UNITED NATIONS ECONOMIC AND SOCIAL COMMISSION FOR
ASIA AND THE PACIFIC

COMMITTEE FOR CO-ORDINATION OF JOINT PROSPECTING FOR
MINERAL RESOURCES IN SOUTH PACIFIC OFFSHORE AREAS
(CCOP/SOP AC)

TECHNICAL BULLETIN

No.3

SYMPOSIUM ON PETROLEUM POTENTIAL IN ISLAND ARCS, SMALL OCEAN BASINS, SUBMERGED MARGINS AND RELATED AREAS

Held in Suva, Fiji, 18-21 September 1979

New Zealand Department of Scientific and Industrial Research, Wellington, 1980

This publication should be referred to as:
United Nations ESCAP, CCOP/SOPAC Tech. Bull. 3

All communications relating to this and other publications of CCOP/SOPAC
should be addressed to:
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Cataloguing in Publication Data

SYMPOSIUM on Petroleum Potential in Island Arcs, Small Ocean Basins, Submerged Margins and Related Areas (1979 : Suva)

Symposium on ... Areas, held in Suva, Fiji, 18–21 September 1979 : [proceedings] — Wellington : DSIR, 1980

(Technical bulletin — United Nations Economic and Social Commission for Asia and the Pacific, Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas, ISSN 0378-6447; no. 3)
ISBN 0-477-06667-4

I. United Nations Economic and Social Commission for Asia and the Pacific. II. Committee for Co-ordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC). III. Series

553.98(265.7:210.7):551.35(265.7:210.7)

Published on behalf of ESCAP, who funded the symposium, by the Department of Scientific and Industrial Research, New Zealand.

Printed in New Zealand by Wright and Carman Ltd, Upper Hutt, on behalf of the Australian Development Assistance Bureau as an Australian contribution to CCOP/SOPAC.

ISBN 0-477-06667-4

FOREWORD

On behalf of the Government of Fiji, I extend to all of you our very warm welcome.

We consider ourselves privileged and honoured to be able to host this gathering of eminent men and women of science; We thank the Committee for Coordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas, better known as CCOP/SOPAC, the UNDP, and ESCAP for their efforts in initiating, organising and funding this Symposium.

The Symposium, entitled 'Petroleum Potential in Island Arcs, Small Ocean Basins, Submerged Margins and Related Areas, marks a milestone in the scientific and economic development of the South Pacific, as I understand that all the areas of the southwest Pacific may be classified as island arcs, small ocean basins, or submerged margins. Because of these geologic features and because 200-mile exclusive economic zones have been delineated by many of our island nations, this Symposium is of special relevance to us. Our small island nations depend exclusively on imported petroleum and are therefore very vulnerable to the inflationary effects of all the various recent price increases. It is our hope therefore that this Symposium will result in increased prospectivity of our region and give us some hope that we too can perhaps become oil-producing countries.

The petroleum potential of geologic structures characteristic of this area has been little understood. Except in Indonesia, and perhaps the Philippines, island arcs and basins have seen very little exploration. The better understanding of plate tectonics, the improved economics of drilling in deep water, recent successes in the Philippines, the impending petroleum shortage, the astronomical price increases, and the political vicissitudes of traditional oil-producing countries have refocused attention on the petroleum potential of this hitherto unattractive and unknown area for exploration. Until recently many companies would not give a thought to exploring in the southwest Pacific, because they considered this area too young and too volcanic. However, systematic geologic mapping has revealed that these areas are not entirely volcanic but have large areas of marine sedimentary rocks, source rocks, and the heat necessary for the hydrocarbon transformations. The discovery of a genuine oil seep in Tonga in 1968 and known seeps and gas discoveries in Papua New Guinea have further increased interest in this region.

It is the policy of the Fiji Government to encourage petroleum exploration. Towards this end, it enacted in 1978 a new Petroleum (Exploration and Exploitation) Act that replaces the outdated Oil Mining Act of 1915. While still catering for onshore exploration with which the 1915 Act was mainly concerned, the 1978 Act provides the legal framework for the regulation and control of offshore petroleum activities.

We are aware that in many other countries Governments are attempting to play a bigger role in the petroleum industry. While we are following these developments with interest, we do not feel that Fiji at the moment is in a position to undertake such an enlarged role. Recognising that our economy is small, that we lack the necessary expertise, and that we have other development priorities, we prefer to see private companies undertake oil exploration, for which we have granted them generous concessionary terms. We have backed these terms by our record since independence of a stable political system, a stable Government, and a stable economy. Having become independent towards the end of the development decade of the 1960s, and not professing any particular political or economic ideology except the Pacific way of moderation and compromise, we in Fiji are prepared to be pragmatic about oil exploration.

Four oil exploration licences over a total of 31,000 km^2 have been granted to overseas companies on attractive terms. These terms include a partial tax holiday while investment is being recouped, guaranteed stability of tax rate, waiver of export tax, interest withholding tax, and customs duties on exploration equipment, liberal foreign exchange regulations, and other Government undertakings. While other countries receive up to 21 % on sales as royalties, our legislation has fixed this at a moderate amount of from 10 to 12%.

While other proven petroleum province countries are able to demand a 60% tax rate from the first year of operations, we in Fiji would be pleased with a 60% return after the companies have recouped their capital investment. Some people have criticised our concessions as being too generous to the exploration companies. But we believe that until discoveries are made it is to our benefit to encourage active exploration and early drilling - and this cannot be done by onerous terms. However, we are not just giving away licences on soft terms, but prefer to give careful and serious consideration to each application on merit depending on the work programme and conditions offered to Government. Seven applications for four other areas are under consideration at the moment.

We are cautiously optimistic that the excellent exploration programmes being carried out at present will continue to produce positive results. We are aware that some innovative new techniques such as Curie Point isotherm interpretation and offshore geochemical sampling have been used to high grade our prospects, and we are looking forward to a drilling programme in our offshore areas in early 1980.

The 1980 drilling programme cannot avoid highlighting the recent Gulf of Mexico blowout, which followed by a decade the Santa Barbara blowout. In spite of these calamities, which underscore the weaknesses of man and nature, the oil industry has to be congratulated for its excellent record in ensuring the safety of rigs and petroleum operations at sea over the past 30 years. And I would not be remiss in expecting that the companies will do their utmost to maintain their enviable safety record when planning drilling programmes in our south seas islands.

The excitement of a seismic ship negotiating our reef-infested waters and of a first-ever drill ship in our waters should not obscure the fact that much of the critical work is done in quiet far-off laboratories by men and women like yourselves. For it is upon your design and techniques and your analysis and interpretation of data that the major decisions depend. If it were not for the refinements in the state of your science, it is quite certain that jungles, polar climates, deep seas, apparently sterile basins, and difficult overthrust belts would not be the foci for the intensive exploration that they are now experiencing and for the renewed hopes for fossil fuel upon which the world is so dependent.

For so long, the South Pacific has been the subject Of the stereotyped image of balmy beaches, guitar-strumming males, and comely South Pacific beauties. What has been less popularised is the growing concensus among scientists in every discipline that the South Pacific is one of the new exciting frontiers of science. It is our every hope that this petroleum symposium will underscore this growing consensus and will contribute to an enhanced understanding of the petroleum potential of our island arcs, small ocean basins, submerged margins, and related areas.

With these words I hereby declare the Symposium open.

William J Clark

Minister for Lands and Mineral Resources, Fiji
Suva, 18 September 1979.

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EXPLORATION FRONTIERS — A QUARTET OF CHALLENGES

A V MARTINI

This meeting in Fiji is a sign of the times. To find new oil and gas producing provinces today, we must go further afield, to more remote areas, examine more complex and unattractive geology, apply more advanced technology. In the last six years we have seen worldwide shortages of crude oil, rapidly escalating prices, and confusion and dismay on the part of the peoples and governments of the world, which had become accustomed to a seemingly limitless supply of cheap energy.

In retrospect, the reasons for this apparently abrupt change in the energy supply of the world seem clear. Figure 1 shows the rate of world oil production and reserves discovered since 1930. In the years immediately before and after World War II, tremendous reserves of new oil and gas were discovered and developed. In the early 1950s, discovery rates exceeded production by factors of 4 or 5. Given the large reserves which could be developed at relatively low cost, petroleum rapidly displaced traditional fuels such as coal from existing markets and created new markets of its own.

The rate of growth of production has been phenomenal. In the last 10 years we have produced more oil than in the 100 years before. By the early 1970s, however, annual production exceeded reserves discovered, and total reserves began to fall. In the largest producing and consuming country of the time, the United States, production itself began to fall, and for the first time, the U.S. became a significant importer. As the expansion of producing capacity became more difficult with falling reserves, the surplus production which had characterized the market for decades disappeared. The world petroleum industry moved from a buyer's to a seller's market.

As to prices, Adam Smith and supply and demand won again. So long as there was a surplus, nothing could move prices up, including the formation of OPEC in the late 1950s opposing a move by the producing companies to reduce posted prices. In my view, nothing will stop further increases in the price of oil until it reaches a level where high-cost energy sources, such as coal, tar sands, and oil shale can compete.

It seems to me that as earth scientists our role and duty must be to accurately assess the potential for further discovery of oil and gas and to find them as quickly as possible—this to ease the world's way over the next 20 years as our civilization moves to new energy sources. To do this successfully we must advance simultaneously our knowledge and effectiveness on four frontiers:

1. In understanding how oil and gas have been formed and trapped.
2. In applying this knowledge to the assessment of the petroleum potential of the world's sedimentary basins.
3. In learning to recover more of the oil that we have discovered and will discover.
4. In educating people and governments on the occurrence of oil and gas and the economic and political conditions necessary for exploration to be effectively carried out.

We are fortunate that beginning in the mid 1960s a series of significant advances in the science and technology of exploration supplied the tools for the job. The most important of these in my view are:

1. Advances in geochemistry.
2. Digital seismic recording and processing.
3. Development of the concepts of plate tectonics.

A few years ago one of our geologists wrote this equation to help keep our thinking focused on the important things in the formation and accumulation of oil and gas.

(Adequate Source × Proper Thermal History) × (Effective Migration) × (Adequate Reservoir) × (Sufficient Traps) = Commercial Accumulation.

Note that if any one factor is zero, the size of the accumulation is zero.

For many years the detection and prediction of reservoirs and traps have occupied a major part of the explorationist's time.

Today, migration is the process least understood and is the one on which we can make most progress. Advances in the understanding of source rocks and in the significance of their thermal history have tremendously improved the efficiency of exploration, particularly in relatively unexplored frontier areas. These advances supplied one of the missing links in understanding the complex chain of processes leading from petroleum generation to migration and entrapment.

Recognition of variations in the type of organic material (organic facies) and development and application of the techniques for measuring paleo temperatures have allowed the prediction of the type of oil and gas that can be expected in a new basin, the depth at which it was generated, and perhaps most important, the depth at which it can no longer be expected.

Figure 2 illustrates the interrelationship between type of organic material, temperature, and hydrocarbons. The original organic source material can be divided into four types or facies. Facies 'A' is typically organic matter deposited in stagnant lakes and yields high wax crudes. Facies 'B' is algal, marine material deposited under anoxic conditions (and incidentally is considered responsible for over 90 per cent of the world's crude oil). Facies 'C' is terrestrial plant debris deposited in nearshore marine, mildly oxic conditions, and 'D' is high carbon, coal-like material generally found in well-oxygenated, open marine sediments.

