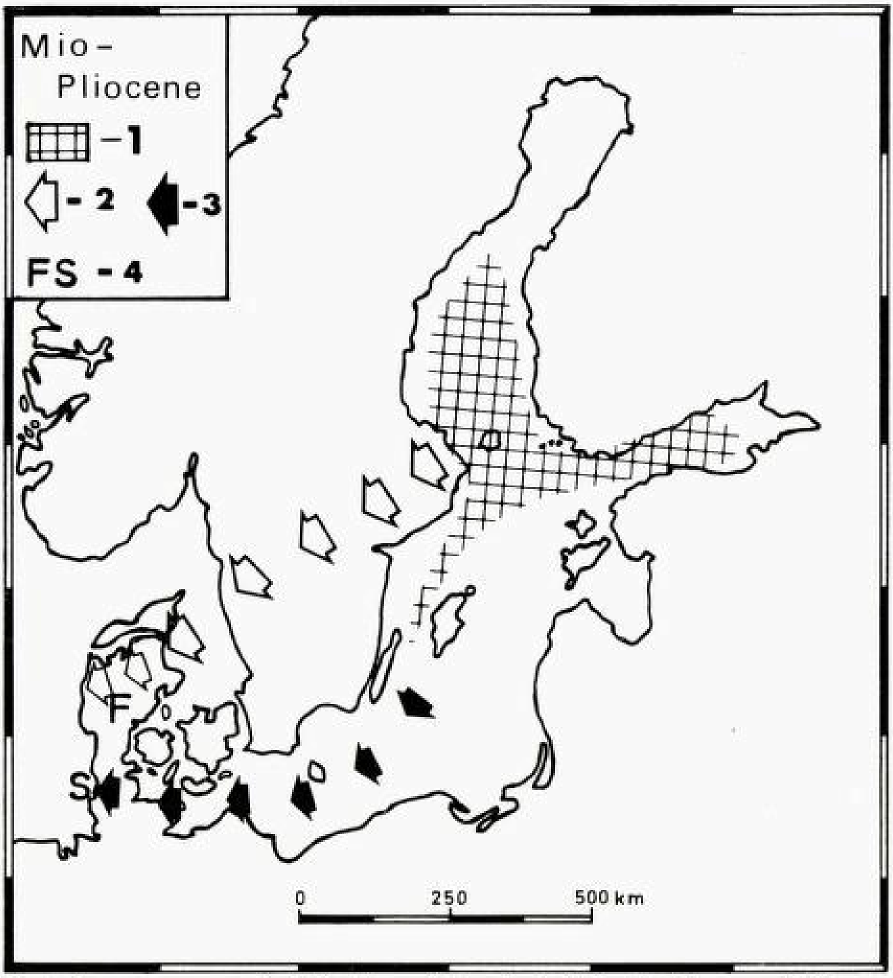

Fig. 7. Transport pattern of silicified lower Palaeozoic fossils in the Miocene and Pliocene.
1. Source area for the silicified Middle Ordovician to Lower Silurian fossils.
2. Suggested transport direction for the silicificates found in the Miocene sands in the North Jutland Basin.
3. Pliocene transport direction for the derived Ordovician and Silurian fossils in the Pliocene of Sylt.
4. Localities (F – Fasterholt, S – Sylt).

the central part of the Baltic Area (shaded area in Fig. 7) being transported by rivers flowing towards the south-west over central Sweden, deposited in the Miocene North Sea or one of its estuaries, and reworked into a transgressional sand with transport directions from the W and NW.

Material of the same age, but in coarser blocks, partly locally decalcified, have been found in the Pliocene of Sylt, in N. Germany, just south of the Danish border (Gripp 1967, 1968). This indicates an increased relief, or a different transport mechanism, and that the route of transportation was deflected towards the south. This may be related to a final phase in the oblique uplift of the western edge of the Fenno-Scandian Shield.

This oblique uplift was first described by Holtedahl (1934, 1960) and is a major tectonic feature in the North Sea area, and to the north along the Norwegian coast. As mentioned above, the tectonic mechanism is not yet unravelled, but it seems to be connected with the opening of the North Atlantic and rifting in the foreland. The chronology of the uplift is not well known, but the available evidence indicates that the major part of it was late, starting in the Oligocene and reaching its maximum after the Upper Miocene, even if movement had been going on along the Holtedahl-line since the Jurassic or even before. This will explain the marked and continuous sharpening of the relief, and the large quantities of fresh terrigenous material transported into the basin in the Miocene, Pliocene and early Pleistocene.

It has been suggested (cf. Gripp 1967) that the boulders found in the Pliocene fluviatile kaolinite sands in N. Germany (Sylt) were related to an early-Pliocene phase in the glaciation of Scandinavia. Such large blocks in a fluviatile environment can be more reasonably explained by transport by winter ice. The climatic conditions necessary for this are not more severe than those of the present day in the area, and it is suggested that rather than indicating a Pliocene glaciation, the climate cooled from that of the Upper Miocene lusitanian to conditions similar to those now prevailing in the North Sea area.

The oblique uplift of the Fenno-Scandian Shield may have triggered off the Pleistocene glaciations, by creating high mountains close to the ocean in the most favourable latitudes, but this can hardly be considered as the primary cause of the glaciations.

Summary

The palaeogeographic picture outlined here starts with an area where special conditions on the Baltic Shield (low relief, run-off evenly distributed during the year in connection with sediment filters in the estuaries) provided an unusually low supply of terrigenous clastics into the Danish part of the North Sea area. These conditions changed in the Oligocene, and there was a great influx of freshly weathered material in the Upper Oligocene and Miocene.

The coastline was probably parallel and close to the present coastline of S. Norway and W. Sweden from the end of the Middle Cretaceous into the Oligocene. In the later Oligocene and Miocene it fluctuated above Denmark, and the area became land in the Pliocene.

Three distinct phases of movements can be identified. The strongest was between the Danian and Selandian, where faulting caused severe disruption of the normal sedimentation pattern, and included the formation of fault-cliffs (probably dominantly submarine, but to a minor degree also on land). The available data give no clear picture of the fault-pattern, but it seems to be connected with the NW–SE trend of the Fyn–Ringkøbing High.

The second phase is expressed as a hiatus between the Eocene and the Oligocene. The younger Eocene and the Middle Oligocene are missing on the

High, and the Lower Oligocene appears to be missing all over Denmark. Because of the plastic sediments deposited in this time interval, it is difficult to define the types of movement, since vertical movements in these rocks are likely to have been taken up plastically and softened out. It is therefore not known if part of the faulting observed in the more brittle carbonate rocks took place in this interval.

The third phase of movements seems to represent the major period of oblique uplift of the western edge of the Fenno-Scandian Shield. The tectonic style and type of movements of this phase are not known, and pose some perplexing problems. In Denmark, these movements are indicated by the strongly changing sedimentary environment (more coarse, freshly weathered terrigenous material) and the drift of the coast-line towards the south-west.

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Late Precambrian Stratigraphy and Structure of the North-Eastern Margin of the Fennoscandian Shield (East Finnmark - Timan Region)

ANNA SIEDLECKA

Siedlecka, A. 1975: Late Precambrian stratigraphy and structure of the north-eastern margin of the Fennoscandian Shield (East Finnmark – Timan Region). *Norges geol. Unders. 316*, 313–348.

Late Precambrian (Riphean and Vendian) rocks occur along the north-eastern margin of the Fennoscandian Shield and the Russian Platform in three zones: 1) the Varanger–Kildin, 2) the Timan–Kanin and 3) the Southern Barents Shelf zones. The last occupies an intermediate position and is entirely covered by the sea.

Varanger–Kildin and Timan–Kanin inter-zonal sedimentary facies comparisons suggest a correlation between the miogeosynclinal sequence of the former zone and the succession of strata of the Central Timans regarded as marginal miogeosynclinal in origin. The internal miogeosynclinal facies of the Timans may be expected to occur in the deep substratum of the Southern Barents Shelf while equivalents of the pericratonic deposits of the Varanger–Kildin Zone may possibly be hidden beneath the Palaeozoic strata of the Russian Platform, assuming that the SW–NE margin-to-basin facies distribution constitutes a regional feature. The linear macro-facies distribution is suggested to reflect the development of a NW–SE trending continental margin to the Fennoscandian Shield at the beginning of the Middle Riphean; the basin has been filled with sediments during the Middle and Upper Riphean.

Geochronological investigations have shown that an early metamorphism of the accumulated deposits in the Kanin–Timan region and of intrusions of gabbroic rocks was essentially completed around 640–620 m.y. Igneous activity extended until ca. 500 m.y. with a second metamorphism around 525–520 m.y. This outline of the sedimentary and structural history of the Varanger–Kanin–Timan Region indicates that the Timanian Fold Belt, embracing the Timans proper, the Kanin Peninsula and its northwestward continuation, has originated during the Baikalian Orogeny pre-dating the Caledonian orogenic movements. The continuation of the internal portion of the orogen is inferred to be hidden beneath the southeastern Barents Shelf, its outer, peripheral portion cropping out in the Varanger–Kildin zone.

The lateral, cross-strike extent of the Timanian Baikalides is as yet uncertain. Alternative structural models suggested by Russian workers for the area embracing not only the Timans but also the Northern and Polar Urals and the Barents Shelf are essentially dependent upon the interpretation of the nature of the Pechora Basin substratum. According to one model, the miogeosynclinal Riphean beds of the Timans have their eugeosynclinal equivalents in the substratum of the Pechora Basin and in the Urals and are thought to have been deposited in an extensive Timanian–Uralian geosyncline. A second model, accepted here, suggests the substratum of the Pechora Basin to be pre-Riphean in age, thus inferring the occurrence of two depositional basins, Uralian and Timanian, separated from each other by an older craton.

The narrow orogenic belt of the Timanian Baikalides is oriented roughly transversely to the Scandinavian Caledonides in the west and to the Uralides in the east. In their relatively simple and short history the Timanides contrast with these major Caledonian and Uralian orogenies which were related to two extensive ancient oceans. Both the Proto-Atlantic and Uralian Oceans opened sometime in the Upper Precambrian; the opening of the Timanian basin probably occurred at approximately the same time, although this basin was infilled and then closed by about the end of the Precambrian. The Timanian basin has been designated in Russian literature as the *Timanian Aulacogen*, a sedimentary basin genetically related to and branching off the pre-Uralian geosyncline. In the west, the near-perpendicular trend of the Timanian Aulacogen relative to the Caledonides is also striking, and it is thought that the part of the Timanian Aulacogen related to the Uralian geosynclinal area extended longitudinally northwestwards towards the east margin of the northern Proto-Atlantic. It is further suggested that the presence of a triple junction of intracontinental spreading led to the development of the Timanian–Uralian dynamic system, with subsequent rapid activity and opening along two of the three spreading zones so producing the Uralian basin. On the other hand, active dilation in the third arm was slower and diminished fairly quickly, and this resulted in the preservation of a comparatively undeveloped spreading zone – the Timian Aulacogen.

A. Siedlecka, Norges geologiske undersøkelse, P.O.Box 3006, N–7001 Trondheim, Norway

Introduction

Late Precambrian (Riphean and Vendian) rocks crop out in Finnmark, Northern Norway, where they are most completely preserved on Varanger Peninsula. Further east, in Russia, they underlie the Rybachiy–Sredniy Peninsula and the Aynov and Kildin islands, and also appear on Kanin Peninsula and in the Timan Mountains (Fig. 1). Their subsurface occurrence has been detected in the Southern Barents Sea and in the Pechora Basin. In general, they appear to form an extensive NW–SE trending belt fringing the Fennoscandian Shield and its southeastern subsurface continuation. This belt is bordered by the Northern Scandinavian Caledonides in the north-west, while in the south-east it terminates against the Northern Urals.

The present paper is essentially a review of recent knowledge of the Late Precambrian geology of this *East-Finnmark–Timan Region*, its origin, stratigraphy and structure, with special attention paid to the Riphean rocks. It is also intended to be a partial guide to the Russian literature, partial because the literature survey is incomplete. Information has often had to be obtained from summaries and reviews in which ideas, rather than basic data, have been reproduced. The possibility of appraising the ideas has thus often been restricted, and consequently some of the presentations and conclusions may not be entirely correct. I have, however, tried to select and present the most important data and ideas of various workers and have attempted to outline the main geological events which have occurred in this region in Late Precambrian time.

For convenience of description the East Finnmark–Timan region will here be subdivided into three geographical zones: 1) the Varanger–Kildin Zone, 2) the Kanin–Timan Zone and 3) the Southern Barents Shelf Zone.

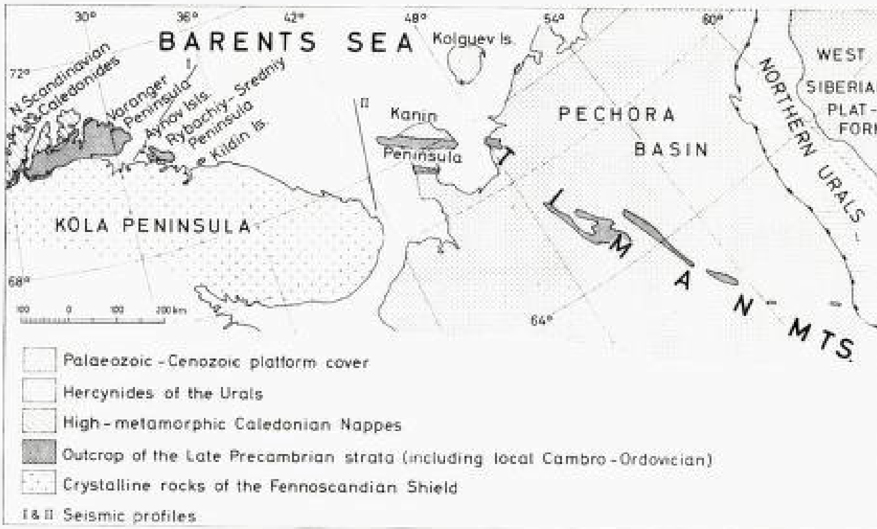


Fig. 1. Location of the East Finnmark–Timan Region (geology mainly after Atlasov 1964 and Markov 1966).

The Varanger–Kildin zone

The Varanger Peninsula, Aynov Islands, Rybachiy–Sredniy Peninsula and Kildin Island constitute a common geological region underlain by Riphean and Vendian (Eocambrian) sedimentary sequences, which contrast sharply with their substratum, i.e. with the igneous and metamorphic rocks of the adjacent northern margin of the Fennoscandian Shield (Fig. 2). Lithological similarities between the strata underlying the various areas of the Varanger–Kildin Zone have previously been pointed out by several authors and attempts at lithostratigraphic correlation have been made (e.g. Holtedahl 1918, 1960; Polkanov 1934; Keller et al. 1963, p. 605; Siedlecka & Siedlecki 1967, 1972; Bekker et al. 1970; Chumakov 1971). Detailed correlation has been and still is difficult to attain, however, mainly because of differences of terminology and the fragmentary character and often insufficient accuracy of lithostratigraphical descriptions.

Varanger Peninsula and adjacent areas of East Finnmark have been studied by many geologists (e.g. Holtedahl 1918; Føyn 1937, 1960, 1964, 1967, 1969; Rosendahl 1931, 1945; Reading 1965; Reading & Walker 1966; Beynon et al. 1967; Bjørlykke 1967; Siedlecka & Siedlecki 1967, 1969, 1971, 1972; Røe 1970; Edwards 1972; Roberts 1972; Siedlecka 1972, 1973; Teisseyre 1972; Banks et al. 1972, 1974) and our recent knowledge of the main features of the geology of this region may be summarized as follows.

Varanger Peninsula is underlain by Late Precambrian and lowermost Cambrian strata resting with a sedimentary contact on crystalline basement. A large NW–SE trending fault-zone, *the Komagelv–Trollfjord Fault Zone*, divides the peninsula into two geological regions (Siedlecka & Siedlecki 1967). South-west of the disturbance there is a $\leq$ 4000 m thick sequence of fluvial

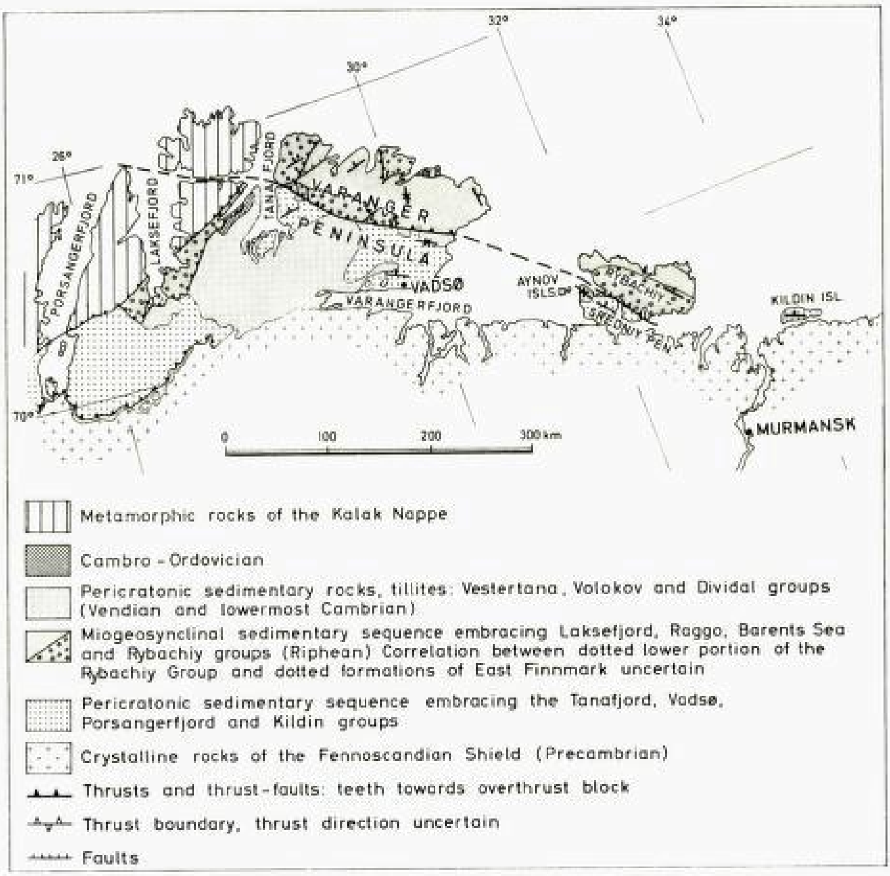


Fig. 2. The Varanger–Kildin Zone (mainly after Agapiev & Vronko, *in* Kharitonov 1958; Bekker et al. 1970, 1970a; Bogdanov, *in* Kharitonov 1958; Siedlecka & Siedlecki 1972).

and shallow-marine Riphean, Vendian and lowermost Cambrian strata. The Riphean rocks (Vadsø and Tanafjord Groups, total maximum thickness ca. 2300 m) are mostly terrigenous, although at the very top they include a carbonate unit containing columnar stromatolites. The Vendian and lowermost Cambrian strata (Vestertana Group) start with Varangian tillites, and there is a 2° unconformity between the Riphean and Vendian rocks, the former being successively removed southwards by a pre-Vendian erosion. Pringle (1973) has reported Rb/Sr whole rock isochron ages (sedimentation ages) of 668 ± 23 m.y. for shale samples from the lower Vestertana Group (Nyborg Formation) and 810 ± 90 m.y. for the Vadsø Group (*not* Tanafjord Group as incorrectly reported by the author mentioned).

North-eastern Varanger Peninsula is underlain by Riphean rocks (younger beds are absent) occurring in two rock-units separated by a tectonic contact and referred to as the Barents Sea and Raggo Groups (Siedlecka & Siedlecki 1967). Strata from parts of these two groups have been correlated (Siedlecka & Siedlecki 1972), and they are thought to represent a continuous, several thousand metres thick sequence of fluvial, shallow-marine, slope and base-of-slope deposits followed by sediments of a large deltaic system which then gave way to shore-zone terrigenous strata and, subordinately, carbonate deposits containing domal stromatolites. A fluvial regime was partly re-established at the top of this sequence. The substratum to this sedimentary succession is nowhere exposed on Varanger Peninsula; however, based on lithostratigraphic correlations (Føyn 1969; Siedlecka & Siedlecki 1972) it is thought that glacial and fluvial deposits of the lower Laksefjord Group, resting on crystalline rocks belonging to the Fennoscandian Shield at Laksefjord (Føyn 1960; Laird 1972), constitute the oldest record of the Late Precambrian deposition of East Finnmark (Siedlecka & Siedlecki 1972). An alternative correlation between glacial deposits of the Laksefjord and Vestertana Groups has been proposed thus suggesting a Vendian age also for the former group (Reading 1965; Laird 1972; Gayer 1973). Although there is no conclusive evidence to support either of the suggested correlations, data obtained from mapping and comparisons with the Rybachiy–Sredniy sedimentary successions seem to support the first of the correlations. The upper surface of this thick pile of sedimentary rocks is erosional and the sequence is unconformably overlain by Riphean rocks of the Tanafjord Group. This unconformity may be observed only in a restricted area located close to the NW side of the Komagelv–Trollfjord Fault Zone (Siedlecki, in prep.).

In general, the Riphean strata of Varanger Peninsula and adjacent areas of East Finnmark appear to have formed an extensive wedge thickening towards the north-east; in the south-west there is an up to 2500 m thick continuous sedimentary sequence representative of a platform cover, while in the north-east the equivalent succession reaches more than 9000 m in thickness, or even more if one accepts the correlation between the Laksefjord, Raggo and Barents Sea Groups, and represents a miogeosynclinal sequence.

Folds and faults showing an approximate SW–NE trend and broadly parallel to the Caledonian front of northern Scandinavia are prominent in the western and northern parts of Varanger Peninsula; to the south-east, in the Vardø area, these folds gradually swing into a roughly N–S trend and later, gentle, ca. NW–SE cross-folds are present (Roberts 1972). NW-trending folds probably belonging to a separate deformation episode from that of the main folding also occur in the vicinity of the Komagelv–Trollfjord Fault Zone (Siedlecka & Siedlecki 1971), and similarly oriented structures are present in the south-eastern part of the peninsula although the tectonics of this area have not been studied in detail (Siedlecki, pers. comm.).

The Komagelv–Trollfjord Fault Zone is not understood in detail: this is a complex major fracture in which reverse faults resulting from SW-directed movement constitute a dominant deformation feature. Evidence of important strike-slip movements has also been presented by Roberts (1972). Displacement along the fault-zone is greatest in the south-east (some 4000 m), while in the north-west sets of minor reverse faults of tens to hundreds of metres displacement are present along with some subordinate strike-slip faults

(Siedlecka & Siedlecki 1967; Siedlecki, in prep.). Younger normal dip-slip movements have also occurred along this fault zone.