These four source-material types are thus originally deposited in different basinal positions. They yield different proportions of oil, condensate, and gas as they undergo increased heating, as shown on the temperature column in Figure 2. The degree to which rocks have been heated can be determined by the rank of coal, the color of spores contained in the rocks, and by measurements of vitrinite reflectance. This permits us to determine from samples from wells whether we have encountered (1) source rocks that have never been locally buried deeply enough to have generated hydrocarbons, (2) source rocks that are now, or have been, at the right temperature for generation, (3) ex-source rocks that have given up all their movable hydrocarbons (they are beyond the oil and gas deadline), or (4) rocks with no latent, present, or past generative capacity. A key last step in using this information is to determine, usually from seismic data, how typical the data from a given

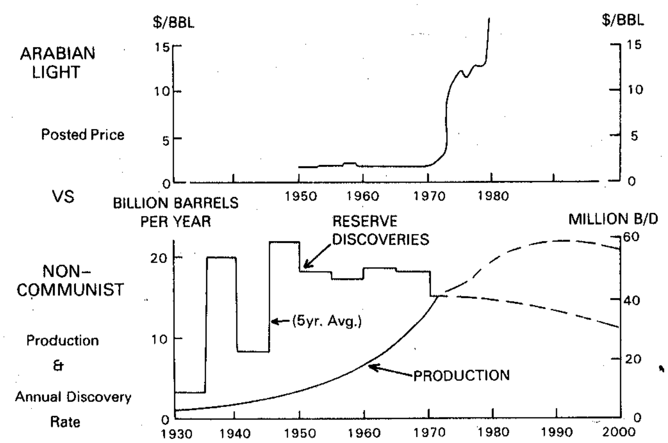

Figure 1. Price of crude oil vs rate of production and discovery.

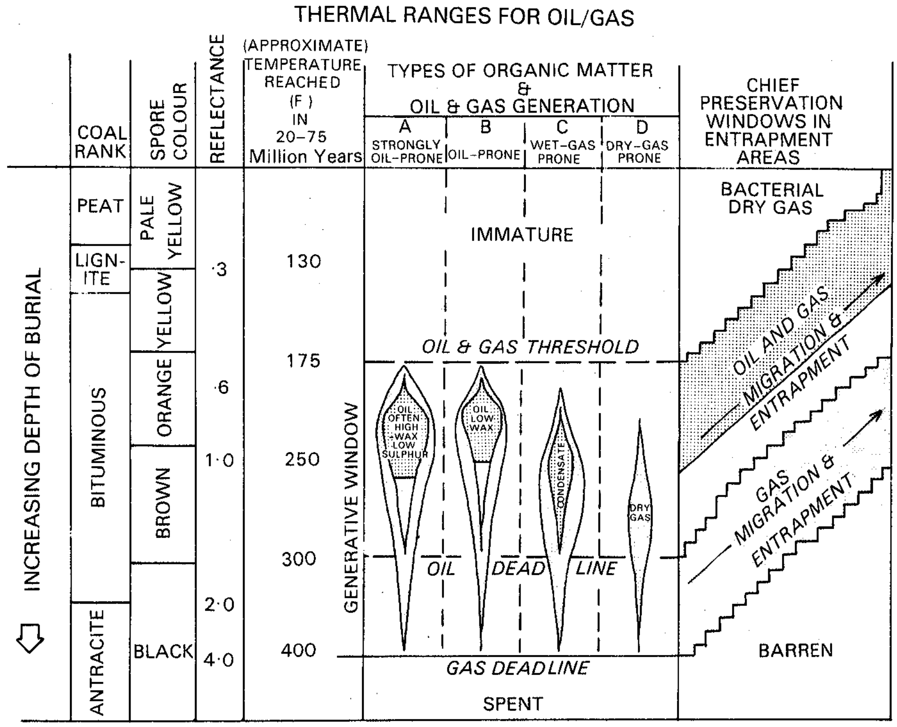

Figure 2. Maturation ranges for oil/gas generation. For explanation of facies A,B,C,D, see text.

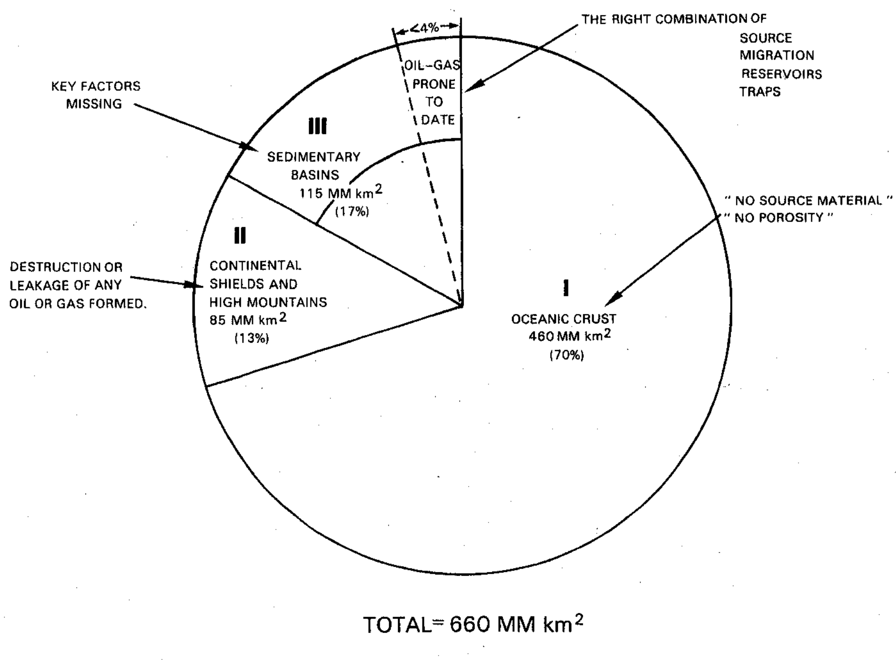

Figure 3. Hydrocarbon potential of the world's surface.

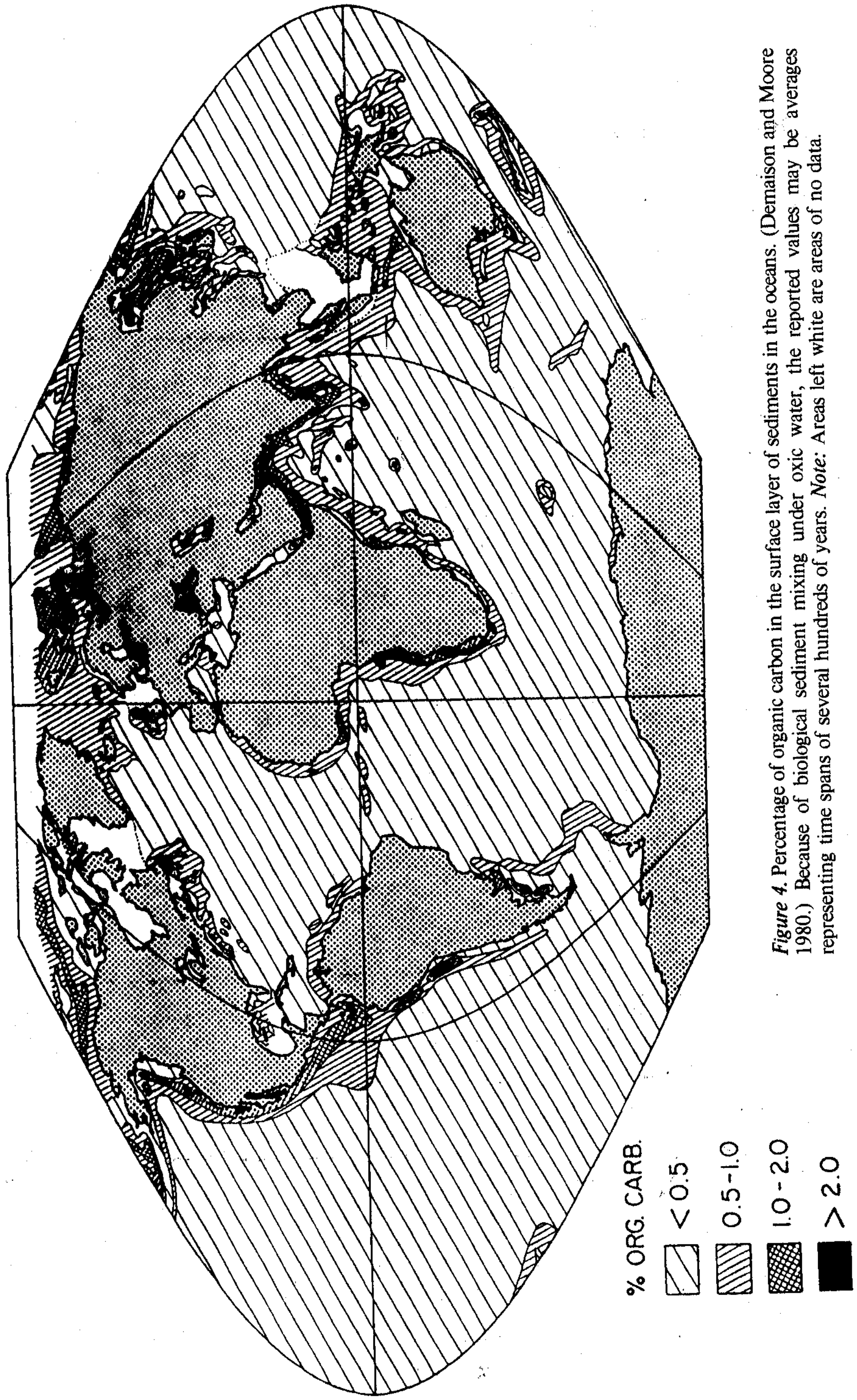

Figure 4. Percentage of organic carbon in the surface layer of sediments in the oceans. (Demaison and Moore 1980.) Because of biological sediment mixing under oxic water, the reported values may be averages representing time spans of several hundreds of years. *Note:* Areas left white are areas of no data.

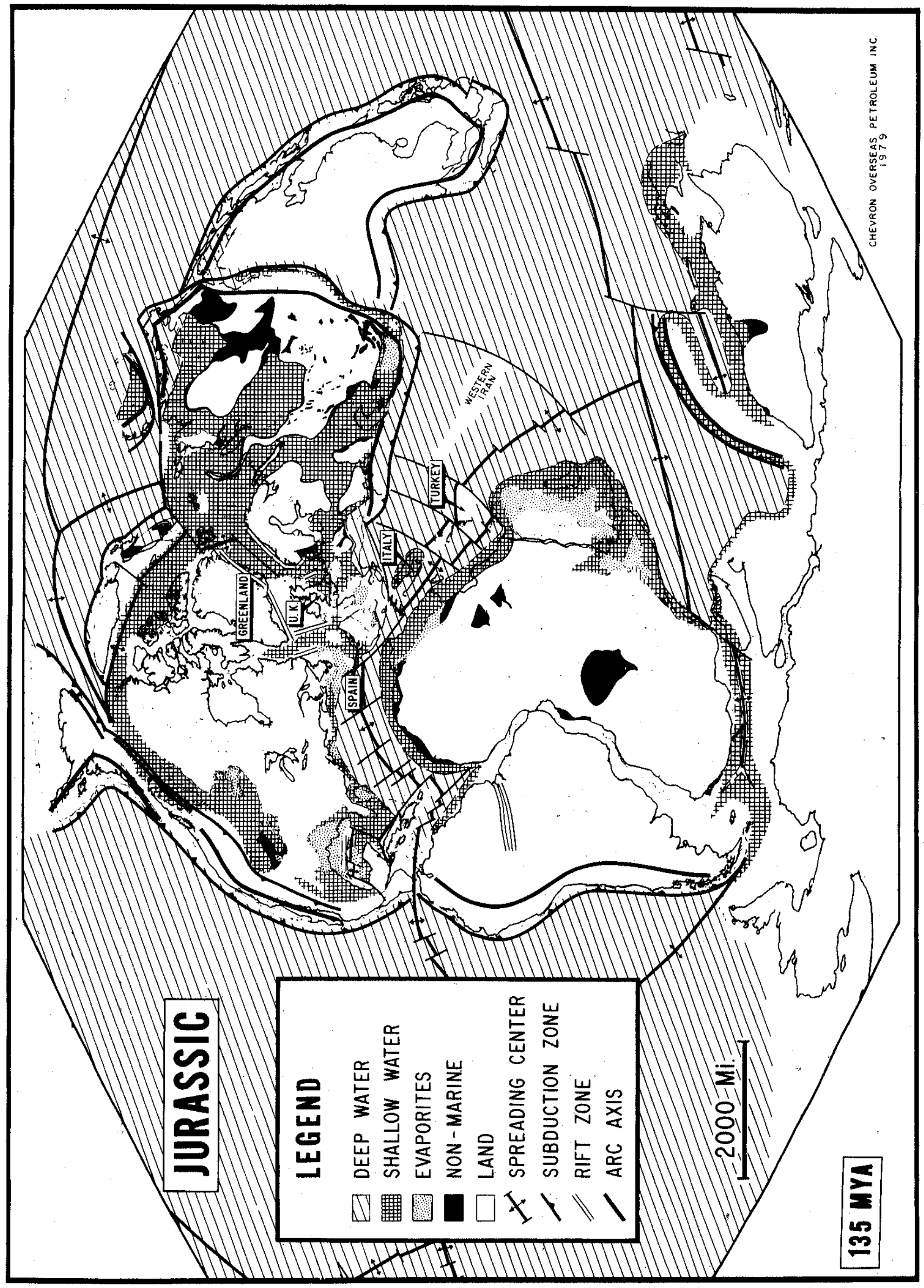

Figure 5. Plate tectonics. Position of continents during Jurassic time (135 m.y. ago).

well are of the basin. Will immature source rocks be buried deeper at another location, or is there another stratigraphic unit not penetrated by the well in question that might contain source rocks?

Digital seismic recording and the mathematical magic of the related processing have brought us far beyond the recording of structural attitude of a few years ago. Increasingly sophisticated measurement by seismic means of the physical properties of rocks is leading us rapidly towards direct measurement of porosity, composition, and fluid content of rocks, and the direct detection of hydrocarbons. Computer-generated information that such technology makes available to the explorationist has, in the last few years, tremendously increased the confidence factor in assessing the petroleum potential of an area.

Plate tectonics and continental drift, although postulated 55 years ago, have come of age in the last 10 or 15 years. They have been a unifying concept providing a simple and consistent solution to geological questions of the past and a framework for new questions whose answers will further improve our understanding.

Given these new tools, our most important task is to improve and apply them in systematic and imaginative ways to define the world's remaining exploratory potential.

If we consider the total surface area of our globe (Figure 3), we find that 70 per cent is made up of oceanic crust. Data from the deep-sea drilling projects and widespread geophysical surveys show it to be an area of very thin sediments and minimal porosity—its potential for petroleum virtually nil. An additional 13 per cent of the world's surface is made up of continental shields and high mountain areas where any oil or gas once formed and trapped has been destroyed by high temperatures or leaked away as traps have been destroyed by uplift and erosion. The remaining 17 per cent of the world's surface is the portion covered by sedimentary basins. To date about a quarter of the basin areas (or 4 per cent of the total world's surface) have been proved to be basins that have the right combination of source, migration, reservoirs and traps and are productive. We must organize our knowledge about these basins and, most important, identify those factors that must be established before we can be reasonably confident of their petroleum potential, or lack thereof:

How many basins are there?
How much sediment do they contain?
What kinds of sediments?
Do they have potential for source? Reservoir? Seals?
What is their thermal history?
What types of traps might be present?
When did traps form in relation to the time of generation and migration?

In many ways the magnitude of the task might appear overwhelming, but, with some imagination and careful analysis, it should not be. As an example, Figure 4 shows the distribution of organic carbon in the modern sediments of the world, a subject for study by oceanographers. An understanding of the reasons for this modern distribution applied to the oceans of Jurassic time, as shown in Figure 5, could lead to a high degree of success in predicting the distribution of significant source rocks in that time period. Similar studies of heat flow, types of reservoirs and traps associated with different types of basins, and other key factors can tell us where to go to get the critical pieces of information that we need to assess the potential for future discovery.

One of the most important elements in determining the ultimate amount of petroleum which will be produced in the world is the recovery factor. Oil discovered to date is approximately 1 trillion barrels. We estimate that the world's present weighted average recovery factor is near 30 per cent. This implies that the oil in place discovered to date is 3.3 trillion barrels. An increase of recovery to 40 per cent of oil in place would add 330 billion barrels, or about as much oil as has been produced in the world to date. While recoveries have improved significantly over the life of the oil industry, infill drilling in old fields, brought about by higher prices, and the disappointing results of many recent trials of new recovery techniques, have shown that there is a great deal of room for improvement.

A major challenge is to get more geology into our reservoir engineering. When we make a discovery, we must determine what stratigraphic model we are dealing with, what sediment components control porosity and permeability, what shape, extent and continuity the reservoir can be expected to have. Differences in reservoirs should be expected if we are dealing with turbidites versus long-shore bars, shelf-patch reefs versus shelf-edge barrier reefs, calcarenite sheets versus surge-channel oolite beds. We must recognize these differences early and plan accordingly.

The development of different geologic types of reservoirs requires different sequences, placements and programs for development wells, depending on the variability that can be expected in the reservoir. The newer, more complicated and more variable a reservoir is, the more time and money will have to be spent on careful gathering and analysis of data by coring, logging and testing. To a much greater degree than in the past, we must use the new seismic techniques to assist in defining our reservoirs at an early stage.

The challenge is to understand the reservoir as completely as possible in order to develop an enhanced recovery program that is the right one. This means that it is truly compatible with the geology and fluids involved physically and chemically. Of course, it must be sensible operationally and viable economically. Once put into effect it is of paramount importance that it be monitored properly so that it can perform at the desired levels and be modified as surprises occur, as they always do.

The fourth frontier which is becoming more time-consuming, but also vastly more important – is that of communication between explorationists and the people of the world and their governments. We must recognize that we have made available cheap energy, which has become a vital part of human existence. When cheap energy is no longer available, people deserve and rightfully demand an explanation. People are suspicious of that which they don't understand, especially when it has a direct economic and social effect on their lives. As earth scientists, both within and without the petroleum industry, we have a responsibility to understand, and to explain to others, the potential for new discoveries and additions to reserves, their possible size and their significance in the context of worldwide producing rates. We also must explain the method and rationale behind our projections. Given the risk nature of the exploration business, which is almost incomprehensible to those outside, this is no easy task, but who is more qualified to try it? Only when people and governments have some understanding of the problem and a realistic grasp of the possibilities, can they be expected to deal with the industry in a realistic and constructive way.

Wallace Pratt said, 'Oil is found in the minds of men'. In these times, all of us must accept the challenge to use our minds and imagination in new and creative ways to accomplish this task. Both the spirit and the content of this Symposium seem dedicated to that end.

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UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 15-23

NATIONAL DEVELOPMENT POLICIES AND LEGAL FRAMEWORK FOR PETROLEUM EXPLORATION IN THE SOUTHWEST PACIFIC

P HOHNEN and R N RICHMOND

Three broad factors can be identified which underlie and give special relevance to this Symposium. These are:

- Mankind's continuing dependence on petroleum;
- The recognized petroleum potential of offshore areas in the South Pacific region and the technical ability to develop encouraging prospects;
- The existence of a legal regime which vests in coastal states the right to control petroleum exploration and exploitation activities in a broad margin around their coastline.

The first (the need for petroleum) does not require special mention. The dependence of all national economies on continuing supplies of petroleum, whether for fuel or chemical uses, and the alarming increases in price over the last six years are eloquent of the need for new sources of supply. The second factor (offshore potential) will be the focal point of detailed technical discussions this week and is an area best left to the geologically trained. It is on the third factor however (the legal regime) and recent national legislative developments in the South Pacific region that we should like to address you for a few minutes in an attempt to provide a description of the legal and policy framework for petroleum exploration in the Southwest Pacific.

What then is the legal regime applicable in the offshore area, and what is its significance to petroleum exploration in the region?

Since its commencement in 1974, the United Nations Conference on Law of the Sea (UNCLOS) has been endeavouring to codify and develop the Law of the Sea to reflect modern-day needs and realities. Despite some differences of views, substantial progress has been made towards producing a comprehensive Law of the Sea Convention.

In the Conference there was early acceptance of the right of coastal states to exercise sovereign rights through a much wider coastal margin than was formerly the case in international law, and informal agreement was reached on the concept of an 'Exclusive Economic Zone' (EEZ) enabling coastal states to exercise sovereign rights within a 200 nautical mile zone around their coasts. As defined, these EEZ rights specifically include the right to regulate the exploration, exploitation, conservation and management of all living and non-living natural resources in the zone (Art. 56 ICNT) and to regulate the construction, operation and use of any installations and structures in the zone (Art. 60 ICNT). While a number of problems continue to face the ongoing negotiations, it is probably safe to say that whether or not an agreed comprehensive Convention emerges, the concept of the EEZ has already become widely accepted as part of international law. By way of illustration of state practice, more than 90 countries have now asserted some form of 200-mile jurisdiction, about half of that number since 1977.

There have also been intensive negotiations concerning the breadth of the continental shelf. Here too, the trend has been towards extending the jurisdiction of coastal states. The Informal Composite Negotiating Text (ICNT) used by the Conference provides a legal definition for a continental shelf at least 200 n.m. wide (Art. 76 ICNT) and there is some possibility that a wider continental shelf (up to 350 n.m.) may ultimately be agreed. In language similar to that applying to the EEZ, the ICNT also provides that the coastal state has exclusive rights to explore and exploit its continental shelf and to authorize and regulate drilling operations for any purpose (Art. 80 ICNT).

The relevance of these developments to petroleum operations is readily apparent. While advances in modern technology permit the operation of drilling rigs at depths up to 6000 ft and more than 200 miles from the coast, it is expected that the great majority of offshore mining operations will fall under coastal state EEZ or continental shelf-type jurisdiction. Indeed the recent developments in international law can be seen as an attempt by coastal states to secure full authority to regulate and benefit from all economic activities in their offshore areas.

The potential impact of this new legal regime will probably be greater in the Southwest Pacific than in any other region of the world. In terms of area alone, in the case of the developing island countries which are members of the South Pacific Bureau for Economic co-operation (SPEC), the EEZ concept places more than 14 million km^2 of sea and seabed under the jurisdiction of 10 countries, whose combined population is only 4 millions. To take another example, geographically SPEC's smallest member, Tuvalu, with a population of 7500 and a land mass of 26 km^2 has an EEZ potential of 900 000 km^2, almost twice the size of Papua New Guinea.

To the developing island countries of the South Pacific, who are typically small, geographically far-flung, often poorly endowed with commercially exploitable natural resources and traditionally dependent on exports of a limited number of tropical agricultural products and onshore mineral deposits such as phosphate for earning foreign exchange, the potential benefits of exploiting the living and non-living marine resources within 200 miles of their reefs are obvious and enormous.

In the light of the developments in the Conference on Law of the Sea, it is not surprising that most of the states in the South Pacific have been quick to declare their intention to exercise to the full the extended maritime jurisdiction and, where appropriate, to do so in a cooperative regional spirit.