Doleritic dykes constitute the only trace of igneous activity on Varanger Peninsula. A SW–NE trending swarm of dykes is present in the north-eastern region of the peninsula; in its south-western part there occur only a few N–S oriented post-folding dykes. Roberts (1972) has suggested an early-Caledonian age for one episode of syn-folding dykes in the Barents Sea Region. Preliminary K/Ar age determinations for the dykes from both sides of the Komagelv–Trollfjord Fault Zone, however, fall into three groups: (A) 340 m.y., (B) ca. 542 m.y. to 640 m.y. and (C) $>$ 1000 m.y. (Beckinsale et al., in press). Further investigations are needed for a precise understanding of these figures which, according to Beckinsale et al., suggest the occurrence of a pre-Caledonian deformation and an Upper Devonian intrusive phase and, in general, are indicative of different geological histories for the south-western and north-eastern regions of the Varanger Peninsula.

Riphean and Vendian strata of the south-western Varanger Peninsula (the platform cover type) continue towards the west, and their lithostratigraphic correlatives may be traced into West Finnmark; the thick miogeosynclinal succession underlying the north-eastern part of the peninsula cannot be recognized with any certainty west of Laksefjord.

The Rybachiy–Sredniy Peninsula has been investigated, and its geology discussed, by some Finnish and numerous Russian geologists (e.g. Fieandt 1912; Lupander 1934; Polkanov 1934, 1936; Tenner 1936; Agapiev & Vronko, *in* Kharitonov 1958; Negrutsa, *in* Shamoida 1968; Keller & Sokolov 1960; Keller et al. 1963; Polkanov & Gerling 1968; Bekker et al. 1970). As on Varanger Peninsula, the Rybachiy–Sredniy is divided by a NW–SE trending fault zone into two separate geological regions. The south-western (Sredniy) region is underlain by Riphean and Vendian strata resting with a sedimentary contact on an erosional surface of the crystalline basement. A suggestion that this contact is tectonic (Fieandt 1912; Holtedahl 1918; Polkanov 1934, 1936; Atlasov 1964, and others) has not been confirmed. It may be noted that observed relationships at the head of Varangerfjord are indicative of a sedimentary contact, and new aeromagnetic data reject the idea that a tectonic discontinuity or fault may be present along the Varangerfjord (K. Åm, pers. comm.). On Sredniy, observations by Lupander (1934), Agapiev & Vronko (*in* Kharitonov 1958, and *in* Keller et al. 1963) and Keller & Sokolov (1960) revealed a sedimentary contact between the basement gneisses and the basal conglomerate of the Riphean Kildin Group.

The Kildin Group, 1500 m thick, is composed predominantly of terrigenous deposits and contains a carbonate unit at the top. On the basis of available descriptions, strata of this group may be thought to have originated in the shallow-water environment of a sea encroaching upon the crystalline rocks of the Fennoscandian Shield. K/Ar radiometric age determinations from glauconite present in the lower Kildin Group have given figures ranging from around 1000 m.y. to 619 m.y. (Polkanov & Gerling 1960; Bekker et al. 1970a). These Riphean strata are overlain by the Volokov Group, up to ca. 500 m thick, which is composed of arenaceous terrigenous rocks with a basal conglomerate. There is a transgressive contact between the two groups of strata, yet no angular unconformity has been reported. A K/Ar whole rock age for a dolerite dyke cutting the Volokov Group sediments has given a figure of ca. 600 m.y. (Bekker et al. 1970a). Lithologies of the Kildin and Volokov Groups also appear on the *Aynov Islands* where terrigenous and carbonate rocks of the upper Kildin Group (45 m) are overlain with a slight angular unconformity by 240 m of shallow-marine (? and fluvial) terrigenous deposits of the (Vendian) Volokov Group (Bekker et al. 1970).

The north-eastern part of the Rybachiy–Sredniy Peninsula, i.e. the Rybachiy Peninsula proper, is underlain by a several thousand metres thick, presumably continuous pile of terrigenous deposits, *entirely different* from those of Sredniy and the Aynov Islands. The sequence of strata, the Rybachiy Group, starts with a coarse polymict diamictite (Agapiev & Vronko, *in* Kharitonov 1958), the Motov Formation of Keller & Solokov (1960), and Keller et al. (1963). This formation underlies the isthmus between the Sredniy and Rybachiy parts of the peninsula and is strongly influenced by the tectonic disturbance present along this depression. The origin and the stratigraphic position of the diamictite have been very controversial: the idea of a glacial origin has been forwarded by Wegmann (1928) while other geologists (e.g. Keller & Sokolov 1960) have been inclined to interpret it as a grain-flow deposit. According to Agapiev & Vronko (*in* Kharitonov 1958) the Motov Formation rests with a sedimentary contact *on* the Volokov Group. Consequently, the whole Rybachiy Group was regarded as younger than the Sredniy Group, occurring in the ca. 350 m downfaulted Rybachiy block. This suggested situation was shown on a cross-section (*in* Kharitonov 1958), but it does not appear on the geological map produced by Agapiev & Vronko; since no further observations supporting this idea have been reported it is here considered only as an alternative and hypothetical interpretation. Correlation with the Volokov Group of Sredniy has also been suggested (Keller & Sokolov 1960; Keller et al. 1963), but this is not supported by any firm data. The Motov conglomerate, ca. 250–470 m thick, is overlain by 3600 m of coarse arenaceous and conglomeratic sediments with diamictite lenses followed by some 900 m of greywacke and shale (a flysch sequence according to Sergeeva 1964). Arenaceous and clayey shale, strongly cleaved and ca. 500 m thick, constitutes the uppermost unit. Lupander (1934) recorded one basic dyke dissecting this formation.

The nature of the tectonic disturbance separating Rybachiy from Sredniy is not quite clear: most authors, however, have interpreted it (or accepted this interpretation) as a reverse fault zone which resulted from a south-west directed stress, the Rybachiy having been lifted up and pushed against the Sredniy (Wegmann 1928, 1929; Lupander 1934; Polkanov 1934 ,1936; Keller & Sokolov 1960; Keller et al. 1963, pp. 103–113).

Strata underlying the Sredniy–Rybachiy Peninsula have been only slightly folded; they dip gently northeastwards over most of the area. Open cylindrical folds are present, especially on Rybachiy, and many of them are asymmetrical, their axial planes dipping steeply towards the north-east. A somewhat more advanced deformation with a pervasive cleavage has been reported from the northeastern part of Rybachiy.

Kildin Island is underlain by a 1000 m thick sequence of sediments, the Kildin Group. This is composed of terrigenous deposits with subordinate carbonate lithologies appearing in its lower part. The carbonate rocks contain columnar algal stromatolites (*Gymnosolen ramsayi* Steinm., *Collenia buriatica* Maslov) and oncolites (Artemiev 1933; Bogdanov 1933; Gurbich 1934; Polkanov 1936; Kharitonov 1958; Krylov 1959: Keller et al. 1963; Bekker et al. 1970a). An erosional surface is thought to be present in the lower–middle part of the Kildin Group (Bekker et al. 1970). K/Ar radiometric ages obtained from glauconite collected from different horizons in this group range from 849 to 759 m.y. (Bekker et al. 1970a). The lithologies occurring on Kildin Island seem to be of shallow-marine origin although the presence of some fluvial deposits cannot be excluded. The beds dip gently northeastwards, and neither the bottom nor the top surface of the Kildin Group is exposed.

CORRELATION

The Tanafjord–Varangerfjord region of Varanger Peninsula, the Aynov Islands, the Sredniy region and Kildin Island are all situated close to the margin of the Fennoscandian Shield, and are underlain by some 2000–4000 m of fluvial and shallow-marine sediments resting with an erosional contact on the crystalline basement; glacial deposits of the 'Varangian Ice Age' are widespread west of the Aynov Islands. These strata are not metamorphosed and are only gently folded over most of the region with the exception of the north-westernmost area close to the Caledonian front. Along with the fluvial and shallow-marine origin of the sediments, this indicates that they were accumulated in a *pericratonic zone* of the Late Precambrian sedimentary basin on a relatively rigid substratum and are representative of platform-cover type deposits. Palaeocurrent directions (data from the south-western part of Varanger Peninsula only), thicknesses and facies distribution indicate that the basin was located to the north-east or east.

The few radiometric age determinations so far available indicate an Upper Riphean and Vendian age for the bulk of the strata of the pericratonic zone. The Riphean/Vendian boundary is usually placed at the interface between the Kildin and Volokov Groups and between the Tanafjord and Vestertana Groups, although this is not quite consistent with the radiometric age boundaries introduced by the Russian workers (Garris et al. 1964). This is, however, an important geological boundary marked by erosion (Sredniy) and an angular unconformity (Aynov Is. and East Finnmark), and by the appearance of Varangian tillites in Norway. The tillites are absent in the Vendian Volokov Group on the Aynov Is. and on Sredniy; these areas probably constituted topographic highs subjected to erosion during the Varangian glaciation.

The Kildin Group on Kildin Island contains carbonate rocks with columnar stromatolites in its lower part, not at the top of the sequence as in the case on Sredniy and in the Tanafjord Group on Varanger, and this seems to have been the result of a facies change. An age correlation between these 'upper' and 'lower' carbonates does not seem to be possible especially as the K/Ar age (on glauconite) of the carbonate-bearing part of the sequence of Kildin Is. is 849 m.y. and the ages of the terrigenous sediments overlying these carbonate rocks exceed 750 m.y. (Bekker et al. 1970). These former strata thus cannot be correlated with the Vestertana and Volokov groups. It is possible that a carbonate unit may have been present at the top of the Kildin Group on Kildin and has since been eroded. Alternatively, such a unit may still be present beneath the surrounding sea-covered areas.

The north-eastern (Barents Sea) region of Varanger Peninsula and the Rybachiy region have a similar geological setting being located outside the marginal zone reviewed above, and are underlain by several thousand metres thick terrigenous sequences (Siedlecka & Siedlecki 1967). These rock sequences represent a *miogeosynclinal zone* of the Late Precambrian sedimentation basin. The presence of flysch-like sequence in both regions is very helpful for comparison purposes, but nevertheless two alternative lithostratigraphic correlations may be suggested depending on the genetic interpretation of the Motov diamictite and conglomerate lenses of the lower Rybachiy Group:

1) If the Motov diamictite and the slightly younger conglomerates (in the Ejna Formation) are glacial (and glacifluvial) in origin, they could be correlated with the lower Laksefjord Group of East Finnmark (the Ifjord Formation). A thick sequence of fluvial and shallow-marine sediments subjacent to the flysch-like deposits of East Finnmark would then be absent or considerably reduced on Rybachiy. This correlation would suggest the occurrence of a widespread glaciation older than the Varangian Ice Age.

2) If, on the other hand, the Motov diamictite and the conglomeratic lenses are representative of grain flow deposits, together with the rest of the Rybachiy Group they should be comparable to the flysch-like strata of the north-eastern region of Varanger Peninsula. This correlation would imply that subaqueous mass movement processes were much more active in the Rybachiy region than further to the north-west. Except for the Løkevik Tilloid of suggested composite glacial and mudflow origin (Siedlecka & Roberts 1972), coarse polymict conglomerates are unknown in the flysch-like sequence on Varanger Peninsula.

The above correlation alternatives, based on structural setting and on comparisons of the sedimentary succession of Rybachiy with those of East Finnmark are different from the previously suggested stratigraphic relationship between the groups of rocks of Srednyi and Rybachiy (see p. 319). The interpretations of the Rybachiy–Sredniy stratigraphic relationship suggested by Agapiev & Vronko (*in* Kharitonov 1958) and by Keller & Sokolov (1960) imply Vendian and younger ages for the Rybachiy Group. This has not been confirmed by radiometric age determinations (Polkanov & Gerling 1960,

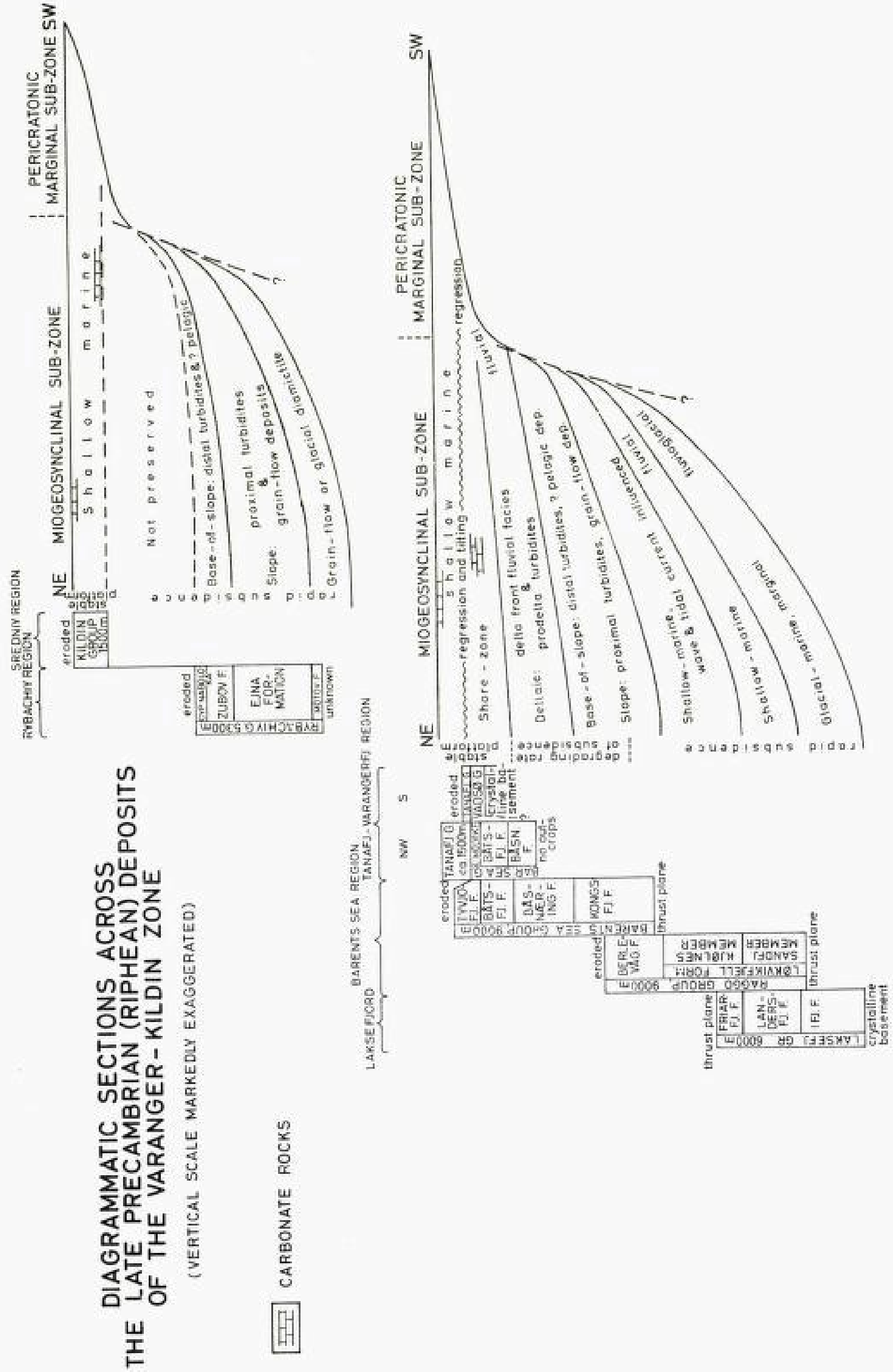


Fig. 3. (Explanation written on the illustration).

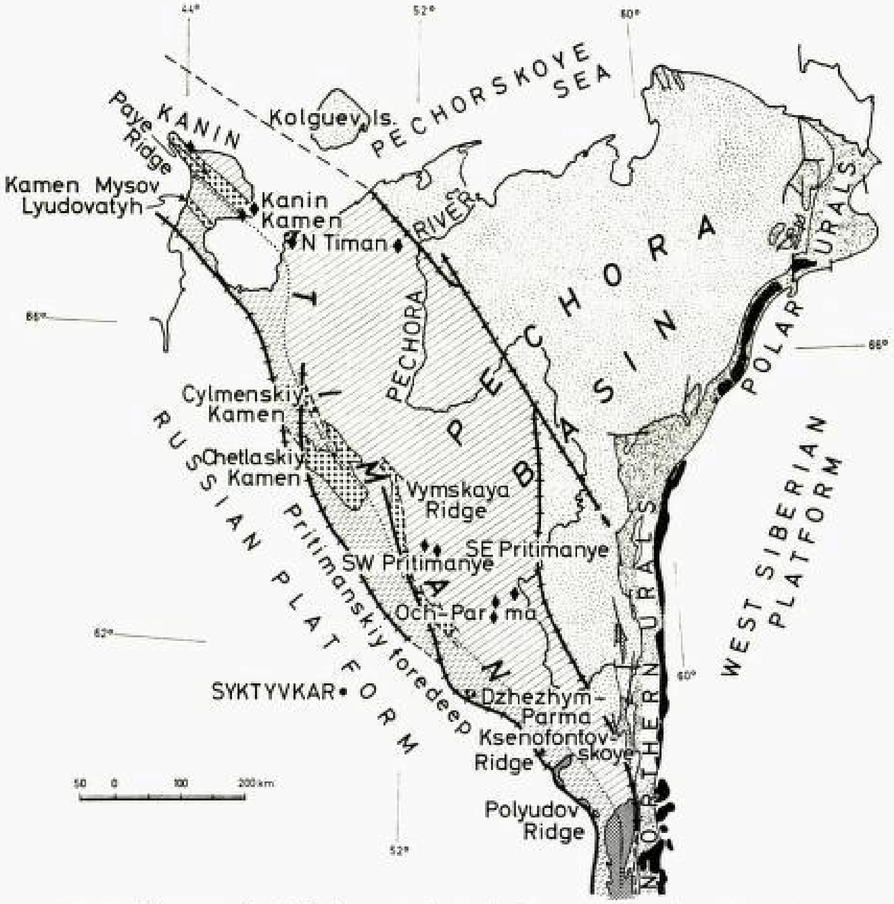
1961) which, although uncertain and ambiguous (Bekker et al. 1970a) give figures ranging from 900 to 686 m.y. (K/Ar whole rock ages).

The correlation alternatives (1) and (2) have been indicated on Fig. 3, which also shows diagramatically the supposed relationship between the marginal and internal zones of Riphean sedimentation within the Varanger–Kildin Zone. This correlation is a matter of broad geological interpretation rather than a simple facies comparison. The sub-zones are now in tectonic contact along the Trollfjord–Komagelv–Rybachiy Fault Zone and the transitional strata which are assumed to have existed are not exposed. This structural situation is the main reason for uncertainty over the marginal-to-basinal facies relationships. The sedimentary contact and slight angular unconformity between the Barents Sea and Tanafjord Groups observed in north-western Varanger Peninsula (Siedlecka & Siedlecki 1972; Siedlecki, in prep.) provide the most important data on this topic, in particular if one remembers that this same Tanafjord Group in the marginal zone is underlain only by the ca. 800 m thick Vadsø Group. This suggests that during most of the time of deposition of the basinal facies, the marginal sub-zone constituted an erosional and/or non-depositional area across which sediment supply to the basinal zone has occurred. Analysis of the overall facies succession in the miogeosynclinal zone indicates a particularly rapid basin subsidence in the first stage up to the development of a prominent slope. The true nature of this slope is unknown, and the fault indicated in Fig. 3 is a purely hypothetical feature. However, suggestions of the existence of a structural, fault-bound margin are to be found in the Timans, where deep-seated fractures delineate the margins of the basin and/or different zones of Riphean sedimentation (see e.g. Zhuravlev 1972 and pp. 333–335 in this paper).

The rate of subsidence of the depositional basin gradually decreased in the *Varanger–Kildin Zone*. The basin was filled in and a paralic sedimentation was established throughout the area of both sub-zones, which, together, then represented a broad depositional shelf. This was followed by some intra-Riphean movements which resulted in a regional regression and erosion and also in a tilting of the western part of the basinal deposits; data on the existence of this tilting further to the east are lacking. A new transgression then re-established the paralic conditions in which the Riphean sedimentation had been completed. This was subsequently followed by regression and tilting and by glaciation initiating the Vendian period.

The Timan–Kanin zone

In this zone Riphean rocks crop out in anticlines protruding through the younger platform-type cover and in elevated, fault-bound blocks arranged in a chain running from the Paye Ridge and Mysy Lyudovatye on the Kanin Peninsula in the north-west to the Ksenofontovskoye and Polyudov Ridges in the south-east (Fig. 4). A sharp angular unconformity occurs between the folded and metamorphosed Riphean rocks and the cover, the latter starting with



Miogeosynclinal Riphean rocks in the Timan Mts. and on Kanin Peninsula

Miogeosynclinal Riphean rocks in the Polyudov Ridge and in the Urals

Boundary between the western (marginal) and the eastern (internal) zone of the Riphean miogeosyncline

Subsurface occurrence of Riphean rocks of the marginal zone of the miogeosyncline

Subsurface occurrence of Riphean rocks of the internal zone of the miogeosyncline

Riphean rocks in the Northern and Polar Urals (their precise nature is not finally established)

Subsurface occurrence of metamorphic rocks interpreted either as Riphean eugeosynclinal rocks or as an older craton

Riphean (Baikalian) intrusions in the Timans, on Kanin and subsurface in the Pechora Basin

Gabbro and peridotite (Hercynian) - main Uralian suture

Deep-seated tectonic discontinuity

Main surface faults in the Timans

Fig. 4. Sketch-map of the Riphean of the Timans, the Pechora Basin and the Urals (mainly after Zhuravlev & Osadchuk 1960; Zhuravlev 1972).

Ordovician, Silurian or Devonian (Shamoida 1968, p. 153; Zhuravlev 1972; Getsen & Naumov 1973). The Riphean strata are cut by numerous pre-Palaeozoic bodies of igneous rocks, more especially in the northern areas.

The Riphean of the Timan Mountains and Kanin Peninsula is in general represented by metasedimentary, mostly terrigenous rocks, with subordinate limestone and dolomite. The total thickness is about 13,500 m (Zhuravlev et al. 1966, p. 54); neither the bottom nor the top surface of this Riphean sequence is known. Numerous sections have been measured and lithostratigraphic units erected in the various parts of this Timan–Kanin zone (see Zhuravlev 1972, p. 21 for a review of older publications), and different correlation schemes between these measured sections have been proposed (e.g. Zhuravlev & Osadchuk 1960, 1962, 1963; Raznitsyn 1962, 1962a, 1965, 1969; Kochetkov 1963; Zhuravlev et al. 1966; Volochayev et al. 1967). These have subsequently been correlated with sections through the miogeosynclinal Riphean of the Urals.

Vendian strata seem to be absent from most of the Timans; there is still some uncertainty here, however, and some pre-Palaeozoic strata of the Central Timans and Polyudov Ridge have been considered Vendian in age (Borovko et al. 1964, *vide* Zhuravlev 1972; Getsen 1972). The presence of Vendian tillites has been reported from the Polyudov Ridge by Borovko (1967, *vide* Chumakov 1970). Getsen & Naumov (1973) have suggested a Vendian age for some slates present on Kanin Peninsula; there is, however, no firm evidence to support this suggestion.