The clearest example of this determination was the Declaration made by the South Pacific Forum in Port Moresby in August 1977, which called on Forum Governments to establish 200-mile fisheries zones before 31 March 1978 and agreed on the principles for the establishment of a regional fisheries agency. All 12 Forum Governments have now proclaimed, or have the necessary enabling legislation to proclaim, 200-mile fisheries or EEZs, and the South Pacific Forum Fisheries Agency has been established in Honiara, Solomon Islands, to assist member countries in securing maximum benefits from their living marine resources.

In parallel with these fisheries developments, Governments have also been focussing on the implications of changes in the Law of the Sea concerning the exploration and exploitation of mineral resources on and under the seabed. For example, most Forum countries have already introduced EEZ or continental-shelf legislation enabling them to regulate the prospecting for and exploitation of offshore minerals and hydrocarbons. One step further, a number of countries, including Fiji, Kiribati, Papua New Guinea, Tonga and New Zealand, have promulgated detailed regulations governing minerals and petroleum exploration in offshore areas under their jurisdiction. Other countries are in the process of preparing such legislation.

Petroleum legislation applying in the region, contains, in the first instance, basic provisions that might be found elsewhere in the world governing the issue of prospecting and mineral licences. As is the trend, it is not unusual to find lengthy and detailed conditions attached to the issue of licences including not only the payment of licence fees but continuing obligations such as the provision of accurate records of prospecting operations and their geological interpretation, which are important information inputs for governments of developing countries in particular. There may also be stipulations as to the period within which exploratory work must be commenced to ensure that potential resources are identified early.

In addition to these considerations, South Pacific Governments have also shown an understandable desire to have their special circumstances and interests recognized. Consequently, one can find regulations seeking to preserve the unique traditions, customs and history of the islands by protecting traditional fisheries, archaeological sites, historic wrecks or even war graves. Strict environmental regulations governing each stage of exploration and exploitation are also common to ensure that the South Pacific's famous beaches, reefs and waters are adequately safeguarded.

Once a prospect has been identified and the exploitation phase commences, other regulations governing foreign investment, taxation and royalties come into play. The nature of these varies from country to country, but they usually include provisions concerning local incorporation, national equity participation (if any), taxation and royalties, and the establishment of local infrastructure and training of nationals in relevant technical skills.

It is implicit in the foregoing provisions that a large role is intended for the private sector in the exploration and exploitation of the region's offshore resources. With the exception of New Zealand, which has formed a national petroleum company to participate directly in offshore exploration, Island governments have shown a preference to leave the actual exploration, and its inherent risks, to private companies, under the terms of an appropriate agreement. Specific statements on policies of the New Hebrides, Papua New Guinea and the Solomons are appended to this paper.

At the same time, both at the national and inter-governmental levels, Island governments are increasing efforts to develop their own ability to explore their offshore areas. Nine South Pacific countries participate in the activities of the Committee for the Coordination of Joint Prospecting for Mineral Resources in South Pacific Offshore Areas (CCOP/SOPAC) which is assisting Island countries to investigate their potential offshore mineral resources and to help them develop their capabilities in this field. On the legal side, at the request of its members, SPEC is also playing a role in providing governments with information on Law of the Sea developments concerning seabed resources.

The countries of the South Pacific have now reached the point where they have demonstrated their intention to explore and exploit fully and responsibly their offshore mineral resources. National policy and legislation have given recognition to the vital role which foreign private capital and technology can play in this exciting process.

CCOP/SOPAC, UNDP and the Fiji Government are to be congratulated on their sponsorship of this Symposium, which will provide an invaluable opportunity for governments, potential investors and other interested parties to conduct a full exchange on all aspects of offshore petroleum potential in the Southwest Pacific and to establish, in the Pacific way, a spirit of cooperation which will set the tone for future, and it is hoped mutually enriching, relations.

APPENDIX

STATEMENTS OF THE NEW HEBRIDES, PAPUA NEW GUINEA AND SOLOMON ISLANDS GOVERNMENTS

Offshore Hydrocarbon Prospecting in the New Hebrides: Statement by the Government of National Unity

The Government is aware of the interest of oil companies in the South Pacific region and recognises the potential of the New Hebrides offshore waters for hydrocarbons and production. It has decided to take the necessary action to facilitate and control offshore exploration and production. It proposes to introduce legislation which will establish a licensing system and financial regime comparable to those applying in other South Pacific states.

Papua New Guinea Mineral Resources Development Policy

The Government considers the development of mining and petroleum projects to be of the highest priority and welcomes foreign investors in undertaking exploration.

The Government has adopted the following basic principles relating to the exploitation of mineral (including petroleum) resources:

Mineral resources belong to the people of Papua New Guinea, and the Government and the people must receive a fair price in return for extraction of the minerals.

Foreign enterprises exploiting Papua New Guinea's mineral resources deserve a reasonable return on their investment, but extraordinary gains above a reasonable return on investment will go in large part to the Government.

The Government has the right to regulate extractive enterprises so as to maximise the benefits to the local community while minimising the potentially harmful social and economic costs.

Prospecting concessions are granted by the Minister for Minerals and Energy, who is advised by the Mining Advisory Board and the Petroleum Advisory Board. Areas available for prospecting concessions may be advertised internationally.

Exploration for Minerals and Mining Development

As a general rule foreign investors will not be allowed to establish small-scale alluvial gold-mining enterprises.

For large-scale exploration, Prospecting Authorities may be granted for up to two years, with renewal dependent on satisfactory completion of agreed work programmes or expenditure of money on prospecting activities, and satisfactory proposals for additional work.

For mining purposes, a Gold-Mining Lease (maximum area 20 ha) or Mineral Lease (maximum area 100 ha) may be granted, both for an initial term not exceeding 21 years; a Special Mining Lease (maximum area 60 km^2) may be granted (for an initial term not exceeding 42 years) in respect of a very large deposit.

When the existence of a very large deposit is demonstrated, the Government is prepared to enter into firm commitments to ensure the satisfactory establishment of a mine by way of a Mining Agreement with the investor covering the arrangements for each project and providing for the issue of a Special Mining Lease.

A Mining Agreement may comprise some or all of the following major elements:

A firm commitment to proceed with development as rapidly as possible.

The developer to pay for required infrastructure.

Satisfactory arrangements with respect to localisation of labour, training, environmental control, and other social issues.

Proposals for Government equity participation.

Proposals for processing of minerals within Papua New Guinea.

A Mining Agreement assures the investor of continuing rights over an ore body in exchange for a definite commitment by the investor to a phased work programme leading to development of a mine; this programme is to be completed within a specified period and, if a mine is not developed, the investor must withdraw without further rights. Since the phased work programme culminating in the development of a mine cannot be defined until very substantial exploration expenditures have been incurred, and since the Government will under no circumstances give an assurance of continuing rights over an ore body without a firm programme leading to development of a mine, finalisation of a Mining Agreement must be deferred to a very late stage, near the end of the exploration phase.

Regarding Government equity in large mining projects: The Government accepts that to settle its level of participation in equity after exploration is almost completed (as part of the Mining Agreement) creates uncertainty, in that a key element of the financial package is left unresolved at the time when decisions on major exploration expenditure have to be taken. The Government is therefore prepared to enter into an agreement on equity at an earlier stage, before the investor undertakes major expenditures on exploration. In exchange for agreement on equity, the Government will require a firm commitment by the investor to an acceptable phased work programme involving significant exploration expenditure, the size of which shall be determined on the merits in each case.

The Government is not prepared to settle equity prior to the granting of a prospecting authority or at the early exploration stage of minimal expenditure against which many prospecting authorities are granted.

The Government's maximum participation in equity will be 30% and the Government's desired share could often be considerably less. The Government is not concerned with obtaining a majority shareholding nor does it aspire to be the largest partner in any project. In most instances the Government will seek an option to participate at par at the commencement of construction in future production expenditures. A carried interest involvement must always remain an alternative for the Government but in such an exceptional case the interest rate for carrying purposes would adequately compensate the investor for the use of his funds.

Regarding taxation: Under the Income Tax (Mining and Petroleum) Act 1978, all major mining projects (where a Special Mining Lease is involved) are subject to special tax provisions similar to the existing Ok Tedi and Bougainville Projects arrangements. *The Act does not affect new small scale mining ventures.* Major provisions in the Act include:

Exploration expenditure incurred within 11 years before the grant of a Special Mining Lease is deductible at a rate determined by dividing residual exploration expenditure by the then estimated life in years of production from the Special Mining Lease or by five, whichever is less.

Exploration expenditure within a Prospecting Authority out of which a producing mine is developed will be deductible immediately. Exploration on another Prospecting Authority will be deductible when the concession is fully relinquished.

Capital expenditure spent on developing the mine is similarly deductible but at a rate determined each year by dividing the residual capital expenditure by the then estimated life in years of production from the Special Mining Lease or by ten, whichever is the less.

In the initial years of operation accelerated deductions of exploration and capital expenditure are allowed if the mine has a cash flow of less than 25% of the initial investment. This helps ensure a payback period of four years.

A deduction at the rate of 25% of cost is allowed for new equipment in the year in which it is installed. This encourages further investment in the mine once it is operating.

An additional tax is imposed on the net cash flow of the mine once the initial investment plus a specified rate of return has been recouped. The additional tax is imposed at a rate equivalent to the difference between the normal tax and 70%. For the purposes of this provision the rate of return is either 20% per annum or the USA Prime Rate plus 12% per annum as each company elects within 3 months after the grant of a Special Mining Lease.

During the investment recovery period the rate of income tax on mining income derived from a Special Mining Lease shall not exceed 35% per annum in the case of a company incorporated in Papua New Guinea or 48% per annum in any other case.

The mining-tax provisions are aimed at increasing the return to government from major mining projects if they are extremely profitable and at the same time, by allowing depreciation advantages, reducing the risk of slow payback from a marginal mine.

In addition to taxation, and legislated under the Mining Act (Amalgamated) 1977, a *royalty* of 1.25% of the f.o.b. or net smelter value of all mine products will be paid to the Government.

Thus, with the mining tax provisions legislated under the Income Tax (Mining and Petroleum) Act 1978, together with the royalty provisions of the Mining Act, the potential investor is provided with full knowledge of the non-equity financial package for investment in exploration and mining development in Papua New Guinea.

Exploration for, and Development of, Petroleum and Natural Gas

The Government's policy in respect of

petroleum resources is constructed around four specific objectives:

To ensure the maximum financial benefit for Papua New Guinea consistent with allowing a reasonable return to the foreign company.

To provide for direct participation by the Government in oil and gas operations.

To ensure effective control over the industry through appropriate financial and technical regulations.

To provide for access by the Government to part of any oil production.

Petroleum exploration and production will, as a matter of Government policy, be undertaken in the private sector, and revenue to the Government will be derived largely from income tax and an additional profits tax charged on cash flows after certain threshold returns have been achieved. The Government will seek a carried interest for its negotiated share of oil produced.

Petroleum Prospecting Licences, for onshore or offshore areas, may be granted under the Petroleum Act 1977 for an initial period of 6 years. One extension period of 5 years may be granted, dependent on satisfactory completion of agreed work programmes and satisfactory proposals for further work, and over not more than 50% of the area.

A *Petroleum Agreement* will be negotiated between the Government and the licensee setting out details of how a project would proceed in the event of a commercial discovery. Relations between the Government and the licensee will be governed by the Petroleum Agreement; the licensee will of course be subject to the normal laws of Papua New Guinea including the specific laws on petroleum taxation and licensing. A Petroleum Agreement provides for the Government's equity participation in future development and will also include provisions relating to training and localization, local purchasing and business development, use of facilities by third parties, currency and exchange control, and disposal of the Government's production share.

Regarding Government equity: The Government will take a minotiry carried interest equity participation in all petroleum developments.

The exact percentage will be negotiated in each case between the licensee and the Government.

The Government's share of exploration and development costs will be provided by the licensee, who will be repaid with interest from the Government's share of production.

Regarding taxation: Major provisions, legislated under the Income Tax (Mining and Petroleum) Act 1978, include:

The licensee will pay a petroleum income tax of 50% of taxable income. This tax, for producers of petroleum, will be in lieu of current company income tax and dividend withholding tax.

An additional profits tax of 50% will be imposed once the initial investment plus 27% per annum has been recouped.

Exploration expenditure incurred within 11 years before the issue of a Petroleum Development Licence will be deductible at a rate determined each year by dividing residual exploration expenditure by the then estimated life in years of production from the Petroleum Development Licence or by four, which ever is the less.

Exploration expenditure within a Petroleum Prospecting Licence out of which a producing field is developed will be deductible immediately. Exploration expenditure on another Petroleum Prospecting Licence will be deductible, when the concession is fully relinquished.

Capital expenditure will be deductible at a rate determined each year by dividing residual capital expenditure by the then estimated life in years of production from the Petroleum Development Licence or by eight, whichever is less.

There is provision for accelerated deductions of exploration and capital expenditure during the early years of a project if cash flow falls below 25% of initial investment.

Operating expenditures will be deductible in the year they are incurred and operating losses may be carried forward for up to 7 years but not carried back.

Deductions for interest on loans will be limited to an amount that would be allowable if market rates were charged on the loans.

Natural gas operations will be treated the same as petroleum operations.

In addition to taxation, and legislated under the Petroleum Act 1977, a *royalty* of 1.25% of the wellhead value of all petroleum production will be paid to the Government.

Under the two-tiered taxation system, with an additional profits tax, the burden of tax is related directly to field profitability. The system is more flexible than other systems such as production sharing. With economically marginal fields the licensee will retain a greater proportion of the cash flow, while with very profitable fields the Government's share will increase. The overall impact of the financial package, with, e.g., 30% Government participation, is to make it rather more generous than the 70:25 production-sharing contract, commonly entered into elsewhere.

Regarding marketing and pricing: Prices will be determined by the Minister on the basis of current arm's length market prices. Exact guide-lines for the Minister to follow are laid down in Schedule 2 of the Petroleum Act 1977. The Minister's determination is subject to arbitration.

Regarding licences: As well as Petroleum Prospecting Licences for exploration there will be Petroleum Development Licences for production.

Petroleum Prospecting Licences will be issued for an initial term of 6 years, and may be extended in respect of no more than 50% of the initial area for an additional 5 years. As a general rule they will be limited in area to 60 blocks of 5 minutes of longitude by 5 minutes of latitude, but in certain cases Licences may be issued over an area of up to 200 blocks. Companies may hold more than one Licence.

Petroleum Development Licences will be issued in respect of not more than nine blocks within a Prospecting Licence in which there has been a commercial discovery. A Petroleum Development Licence will be granted for an initial 25 years and may be extended for up to an additional 20 years.

Instruments of consent may also be issued for seismic and other survey work in the course of scientific investigations over areas not currently held under licence. Such instruments will carry no automatic rights to Licences.

The Petroleum Act 1977 allows for automatic progression from a Petroleum Prospecting Licence to a Development Licence subject to the State's approval of the development programme and associated matters. Such approval is subject to arbitration if agreement cannot be reached.

Environment and Conservation Considerations

While the Government wishes to proceed rapidly with economic development, it is also concerned with protecting the environment from undesirable effects of uncontrolled or inappropriately regulated development. Any development project, including mining and petroleum developments, must take into account the impact on the natural environment and human communities, and final decisions about a project will be made not only on the basis of economic and technical considerations but also on environmental considerations.

Guidelines for Feasibility Studies are available and in conjunction with Acts of Parliament — the Environmental Planning Act 1978, the Environmental Contaminants Act 1978, and the Conservation Areas Act 1978 — provide the legal framework for the implementation of Papua New Guinea's environmental policies.

The Environmental Planning Act 1978 establishes proper environmental considerations as part of policy and project planning. Proponents of mining and petroleum developments, and other development projects, are required to identify and evaluate the ecological and social implications of the proposed development and to adequately plan appropriate management and protective measures. The Ministry of Environment and Conservation together with other Government departments and agencies determines the adequacy of the environment planning and may suggest modifications and new approaches. Clearly, the requirements vary from project to project and a realistic stance, consistent with practicably available procedures and technology and the particular characteristics of the project area, will always be adopted.

The Environmental Contaminants Act 1978 is the basic Act controlling pollution of air, land, and water and is similar to pollution-control acts in other countries. A primary aim of the Act is to ensure that pollution does not become *unreasonable* as a result of development.

The Conservation Areas Act 1978 provides for the protection of certain sites, lands, and landforms, etc., which may be considered as part of the National Heritage of Papua New Guinea.

The Government of Papua New Guinea considers the development of its mineral and petroleum resources to be of the highest priority. It realizes that foreign expertise and capital are essential for the development of these resources and in this context welcomes foreign investment. A coherent policy with clearly defined rules and regulations for such investment is established. The Mining Act (Amalgamated) 1977, the Petroleum Act 1977, the Income Tax (Mining and Petroleum) Act 1978, and the Government's willingness to negotiate its level of equity participation in mining projects before large amounts are expended on exploration together with environment and conservation legislation are all facets of this total policy. Policies are fixed, have been endorsed by large mining and petroleum companies, and by large banks, and permit speedy investment negotiations. The 'rules of the game' are firmly established and 'players' are invited to participate.

Solomon Islands Petroleum-Exploration Policy

The Solomon Island development strategy requires that the natural resources of the country, including its offshore areas within the 200-mile economic zone, be explored and evaluated as quickly as possible.

The Government is particularly interested to receive proposals for commercial investment in assisting to develop natural resources, and prospecting for hydrocarbons in the offshore will be encouraged. An offshore Petroleum Act is now at an advanced stage of preparation and will establish the framework to enable orderly prospecting and development to be carried out by reputable commercial companies. It is to be hoped the Act will be extended to onshore. Within the terms of the Act it is proposed to enter into detailed agreements between individual companies and the Solomon Islands Government concerning work programmes, expenditure levels, the extent and nature of Government participation and fiscal arrangements.

The Act, which has been prepared for Solomon Island Government by the Technical Assistance Group of the Commonwealth Secretariat, is an amalgam of the Fiji, Papua New Guinea legislature and the legislation of other Commonwealth countries.

The main features of the Act are as follows:

1. Petroleum Prospecting Licences will be granted under a 5-minute graticular block system, with up to 60 contiguous blocks being granted under any one licence.
2. Exclusive licences will be granted for an initial period of 6 years with provision for a 5-year extension subject to normal relinquishment provisions.
3. Where petroleum is discovered in a licence area within the period of 2 years before the date of expiration of an extended petroleum licence, a further extension of up to 3 years may be granted.
4. In the event of a discovery, provision will be made for the retention of the discovery block and up to eight adjacent blocks. Upon declaration of the discovery, a period of not less than 2 years will be given to the licence to assess the feasibility of the construction, establishment and operation of an industry for the recovery of petroleum from the location.
5. It will include adequate assurance that a company which expends substantial funds on exploration and discovery will have the right to develop that

discovery subject to acceptable development proposals.

6. Petroleum production licences will confer exclusive rights to carry on prospecting, mining and marketing operations for a period of 25 years with provision for a 20-year extension.
7. Provision for pipeline licences.
8. Provision for transfer, surrender and cancellation of licences.

The Government of Solomon Islands is now considering preparing a statement on its petroleum policy and legislation in which it will determine such matters as state participation, taxation and financial regimes, pricing and marketing and other allied issues.

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GLOBAL EOCENE PLATE REORGANIZATION: IMPLICATIONS FOR PETROLEUM EXPLORATION

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ABSTRACT

Evidence of a global plate reorganization during Eocene time is manifested in a number of regions throughout the world. The pattern is not wholly consistent in all cases, but suggests a general change from large N–S components of relative plate motion to large E–W components. In some cases, the change is also accompanied by deceleration of spreading rates, and obduction of ophiolites. The reorganization is the probable result of an increase in resistance along the global plate system. It is generated by an increase in the length of E–W trending collisional plate boundaries during the interval 55–40 m.y.b.p., forcing the system to reorient along lines of less resistance.

This model of plate reorganization provides a variety of environments favourable for the generation and entrapment of hydrocarbons. Examples include collisional environments of the Himalayan, New Guinea and Caribbean areas; rift environments of the Australia-Antarctic margins; and environments with a more complex history of collision and subsequent strike-slip, such as in Central America.

INTRODUCTION

In recent years, plate tectonics has proved to be a useful tool in the field of petroleum exploration. Perhaps one reason for this is that the plate-tectonic history of an area can be a key in isolating tectonic events which may have provided a favourable environment for the generation and entrapment of hydrocarbons.

Major accumulations of hydrocarbons are without a doubt rare features within the earth's crust and the existence of many of them may be the result of unusual structural events in the earth's history. An examination of regional plate motions has led us to speculate concerning a global plate reorganization during Eocene time, which may be a prime example of such anomalous tectonic events.

REGIONAL PLATE MOTIONS

In examining regional plate motions a global circuit of plate boundaries is made from the Indian Ocean to the Pacific, Arctic, Atlantic and Caribbean regions (Fig. 1). For simplicity, the text describes only major events. Pertinent information and references are summarized in Table 1.

Indian Ocean

From the beginning of India's northward migration 140 m.y. ago, sea-floor spreading proceeded in a north-south direction. Then, during early Eocene time, spreading rates slowed drastically and spreading directions and fracture zones were reoriented NE–SW. This major change in Indian Ocean tectonics took place very close in time to the arrival of the Indian Shield at the southern margin of Asia, which began the Himalayan Orogeny.

In the southeastern Indian Ocean the oldest known sea floor is Lower Eocene in age (magnetic anomaly 22). This dates the breakup of Australia from east Antarctica, an event which appears to be synchronous with the change of plate-motion direction in the rest of the Indian Ocean.

Shortly after the breakup between Australia and Antarctica, the Australian Shield reached a subduction zone south of New Guinea. As a result of this collision, differential motion between Australia and New Guinea ceased and a new subduction zone was formed along New Guinea's northern coast. This collisional event has been dated as Middle Eocene (46 m.y.b.p.).