Central Timans. The most complete section through the Riphean sequence is exposed in the Chetlaskiy Kamen Anticline (Fig. 4), where the lithostratigraphy of the Riphean strata of the Timans was first established. This stratigraphy was then extended to other sections in the Central and Southern Timans and correlated with that in the Northern Timans and Kanin Peninsula. Five formations have been differentiated (Zhuravlev & Gafarov 1959; Zhuravlev & Osadchuk 1960, 1962: Zhuravlev et al. 1966): The lowermost, the 1850 m thick *Svetlinskaya Formation*, consists of grey quartzite with subordinate interbeds of siltstone and quartz–sericite slate. This is overlain with a slight angular unconformity by conglomerates of the *Chetlaskaya Formation* (up to 2700 m), which also contains grey quartzitic sandstone, slate and some calcareous sandstone. Succeeding this with an erosional boundary (Zhuravlev & Osadchuk 1960, p. 93) is the *Dzezhymskaya Formation* which consists of alternating quartzite and siltstone with lenses of feldspathic sandstone and conglomerate; this formation is characterized by its variable thickness (250–> 850 m) and by facies changes which result in the sequence being more conglomeratic in sections situated closer to the edge of the Fennoscandian Shield. The next unit, the *Bystrynskaya* (*Bystruhynskaya*) *Formation* (up to 4200 m) is composed of carbonate rocks in its lower part; bedded and stromatolitic dolomites are present, these being overlain by multicoloured slates containing some lenses of carbonates. The fifth lithostratigrapic unit, the

Kislorucheskaya (Och–Parminskaya) Formation (ca. 4000 m) consists essentially of phyllite with subordinate quartzite layers and is considerably more metamorphosed (biotite–chlorite subfacies of the greenschist facies according to Zhuravlev 1972, p. 27) than all the other formations. The stratigraphic position and extent of the Kislorucheskaya Formation remains controversial. Originally its lithologies were thought equivalent to the Chetlaskaya Formation (Kalberg 1948) and subsequently considered by Zhuravlev & Osadchuk (1960) as a facies equivalent of the rest of the Riphean strata, deposited in a deeper basinal zone. Later these same authors (Zhuravlev & Osadchuk 1962) regarded it as the youngest unit and the facies zonation model was replaced by one of tectonic zonation. Raznitsyn (1965) suggested that the Kislorucheskaya Formation is older than all other Riphean formations of the Central Timans. Neither of the two last-mentioned suggestions seems to have been accepted, and the more recent models in general support the original idea (e.g. Volochayev et al. 1967; Raznitsyn 1969). Zhuravlev (1972), however, has retained his tectonic zonation model.

Northern Timans and Kanin Peninsula. Riphean rocks of the Northern Timans, the Barmin Formation of Barkhatova (1936), are represented by a several kilometres thick series of dark-coloured, greenish-grey phyllite and schist. These lithologies have more recently been studied in some detail by Getsen (1968), who introduced a new lithostratigraphy (Malochernoretskaya and Yambozherskaya Formations) and reported the presence of widespread graded bedding in these metasedimentary rocks. An analogous metamorphic series of Riphean quartzite and schist underlies the Paye Ridge on Kanin Peninsula, while at Mysy Lyudovatye there are metamorphosed stromatolitic dolomites and some slaty argillaceous lithologies (Tschernyschev 1901; Ramsay 1911; Lyutkevich 1953; Malkov & Puchov 1963; Malkov 1966; Getsen 1968, 1971, and others). Getsen (1971) measured a coastal section of the Riphean rocks at Kanin Kamen (Fig. 4) and differentiated two groups of rocks. The lower, Tarkhanov Group (4400–5000 m), consists of schists and quartzites (the epidote–amphibolite metamorphic facies of Russian workers), which in spite of metamorphic alteration have retained a variety of sedimentary structures and have been defined by Getsen (1971) as 'flyschoid'. This unit is overlain by the somewhat less metamorphosed Tabuyev Group (3800 m) composed of feldspathic and oligomict sandstone in the lower part and arenaceous and argillaceous graded-bedded lithologies higher up. The latter contains subordinate limestone, the amount of which increases upwards. Getsen (1971) suggested a correlation between this upper carbonate-bearing portion of the Tabuyev Group and the lower part of the Mysy Lyudovate Dolomite (1450 m). The total known thickness of the Riphean strata on Kanin has been estimated by Getsen (1971) at 8500–9000 m.

The general stratigraphy of the Riphean rocks of the Timans is essentially based on (1) lithostratigraphic correlations between the separate outcrops within the Timan–Kanin zone, (2) lithostratigraphic correlations with classical profiles of the miogeosynclinal Riphean of the Urals and (3) algal stromatolites, oncolites and catagraphs. Results obtained by the adoption of these methods have been checked by radiometric age determinations. Thus, the Mysy Lyudovate Dolomite and the Bystrynskaya Formation have been considered as chronostratigraphic equivalents and a study of stromatolites has indicated an Upper Riphean age for the carbonate formations (Raaben & Zhuravlev 1962; Raaben 1964; Zhuravlev et al. 1966; Raaben 1968). However, a K/Ar age determination of a diabase sill cutting the shaly portion of the Bystrynskaya Formation gave a figure of 1220–1230 m.y. (Malkov 1969)! On the other hand, the Barmin Formation of the Northern Timans has been tentatively compared with the Kislorucheskaya Formation (Zhuravlev & Osadchuk, *in* Keller 1963, p. 231), this being quite different from Getsen's (1971) correlation between the Northern Timan and Kanin sections (see p. 326). In spite of these and other existing controversies and uncertainties, it seems almost indisputable that Middle and Upper Riphean strata are represented in the Timan–Kanin zone.

TECTONIC DEFORMATION OF THE TIMAN–KANIN RIPHEAN

Many geologists have written about the structure of the Timans and Kanin, and the following brief review of the pre-Palaeozoic deformation of this region is based mainly on summaries given by Raznitsyn (1968, 1969), Getsen (1972) and Zhuravlev (1972).

The Riphean strata of the Timan–Kanin Zone are exposed in separate fault-bounded blocks and are folded, the fold axes exhibiting regional north-westerly trends (Fig. 4). In the Central Timans, the structure of the Chetlaskiy Kamen horst is dominated by an open asymmetrical anticline and the degree of metamorphic alteration of the Riphean strata is low. The anticline plunges north-west, and its axial plane dips to the south-west. The steeper NE limb of the fold is dissected by a set of NW-trending pre-Palaeozoic faults and the Riphean strata adjacent to this fault-zone in the north-east (the Kislorucheskaya Formation) are characterized by a more intense tectonic deformation and metamorphism than the bulk of the rocks of the Chetlaskiy Kamen. Numerous dykes of basic alkaline rocks are associated with this fault-zone. South-east of the Chetlaskiy Kamen (SW Pritimanye) and in Dzhezhym–Parma, weakly metamorphosed Riphean strata dipping gently southwestwards have been reported from subsurface investigations. The Och–Parma and Vymskaya ridges are dominated by isoclinal folds with NE-dipping axial planes and the Riphean sedimentary rocks are altered into slates. In the north, deformation of Riphean strata at Mysy Lyudovate may be roughly compared to that of the Chetlaskiy Kamen while rocks underlying the Paye Ridge and cropping out in the Northern Timans are altered into slates and schists, folded in steep tight or isoclinal folds, characterized by a variety of micro-tectonic structures and dissected by numerous igneous intrusions. However, the degree of metamorphism does not appear to have exceeded upper greenschist facies.

In general, two contrasting degrees of tectonic deformation and metamorphism may be differentiated in the Timan–Kanin Zone: 1) slight deformation and low metamorphism, as exemplified by the Chetlaskiy Kamen structure; 2) stronger deformation and metamorphism, this being most conspicuous in the north. Exposed areas of the (1) weakly and (2) strongly deformed Riphean strata seem to be arranged in two parallel NW–SE belts (Fig. 4) related to two different sedimentary and tectonogenic regimes.

Further data on the subsurface structure of the Timan–Kanin Zone and neighbouring areas have been provided by drillings and by geophysical work. Several drill-holes located east of the central and southern Timans have penetrated Riphean slates and schists (and associated intrusions), while the distribution of magnetic and gravimetric anomalies has indicated the existence of deep-seated NW–SE trending fracture zones bounding the belt of Riphean occurrences in the south-west and north-east (Fig. 4). Ideas on the origin and structural history of this belt have been and still are developing following two basic alternative models: 1) a model of a Timan–Ural geosyncline with subdivision into Timanian miogeosynclinal and Uralian eugesynclinal parts; and 2) a model of two independent sedimentary basins separated by a craton. The development of these controversial concepts will be reviewed later in this paper together with ideas on the broad regional framework of the north-eastern border of the Fennoscandian Shield and its eastern subsurface continuation.

PRE-SILURIAN IGNEOUS ACTIVITY AND METAMORPHISM IN THE TIMAN–KANIN ZONE

The intrusion of gabbroids, particularly common in the Northern Timans and on Kanin, represents the oldest period of regional igneous activity. Radiometric ages (K/Ar, mostly whole rock) of the metamorphism of the gabbroic rocks are around 640–620 m.y. (Malkov & Puchkov 1963; Malkov 1966). On Kanin the gabbroids are dissected by syenites and granites which give ages of 550–480 m.y. (Polkanov & Gerling 1961; Ivensen 1964). According to Malkov & Puchkov (1963) and Malkov (1966) these younger ages must be due to greisenisation since the age of a pegmatite dyke dissecting the granite is 640 m.y. The granite is considered comagmatic with granitic bodies detected in the subsurface in the Southern Timans (Malkov 1966). A younger group of igneous rocks comprises essexite gabbro of the Northern Timans and later granites and syenites dissect this particular gabbro. A great variety of basic dykes (e.g. lamprophyres, camptonites, picrites, porphyrites) is genetically related to the essexite gabbro. This suite of basic rocks is metamorphosed, and dykes boudined and amphibolitised: the age of their metamorphism is around 525–520 m.y. (Malkov 1966). The granites which dissect these gabbroic rocks gave intrusive ages ranging from 540 m.y. to 534 m.y. This inconsistency with the geological evidence may perhaps be due to different material being used for the determination of radiometric ages (K/Ar whole rock analyses for the basic rocks; biotite, muscovite or whole rock samples for the acid varieties).

Numerous syenite–aplites and granite–aplites transect the granites and syenites of this younger group: these give quite variable ages of 550 m.y., 520 m.y. and 445 m.y. (Malkov 1966). Finally, a swarm of SW–NE trending diabase dykes cross-cuts the older igneous bodies in the Northern Timans; only two ages have been reported, 516 m.y. and 505 m.y. (Malkov 1972), but no analytical data are given.

In a summary and interpretation of radiometric age data for the Riphean rocks of the Timans given by Raznitsyn (1965), two older periods of metamorphism have been indicated for the Southern Timans in addition to the two above-mentioned metamorphisms: an older one at 1130 m.y. (uncertain) and a second characterized by figures of 790 m.y., 768 m.y., 700 m.y. and 687 m.y. Other age determinations have been obtained which do not readily fit in with the age patterns outlined above: an intrusive age of 1300 m.y. has been indicated for granite in the Northern Timans (Malkov & Puchkov 1963); while a diabase sill cutting the Bystrynskaya Formation in the Central Timans provided a rather anomalous age of 1220–1230 m.y. (Malkov 1969).

Getsen (1970, 1971, 1971a), summarizing data on the radiometric ages of the Timans, has differentiated: a) intrusions with an age range of 665–445 m.y. and b) an age of metamorphism(s) of 620–483 m.y. Difficulty arises, however, in understanding the precise significance of the ranges of age given in Getsen's papers.

The southern Barents Shelf zone

This embraces the offshore area located between the Varanger Peninsula in the north-west and the Kanin Peninsula in the south-east (Fig. 1). Of interest here is the possible occurrence of Late Precambrian rocks and structures forming a link between the Varanger–Kildin and Timan–Kanin zones.

Published information on the structural framework of the Barents Shelf is based on sparse geophysical studies combined with data on the stratigraphy and structure of the bordering land areas and of islands located along the western and northern edge of the shelf. The geophysical data show that the continental crust of the Barents Shelf is ca. 30–40 km thick and is composed of a basement with a sedimentary cover (e.g. Beliayevsky et al. 1968; Demenitskaya et al. 1968; Litvinienko 1968; Demenitskaya & Hunkins 1970; Eldholm & Ewing 1971; Emelyanov et al. 1971). According to the geological information the basement/sedimentary cover interface reflects the boundary between 1) the Caledonian and older cratonised substratum exposed in Devonian time, and 2) the Devonian and younger strata. The thickness of the cover has been roughly estimated as some 4–5 km (e.g. Harland 1967; Eldholm & Ewing 1971). An uneven subsidence of the basement, inferred on the basis of geological data (Frebold 1951; Harland 1960, 1967) has been confirmed by differences in cover thicknesses under the sea areas (e.g. Eldholm & Ewing 1971), and a subdivision of the cover has been suggested on the basis of seismic velocity changes combined with geological information (e.g. Eldholm

& Ewing 1971; Emelyanov et al. 1971; Vogt & Ostenso 1973; Sundvor 1974). This same sedimentary cover continues into the Pechora Basin, where numerous gas and oil reservoirs have been discovered, in particular in the middle–upper Palaeozoic strata. Analogous horizons and structures may be expected to occur in the northwestern submarine continuation of this cover (e.g. Krems et al. 1968; Semenovich et al. 1973).

The nature and structural framework of the basement is much more uncertain; seismic records (DSS) reaching the deep portions of the crust are scarce and interpretations of these much more difficult. Demenitskaya et al. (1968, their Fig. 2.1) have suggested that the crust is composed of a 'basaltic' layer overlain by a 18 km thick pile of sedimentary rocks which in its lower part is representative of the Riphean complex. The supposed Riphean portion of the strata is characterized by seismic velocities of around 5.5 km/sec., while a velocity of 4.7 km/sec. marks the bottom surface of the younger (i.e. post-Riphean) sedimentary cover. A 'granitic' layer indicated in this interpretation south of the Barents Shelf wedges out rapidly north of the southern shorelines of the Barents Sea. Belayevsky et al. (1968) have in the generalized trans-continental section of crust included the Riphean complex into the 'granitic' layer. The model of Demenitskaya et al. (1968) assumes that the Riphean complex underlies the Barents Shelf from Kola Peninsula in the south to Franz Joseph Land in the far north. This is probably based on one of two alternative geological models of the Timan–Ural geosyncline(s), discussed later in this paper (p. 333).

If one now considers the southern Barents Shelf, keeping in mind the 'Riphean' 5.5 km/sec. velocity, the results of the seismic work of Litvinienko (1968), Eldholm & Ewing (1971) and Emelyanov et al. (1971) appear to be of particular interest. The SW–NE trending seismic crustal profile of Litvinienko (1968) crosses the Kola Peninsula and southern Barents Shelf between the Varanger and Rybachiy–Sredniy peninsulas (Fig. 1). In this profile the crystalline rocks of the Fennoscandian Shield are characterized by velocities of 6.0 km/sec. A distinct 5.5 km/sec. refractor appears beneath the sea-covered areas, this being followed by horizons characterized by 4.7, 3.2 and 2.2 km/sec. The 5.5 km/sec. horizon comes close to the sea bottom near the Rybachiy–Sredniy Peninsula indicating that this particular velocity is indeed characteristic of the buried Riphean complex. In addition a 6.0–6.1 km/sec. horizon appears some 150 km NE of the shore-line suggesting a shallow occurrence of crystalline basement beneath the post-Riphean cover, characterized in this profile by 4.7 km/sec. and lower velocities. Figures obtained by Eldholm & Ewing (1971) from their refraction work in the southern Barents Shelf are in agreement with those published by Litvinienko (1968). Velocities in the range 5.0–5.5 km/sec. recorded in the western portion of the surveyed area may, according to Eldholm & Ewing (1971), be related to offshore continuation of the Caledonian orogen, while those from the eastern part of the area are probably associated with the Riphean structural complex. A distinct change in the magnetic and gravimetric anomalies appears in the area of the possible junction between the Caledonian and Riphean structural complexes (Eldholm & Ewing 1971).

The paper by Emelyanov et al. (1971) constitutes the most recent Russian summary of the geology of the Barents Sea. Of new data published in this article a seismic profile located north-west of the Kanin Peninsula is of particular interest for an understanding of the deep structure of the Southern Barents Shelf Zone. An approximately 500 m high structural elevation located along the NW extension of the Riphean Paye Ridge on Kanin Peninsula is indicated in this profile, and there also occurs a distinct structural depression south-west of the elevation. Along this same profile the post-Riphean sedimentary cover of Devonian to Quaternary strata reaches the considerable thickness of some 3.5 km. This profile subdivision has presumably been based on the stratigraphy of the platform cover of the Pechora Basin and the Timan–Kanin Zone.

Summarizing the above information, the following points should be emphasized with regard to the structure of the Southern Barents Shelf zone:

1) There is no direct geological evidence available on the structure and lithostratigraphy of the Barents Shelf.
2) Geophysical and indirect geological information is indicative of the existence of a 4–5 km thick platform-type Palaeozoic–Cenozoic sedimentary cover starting with Devonian or Carboniferous strata. Considerations of the stratigraphy and structure of this cover are outside the scope of this paper.
3) The structure of the basement is uncertain. A seismic velocity of around 5.0–5.5 km/sec., intermediate between velocities characteristic for the Fennoscandian Shield and those for the cover, can be related to different structural complexes; these are indicative either of Caledonian or of Riphean (Baikalian) basement.
4) The thickness of the Riphean complex underlying the Palaeozoic and younger cover and resting on the 'basaltic' layer may be estimated at ca. 12–13 km, assuming the thickness of the sedimentary cover as being around 4–5 km. The recently reported 7 km thickness of the Riphean complex in the area of Varanger Peninsula (K. Åm, this volume) presumably refers to the depth down to a major surface of tectonic discontinuity beneath the Barents Sea Region.

Outline of inter-zonal correlation

The occurrence of an extensive NW–SE trending belt of Late Precambrian rocks bordering the NE margin of the Fennoscandian Craton seems to be now well established, and the available data are also indicative of a differentiation of sedimentary environments and structural and metamorphic events within this belt. There is unfortunately insufficient evidence of a detailed inter-zonal, i.e. longitudinal, stratigraphic correlation, and difficulty is also encountered in determining the stratigraphic relations across the belt both in the Varanger–

Kildin and in the Timan–Kanin zones. However, if the above-mentioned internal differentiation of the belt of Late Precambrian rocks is taken into account, in particular with respect to thicknesses of the sedimentary sequences and facies development, structure and metamorphism, an inter-zonal 'en gros' correlation may be successfully reached.

In the Varanger–Kildin zone the Riphean deposits are related to two sedimentary regimes; one with a relatively stable rigid substratum and the other with a subsiding mobile substratum. The thick sequence of strata associated with the mobile subzone did not suffer any pronounced pre-Caledonian structural deformation or metamorphic alteration, and in its overall appearance it is similar to the Riphean sequences cropping out in the Chetlaskiy Kamen in the Central Timans and to equivalent sequences of Dzhezym Parma, the Ksenofontovskoye Elevation and Kamen Mysov Lyudovatych on Kanin (Fig. 4). Nowhere in the Varanger–Kilin zone are there rocks and structures which could be compared to the strongly disturbed rocks of the Paye Ridge or Northern Timans, nor are there known intrusions other than diabase dykes. Equivalents of this strongly disturbed Riphean basement, if they exist, would have to occur beneath the Barents and Pechora Seas and the obtained offshore seismic velocities interpreted as related to the Riphean Timanian complex would have to be associated with these considerably disturbed and intruded Riphean strata. The general structural trend is also suggestive of this situation.

The pericratonic deposits of the Varanger–Kildin zone, accumulated on a relatively stable substratum, do not seem to have any direct equivalents in the rock succession of the Timans. South-west of the Timans, however, beneath the Moscow Syneclise, very slightly disturbed Late Precambrian (Riphean and Vendian) strata have been detected resting on the pre-Riphean crystalline substratum and regarded as belonging to the platform cover of the Russian Platform (e.g. Bruns, *in* Keller 1963). These strata could possibly be equivalent to the epicontinental marginal sequence of the Varanger–Kildin Zone. If the upper part of the Riphean portion of these strata has once extended towards the Timans overlying the miogeosynclinal Riphean succession, as is the case on Varanger Peninsula, it would have had to have been completely eroded and removed in pre-Ordovician time.

The comparisons outlined above suggest that a regionally extensive longitudinal zonation of the sedimentary basin occurred in Riphean times with the different sedimentary and structural regimes forming belts along the edge of the craton. This is a development and modification of some earlier ideas (Zhuravlev & Osadchuk 1960; Raznitsyn 1962), a modification in which strata of the Varanger–Kildin Zone are shown to be related not to one but to two sedimentary regimes, only one of these regimes being comparable to the bulk of the Riphean strata of the Timan–Kanin Zone. Consequently, the East Finnmark–Timan Region of Riphean sedimentation embraced not two but three, main, depositional regimes: (1) external (pericratonic), (2) intermediate and (3) internal. Sediments associated with the external and intermediate belts crop out extensively in the Varanger–Kildin Zone, while those of the intermediate and internal belts appear in the Timan–Kanin Zone.

Ideas on the structure of the Pre-Palaeozoic basement of the Timan - Pechora - Ural Region and its extension beneath the Barents Sea

Investigations of the Riphean rocks and structures of the Timans have been fundamental not only for the notions of facies and tectonic zonation but also in providing some general ideas on the structure and age of the Timan–Pechora–Northern Urals region. These ideas and interpretations reviewed briefly below, have been extended by various Russian workers to the deep structure of the Barents Shelf.

Two alternative models have been forwarded for the basement structure of this region and both seem to have been originally suggested by Schatsky (1935, 1964). In general terms, the main features of the models are as follows: 1) the development of one broad Riphean geosyncline oriented NW–SE and embracing the Timans, the substratum of the Pechora Basin and the Urals; 2) the existence in Riphean time of two independent depositional basins: Timanian and Uralian. Subsequently, two schools of thought have developed among Russian geologists supporting one or other of the models, and the problem is still far from being solved.

Zhuravlev, Osadchuk, Gafarov and others in a series of papers have developed the model of the *Timan–Ural Riphean Geosyncline* (e.g. Zhuravlev & Gafarov 1959; Zhuravlev & Osadchuk 1960, 1962; Gafarov 1966; Zhuravlev 1964, 1972; Zhuravlev et al. 1965; Vasserman et al. 1968; Morkrushin & Tarbayev 1973). South-western miogeosynclinal and north-eastern eugeosynclinal parts have been differentiated in this broad depositional basin, which is thought to be present beneath the Pechora Basin and the Pechora and Barents Seas (Figs. 4; 5a).