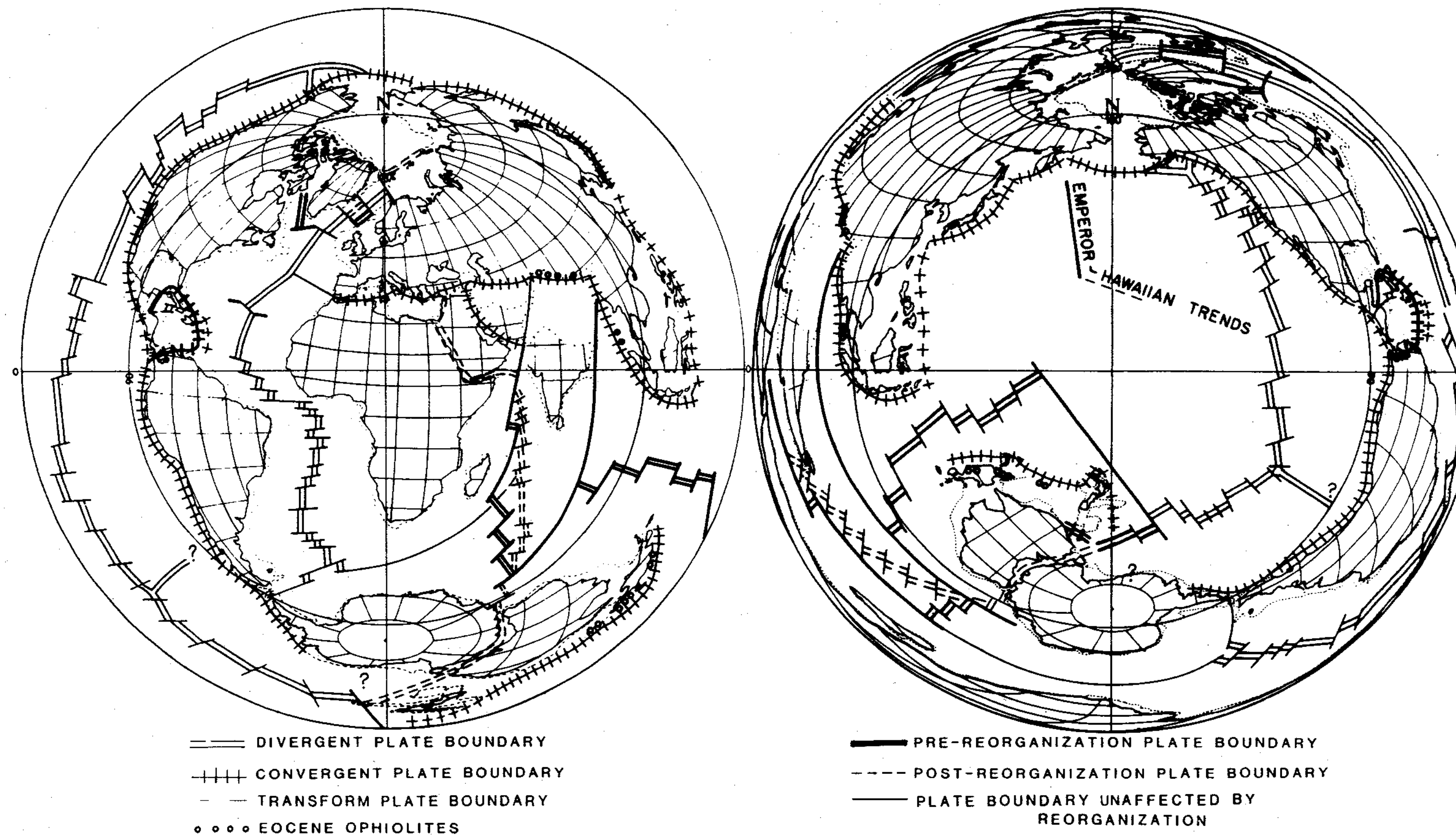

Figure 1. Plate boundaries before and after Eocene plate reorganization plotted on Lambert equal-area projection showing Eocene (50±5 m.y.b.p.) distribution of continents.

Pacific Ocean

Eocene also is the suggested age for a change in the western Pacific, where N–S trending fracture zones became the site of the new arc-trench systems of the Philippines, Marianas, Bonin, Palau, Yap and Tonga.

In the north-eastern Pacific a bend in fracture zones north of the Pioneer Fracture Zone and a Z pattern of magnetic anomalies (between 21 and 22) indicate that a change in plate-motion direction occurred in early Eocene time.

To the southwest the bend in the Hawaiian-Emperor Seamount chain corresponds to a time interval of 41 to 43 m.y. ago. This is further evidence of a major change of plate motion in the Pacific.

Arctic

Sea-floor spreading on the Nansen Ridge began about 57 m.y. ago (actually latest Paleocene). This event was responsible for breaking apart the Lomonosov Ridge from the Eurasian Arctic shelf. Spreading also began in the southern Norwegian Sea at this time, but by the end of the Eocene the ridge had jumped to its present site directly north of Iceland.

North Atlantic

In the central North Atlantic the most obvious change of plate motion occurred in Late Cretaceous time (80 m.y.b.p.). A less obvious change, however, can be seen by examination of fracture-zone trends in greater detail. On the Kane Fracture Zone a major bend occurs between magnetic anomalies 21 and 24 (corresponding to 50–57 m.y.b.p.).

Caribbean

The Caribbean region is enclosed on all sides by plate boundaries. Its history is marked by numerous tectonic events. In Cretaceous time N–S compression dominated its northern margin. During Eocene time N–S compression was replaced by E–W strike-slip movement. This change was accompanied by a jump in the plate boundary from north of Cuba to the Cayman Trough. This resulted in transfer of Cuban lithosphere to the North America plate. Almost all volcanism ceased in the north Caribbean, and subduction moved east from the Aves Swell to the Lesser Antilles. Again, in Cuba this change is recorded by a Middle Eocene unconformity throughout the island.

GLOBAL PLATE REORGANIZATION

It may be noted that many of the events mentioned above are not synchronous, and in fact they may not be. But it should be kept in mind that a certain amount of inaccuracy may be due to the various dating techniques. On land, paleontologic and radiometric dates are widely used. On the sea floor, use of the magnetic reversal time scale is made. The Heirtzler time scale was published in 1968 (Heirtzler *et al.*). LaBrecque *et al.* (1977) revised this scale, shifting magnetic anomaly ages by as much as 7%. This shifted the date for the change in Indian Ocean tectonics from 53 to 50 m.y.b.p., thus bringing it a little closer to the time of the Middle Eocene collision in the Himalayas (Molnar and Tapponier 1977).

The nature of the plate interactions that produced the observed patterns of the Eocene reorganization may be deduced from an assessment of global plate boundaries. The lengths of divergent and convergent plate boundaries before and after the plate reorganization were measured and totalled (Table 2). Convergent boundaries were divided into subduction (of oceanic crust) and collision (continental).

It was found that the greatest change in length occurred at E–W trending collisional boundaries. Collision of the Australian Shield with the New Guinea island arc occurred along an E–W boundary. The Bahama Banks collided with the Greater Antilles along an E–W trending zone, and the change from E–W strike-slip to N–S collision in the Mediterranean occurred along an E–W zone.

If momentum of plates can be neglected (McKenzie and Sclater 1971) then it is logical to assume that any change in their motion must involve a change in either the driving forces or the resistive forces. The evidence presented in this paper clearly supports the interpretation that the plate reorganization from Mesozoic patterns to Cenozoic patterns of plate motion was primarily a response to the global distribution of resistive forces. At the

TABLE 1

Regional Plate Motions (Rona and Richardson 1978) (See Fig. 1)

Region	Event	Time given in m.y.b.p. (Evidence in parentheses)	Reference
Indian Ocean	Beginning of continental collision between India and Eurasia	55 to 40 (structural and stratigraphic relations)	Gansser 1959, 1964 Le Fort 1975 Molnar & Tapponier 1977 Sahni & Kumar, 1974.
	Decrease in seafloor-spreading rates between India and Antarctica-Australia by a factor of approximately 2, from >8 cm/y to <4 cm/y (half rate)	53 to 40 (magnetic anomalies 21 and 22)	Sclater & Fisher 1974 Johnson et al 1976 Pierce 1977
	Subduction of nearly E-W trending oceanic ridge between Australia and Eurasia at convergent plate boundary along the southern margin of Eurasia	53 to 32 (magnetic anomalies 11 and 22)	Sclater & Fisher 1974
	Reorientation of oceanic ridge system from nearly E–W to NW–SE with concomitant change in seafloor-spreading direction from nearly N–S to NE–SW	53 to 50 (magnetic anomalies 21 and 22)	Sclater & Fisher 1974 Pitman et al 1974
	Ending of sea-floor spreading with large E–W component in Tasman Sea	59 to 56 (after magnetic anomaly 25)	Hayes & Ringis 1972
	Beginning of sea-floor spreading with large N–S component between Australia-Broken Ridge and Antarctica-Kerguelen Plateau	53 to 50 (magnetic anomalies 21 and 22)	Weissel & Hayes 1972
	Decrease in rate of relative motion between India and Eurasia plates by a factor of approximately 2, from about 10 cm/y to 5 cm/y	50 to 38 (magnetic anomalies)	Johnson et al 1976
South Pacific	Decrease in rate of sea-floor spreading by a factor of approximately 2, from about 5 cm/y to 2.3 cm/y	56 (magnetic anomaly 24)	Pitman et al 1968 Heirtzler et al 1968
	Collision between Australian shield and New Guinea Island arc	49 to 43 (Middle Eocene; structural and stratigraphic relations)	Brookfield 1977
	Beginning of sea-floor spreading, west of Scotia Sea	54 to 30 (magnetic anomalies)	Herron 1974
	Rifting of South America from Antarctic Peninsula	54 to 30 (magnetic anomalies)	Herron 1974
	Laramide Orogeny predominantly E–W compressive deformation and plutonism in western North America and northwestern South America	80 to 40	Coney 1971
	Change in direction of sea-floor spreading from ENE–WSW to E–W evidenced by Z pattern of magnetic anomalies and trends of fracture zones	55 to 50 (magnetic anomalies 21 to 23)	Menard & Atwater 1968 Pitman et al 1974
	Beginning of subduction along Philippine, Bonin, Marianas, Yap, Palau, and Tonga trench-arc systems	Shortly after 45 (maximum ages and trends of associated island arcs)	Uyeda & Ben-Avraham 1972 Hilde et al 1977 Uyeda 1977

	Change in direction of plate motion from NNW to WNW with respect to Hawaiian hot spot evidenced by Emperor Seamount - Hawaiian Ridge bend	43 to 41 (radiometric dating)	Clague et al 1975
Arctic	Beginning of sea-floor spreading in an E–W direction about the Gakkel Ridge, resulting in opening of the Eurasian Basin	63 (magnetic anomalies)	Pitman & Talwani 1972 Herron et al 1974
North Atlantic	Duration of sea-floor spreading in the Norwegian Basin	56 to 42 (magnetic anomalies)	Vogt et al 1970
	Beginning of spreading at Iceland - Jan Mayen Ridge	42 (magnetic anomalies)	Vogt et al 1970
	Duration of second episode of sea-floor spreading in the Labrador Sea	56 to 44 (magnetic anomalies 19 to 24)	Mayhew et al 1970 Le Pichon et al 1971
	Beginning of spreading on Reykjanes Ridge	56 (magnetic anomaly 24)	Vogt et al 1969 Herron & Talwani 1972
	Change in spreading direction on Reykjanes Ridge from indeterminate to E–W	40 (magnetic anomalies)	Vogt et al 1969 Herron & Talwani 1972
	Decrease in rate of sea-floor spreading north of the Azores-Gibraltar lineament by a factor of approximately 2, from about 2.5 cm/y to 1.0 cm/y	59 to 50 (magnetic anomalies 21 to 25)	Pittman & Talwani 1972
	Decrease in rate of sea-floor spreading south of the Azores-Gibraltar lineament by a factor of approximately 1.5, from about 1.7 to 1.2 cm/y	59 to 50 (magnetic anomalies 21 to 25)	Pittman & Talwani 1972 Lattimore et al 1974
	Change in relative direction of spreading from nearly E–W to WNW–ESE evidenced by bend in Kane Fracture Zone	59 to 50 (magnetic anomalies 21 to 25)	Rabinowitz & Purdy, 1976
	No apparent changes in sea-floor spreading rate and direction (based on limited data)	72 to 0 (axial magnetic anomaly to anomaly 31)	Dickson et al 1968 Maxwell 1970 Ladd et al 1963
Caribbean	Separation of Caribbean plate from Pacific plate as a consequence of southward extension of the Middle Americas Trench	54 to 38 (Eocene age of earliest volcanics in Central America)	Malfait & Dinkelman 1972
	Jump of subduction zone from Aves Ridge to Lesser Antilles island arc	54 (60 m.y.b.p. radiogenic date of granodiorite from Aves Ridge; 54 m.y.b.p. initiation of volcanism in Lesser Antilles)	Weyl 1966 Malfait & Dinkelman 1972
	Collision of Bahama Banks with Greater Antilles (Cuba)	Post-70 to 43 (Late Cretaceous to Middle Eocene based on structural relations)	Meyerhoff & Hatten 1968 Mattson 1973
	End of N–S subduction at Puerto Rico	49 to 43 (Middle Eocene radiogenic age of youngest volcanic rocks present)	Mattson 1973