As shown in Fig. 4 the miogeosynclinal portion of the Timan–Ural geosyncline has been further subdivided into 1) *western marginal* and 2) *eastern internal* zones based on differences in lithology and the degree of structural deformation, which is much more advanced in the eastern zone. Sequences in these two zones were initially considered as facies equivalents (Zhuravlev & Osadchuk 1960) but later the idea of tectonic zonation was forwarded (Zhuravlev & Osadchuk 1962), the marginal and internal zones being thought to represent diachronous deposits related to stages of tectonic development of the miogeosyncline. Carbonates of the Bystrynskaya Formation (see p. 325) have been regarded as having developed at the junction between the two zones (Getsen 1970, 1971b; Zhuravlev 1972). The miogeosynclinal Riphean structural complex is characterized by distinct negative magnetic anomalies and both magnetic and gravimetric surveys have indicated that is is bordered to the south-west by a system of deep-seated en échelon lines of discontinuity – the marginal suture of Russian geologists – while the so-called East Timan

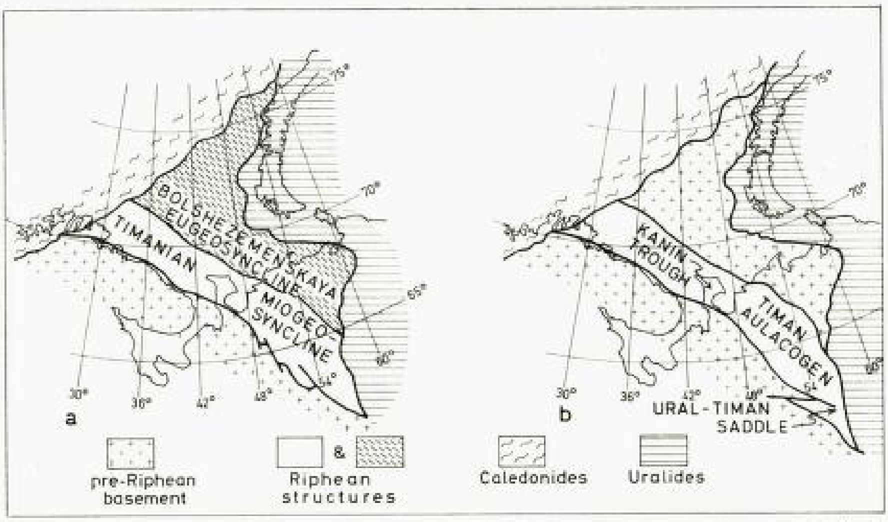


Fig. 5. Alternative interpretations of the structure of the pre-Palaeozoic basement of the Timan–Pechora–Urals Region (after Bogdanov 1965; Gafarov 1966; Siedlecka & Siedlecki 1967; Provodnikov 1970). (See text for detailed explanation.)

deep fracture delineates the Timanian fold belt in the north-east bordering the proposed Riphean eugeosyncline (Zhuravlev & Gafarov 1959; Gafarov 1963, 1966).

According to the second model, the Timan–Kanin Zone represents a post-Karelian, intracratonic, deep trough which has been variously described as an orthogeosyncline, intracratonic geosyncline, miogeosyncline or aulacogen (Stille 1955; Bogdanov 1961, 1965; Offman 1961; Schatsky 1955, 1964; Raznitsyn 1962, 1965, and others (Figs. 4; 5b). More recently Provodnikov (1970, p. 40) has worked out a more detailed structural picture of the Timan–Kanin Zone subdividing it into the Timan Aulacogen, the Kanin Trough, the Ural–Timan Saddle and the Timan Foredeep and has suggested that this structural framework is Karelian in age – a view which is not consistent with geological data. His structural model is in line with Schatsky's (1964) general idea, and this constitutes a contribution to the principal discussion on the structure of the basement of this region; his suggestions on age must obviously be rejected, however.

The present author is in favour of the second model for the following reasons: 1) Some new data on the nature of Riphean strata from the western part of the Northern Urals indicate that they are representative of trough (or miogeosynclinal) deposits rather than of a eugeosynclinal facies (e.g. Provodnikov 1970; Puchkov & Raaben 1972), and thus the postulated continuation of the 'eugeosynclinal' Riphean of the Urals beneath the eastern Pechora Basin (the Bolshezemenskaya Tundra) does not seem to be applicable any longer; 2) The Timan–Kanin Zone has represented a tectonically active (mobile) belt over a relatively short time-span in comparison with the Uralian fold belt. This situation is easier to explain if one assumes the existence of two separate zones of different mobility, and the known dynamic relations between the geosynclines and associated aulacogens, for example from the substratum of the Siberian Platform, (e.g. Salop & Scheineman 1969), appear to provide a satisfactory comparative model; 3) The field of positive magnetic anomalies characteristic of the Pechora Basin separates the Urals from the Timans which both represent quiet magnetic zones (see Demenitskaya 1967); 4) An explanation for the prominent deflection of structures of the Polar Urals towards Novaya Zemlja becomes simpler if one accepts the presence of an irregular rigid plate west of the Polar and Northern Uralian orogenic belt rather than the occurrence of a common broad Timan–Ural eugeosyncline; 5) Seismic data published by Litvinienko (1968) also support the idea of the existence of a pre-Riphean craton north of the Riphean complex detected in the southern Barents shelf (see p. 330 in this paper). It should be said, however, that none of the above points dismisses the first hypothesis and that conclusive evidence in favour of one or other of the theories is still lacking.

Returning to the subject of linear zonation of the sedimentary facies and structure development of the East Finnmark–Timan Region, a more or less symmetrical zonal arrangement of the depositional regimes has to be assumed to be present at depth beneath the Pechora Basin if one accepts the hypothesis of the occurrence of an independent Timanian sedimentary basin, although some asymmetry in this theoretical distribution pattern may of course be allowed. On the other hand, in moving from the continental margin located close to the Timans towards the Uralian basin one should expect an asymmetrical facies distribution if one accepts the first hypothesis.

Relationship of the Timanian Baikalides and Scandinavian Caledonides

Ideas on the occurrence of a major orogenic belt along the north-eastern margin of the Fennoscandian Shield have developed since the end of the nineteenth century. Ramsay (1899, 1911), Reusch (1900) and Tschernyschew (1901) have already suggested the presence of an old orogen, the Timanian Fold Belt, stretching from the Timans to Varanger. Schatsky (1935, 1958, 1960) later elaborated on the geological implications of this idea of the early explorers: he defined the Timanides as a Riphean Fold Belt deformed during the Baikalian Orogeny which predated the Caledonian diastrophism. Several Russian geologists have also suggested alternative interpretations for a further westward continuation of this orogenic belt, i.e. its relationship to the Scandinavian Caledonides. Some (e.g. Polkanov & Gerling 1960) have suggested that the Timanian Baikalides bend south-westwards and have in general the same trend and extent as the superimposed Scandinavian Caledonides. There is, however, no field evidence to support this hypothesis and it has been completely abandoned by later Russian students, who in turn have forwarded an alternative interpretation according to which the NE extension of the Scandinavian Caledonides crosses the Timanian Baikalides beneath the platform cover of the Southern Barents Shelf (e.g. Atlasov 1964; Gafarov 1966; Emelyanov 1971; Zhuravlev 1972). Raznitsyn (1962) has independently suggested an intermediate model.

Geologists dealing with problems of Late Precambrian geology in Scandinavia have not been very much involved in the discussion of a possible continuation of the Timanian Baikalides within and off the coast of northern Norway. Holtedahl (1918, 1960) has been inclined to believe that some structural features of eastern Varanger Peninsula might have constituted relics of the Timanian structural trend, and he was aware of the overall lithological similarity between the Varanger, Rybachiy–Sredniy and Kildin succession. This is in full agreement with the recent geological data collected by Siedlecka & Siedlecki (1967, 1972).

In the Varanger–Kildin Zone distant Baikalian movements have caused periods of tilting and erosion. These are recorded by the angular unconformity within the Riphean sequence (on Varanger Peninsula) and that occurring between the Riphean and Vendian successions (Varanger and Aynov Islands). As mentioned earlier (p. 317), although the NW–SE oriented cross-folds of the Vardø area are, according to Roberts (1972), essentially Caledonian, similarly trending folds in the central and south-eastern parts of Varanger Peninsula have not been studied in detail. It could be suggested, therefore, that some of these open NW–SE structures may be related to the Baikalian folding, which is quite well developed on Rybachiy–Srednyi. The high K/Ar ages of some dolerite dykes on Varanger Peninsula (Beckinsale et al., in press) should also be kept in mind in considerations of the tectonic deformation of the area, but a final solution of this problem must wait until more data are available. In general, the pre-Caledonian structure of the Varanger–Kildin Zone is not especially complex and the zone may therefore be regarded as representing the peripheral belt of the Timanides, the central more strongly deformed zone of which probably lies buried beneath the Barents Shelf.

The Trollfjord–Komagelv–Rybachiy Fault Zone is considered by the present author as a structure genetically related to a primary fracture zone, which helped to initiate the Upper Precambrian depositional basin, this idea being based on comparison with the deep structure of the Timans. Assuming that this fracture has been a zone of weakness in the crust, subsequent upthrusting, down-faulting and strike-slip multiphase movements may be envisaged to have occurred along this essentially ancient feature. Some of the strike-slip movements, suggested on the basis of regional geological and structural data (Laird 1972; Roberts 1972; Harland & Gayer 1972), could have occurred very late (? Cenozoic), at much the same time in fact as the parallel fault zones detected offshore which follow the ancient structural trend (e.g. Klenova 1960; Emelyanov et al. 1971; Åm, *vide* Roberts 1972).

Harland & Gayer (1972) and Gayer & Roberts (1973) have suggested the name 'Southern Barents Sea Caledonides' or 'Barents Sea Caledonides', respectively, for NW–SE oriented structural complex underlying the north-eastern parts of the Varanger and Rybachiy Peninsulas. These 'Southern

Barents Sea Caledonides' in general trend and extent correspond to what has been considered as a continuation of the Timanides and later, more specifically, as an outer structural portion of the Kanin–Timan Baikalian Orogenic Belt (e.g. Atlasov 1964).

The two principal orogenic belts — the older Baikalian Timanides and the younger Scandinavian Caledonides — have resulted from several pulses of crustal deformation, and by definition there exists a certain overlap of movements in time. The Baikalian orogeny has been assumed by Schatsky (1960) to have been completed by Middle Cambrian, while Harland (1965) has suggested that the duration of the Caledonian diastrophism might extend from 800 ± 100 to 350 m.y., although usually the Caledonian orogeny is understood as a lower Palaeozoic period of crustal movements. In north-western Norway there is now good evidence that the main *Caledonian* metamorphism and folding occurred in late Cambrian to early Ordovician times, 530–500 m.y. (Sturt et al. 1967; Pringle & Sturt 1969; Pringle & Roberts 1973), with a second metamorphic event in the Silurian, 420–384 m.y. There are as yet no data on the age of the Caledonian deformation in the north-easternmost part of Norway; Roberts (1972) has argued that although Tremadocian fossils are present in one area such that the main deformation may possibly be regarded as Silurian, structural continuity towards the south-west suggests that the main folding and metamorphism in the Barents Sea Region could very well be late Cambrian/early Ordovician, as in west and central Finnmark.

On the other hand, the K/Ar ages of syn-tectonic and partly post-tectonic basic dykes of the Barents Sea Region (Beckinsale et al. in press) fall generally within the range 542 m.y. to 1000 m.y. These data, although only preliminary, are indeed indicative of some pre-Caledonian, early-Baikalian disturbances.

The last *Baikalian* metamorphic event in the Timans and on Kanin took place around 525–520 m.y., i.e. contemporaneously with the peak of Caledonian orogeny in north-western Norway. The angular unconformity at the Riphean/Vendian interface in north-eastern Norway must be thus thought either as the last definite trace of the Baikalian orogeny or as reflecting an initial stage of the Caledonian diastrophism. It is evident that there is no time gap between the Caledonian and Baikalian orogenies to make these positively distinguishable, nor are they spatially distinctly separated. The precise character of the junction between the two fold belts, the Scandinavian Caledonides and the Timanian Baikalides, is not known because the critical land areas of East Finnmark embrace only the peripheral parts of these fold belts and require further detailed investigation. The overlap in time, no matter what is the structural pattern of the junction between the more central parts of the two discussed orogenic systems (i.e., either a transection of structural trends or a bending), is indicative of: (1) the occurrence, in the vast Finnmark–Timans region, of several pulses of crustal movement which have been somewhat artificially assigned to two separate orogenies and, (2) a gradual shift of intensity of movements towards the west and south-west (the Timans→Northern-Scandinavian Caledonides→Central Scandinavian Caledonides), this reflecting changes in the pattern of mobile and stable areas of the crust.

Suggested genetic relationships between the Uralian, Timanian and Scandinavian orogenic belts

Because the problem of the deep structure of the SE Barents Shelf and the Pechora Basin has not been finally solved (see pp. 333, 335), in these further general consideration I have assumed that one of the two existing alternative models is more credible than the other. For the reasons presented above (p. 334) I have accepted the reconstruction in which the Timanian and Uralian basins have been independent palaeogeographically and the Fennoscandian Shield has been separated from the Barents Shield by the Timan Aulacogen (Fig. 5b). Following from this, assuming continental drift to be a phenomenon of general application I have then attempted to translate the relationships between Scandinavian Caledonides, Uralides and Timanides into the language of plate tectonics.

Although controversial models exist, the concept of continental drift has been shown to be applicable to the early history of the North Atlantic Ocean (Wilson 1966; Harland 1967, 1969, 1973). The proto-North Atlantic is considered to have opened probably sometime in the Precambrian (?Middle/Upper Precambrian) and to have existed through to late Palaeozoic time. Modifications of the ocean topography and of the distribution of geosynclines have occurred due to changes of relative plate movements, and zones of plate convergence are recorded by fold-belts.

Harland & Gayer (1972), in an analysis of the early stages of development of the proto-North Atlantic, have suggested the occurrence of rifting between the Baltic and Barents cratons (their Fig. 2a) in late Precambrian time. This initial rifting developed in the late Precambrian/early Palaeozoic time into a zone of oceanic spreading, the Iapetus Ocean (their Fig. 2b), which in this interpretation is located in a position intermediate to the structural units inferred by Russian geologists to occur beneath the Barents Shelf: the Timanian Baikalides and the Norwegian–Barentsian Caledonides (e.g. Atlasov 1964; Gafarov 1966).

In order to fit the model of oceanic spreading suggested by Harland & Gayer (1972) with more recent Russian structural maps of the Barents Shelf, I would suggest that the Iapetus Ocean in fact coincides with the Norwegian–Barentsian Caledonides and is independent of the Timanian fold belt. Furthermore, the deep structure of the Barents Platform is here differentiated into the Barents Craton in the north-west separated by the Norwegian–Barentsian Caledonides (Iapteus Ocean) from the *Pechora Craton* to the south-east. The Pechora Craton is a new term which I have introduced for the deep substratum of the south-eastern portion of the Barents Shelf and the Pechora Basin. I retain the term Barents Craton for the north-western portion of the Barents Shelf substratum: see also Glossary, p. 348). The Pechora Craton would be separated from the Fennoscandian Shield by the Timanian Baikalides (the Timan Aulacogen). An outline of this suggested distribution is shown in Fig. 6.

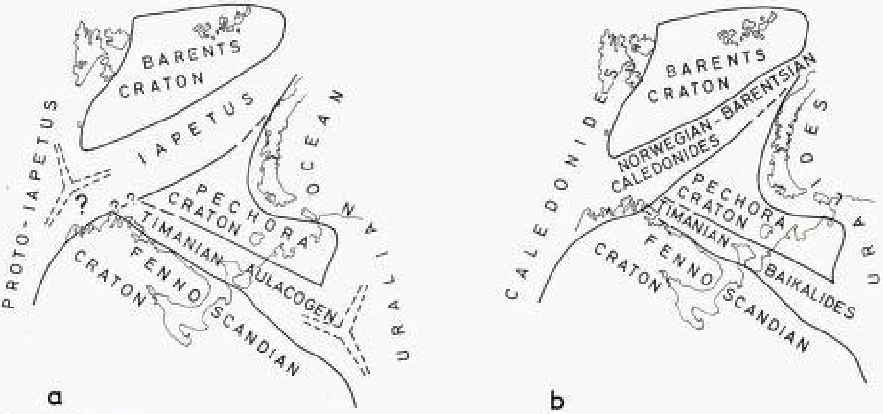


Fig. 6. Palaeogeographic (a) and structural (b) setting of the Timanian belt in the basement of the Barents Shelf (see text for detailed explanation).

To the east of the Fennoscandian, Barents and Pechora cratons there existed the Uralian Ocean, which, as with the proto-North Atlantic, persisted throughout Palaeozoic time. The opening of this basin presumably occurred sometime in the Middle/Upper Precambrian and the thick eugeosynclinal and miogeosynclinal Riphean sequences were then deposited. Analysis of the geological and geophysical information has suggested the existence of an earlier eastward-dipping subduction zone (between the Russian Plate and an inferred island arc) and of a later developing Benioff zone along which oceanic crust has been disappearing until the Russian and Siberian plates collided in Permian/Triassic time (Hamilton 1970). Although the Precambrian/Cambrian history of the Urals is somewhat uncertain and difficult to interpret in terms of plate tectonics, it would seem that the first period of converging plate movements has produced the Baikalian (? and Caledonian, see Raznitsyn 1962, 1972) pre-Uralids and the second the Hercynian Uralids of the Russian geologists (e.g. Bondarev et al. 1973).

In the above crude simplification of the suggested history of the proto-North Atlantic and Uralian oceans I have attempted to show that since ? Middle/Upper Precambrian time the Fennoscandian Shield and the Barents and Pechora cratons have been bordered by two geographically and temporally extensive oceans in which shelf deposits and thick geosynclinal sequences were accumulated. Oscillatory movements of plates underlying and bordering the two respective oceans were not synchronized, although both oceans closed at about the same time. In the light of all this the history of the Timan Aulacogen appears much shorter and simpler; it is a sedimentary basin which probably opened sometime in the Middle/Upper Precambrian and ceased to exist by the very end of the Precambrian. This narrow elongated basin probably resulted from a restricted intracontinental spreading down-warping and faulting followed by folding and upthrusting of the accumulated sediments without any subduction zone development. Rifting was probably shallow in contrast to equivalent phenomena in the Uralian and proto-North Atlantic oceans. Thus, although the initial opening of all three basins probably occurred more or less simultaneously in Precambrian times, their subsequent histories have been quite different, the Timanian basin being a zone of comparatively weak tectonic activity of short duration. Data on the suite of igneous rocks of the Timans (p. 328, Fig. 4) are not very helpful for purposes of interpretation. The variety of intrusive bodies indicates that they might have been related to periods of mantle upwelling and to phases of relaxation post-dating compressive stresses.

Aulacogens are, by definition, deep linear troughs genetically related to extensive oceanic basins or geosynclines, arranged transversely to the margins of the platforms and dying out towards the centre of a craton. Since the Timanian Aulacogen is oriented more or less perpendicular to the main trend of both the proto-North Atlantic and the Uralian basins, an important question immediately arises; did two separate genetic junctions exist, 1) the Uralian/Timanian junction and 2) the proto-North Atlantic/Timanian junction, or was there only one and, if so, where was it located? Theoretically both possibilities are equally probable; however, sedimentary facies, stratigraphy and structure of the Riphean of the Urals and the Timans are strongly suggestive of a Timan–Ural palaeogeographic interconnection, and some further suggestions may be made taking into account data on current directions in the Riphean strata of north-eastern Norway. These are indicative of a south-westerly to westerly source of supply, a fact which suggests the occurrence of a topographic high located to the W–SW at least during a part of Riphean time, with a roughly eastward longitudinal transport along a basin inclined towards the Uralian Ocean. If this is assumed as a general feature, not only restricted to north-eastern Norway, the Timanian Aulacogen would not have had a direct connection with the proto-North Atlantic Ocean. On the other hand, the great thickness of the basinal Riphean strata of Varanger Peninsula is very suggestive of a considerable north-westward extension of a palaeogeographically important sedimentary basin related to the proto-North Atlantic Ocean.

Summing up the various evidence, two alternative solutions may be offered to explain the evolution of the Timanian basin: 1) the Timan Aulacogen was genetically related only to the Uralian Ocean, and extended north-westward towards the proto-North Atlantic; or 2) two separate aulacogens developed simultaneously related to the proto-North Atlantic and to the Uralian Ocean, these merging either temporarily or permanently into one. The second idea of two independently created aulacogens seems rather difficult to imagine, and I am therefore inclined to believe in the first possibility, the development of the Timanian Aulacogen related to the Uralian Ocean. In this model, the fairly rapid sagging and NW longitudinal extension of the aulacogen is seen as resulting in a connection and eventual merging with the proto-North Atlantic Ocean.

In my attempt to combine the notion of aulacogen development with the concept of plate tectonics I have found the suggestions made by Hoffman

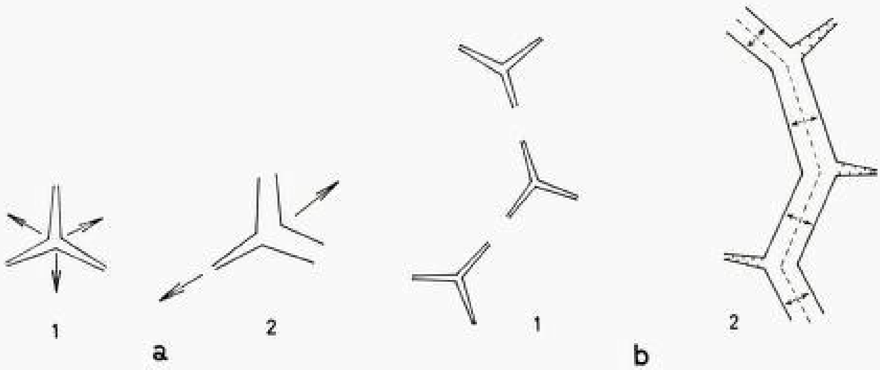


Fig. 7. Diagrammatic representation of (a) the development of an aulacogen from a triple junction and (b) the development of an ocean and associated aulacogens from a set of triple junctions.

(1972)* for the Precambrian of the NW Canadian Shield to be the most applicable in the present situation. Adapting this, I have envisaged a primary intracontinental junction between three spreading zones (the RRR type of McKenzie & Morgan 1969) at the Timanian/Uralian junction (Fig. 6). I have then assumed that the initially similar velocities of spreading could have subsequently changed with two of the three fractures becoming particularly active. As a result of this high mobility the Uralian Ocean would have developed rapidly, whereas the third branch, the Timanian Aulacogen, would have opened relatively slowly; active dilation in this third arm diminished fairly quickly and an underdeveloped abandoned basin, the aulacogen, was produced. The development of such a dynamic system is shown schematically in Fig. 7**. The early Proterozoic structure of the Siberian Platform may serve as a particularly good example of such a theoretical model. An extensive ocean with large geosynclines was developed there south of an ancient shield, and three aulacogens radiating northwards from the northern oceanic margin subdivide the shield into several cratonic blocks, the Angara and Aldan cratons being the most extensive (Salop & Scheinman 1969: Leytes et al. 1970).

The interpretations and ideas as presented in this paper by no means solve the problem of the deep structure and crustal history of the Barents Shelf and of the nature of the north-eastern continental margin of the Fennoscandian

* A modified and extended version of Hoffman's paper appeared in print after this manuscript had been submitted for publication: Hoffman, P., Dewey, J. F. & Burke, K. 1974: Aulacogens and their genetic relation to geosynclines, with a Proterozoic example from Great Slave Lake, Canada. *In* R. H. Dott Jr. & R. H. Shaver (Eds.): Modern and ancient geosynclinal sedimentation. *SEPM Spec. Publ.* No. 19, 38–55.

** The hypothesis presented for the formation of the Timanian Aulacogen has parallels in a recently published model (Burke & Dewey 1973) of plume-generated triple junction development according to which the opening of two of the three rifts, the third being an inactive failed arm, is suggested to be a general feature of intracontinental triple junction evolution. The actual number of the *Journal of Geology* (Vol. 81, No. 4) in which Burke & Dewey's (1973) paper appeared was not received by the NGU library until January 16th 1974, after the draft of my manuscript had been written and the main ideas presented at the Bergen University Oil Conference, December 15th, 1973.

Shield. The model of an aulacogen related to the Uralian Ocean and possibly also to the proto-North Atlantic, although admittedly based on relatively sparse evidence, is, however, surely one of several possible approaches which, it is hoped, will lead to our eventual understanding of this part of the earth's crust. In writing this paper I have intended to attract the attention of other students to the problems of the Late Precambrian structure of the north-eastern margin of the Fennoscandian Shield and adjacent fold belts and to initiate interest and discussion; if this is achieved a principal aim will have been fulfilled.

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Abbreviations:
AN SSSR A. Sci. USSR – Academy of Sciences of the Union of Soviet Socialist Republics.
AAPG – The American Association of Petroleum Geologists.
GSA – The Geological Society of America.
Izv(estia) – Proceedings.
MOIP – Moskovskoye Obshchestvo Ispitateley Prirody – Moscow Society of Naturalists.
GIN – Geological Institute.

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GLOSSARY

Upper Proterozoic: ca. 1600–560 m.y. (Garris et al. 1964). The term embraces Riphean and Vendian and corresponds approximately to Middle & Upper Proterozoic of the Canadian scheme (Semikhatov 1973).

Upper Precambrian = Upper Proterozoic.

Late Precambrian: corresponds to Upper Proterozoic as defined above, used rather informally.

Riphean: term introduced by Schatsky (1945) for the sedimentary Precambrian succession of the Bashkirian Urals. Embraces most of the Upper Precambrian: ca. 1600–650 m.y. Subdivided into Lower R. (1650 $\pm$ 50–1350 $\pm$ 50 m.y.), Middle R. (1350 $\pm$ 50–950 $\pm$ 50 m.y.) and Upper R. (950 $\pm$ 50–650 $\pm$ 50 m.y.) on the basis of stromatolites, onkolites and catagraphs (Garris et al. 1964).

Vendian: Term introduced by Sokolov (1952, 1958) for the youngest stratigraphic unit of the Upper Proterozoic: 650 $\pm$ 50–560 $\pm$ 50 m.y. Some beds assigned to Vendian exhibit higher ages (see e.g. Sokolov 1973). Characterized by a widespread glaciation. Equivalent to the Eocambrian of Norwegian geologists.

Varangian Ice Age: term introduced by Harland (1965a). Represented by glacigene deposits of Vendian age.

Aulacogen: term introduced by Schatsky (1964) to designate large, fault-bounded, graben-like downwarps genetically related to geosynclines. Aulacogens radiate inwards from platform margins, are characterized by gentle to moderate folding and their tectonic activity decreases with increasing distance from the platform/basin margin (Salop & Scheinmann 1969; Bogdanov et al. 1972).

Baikalian Orogeny: tectonic movements which occurred around the time of the Precambrian/Cambrian transition and which were completed in the Middle Cambrian. Initial movements are known from the Lower Riphean (Schatsky 1932, 1935, p. 154; *vide* his papers of 1958, p. 99; 1960).

Barents Platform: comprises Palaeozoic–Cenozoic sequences of the Barents Shelf overlying the Barents Craton, Pechora Craton, Norwegian–Barentsian Caledonides and the Timanian Baikalides.

Barents Craton: basement of the north-western part of the Barents Platform.

Pechora Craton: new term (Bolschemenskaya Zona of Gafarov 1966); basement of the Pechora Basin, Pechora Sea and south-eastern part of the Barents Platform.

The Geology of Varanger Peninsula and Stratigraphic Correlation with Spitsbergen and North-East Greenland

STANISLAW SIEDLECKI

Siedlecki, S. 1975: The geology of Varanger peninsula and stratigraphic correlation with Spitsbergen and north-east Greenland. *Norges geol. Unders. 316*, 349–350.

S. Siedlecki, Norges geologiske undersøkelse, P.O.Box 3006, N–7001, Trondheim, Norway

Based on a study of the various sedimentary successions observed in Varanger peninsula as well as on the suggested correlations between particular units of the three large groups of sedimentary and metasedimentary rocks, the Barents Sea, Raggo and Laksefjord Groups, it is now possible to reconstruct the history and sequence of Late Precambrian sedimentation, not only on Varanger Peninsula, but also in other areas outside the northern margin of the Fennoscandian Shield. It has also enabled Dr. Anna Siedlecka (1973) to introduce a new lithostratigraphic term, the 'Øst-Finnmark Supergroup', which embraces the above-mentioned three groups as well as the younger but still pre-tillitic Tanafjord Group.

A lithostratigraphical equivalent of the Øst-Finnmark Supergroup, deposited in approximately the same basin but later separated by continental drift, is the 'Eleonore Bay Group' of East and North-east Greenland. The Late Precambrian Hecla Hoek Sequence on Spitsbergen, and especially the 'Middle Hecla Hoek', fully developed in eastern New Friesland on Spitsbergen, may also be equivalent in age to the sequences constituting the Øst-Finnmark Supergroup.

Lithostratigraphical correlation between the Øst-Finnmark Supergroup and the equivalent series on Greenland and Spitsbergen is possible only on a large and very approximate scale. The only satisfactory horizon for a chronostratigraphic correlation in the unfossiliferous series below the Cambrian is presented by the tillites. There is no doubt that they represent one major ice age (the Varangian Ice Age of Kulling and Harland) for at least this entire northern region.

Below the tillites one finds several thousand metres thick sequences of sediments with a quite characteristic unit of carbonate rocks, often rich in stromatolite-structures. This is represented by the Grasdal Formation and Porsanger Dolomite in East Finnmark, the Akademikerbreen Group in Spitsbergen and the Nøkkelfossen Formation in East Greenland. It should be emphasised, however, that the carbonate units of Spitsbergen and Greenland are several times thicker than the Grasdal Formation, and it seems likely that they may be equivalent not only to the uppermost part of our Tanafjord

Group, but also that they represent a carbonate facies of the entire Tanafjord Group and of the upper part of our Barents Sea Group. The last two groups (the Barents Sea and Tanafjord Groups) were deposited relatively close to the Fennoscandian Shield, which was an important source area for almost all the terrigenous sediments of Varanger Peninsula.

In the lower parts of the Eleonore Bay Group on Greenland and in the lower part of the Middle Hecla Hoek on Spitsbergen, terrigenous psammitic and pelitic sediments are present. These can be very broadly correlated with the lower parts of the Øst-Finnmark Supergroup. The 'multicoloured' and carbonate-bearing Brogetdal Formation (within the Eleonore Bay Group) can probably be correlated with the Båtsfjord Formation on Varanger Peninsula. As mentioned above, the Båtsfjord Formation is also 'multicoloured' and carbonate-bearing.

Although a more precise correlation of minor lithostratigraphic units still requires further investigation, an important conclusion resulting from the work on Varanger Peninsula is that thick sequences of Late Precambrian sedimentary rocks represent an important element in the 'post-Archean' geology of the North Atlantic continental margin. The Late Precambrian in all probability underlies all the younger sequences in the vast areas of the Barents Sea, and the shelf-margins of Greenland and Svalbard. The detailed mapping of Varanger Peninsula began too recently for this area to be recognized as 'classical' for the Late Precambrian, yet many aspects of the Late Precambrian geology are better seen in Varanger Peninsula than in more familiar or established late Precambrian areas in other parts of Europe.

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Aeromagnetic Basement Complex Mapping North of Latitude 62°N, Norway

KNUT ÅM

Åm, K. 1975: Aeromagnetic basement complex mapping north of latitude 62°N, Norway. *Norges geol. Unders. 316*, 351–374.

The Geological Survey of Norway (NGU) has conducted aeromagnetic measurements over Norway since 1959, and to date approximately 90% of the country has been covered with 435,000 line kilometres of such measurements. Since 1962 NGU has also measured 183,000 kilometres of aeromagnetic profiles over Norwegian shelf areas. The entire shelf between Stad (62°N) and Bear Island (74.5°N) out to a water depth of 2000 m and as far east as 36°E has now been systematically covered with approximately 4 km line spacing. From the measurements over the shelf, magnetic isogam maps with a 20 gamma contour interval have been constructed at a scale of approximately 1 : 350,000 south of 69°N, and ca. 1 : 700,000 north of this latitude. Copies of the maps and the original records have been sold to a number of interested companies. The material is confidential, but NGU has been allowed to publish generalized small-scale maps together with a rough interpretation of the data.

An interpretation in terms of depth to magnetic basement reveals that the shelf north of 62°N is composed of great thicknesses of sediments. Outside western Norway there are maximum thicknesses exceeding 10 km, and the inner part of the Vøring Plateau is underlain by approximately 6 km of sediments. Even in Vestfjorden, a major fjord south of Lofoten, there are more than 4 km of sedimentary rocks.

In the Barents Sea some special problems arise in the interpretation due to the presence of several kilometres of non-magnetic, Late Precambrian–Eocambrian sediments. An aeromagnetic survey over the Varanger Peninsula shows up to 7 km of such sediments. The Caledonian belt of metamorphic rocks is practically non-magnetic, except for some small iron ore-bearing formations and scattered ultrabasic bodies. However, in addition to anomalies from the basement, which is probably the Precambrian Baltic Shield, it has been possible to trace anomalies from a shallow level (?Paleozoic). Contouring of these shallow depths shows that they are in fairly good agreement with acoustic (refraction) basement.

K. Åm, Norges geologiske undersøkelse, P.O.Box 3006, N-7001, Trondheim, Norway.
Present address: Statens Oljedirektorat, N-4000 Stavanger, Norway

Introduction

Aeromagnetic measurements in Norway have been conducted by the Geological Survey of Norway (NGU) since 1959, and up to the present time about 90% of the country has been systematically covered with approximately 435,000 line kilometres of such measurements (Fig. 1). Areas with relatively gentle topography were covered first, mostly with around 500 m line spacing and 150 m ground clearance. In recent years, when areas with more rugged terrain have been surveyed, it has been impossible to maintain a constant ground clearance. In these cases profiles have been flown as low as possible at constant barometric altitude within each region, and the line spacing has been adjusted accordingly.

* Publication No. 54 in the NTNF Continental Shelf Project.

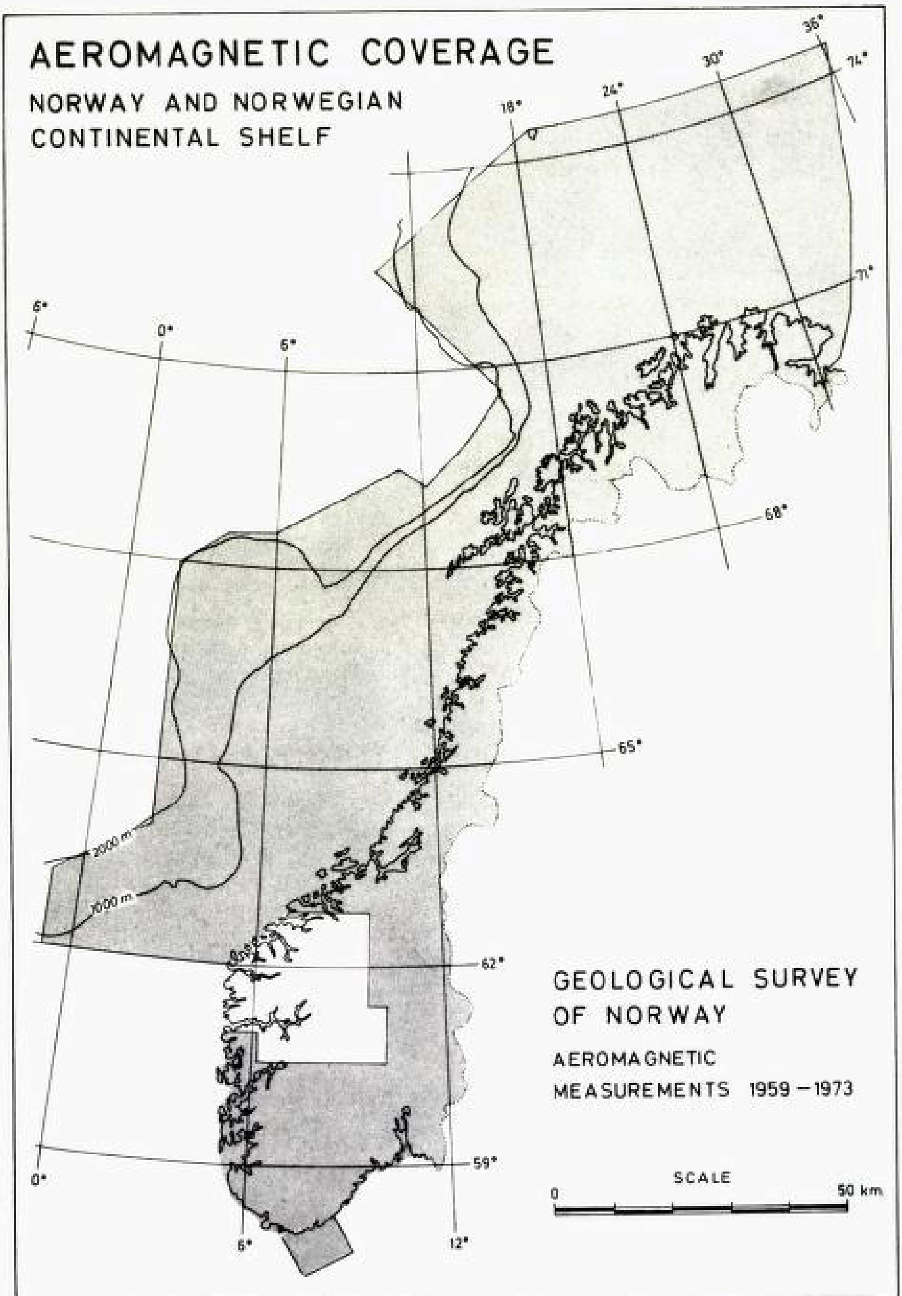


Fig. 1. Areas systematically covered with aeromagnetic measurements by NGU in the years 1959–1973.

From the measurements over land, NGU produces magnetic isogam maps with a 100 gamma contour interval at scales 1 : 50,000 and 1 : 100,000. The maps are available in the form of paper copies at a price of 12 N.kr. per sheet. These maps are further reduced to a scale of 1 : 250,000, redrawn and printed in colour.

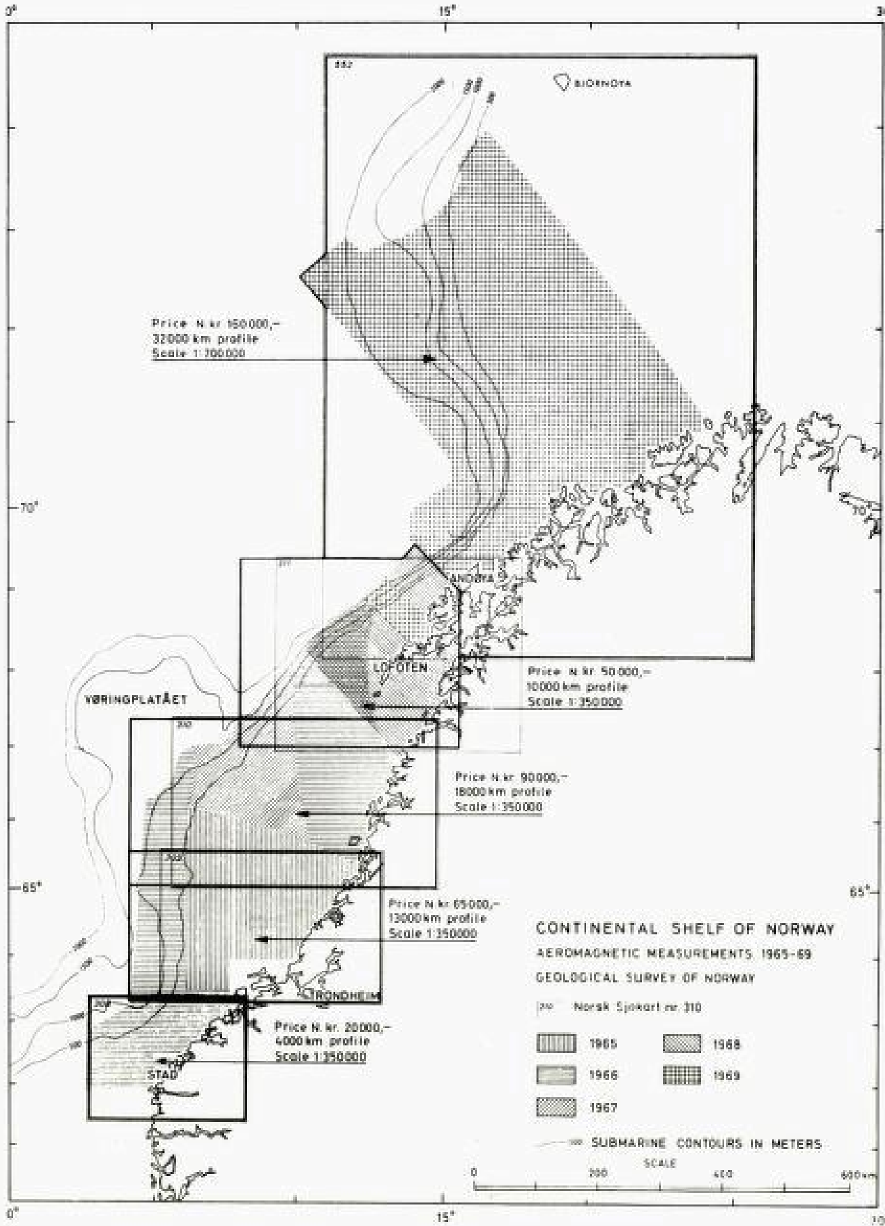


Fig. 2. Aeromagnetic maps from the continental shelf of Norway on sale at NGU, Trondheim.

Since 1962 NGU has also measured some 185,000 kilometres of aeromagnetic profiles over Norwegian shelf areas. After some reconnaissance flights in the Skagerrak in 1962–63 (Aalstad 1963, 1964a, Sellevoll & Aalstad 1971) and between Stad (62°N) and Andøya (69,5°N) in 1963–64 (Aalstad 1964b), systematic suveying started in 1965. To date, the entire shelf between Stad and Bjørnøya (Bear Island, 74,5°N) out to a water depth of around 2000 m and as far east as 36°E has been covered with approximately 4 km line spacing or a total of 160,000 line kilometres (Fig. 1). In addition, a detailed survey was flown in the Skagerrak in 1965 (Sellevoll & Aalstad 1971, Åm 1973), and in 1970 some reconnaissance profiles were measured over Svalbard and surrounding waters (Åm, in press).

Magnetic isogam maps with a 20 gamma contour interval have been constructed at a scale of 1 : 350,000 south of 69°N and 1 : 700,000 north of this latitude. The maps are largely confidential, but copies of some of the map sheets and original records are available at a price of approximately 5 N.kr. per line kilometre; this concerns the area covered up to 1969 (Fig. 2). In addition, NGU has been allowed to publish generalized small-scale maps from the shelf together with a rough interpretation of the data.

In the years to come, NGU hopes to be able to carry out detailed measurements along the coast in order to tie together the separate measurements made over land and sea. At the moment, this is a difficult task because of the wider line spacing used in the profiling of the sea areas.

A general outline of the geology of Northern Norway is presented in Fig. 3. A full account of the geology will be found in Holtedahl (1960). The old Precambrian Baltic Shield to the east is known to contain strongly magnetic rocks, but also large areas of non-magnetic material, mainly quartzites. In the north-east this metamorphic complex is overlain by several kilometres of Late Precambrian–Eocambrian sediments. This unit is completely non-magnetic except for some swarms of diabase dykes. The overlying Caledonian belt of metamorphosed Eocambrian to Silurian rocks is also practically non-magnetic except for some iron ore formations and strongly magnetic gabbro provinces. However, it should be noted that many of the Caledonian gabbros are non-magnetic. The only Mesozoic sediments known in Norway are found in a small down-faulted area on Andøya.

The purpose of this paper is to present some of the results of aeromagnetic mapping over Norway and its continental shelf conducted by NGU. The following subjects and areas will be dealt with:

1. A short section on the interpretation of magnetic measurements.
2. The shelf between Stad and Lofoten including the Vøring Plateau.
3. Lofoten–Vesterålen (granulites).
4. Andøya and Andfjorden (Mesozoic sediments).
5. Kvænangen–Finnmarksvidda (section across the Caledonides).
6. Varangerhalvøya (Late Precambrian sediments).
7. The shelf between Andøya and Bjørnøya.