TABLE 1 *(Continued)*

Region	Event	Time given in m.y.b.p. (Evidence in parentheses)	Reference
	Beginning of present Puerto Rico Trench	Post-43 (post-Middle Eocene based on structural and stratigraphic relations)	Monroe 1968 Khudoley & Meyerhoff 1971 Tucholke & Ewing 1974
	End of N–S subduction at Cuba	49 to 43 (Middle to Late Eocene based on structural relations)	Meyerhoff & Hatten 1968 Khudoley & Meyerhoff 1971
	End of N–S subduction at Hispaniola	Prior to 43 (pre-Late Eocene based on stratigraphic relations)	Nagle 1971
	General change of plate motion from N–S compression to E–W strike slip	54 to 38 (Eocene based on structural relations at northern and southern boundaries of Caribbean plate)	Malfait & Dinkelman 1972
	Beginning of sea floor spreading in present Cayman Trough	49 to 43 (Middle Eocene based on estimated spreading and sedimentation rates)	Holcombe et al 1973 Heezen et al 1973
Mediterranean	Right lateral E–W strike-slip motion along Tethys	80 to 53 (deduced from relative plate motions in the North Atlantic)	Dewey et al 1973
	Change to N–S collision, compression and subduction along Tethys	53 (deduced from relative plate motions in the North Atlantic)	Dewey et al 1973
	Collision of Iberian peninsula with Eurasia	54 to 38 (Eocene based on structural relations)	Boillot & Capdevila 1977
	Beginning of deformation of Betic mountains in southern Spain	49 to 43 (Middle Eocene based on structural relations)	Dewey et al 1973
	Possible reversal of subduction polarity in Swiss Alps	49 to 43 (Middle Eocene based on structural relations)	Hsu & Schlanger 1971
	Meso-Alpine orogeny involving main movements of Penninic and Austroalpine nappes in Swiss Alps	43 to 38 (Late Eocene to Early Oligocene based on structural relations)	Trümpy 1975
	Subduction in Tellian Atlas and Sicilides	54 to 38 (Eocene based on active flysch sedimentation)	Dewey et al 1973
	End of subduction in eastern Alps as a consequence of continental collision	45 to 42 (structural and stratigraphic relations)	Dietrich 1976
	Orogeny in High Atlas and Saharan Atlas.	43 to 30 (post-Middle Eocene and pre-Middle Oligocene based on structural and stratigraphic relations)	Laffitte 1939 Société Chérifienne des Pétroles 1966
Red Sea	Beginning of sea-floor spreading with large E–W component	38 (magnetic anomalies)	Girdler & Styles 1974

onset of this Eocene reorganization, the total length of collisional plate boundaries increased from 2500 to 19 000 km, a change of more than 700% (Table 2). And as a result, the global plate system reoriented itself along lines of less resistance.

IMPLICATIONS FOR PETROLEUM EXPLORATION

It has been noted that most of the world's major sedimentary basins are adjacent to past or present plate boundaries (Halbouty *et al.* 1970; Rona and Neuman 1976). Thus it is logical that a major change in movement along a particular plate boundary may be recorded in some way within the associated sedimentary basin. Such a recorded event may range from a simple change in sedimentation patterns to the introduction of major structural complexities.

Below are cited several examples of sedimentary basins which have been strongly affected by the Eocene reorganization. It is our suggestion that this plate reorganization was responsible for creating in many of the world's basins a variety of environments favorable to the generation and entrapment of hydrocarbons.

Pakistan

The Potwar Basin lies on the northwestern margin of the Indian Shield (Fig. 2), where Eo-Cambrian to Recent miogeosynclinal and epicontinental shelf sediments have accumulated at the southern margin of the former Tethys Seaway. The leading edges of the foredeeps bordering the northern edge of the Indian Shield were heavily deformed beginning in the Eocene and continuing to the present time.

Structural development in the Potwar Basin began in the Mesozoic with the formation of broad, gentle salt pillows arching the overlying sediments. However, it was not until late Eocene time that N–S compression from the Himalayan Orogeny began to create the giant structures seen today. The general structure of the basin is controlled by a series of north-dipping thrust faults (Fig. 3). These faults

TABLE 2

Early Cenozoic Plate Reorganization (see Fig. 1)

	Length in Kilometers*		
All Convergent Plate Boundaries (Subduction and Collision)	Before	After	Net Change
South America and North America	17 000	18 000	+ 1 000
Aleutian-Kurile-Japan	8 000	8 000	0
Philippine-Ryuku-Marianas-Yap	0	9 000	+ 9 000
Zagros-Indus-Indonesia	14 000	14 000	0
New Guinea-Tonga	8 000	11 500	+ 3 500
Caribbean	5 000	2 500	− 2 500
Mediterranean	0	11 500	+11 500
Worldwide Total	52 000	74 500	22 500
Convergent Plate Boundaries: Collision	Before	Transition	After
Zagros-Indus-Indonesia	2 500	7 500	7 500
New Guinea-Tonga	0	8 000	0
Caribbean	0	1 500	0
Mediterranean	0	11 500	11 500
Worldwide Total	2 500	28 500	19 000
All Divergent Plate Boundaries	Before	After	Net Change
East Pacific	16 000	16 000	0
Australia-Antarctic-Indian	3 000	12 000	+ 9 000
Indonesian	10 000	0	−10 000
Red Sea	0	2 000	+ 2 000
Atlantic-Arctic	19 000	20 000	+ 1 000
Tasman	2 000	0	− 2 000
Worldwide Total	50 000	50 000	0

* Measured with estimated 5% accuracy on 40 cm diameter globe.

probably do not penetrate basement but utilize the basal salt as a detachment surface. The resulting environment is known to be favorable to the migration and entrapment of hydrocarbons.

Australia

The formation of the South Australian margin was the result of a continental breakup between Australia and Antarctica. The breakup has been dated as Early to Middle Eocene (Fig. 4) by Weissel and Hayes (1972) and correlates well with the tectonic change in the rest of the Indian Ocean. Along this margin great thicknesses of sediment have been deposited — in places up to 10 km (Boeuf and Doust 1975). In the Great Australian Bight region much of the section is composed of Lower and Upper Cretaceous rocks. Apparently, rifting began in the Early Cretaceous and continued until the initiation of sea-floor spreading (magnetic anomaly 22) in Eocene time. Much of the early rift sedimentation ranges from fluvial to marine-deltaic. These sediments are separated from a thin Tertiary cover by a Paleocene-Eocene unconformity which corresponds well with the 'Breakup Unconformity' of Falvey (1974). As a result of this continental breakup, an environment has been formed which should be very favorable for hydrocarbons. Admittedly, much of the section and structure was developed prior to the onset of sea-floor spreading. But the rift sequence and its setting are directly related to the Australia-Antarctica breakup, which climaxed in Eocene times.

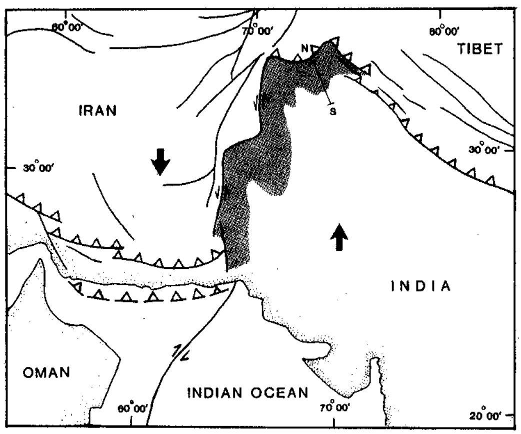

Figure 2. Present-day tectonics of southwestern Asia. Stippled pattern represents fold belt on the northwestern edge of the India plate. N–S line is location of cross-section of Potwar Basin in Northern Pakistan (Fig. 3).

New Guinea

Much of the present-day structure of New Guinea is influenced by a collision between the Australian Shield and the New Guinea island

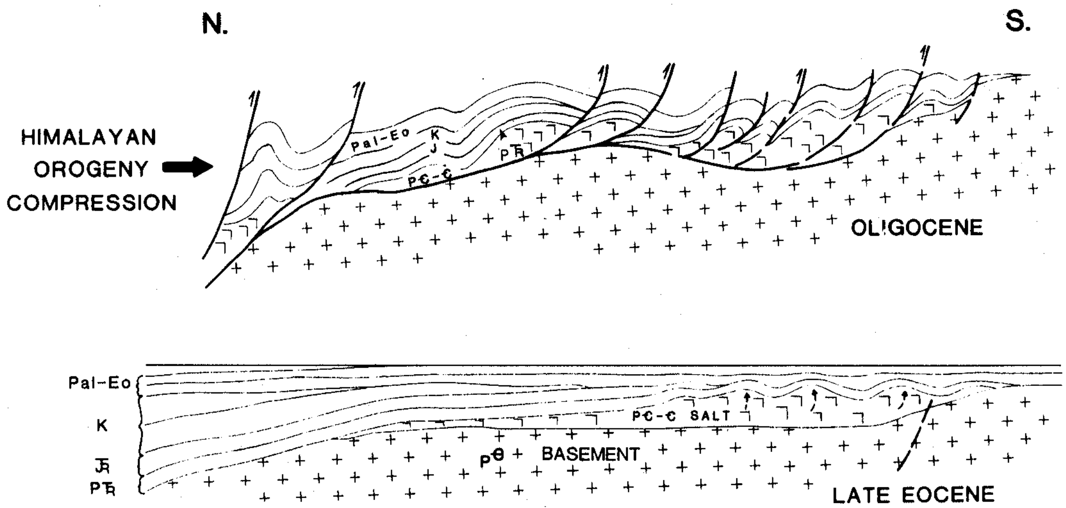

Figure 3. Schematic development of the Potwar Basin in late Eocene and Oligocene times. For location of cross-section, see Fig. 2.

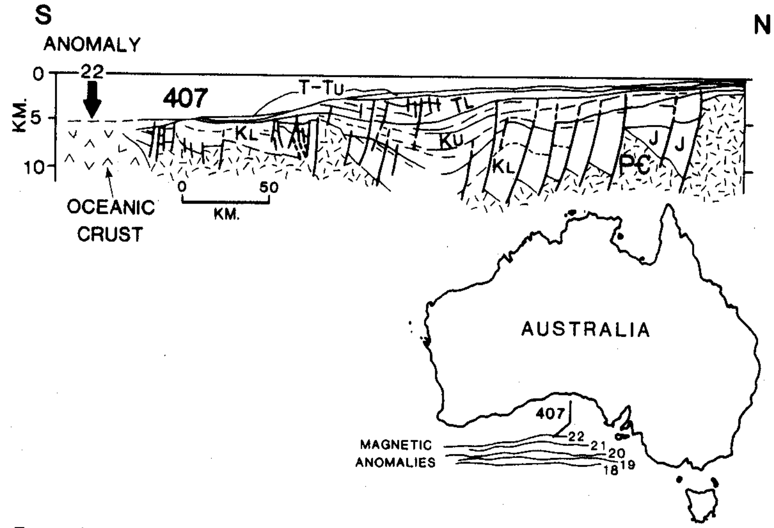

Figure 4. Interpretive line drawing of a seismic section in the Great Australian Bight of Australia (Boeuff and Doust 1975). Magnetic anomalies from Talwani *et al.* (1979).