Interpretation of magnetic data

Magnetic anomalies reflect variations in the content of magnetite in the bedrock, and as the distribution of magnetite is governed by geological laws a magnetic map reveals geological features. In particular, strike directions, tectonic lines and the outline and areal extent of geological provinces are often

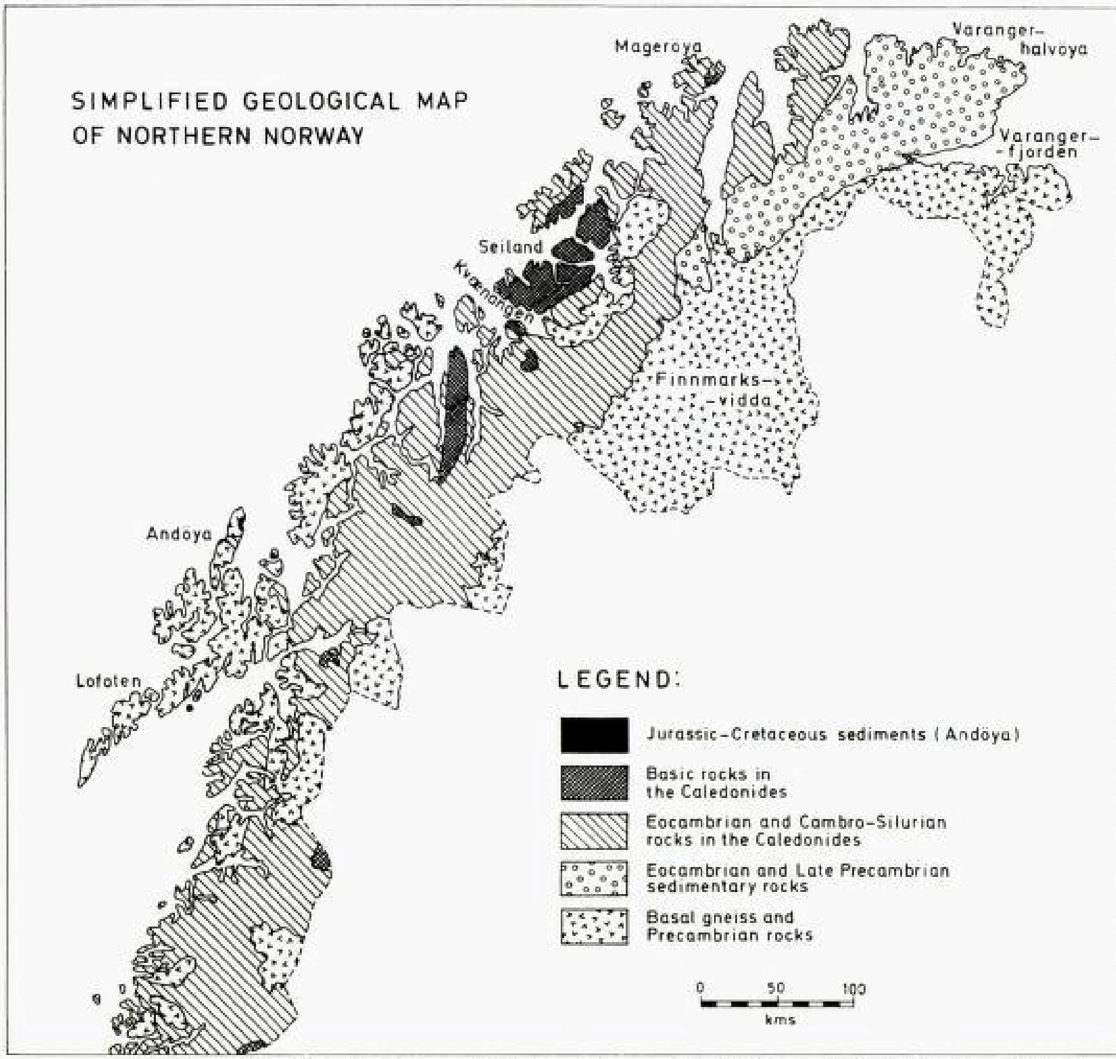


Fig. 3. Geological map of Northern Norway. Modified from Holtedahl & Dons (1960).

clearly seen. From the form of a magnetic anomaly it is also possible to determine the approximate depth to the top of its causative body (Vacquier et al. 1951). In a general way it can be stated that a rugged appearance with steep gradients and sharp anomalies indicates shallow sources, while a smooth picture with gentle slopes and usually broader and weaker anomalies is indicative of a deeper origin. This is well illustrated in Fig. 4, which shows one of the reconnaissance profiles measured over the shelf by NGU in 1963. The sharp and rugged anomalies over land are due to causes actually exposed or close to the surface, whereas it is probably 5–8 km down to the sources of the broad anomalies recognized above the shelf.

It is well known that aeromagnetic surveys over sedimentary basins are very useful for the determination of a general picture of the basement surface. Such

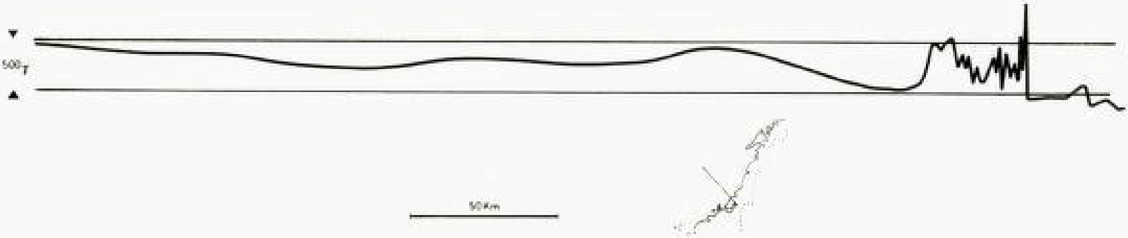


Fig. 4. Reconnaissance NW–SE magnetic profile flown across the shelf by NGU in 1963, after Aalstad (1972). The profile illustrates the difference in magnetic signature over the crystalline rocks on land and over the deep sedimentary basin on the shelf.

interpretations are generally found to be in reasonably good agreement with reality (e.g. Nettleton 1971). The depth determinations are based on two fundamental facts: 1) From the form of a magnetic anomaly it is possible to calculate the depth to the top of the body causing the anomaly. 2) Significant magnetic anomalies are almost exclusively due to magnetization contrast in the basement because the sedimentary cover is generally devoid of larger masses of magnetic material.

Consequently, if it is assumed that the magnetic bodies reach the surface of the crystalline basement, the depths to the tops of the anomalous bodies, determined from a careful study of the anomalies, will give points on the basement surface. When such depths have been determined for all magnetic anomalies in an area, the results are critically examined before a generalized contouring of the basement surface is made, giving less weight to uncertain determinations and to values differing too much from the others.

A review of the various methods for manual interpretation of magnetic data by the use of characteristic points and distances in the anomaly curves has been published by Åm (1972). Three lengths related to the inflection tangents of an anomaly are often used as depth estimators. These are the Straight Slope length, the Peters length and the Sokolov length, illustrated in Fig. 5. To yield the correct depth, the depth estimator used must in each case be divided by a certain factor depending on the form and dimensions of the causative body, which can be found with the aid of a suitable interpretation chart. However, magnetic anomalies generally fall into two classes so that as a rule the Straight Slope length has to be divided by 0.7 or 1.0, the Peters length by 1.5 or 2.0 and the Sokolov length generally by a factor around 2.0.

In the present case the interpretation has been made directly on the original magnetometer records by the use of characteristic points and distances in the anomaly curves. All the three depth estimators mentioned above have been used where possible. Since the original magnetometer records from the Kvænangen–Finnmarksvidda region are curvilinear, it was difficult to use the Straight Slope or the Peters length in that region. However, the Sokolov length, when averaged from two neighbouring profiles flown in opposite directions, yielded results close to those obtained on redrawn profiles. The Solokov length was therefore used throughout in the Kvænangen–Finnmarks-

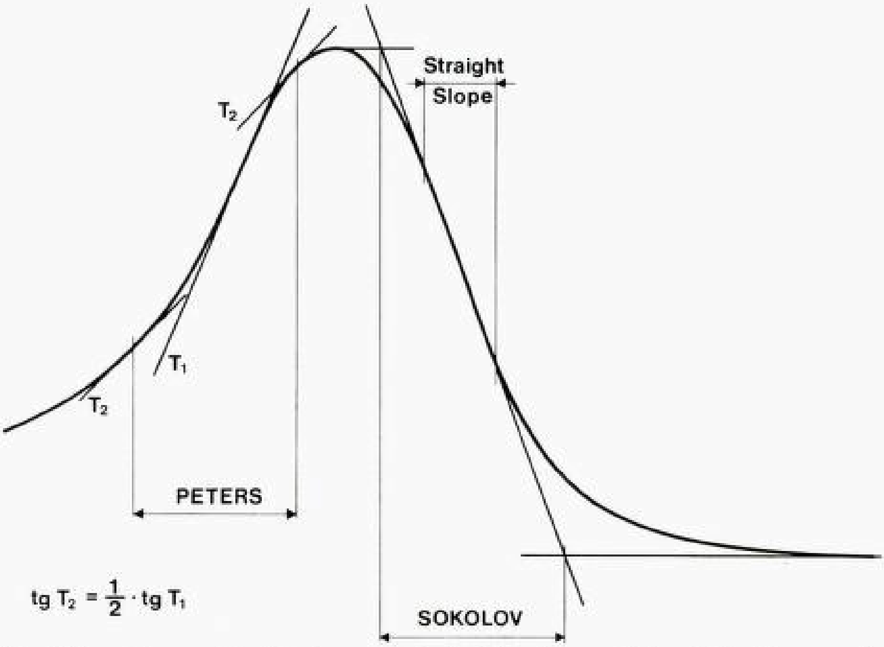


Fig. 5. Magnetic anomaly with three popular depth estimators. When divided by a certain factor, a depth estimator gives the depth to the top of the body causing the magnetic anomaly. In the Figure the vertical represents field strength and the horizontal is distance.

vidda region, together with the interpretation chart in Fig. 16 of Åm (1972). The depths thus obtained have been corrected for strike directions and strike extents where necessary.

The shelf between Stad and Lofoten (62-68°N)

A map of the residual magnetic field between Stad and Lofoten was published in 1970 together with a rough interpretation of the data, mainly in terms of depth to magnetic basement (Åm 1970). The data indicated the existence of a sedimentary basin more than 200 km wide aligned parallel to the coast with its axis 120–150 km from the coastline.

Off the coast of Nordland (65–67°N) the axis of this basin is situated not far from the middle of the shelf with maximum depths to basement exceeding 9 km. A culmination (7–8 km depths) is indicated outside Trøndelag (64°N). Off Møre (63°N) the basin deepens again, reaching depths of probably more than 13 km with the axis situated on the continental rise about 60 km off the shelf edge. There is a pronounced difference in magnetic signature north and south of the culmination (64°N), with strong and irregular anomalies to the south and a weak and featureless picture to the north. The depth to anomaly sources is approximately the same in both regions, so there must be a difference in the type of basement. The strong and irregular magnetic anomalies to the south are not unlike the anomalies observed over the Precambrian gneisses of the adjacent mainland, implying that the magnetic basement south of 64°N is probably Precambrian. The weaker anomalies to the north of 64°N might possibly be due to intrusions of Caledonian age, suggesting that the magnetic basement could be Palaeozoic in this region.

The Vøring Plateau can be divided into two parts on the basis of the magnetic picture seen in the U.S. Naval Oceanographic Office map (USNOO 1967). The inner half of the Vøring Plateau is magnetically quiet and similar in magnetic pattern to the adjacent shelf area. Some very weak magnetic anomalies indicate that the basement is 6–7 km deep in this part of the plateau. The outer part of the Vøring Plateau is characterized by shallow 'volcanic' anomalies similar to those observed over the deep ocean. These anomalies could be due either to (1) shallow oceanic basement or to (2) volcanic material in the sediments. The first possibility is supported by Talwani & Eldholm (1972) who demonstrated the existence of shallow refraction basement under the outer part of the Vøring Plateau, and the second is supported by Hinz (1972), who found that this part of the plateau is a continental fragment. There are also some magnetic indications that the second possibility might be true, e.g. magnetic anomalies continuing into the eastern part of the plateau, where they seem to originate from sources above the basement. It is hoped that an interpretation of the detailed aeromagnetic survey of the plateau carried out by NGU in 1973 will provide an answer to this problem.

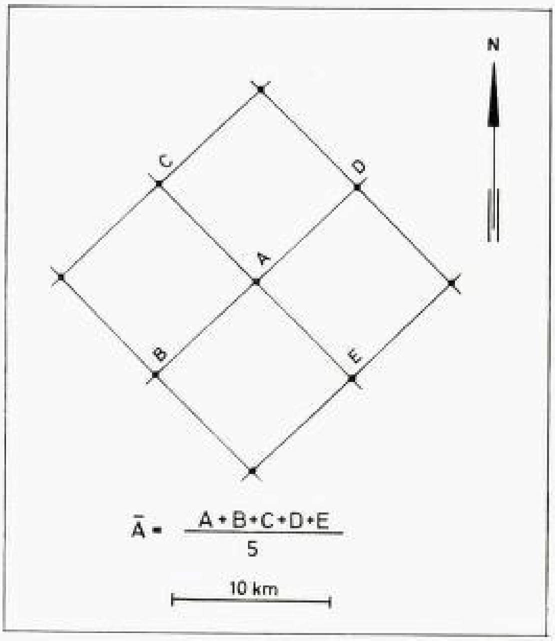
Lofoten–Vesterålen (granulites)

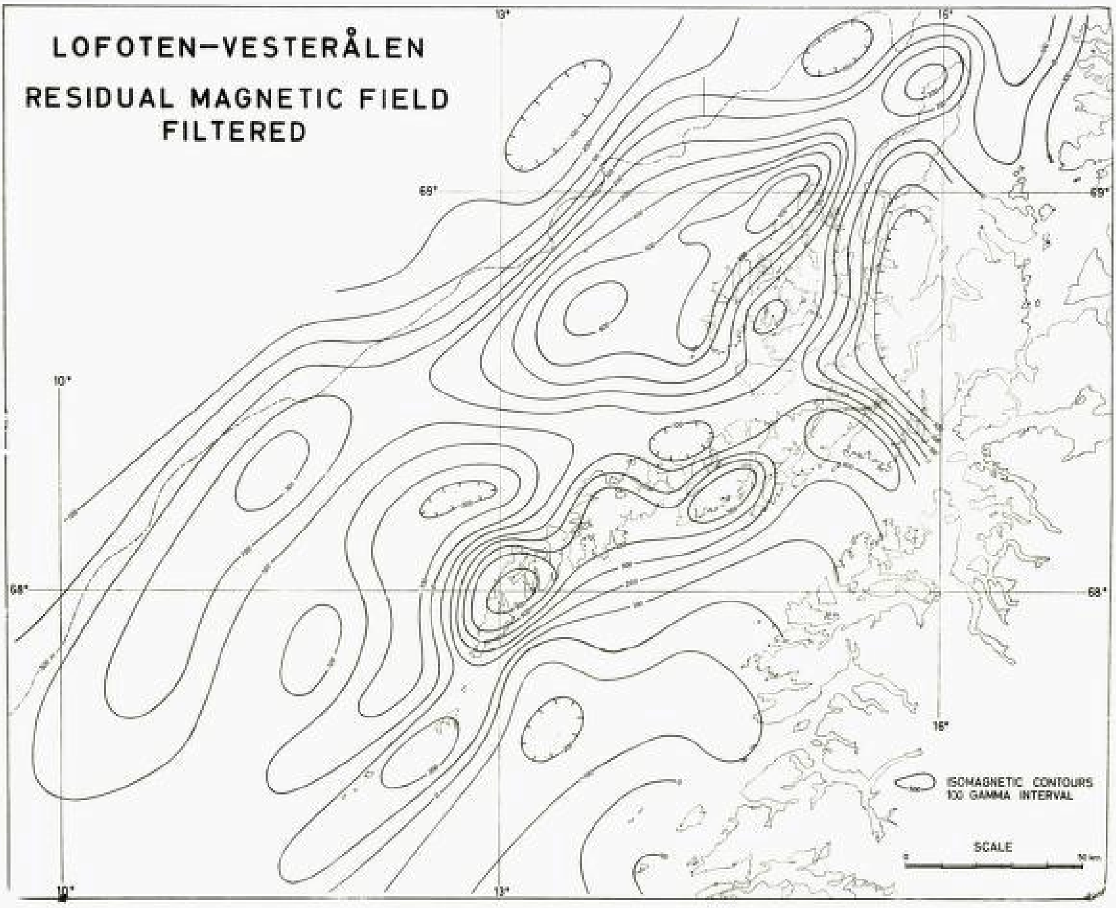
The Lofoten–Vesterålen island group (68–69°N) was covered with aeromagnetic measurements by NGU in 1965. The survey was flown N–S at 1000 m barometric altitude and with a flight-line spacing of approximately 2 km. The surrounding sea areas were flown NW–SE in 1968 and 1969 at a height of 200 m above sea level with ca. 4 km line spacing and Decca navigation. In spite of the different flight elevations over land and sea, it proved to be comparatively easy to tie the maps together. Five point moving averages in a 9 km square grid (Fig. 6) were then taken to filter the map, and the IGRF regional gradient, which is 2.6 γ/km towards NE in this area (Fabiano & Peddie 1969), was removed. The resulting residual map is presented in Fig. 7.

The Lofoten–Vesterålen island group consists mainly of granulite facies rocks (Heier 1960). From the strongly filtered aeromagnetic map (Fig. 7) it can be seen that large magnetic anomalies are associated with this granulite complex. The magnetic anomaly over Lofoten can easily be explained by the magnetic properties of the rocks occurring at the surface, if these extend down to a depth of around 20 km. The average magnetic susceptibility measured on

Fig. 7. Filtered magnetic map of Lofoten–Vesterålen with regional field removed. See Fig. 3, for geology and location.

Fig. 6. Five point moving average used to filter the aeromagnetic map from Lofoten–Vesterålen.





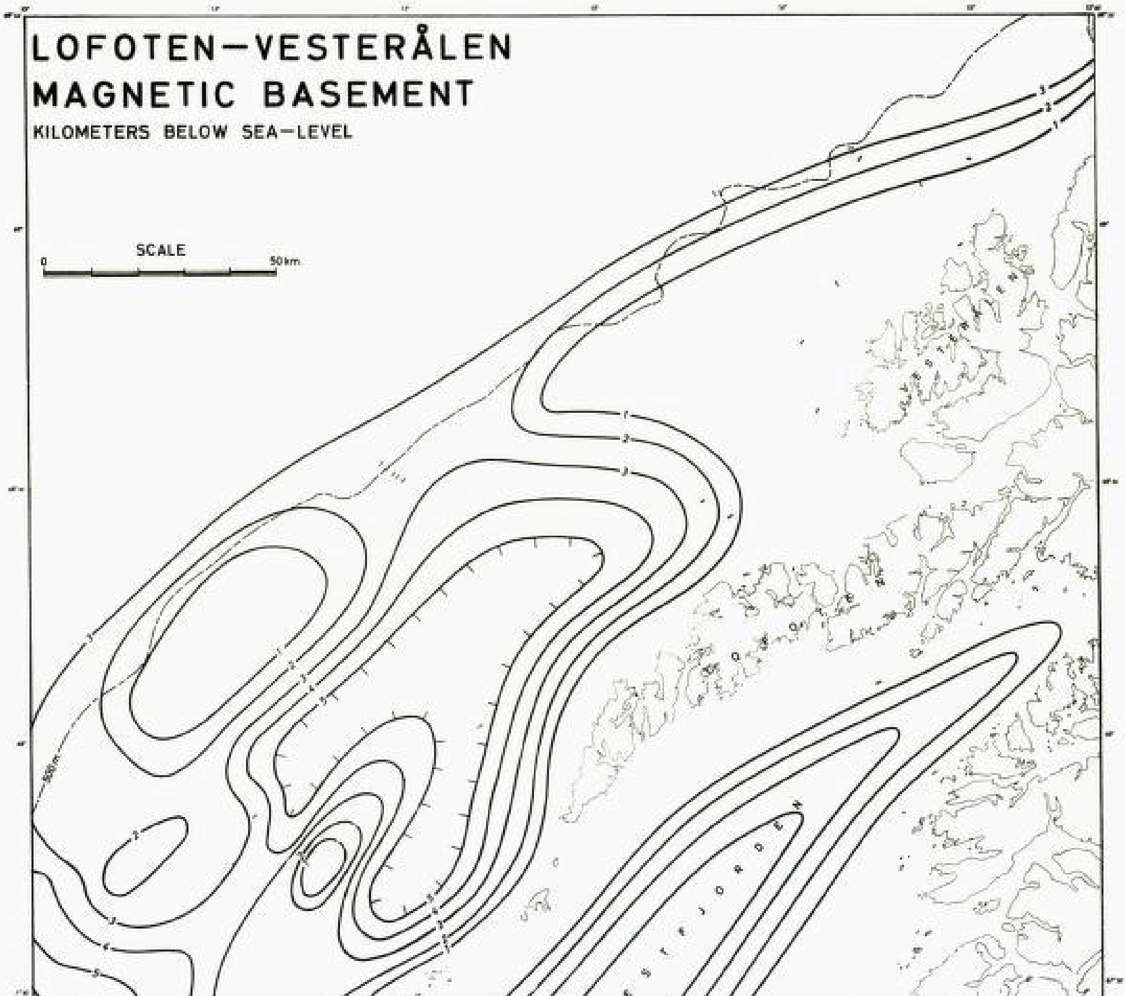


Fig. 8. Tentative magnetic basement map of Lofoten–Vesterålen.

900 rock samples is 0.003 cgs. The anomalies clearly demonstrate that the Lofoten granulites continue towards the south-west. The similar magnetic anomalies to the west show that there is another equally large mass of such rocks lying parallel to Lofoten on the shelf edge, with its top at some depth below the surface. This body is also causing the gravimetric 'shelf high' of Talwani & Eldholm (1972), and it is most likely a southward continuation of the granulites of Vesterålen. In between these two large masses of high-

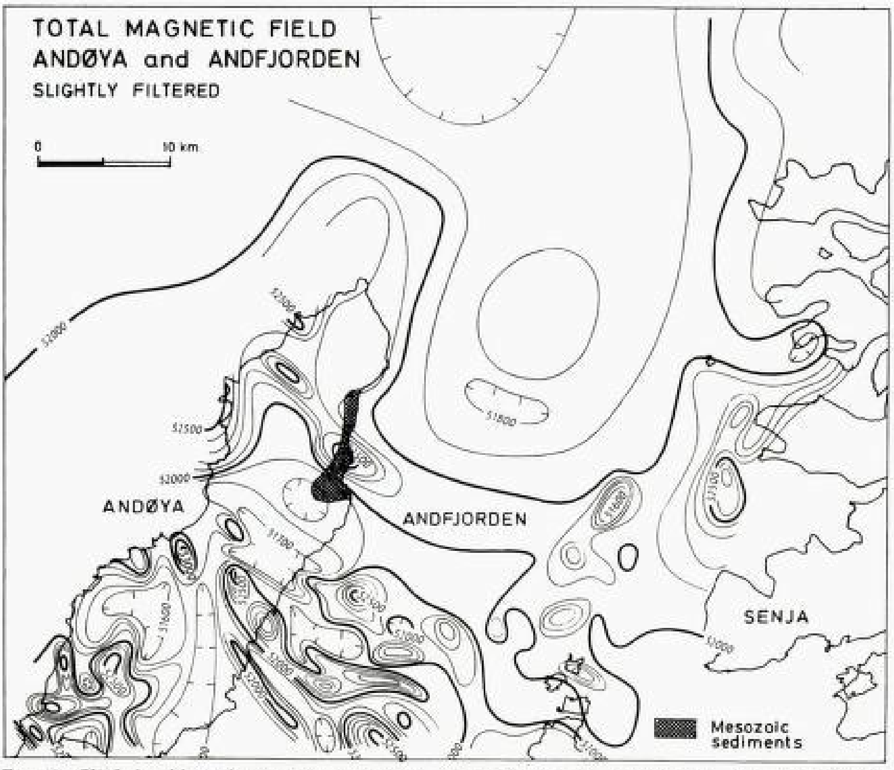


Fig. 9. Slightly filtered aeromagnetic map of Andøya and Andfjorden. See Fig. 3 for location.

grade metamorphic rocks there are indications of a much smaller body at approximately 67°45′N, 11°45′E as outlined by the 100 gamma contour.

A magnetic basement map is presented in Fig. 8. This is based on a careful study of the original records, but as the magnetic gradients seem to be, in part, associated with sources situated far below the top of the basement, the contours must be regarded as tentative. Three refraction profiles shot in the area give considerably shallower basement depth values (Sundvor & Sellevoll 1971). However, the contours in Vestfjorden are based on some good magnetic depth determinations ,and there is little doubt about the position of the 1 and 2 km contours in the rest of the area. The ridge associated with the small central granulite complex cannot therefore be questioned. It is also quite clear that the Lofoten massif and its twin to the west represent horst-like structures.

Andøya and Andfjorden (Mesozoic sediments)

Andøya is the only locality in Norway where Mesozoic sediments have been found. The occurrence has been known for more than a century, and a summary of the geology has been given by Ørvig (1960); see also Dalland (this volume). Upper Cretaceous sediments have also been dredged from the continental rise west of Andøya (Manum 1966). The area around Andøya

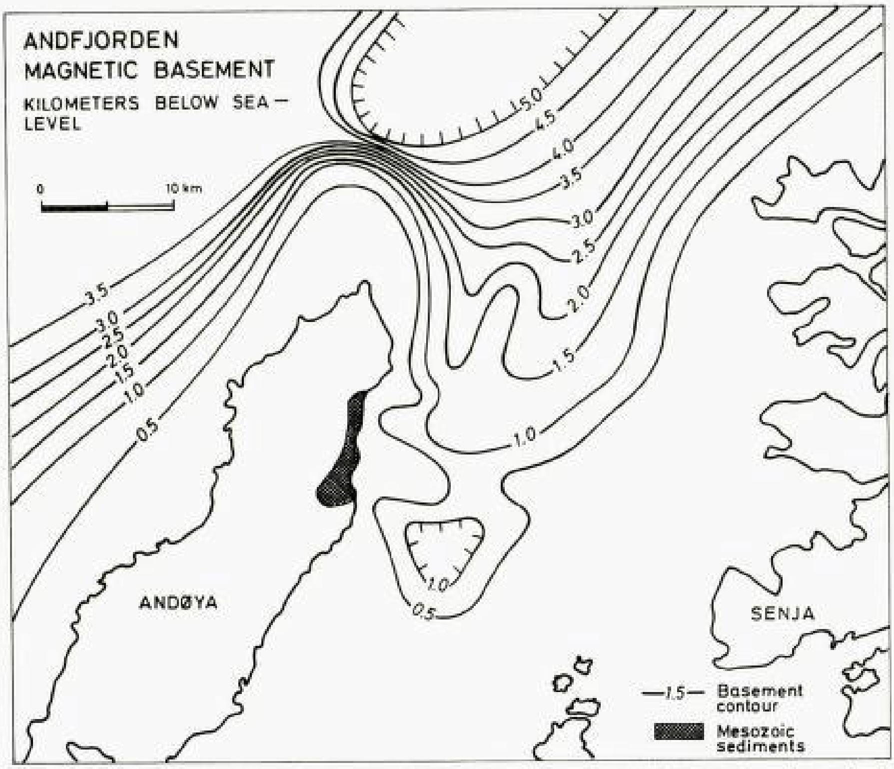


Fig. 10. Generalized contour map of the basement surface in Andfjorden based on depth determinations made on the original magnetometer records.

was therefore a natural target for an aeromagnetic survey after the first reconnaissance profiles had been flown over the shelf in 1962 and 1963.

Andøya and Andfjorden were covered with aeromagnetic measurements by NGU in 1964. The survey was flown E–W with 150 m 'ground' clearance and 1 km line spacing. A part of the resulting map has been published by NGU at a scale of 1 : 100,000 (Aeromagnetisk kart 1233). A somewhat simplified aeromagnetic map of the area is presented in Fig. 9.

One of the more important conclusions that could be drawn from the data was that the area of Mesozoic sediments is divided into two separate parts by a magnetic body with its top at or close to the surface. This ridge-forming body is probably the south-eastward continuation of the Trolltind gabbro, which is exposed to the north-east along the continuation of the anomaly in question (Dalland et al. 1973).

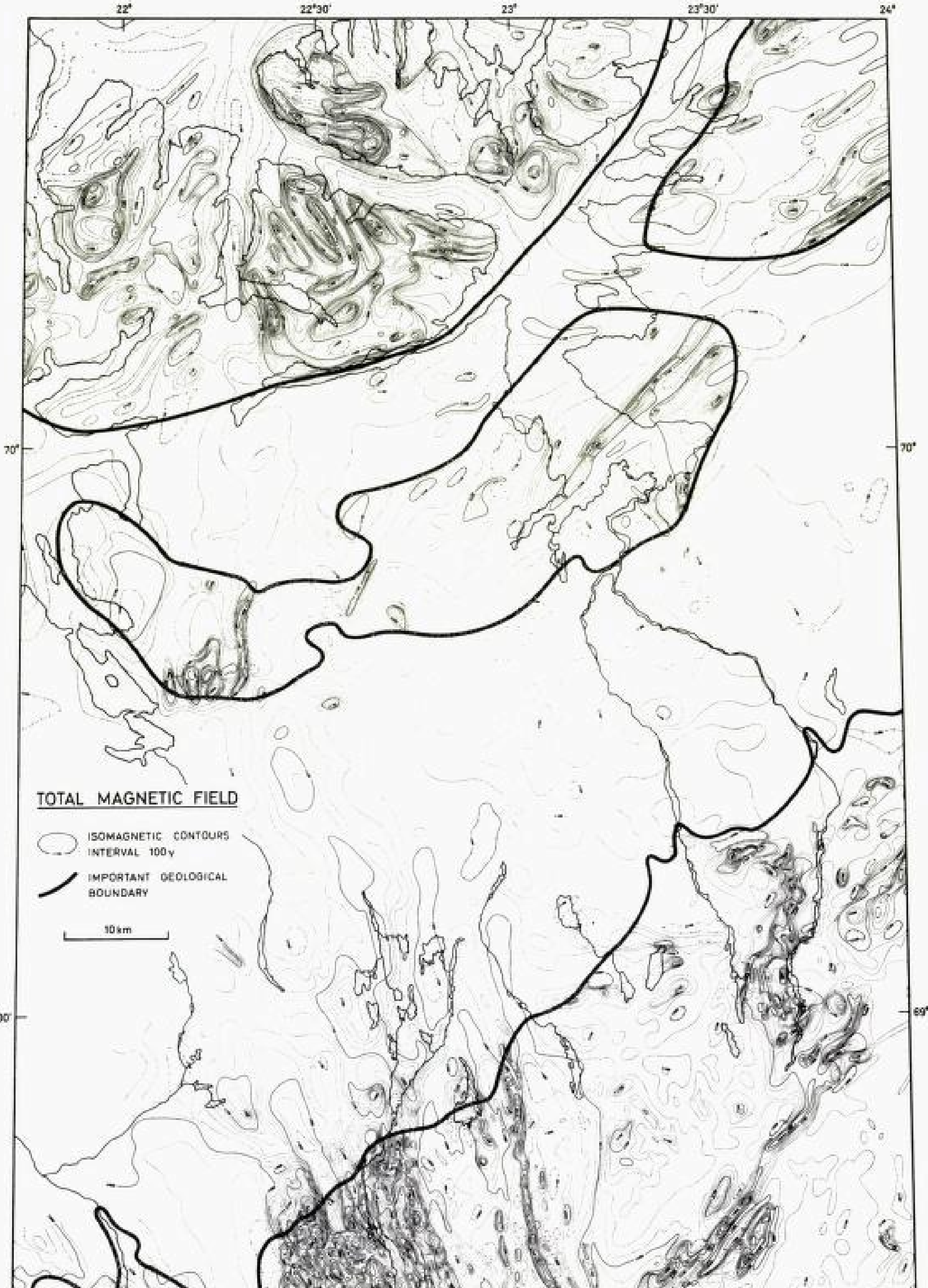
Fig. 10 shows a generalized contour map of the basement surface based on depth determinations made on the original magnetometer records. The depths to the east of Andøya were calculated by I. Aalstad in 1965, while the remainder were determined by the present author in 1970 on records from measurements made by NGU in 1969. The interpretation shows that the basement high associated with the Trolltind gabbro continues across Andfjorden. It also shows that Andfjorden is almost devoid of sediments, the water depth being around 500 m in this part of the fjord. Ten kilometres north of Andøya the depth to basement is more than 5 km. This is in good agreement with a depth of 5.1 km to seismic refraction basement as reported by Sundvor (1971, Profile R1). The rapid change in depth over such a short distance must evidently be due to the presence of a major fault.

Kvænangen–Finnmarksvidda (Caledonides)

As a part of the general aeromagnetic mapping of Norway, the area in question was covered with aeromagnetic measurements in 1959, 1962 and 1965. The survey was flown E–W with 100–150 m ground clearance and approximately 1 km flight-line spacing. The resulting isogam maps have been printed as 1 : 250,000 coloured maps (NGU 1971a & 1971b), a portion of which is reproduced in Fig. 11 at a reduced scale together with the principal geological boundaries between the Precambrian of Finnmarksvidda, the Caledonian rock complexes, the Precambrian windows of Alta–Kvænangen and Komagfjord and the Seiland gabbroic province (see Fig. 3).

The prominent magnetic anomalies to the south and south-east are associated with basic volcanics or 'greenstones' of Precambrian age. These anomalies are seen to continue into the Caledonian region in a weaker and more regular form, indicating that the greenstones continue northwards at some depth under the Caledonian nappe cover. Generally speaking these broad anomalies can be followed continuously until they join up with the sharp anomalies associated with the Precambrian greenstones of the Alta–Kvænangen window. The reason why these 'deep' anomalies can be so readily traced must be that the rocks of the Caledonian cover are practically non-magnetic. If this were not so they would give rise to sharp anomalies which would interfere with and eventually mask completely the deeper effects from the Precambrian greenstones. Over the Seiland gabbroic province there are strong and irregular magnetic anomalies due to magnetic bodies at or close to the surface. Most of this gabbro province, however, is virtually non-magnetic.

Based on this map (Fig. 11) and a study of the original magnetometer records, an interpretation of the form of the basement surface below the Caledonian nappe cover in this region has been produced (Fig. 12). Some major tectonic lines and areas showing signs of magnetic material at or close to the surface in addition to deeper material are also indicated. It should be borne in mind that each depth 'point' gives the approximate depth to the top of magnetic material, which in this case generally means Precambrian 'greenstone'. There is, of course, no guarantee against the greenstones being covered with non-magnetic Precambrian quartzites, and since the points where the Precambrian surface is known ('geologic control' in Fig. 12) sometimes refer to quartzites, the contouring is somewhat inconsistent. The question marks at 69°50′N denote that no anomalies exist so that no depths can be determined. It is therefore impossible to tell whether or not the basement deepens in this area. The maximum depth in the north-eastern part of Fig. 12 is in good



22°
22°30'
23°
23°30'
24°
70°
69°30'
TOTAL MAGNETIC FIELD
ISOMAGNETIC CONTOURS
INTERVAL 100 γ
IMPORTANT GEOLOGICAL
BOUNDARY
10 km

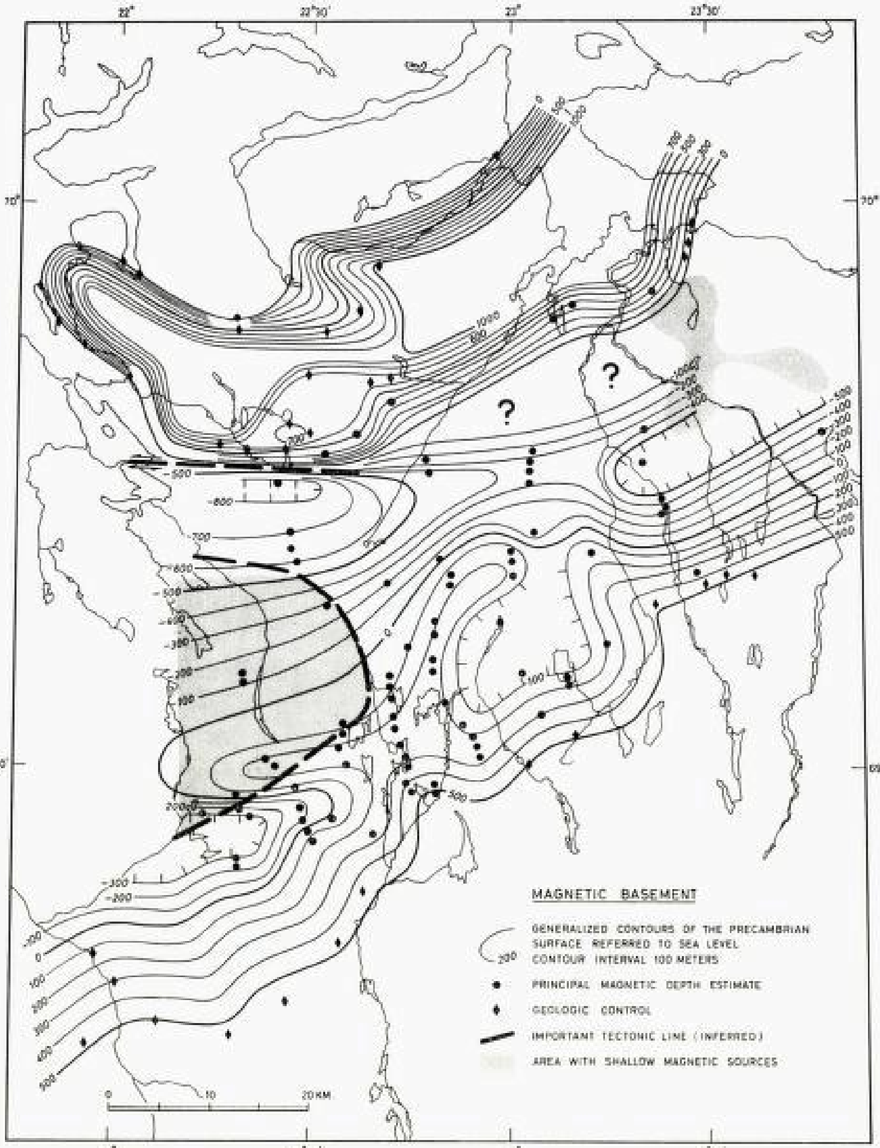


Fig. 12. Interpretation map based on a study of Fig. 11 and the original magnetometer records. The topography in the Caledonides of the region is around 500 m above sea level, meaning that the Caledonian cover is generally less than 1 km thick.

agreement with the tentative depth given for this area by Holtedahl (1918, p. 290).

The important tectonic lines indicated in Fig. 12 have been drawn after a careful inspection of the magnetic map (Fig. 11). The anomalies associated with the greenstones around 22°45′E in the southern part of the map are

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Fig. 11. Aeromagnetic map across the Caledonides in the Kvænangen–Finnmarksvidda region. See Fig. 3 for geological explanation.

seen to continue towards the NNW until they stop against an approximately NE-striking line around 69°30′N. Some weak features, however, are also seen to continue beyond this line. The anomaly at 22°35′E, 69°30′N seems to have been bent or displaced eastwards; it then curves back to resume its 'normal' shape around 22°30′E, 69°40′N. Strangely enough, this important tectonic line, which is based on the magnetic features of the Precambrian basement, has recently been found to coincide with the general tectonic trends within the Caledonian cover (Preliminary map Nabar 1834 III and K. B. Zwaan, pers. comm.).

In the area covered with Caledonian nappe rocks most of the magnetic anomalies originate from the underlying Precambrian basement. However, in some cases sharp anomalies originating from the surface rocks are also seen in addition to the anomalies from deeper sources. The anomalies can be clearly seen in the original magnetometer records, but only a few of them are visible on the contoured map (Fig. 11). These 'shallow' anomalies can generally be ascribed to small, outcropping ultrabasic bodies.

The anomaly at 22°27′E, 69°42′N is caused by a body situated approximately 800–900 m below sea level, while a little farther north the same source body is situated 700 m above sea level. This means that there is considerable vertical displacement over a comparatively short distance, indicating the existence of a large fault with at least 1.6 km of vertical throw. There is also an apparent horizontal displacement of 7–8 m, indicated with arrows in Fig. 12. The actual displacement along this fracture could in fact be explained solely by vertical movement if the greenstone plate is dipping gently westwards.

In conclusion, it can be stated that the cover of Caledonian nappe rocks is relatively thin in this area, the thickness being generally less than 1000 metres.

Varangerhalvøya (Late Precambrian sediments)

Varangerhalvøya was covered with aeromagnetic measurements in 1969. The survey was flown in an east-west direction at 750 m barometric altitude and with 2 km flight-line spacing except for the southernmost part, where the spacing was 1 km. The resulting magnetic map is represented in Fig. 13.

The anomalies are due to magnetic bodies in the Precambrian crystalline basement. The only exception seen on the map is the elongated anomaly in the north-western corner which is caused by a low-grade magnetite-bearing formation in rocks of the Caledonian nappe complex (R. Kvien & S. Siedlecki, pers. comm.). There is no sign in the map (Fig. 13) of shallow magnetic anomalies originating from the Late Precambrian–Eocambrian sediments on Varangerhalvøya. A careful inspection of the original records, however, reveals some weak features originating within 1 km of the surface. These faint anomalies can be traced from profile to profile and they are probably due to basic dykes, which are especially numerous in this area (S. Siedlecki, pers. comm.). The picture which emerges is very interesting (Fig. 14). It indicates

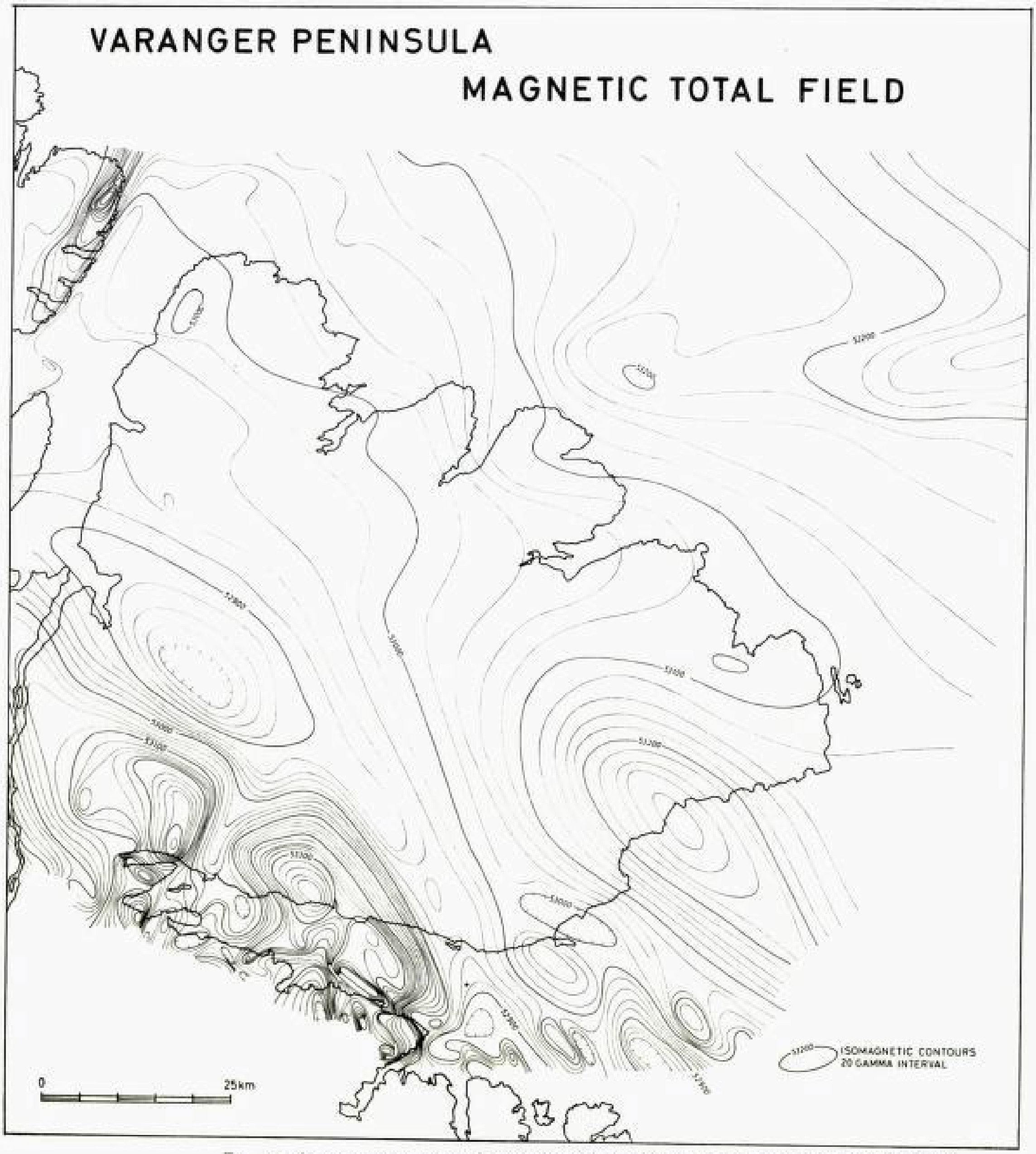

Fig. 13. Aeromagnetic map of Varangerhalvøya. See Fig. 3 for geology and location.

that there has been an eastward movement of the northern part of Varangerhalvøya. This is in good agreement with the conclusion reached by Roberts (1972, p. 37) on the basis of structural work in the area, that the complex Trollfjord–Komagelv thrust-fault mapped by Siedlecka & Siedlecki (1967) has an important dextral strike-slip or partly oblique-slip component of movement.

Since there are practically no magnetic effects from the sediments, it is

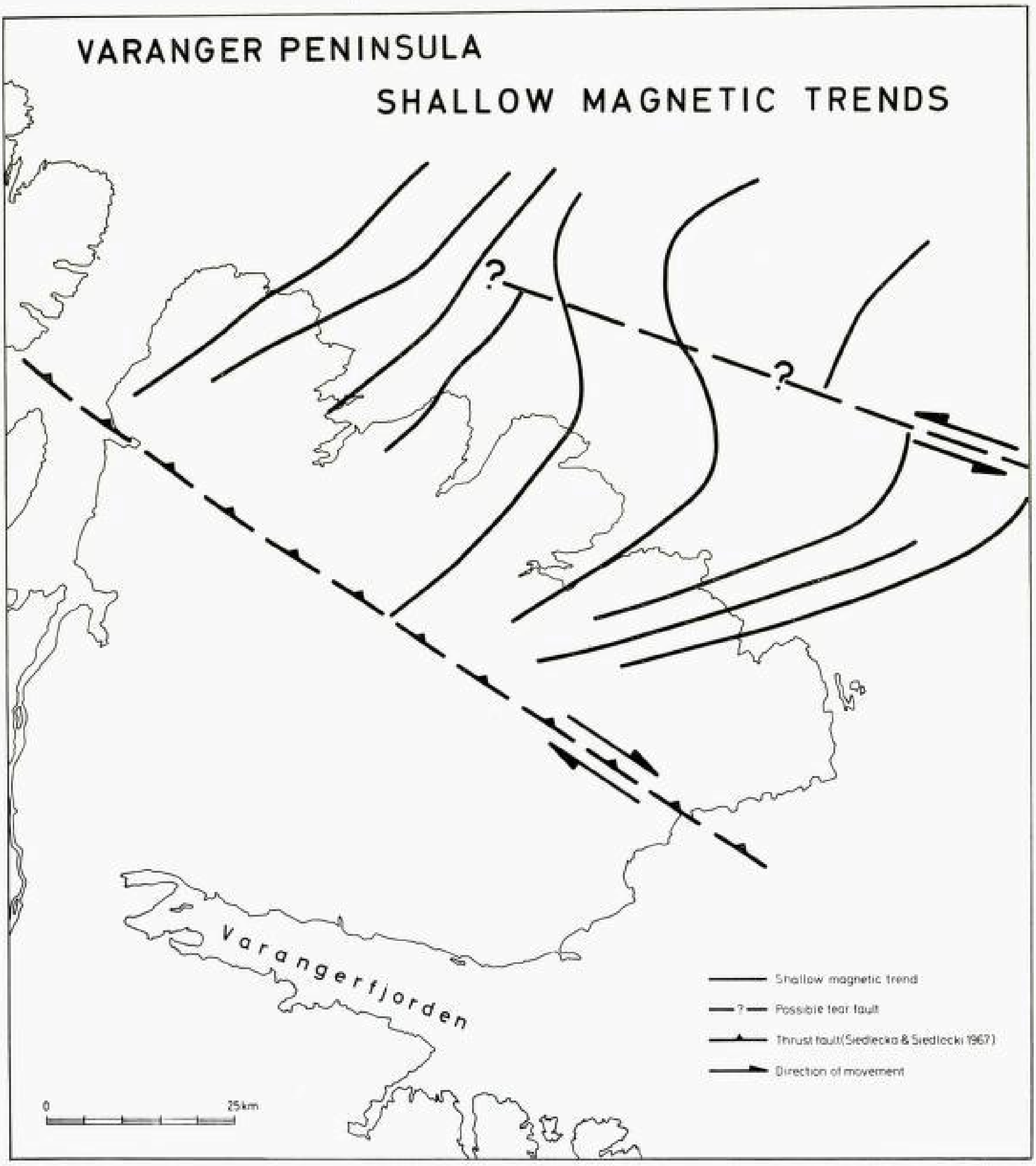

Fig. 14. Shallow magnetic trends over Varangerhalvøya (probably due to basic dykes) as seen from the original magnetometer records.

possible to trace the old Precambrian Shield beneath the sediments of Varangerhalvøya. An interpretation in terms of depth to magnetic basement (Fig. 15) shows that there are up to 7 km of non-magnetic Late Precambrian–Eocambrian sediments on Varangerhalvøya. It also shows that the shield is dipping gently below the sediments without any sign of a major vertical fault

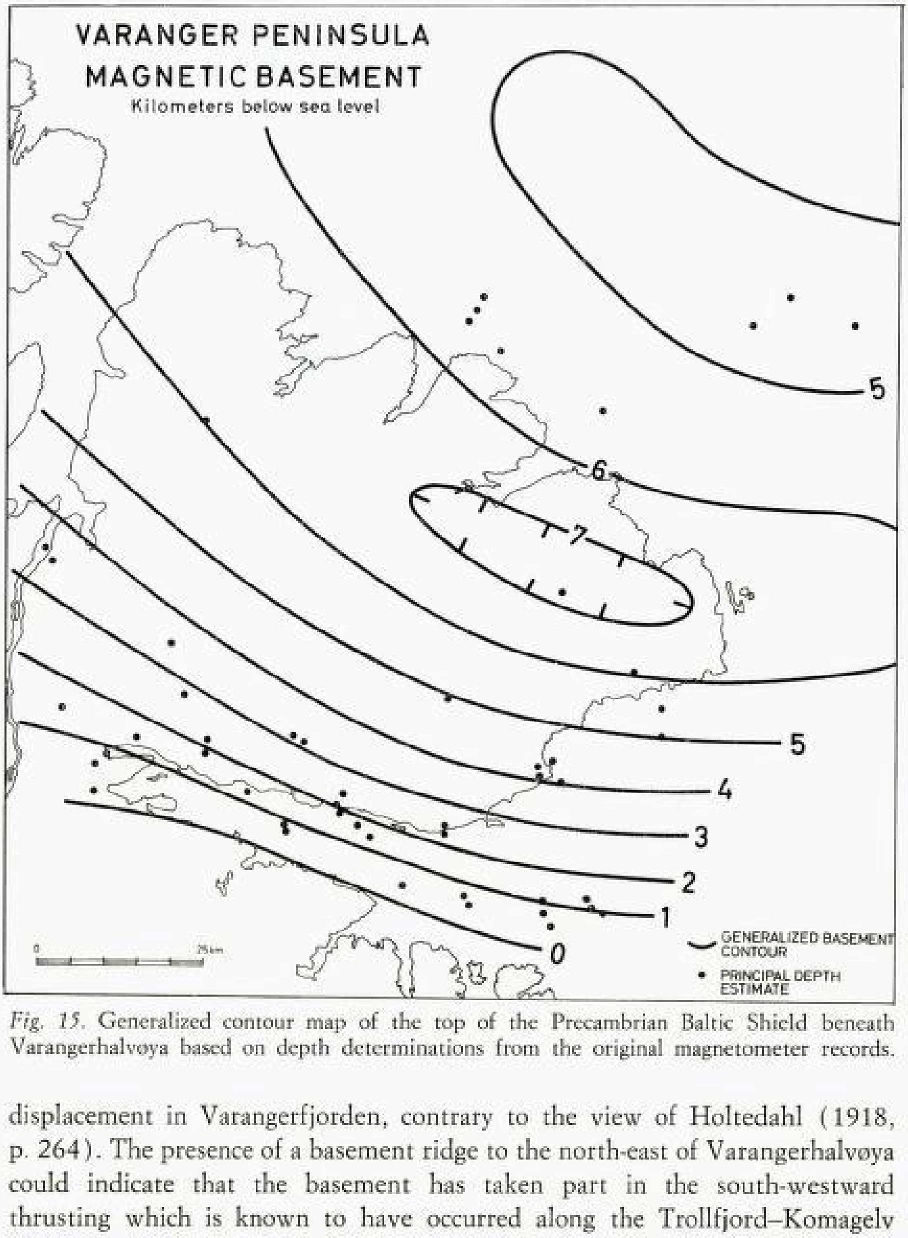

Fig. 15. Generalized contour map of the top of the Precambrian Baltic Shield beneath Varangerhalvøya based on depth determinations from the original magnetometer records.

displacement in Varangerfjorden, contrary to the view of Holtedahl (1918, p. 264). The presence of a basement ridge to the north-east of Varangerhalvøya could indicate that the basement has taken part in the south-westward thrusting which is known to have occurred along the Trollfjord–Komagelv thrust-fault.

The shelf between Andøya and Bjørnøya (69–74.5°N)

A simplified magnetic map of the region with a 100 gamma contour interval is presented in Fig. 16. The area to the west of a line running from 71°N, 24°E to 74°N, 17°E was flown in 1969 at 200 m altitude in a NW–SE direction with 4 km line spacing and Decca navigation. The area to the east of this line was flown in 1970 and 1971 at 500 m altitude. The flight path followed

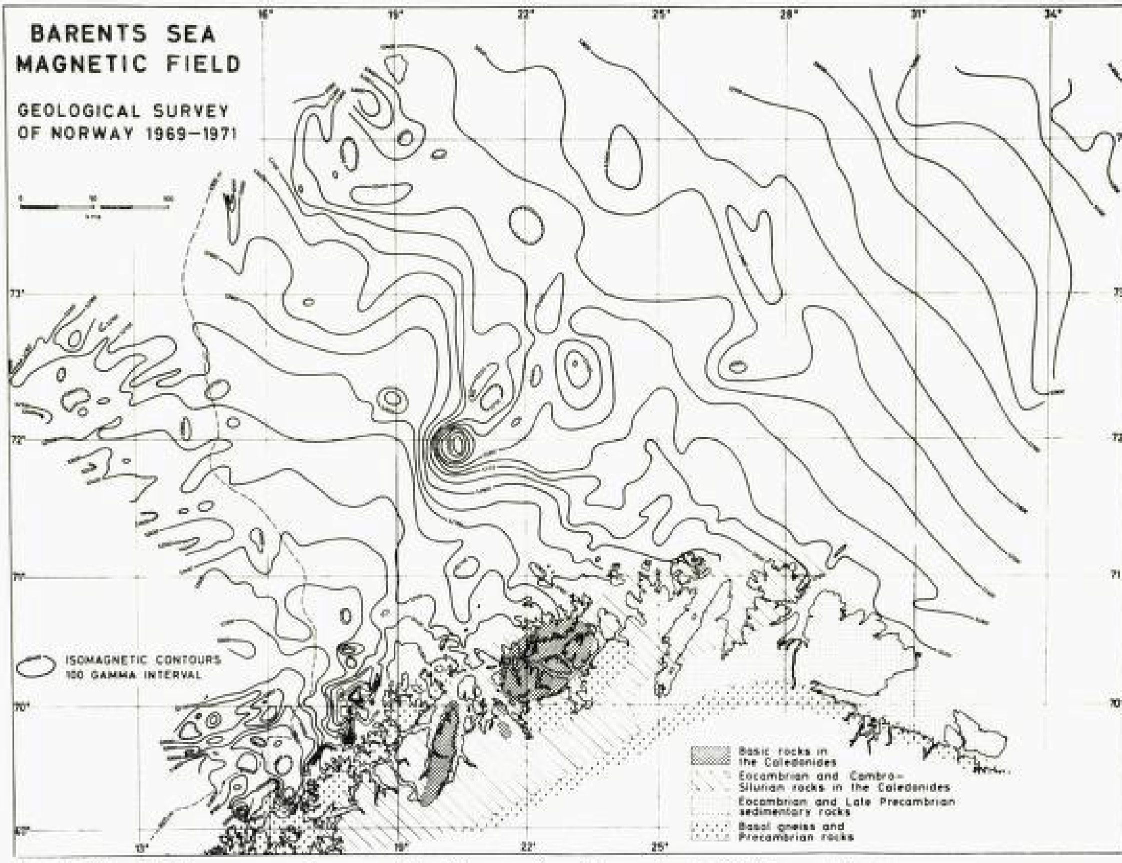

Fig. 16. Simplified aeromagnetic map of the Barents Sea. The regional field has not been removed.

Decca lanes almost perpendicular to the coast, resulting in a flight-line spacing of 3–4 km close to the coast, gradually increasing to up to 15 km at the outer end of the profiles.

The magnetic picture in Fig. 16 shows linear oceanic anomalies to the west of the 1000-metre water depth contour. The anomalies are due to semi-outcropping sources and they stop abruptly against the shelf where there is a discordant line of anomalies following approximately the shelf edge (1000-m water depth contour). This indicates the presence of a major fracture zone along the shelf edge which would coincide with the de Geer fracture zone (transcurrent fault) postulated by e.g. Wilson (1965) and Harland (1969, p. 841). To the east of this there is a magnetic quiet zone over the western part of the shelf before one gets into an irregular magnetic pattern trending in several directions and clearly demonstrating that the basement is of continental type. In the eastern part of the shelf the magnetic picture is extremely quiet, indicating large thicknesses of non-magnetic material. The NW–SE trending contours only reflect the shape of the normal magnetic field, which has not been removed in the present case.

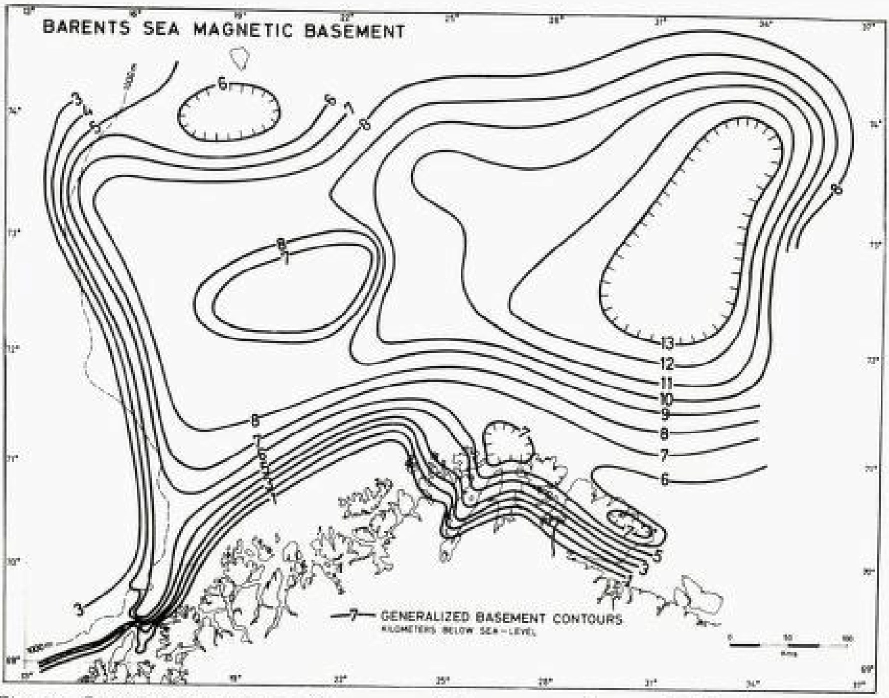

Fig. 17. Generalized contours of the magnetic basement surface, probably the top of the crystalline Precambrian Shield.

It is difficult to tell what is the actual cause of the magnetic anomalies, but from the correlations outlined in the preceding chapters it is considered that magnetic bodies in the Precambrian Shield and some Caledonian gabbros are likely to provide the most important sources of the anomalies measured over the shelf. Some weak effects can also be expected from Eocambrian and Lower Paleozoic rock complexes. In addition, Tertiary intrusions are known from Svalbard and such intrusions would of course give rise to strong anomalies if they happened to occur in the sediments of the shelf.

When trying to map the basement surface in the Barents Sea, some special problems arise in view of the presence of several kilometres of non-magnetic, Late Precambrian–Eocambrian sediments. The Caledonian belt of metamorphic rocks is also practically non-magnetic except for some ore-bearing formations and ultrabasic bodies. However, in addition to the anomalies originating from bodies within the basement, it is evident from maps and especially from the records that there are additional anomalies which must be due to shallower sources. This is also supported by spectral analysis of the data.

Fig. 17 shows the resulting contoured surface of the basement, which is probably the old Precambrian Shield. The general trend is almost east-west with prominent trends striking NNE–SSW and NNW–SSE. The greater depths to the east are based on a few depth estimates in the area of very weak anomalies. The estimates are probably on the shallow side, which would mean that they are in reasonably good agreement with the results published by

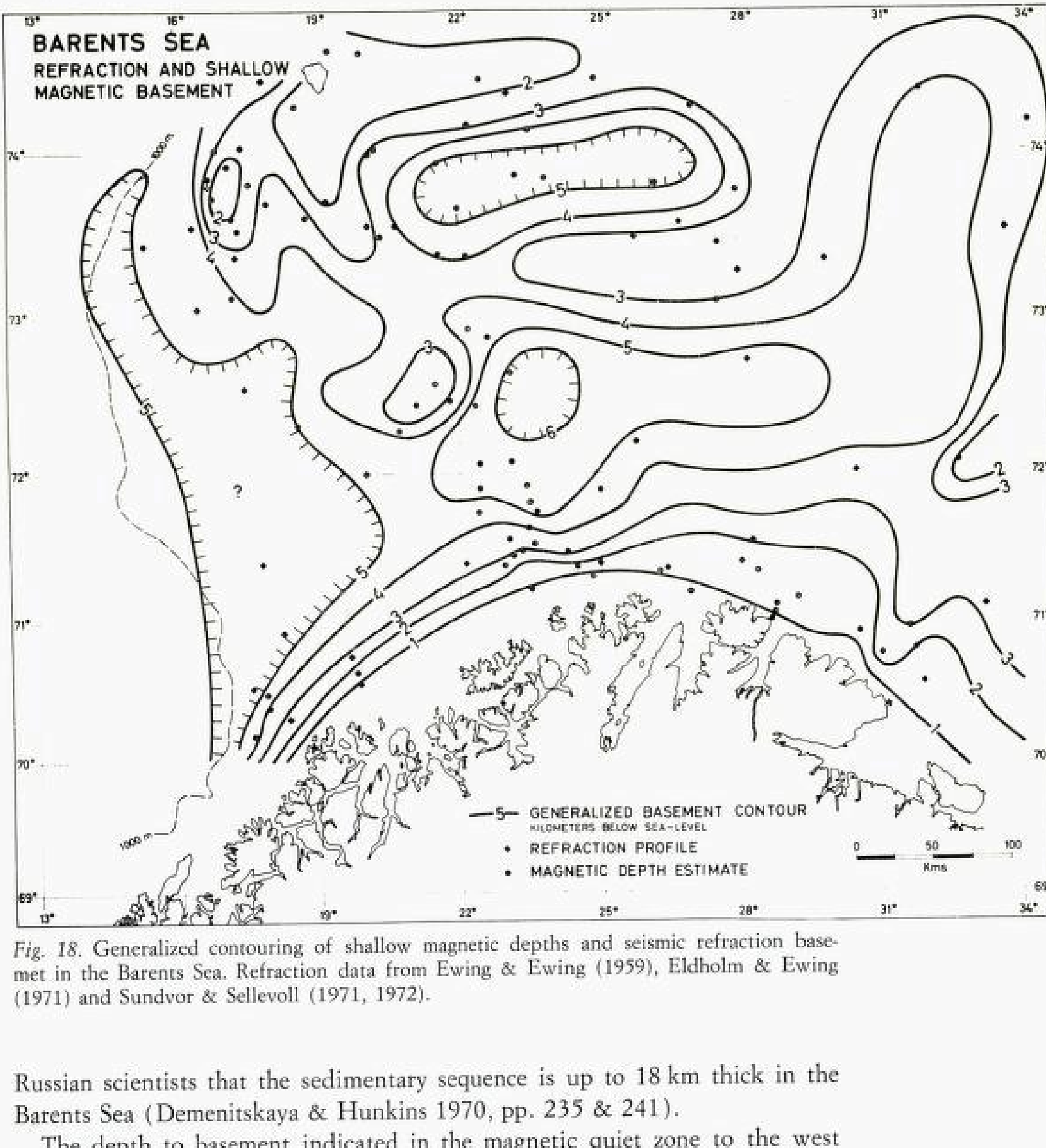

Fig. 18. Generalized contouring of shallow magnetic depths and seismic refraction basemet in the Barents Sea. Refraction data from Ewing & Ewing (1959), Eldholm & Ewing (1971) and Sundvor & Sellevoll (1971, 1972).

Russian scientists that the sedimentary sequence is up to 18 km thick in the Barents Sea (Demenitskaya & Hunkins 1970, pp. 235 & 241).

The depth to basement indicated in the magnetic quiet zone to the west (between the large basement high and the shelf edge) represents a minimum value. The depth is probably much greater than the 8 km shown in Fig. 17.

Contouring of the shallower depths shows that they are in fairly good agreement with depths to acoustic refraction basement (velocities higher than 5 km/sec.) as published by Ewing & Ewing (1959, Profile F3), Eldholm & Ewing (1971) and Sundvor & Sellevoll (1971, 1972). From work on Varangerhalvøya (e.g. NGU Rapport 227, 1959) the Late Precambrian–Eocambrian sediments are known to give velocities around 5 km/sec. It is therefore reasonable to assume that the shallow magnetic sources define a Palaeozoic basement. This basement is presented in Fig. 18, where the published refraction data have also been incorporated.

In the western part of the map (Fig. 18) there is no sign of two basements, possibly meaning that the oldest (Late Precambrian) sediments are restricted to the eastern region. The question mark to the west means that no shallow depths have been found in this region. The shallow depths around 74°N, 17°E could of course be due to Tertiary igneous material in the sediments. The deep basins around 23°E should be examined with some care since they are situated rather close to the deeper basement surface; consequently it will be difficult to differentiate between the two basements in these areas. These basins in the upper basement could, therefore, possibly represent highs in the lower basement instead.

The north-south and east-west trending ridge over Bjørnøya represent a continuation of the basement ridge (Hecla Hoek) along the western coast of Spitsbergen, which has been subject to Caledonian metamorphism (Åm, in press). This indicates a connection between the Caledonides of Spitsbergen and Norway. The NNE–SSW trending basin to the east (32°E) could also represent a continuation of the Norwegian Caledonides in that direction.

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Remarks about Displacements of the Old Rift Systems in the North Sea Area

A contribution arising from discussions following two of the papers presented, including the background for questions about observed lateral displacement in the area

OLE BRUUN CHRISTENSEN

The old rift valley systems of West Europe have been discussed for a long time. Some earth scientists interpret the exposed parts of the rift systems (Oslo Graben, Rhine Graben, etc.) as more or less local structures. Others interpret the grabens as elements in a large-scale framework (e.g. Mittelmeer–Mjøsen Zone), where links have been buried below thick sediment cover or even destroyed. However, the origin of the graben structures is of Variscan age, and there is no reason to believe that later tectonic disturbances in the North–West Europe region were of sufficient force to destroy parts of such large-scale structural elements.

Two of the papers presented (by Dr. Naylor (Whiteman et al.) and Dr. Ziegler, respectively) give new knowledge about rift valley fragments in the Central and Northern North Sea. In addition, these fascinating papers give us splendid illustrations on the geological developments in the North Sea area. It is, however, difficult to imagine that the whole area between the Hercynian Front (Wales–London–Ardennes platform), the Fennoscandian Border Zone, and the continental margins should not be more intensively structured by the Alpine Orogeny at different geological times. The major part of this area is composed of a rather thin granitic layer. Considering the strong disturbances and major block rotations in Southern Europe together with an undulated pattern, e.g. of the Alpine ranges, some neutralizing horizontal dislocations could sometimes be expected into the area in question. Have such extensions between the Alpine Geosyncline and the continental margin in the north and north-west been recognized in any part of the North Sea area?

In connection with our geological and geophysical interpretations along and closely south of the Fennoscandian Border Zone we have, for the best explanation of the development of the region, used a neo-tectonic model acting during the Jurassic and Lower Cretaceous. The model is closely related to that of the Recent wrench-fault tectonics in South-West California. There, the granitic layer is relatively thin, and the distances to the continental margin and to the actual orogenic disturbance are rather short. Furthermore, the structural trends and the general scale of the structures are nearly the same for the two areas. In the North Sea area, the trends of the wrench-faults are demonstrated on a small scale by displacements in many of the salt structures. On a regional scale, horizontal dislocations have moved parts of an originally continuous

Mittelmeer–Mjøsen Rift System as segments into the Central and Northern Sea. These mainly north-western movements of micro-plates seem to have terminated in the lowermost Cretaceous. They were probably activated by the initial Alpine Orogeny from the south or south-east along the Eastern European Platform, and by an initial and partial opening of the North Atlantic in the north-west and west. The model provides reasonable explanations for local problems in connection with the Fennoscandian Border Zone, geological distributions in the North Sea Basin, and paleogeographical features in common with Polish and south-west Russian areas, at least during Permian into Upper Jurassic.

During the Upper Cretaceous and the Tertiary, the Fennoscandian Border Zone was affected by a rather complicated tectonics, mainly appearing as compressional features, the northernmost of which include strong isostatic movements. Close to the Fennoscandian Border Zone, younger erosion has often removed important stratigraphical sequences. The partly tensional Cimmerian features are superimposed by the younger disturbances. Concerning the possible effect of fracture-zones (resulting from sea floor spreading) penetrating into the continental areas during the opening of the North Atlantic in Uppermost Cretaceous and Lowermost Tertiary times, it is remarkable that Cimmerian fractures in the north, along the Fennoscandian Border Zone in Denmark, have not been renewed to any appreciable extent.

Ole Bruun Christensen, Geological Survey of Denmark, Thoravej 31, DK–2400 København NV, Denmark