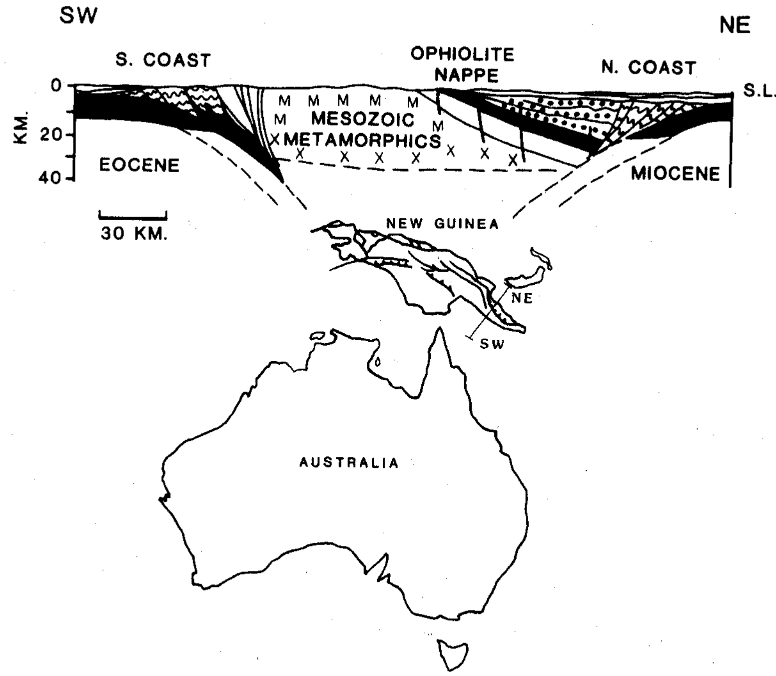

Figure 5. Inferred cross-section across Eastern Papua (Brookfield 1977). Black layer represents oceanic crust.

arc during the Eocene (Brookfield 1977). A thick section of Tertiary sediments accumulated south of Eastern Papua after the collision. By the Late Oligocene, subduction had reversed its polarity and began north of the island (Fig. 5). In northeastern Papua, a thick sequence of graded turbidite sandstones was also deposited. Both of the above-mentioned sediment accumulations are results of the collision of New Guinea with the Australian Shield and the subsequent polarity reversal of the subduction zone.

Cuba

The terminal event of Cuba's Laramide Orogeny is known to be Middle Eocene in age (Khudoley and Meyerhoff 1971). Mattson (1973) suggested that this orogeny was strongly influenced by a collision of the Greater Antilles island arc (of which Cuba is the westernmost member) with the Bahama carbonate platform.

Sediment thickness in the southern Bahamas may be as great as 10 km (Meyerhoff and Hatten 1974). Bahamian-type carbonates also exist in northern Cuba. However, owing to close proximity to the collisional boundary, their structure is well developed. Much of the pre-Middle Eocene section is folded and imbricated by south-dipping thrust faults. Triassic-Jurassic salt diapirs seem to have risen along these inclined planes (Fig. 6).

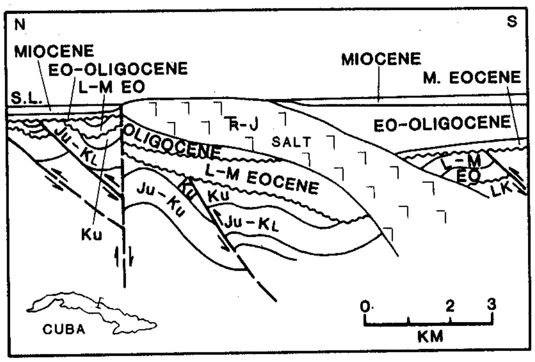

Figure 6. Geologic cross-section across the Punta Allegre diapir, north central Cuba (from Meyerhoff and Hatten 1968).

Fractured carbonate reservoirs overlying mobilized salt may provide excellent sites for oil or gas entrapment. Such conditions are reminiscent of the Reforma trend of Mexico.

Gulf of Honduras

The region surrounding the Gulf of Honduras (Central America) has had a complex geological history. Evidence exists in Guatemala of a Late Cretaceous continental collision between North America and Nuclear Central America along the Motagua fault trend (Donnelly 1977). However, the Motagua and Polochic fault zones also share the present strike-slip boundary between the Caribbean and North America plates (Fig. 7). The major change in this region from N–S collision to E–W strike-slip occurred in Eocene time (Malfait and Dinkelman 1972). The same age has also been assigned to the initiation of left-lateral movement in the Cayman Trough (Holcombe *et al.* 1973).

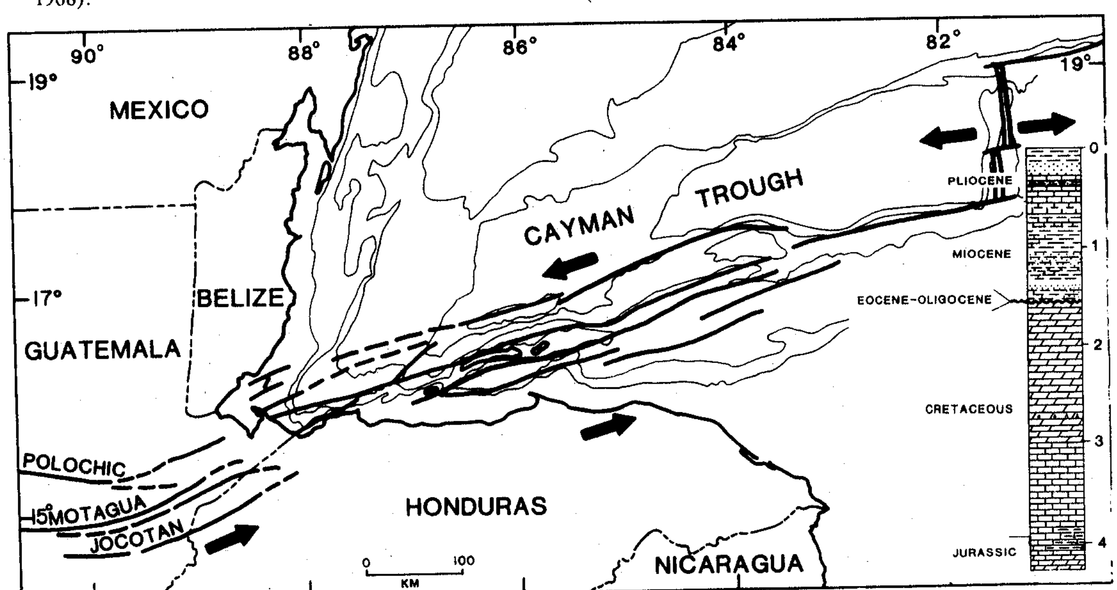

Figure 7. Fault trends in the southern Gulf of Honduras. Arrows indicate relative motion of North America and Caribbean plates. Stratigraphic column is typical of the northwest Honduras shelf. (Numbers on column represent thousands of feet.)

Plate reorganization in this region has produced a number of conditions which favor the occurrence of oil and gas. Mesozoic carbonates on the northern margin of Guatemala and Honduras were strongly affected by Laramide deformation. As a result, fracture porosity is common in many areas, and the top of the carbonates exhibits high relief.

An Eocene-Oligocene unconformity separates the Mesozoic carbonates from an upper unit of Tertiary clastics, which were deposited as a result of subsidence following the change from collision to strike-slip.

This sequence of deformed carbonates, overlain by 1 to 2 km of clastics, could very well serve as a reservoir for major hydrocarbon accumulations.

CONCLUSIONS

Numerous examples have been cited concerning a major global plate reorganization which took place largely during Eocene time. We have demonstrated that this reorganization significantly affected a number of the world's sedimentary basins in a variety of different ways, and in many cases in a manner favorable for the occurrence of oil and gas. It is hoped that this approach to exploration may prove useful, as it narrows the search for petroleum in space and time by allowing the explorationist to focus more closely on significant periods in basin development.

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UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 37-45

AEROMAGNETIC INVESTIGATIONS AND SEA FLOOR SPREADING HISTORY IN THE LAU BASIN AND NORTHERN FIJI PLATEAU

NORMAN Z CHERKIS

ABSTRACT

Recent aeromagnetic investigations flown over the northern Fiji Plateau and Lau Basin between 13°S and 23°S suggest that these features represent back-arc spreading processes. Tracks were flown perpendicular to the normal strike of the topography with a 5 km to 18 km track spacing in a fan-shaped pattern to allow for the arcuate nature of the basins. Residual anomaly profiles suggest that the Fiji Plateau and Lau Basin are presently active marginal basins. The present tectonic setting of the basins is that of a series of active spreading centers that are controlled by transform faults. By this means, extension in these two basins has taken place in an E–W direction on the western part of the Fiji Plateau changing to a N–S direction along the northern margin of the Fiji Plateau and in an E–W direction in the Lau Basin. Current spreading areas are highlighted by high-amplitude anomalies of up to 1000 gammas (peak-to-trough). The magnetic anomaly sequence 1 to 3′ has been identified, indicating an age of very late Miocene to early Pliocene for the opening of the basins. Peggy Ridge, the most prominent bathymetric feature in the Lau Basin, is characterized by a strong, broken, positive anomaly that is closely paralleled by negative lineations of −300 to −500 gammas.

INTRODUCTION

In a joint National Oceanic and Atmospheric Administration and U.S. Naval Research Laboratory project, an aeromagnetic investigative program was undertaken over the northern Fiji Plateau and the Lau Basin in April and May of 1979. The research platform was a Lockheed RP-3-A Orion patrol-type aircraft, into which was mounted a proton-precession magnetometer. The data were acquired at a rate of one data point each 2.5 s at an altitude of 330 m and an airspeed of 250 knots. Navigational control was maintained by two Litton model-72 inertial navigation systems which controlled the aircraft's autopilot and inserted an on-line position for each data point into a Hewlett-Packard 2100 mini-computer system. The average drift rate for the navigation was less than 1 km/h. Data were also stored in digital form on a disk cartridge and magnetic tape, and analog records and printouts were also made for back-up purposes in the unlikely event of primary system malfunction. The raw data were edited on board the aircraft during transits and a computer-generated plot of the data was made prior to departing the aircraft after each flight. In effect, a finished chart of each day's data was in hand for immediate analysis at the end of each flying day, as well as an updated composite of all track data. A total of 47 450 km of data track was flown on the 14 flights (Fig. 1), with an average flight time of 9.5 h per flight. The data presented here are based solely on our airborne magnetic measurements, as there was not sufficient time between the termination of the experiment and this meeting to make comparative studies between the marine data set presented by Weissel (1977) and our data.

BACKGROUND

The Lau Basin and Fiji Plateau are small marginal basins formed as a result of back-arc spreading behind a convergent plate boundary. Marginal basins were recognized by Karig (1971) as basins lying behind volcanic island arcs, in this case the Tonga Ridge (Fig. 2), and in isolated or semi-isolated locations. Additionally, the basins are now believed to have been created by crustal dilation (Lawver *et al.* 1976). Hawkins (1974) used petrologic analyses of dredge hauls to determine that the Lau Basin is floored by young oceanic tholeiites. The same typical oceanic basement can be inferred for the Fiji Plateau inasmuch as

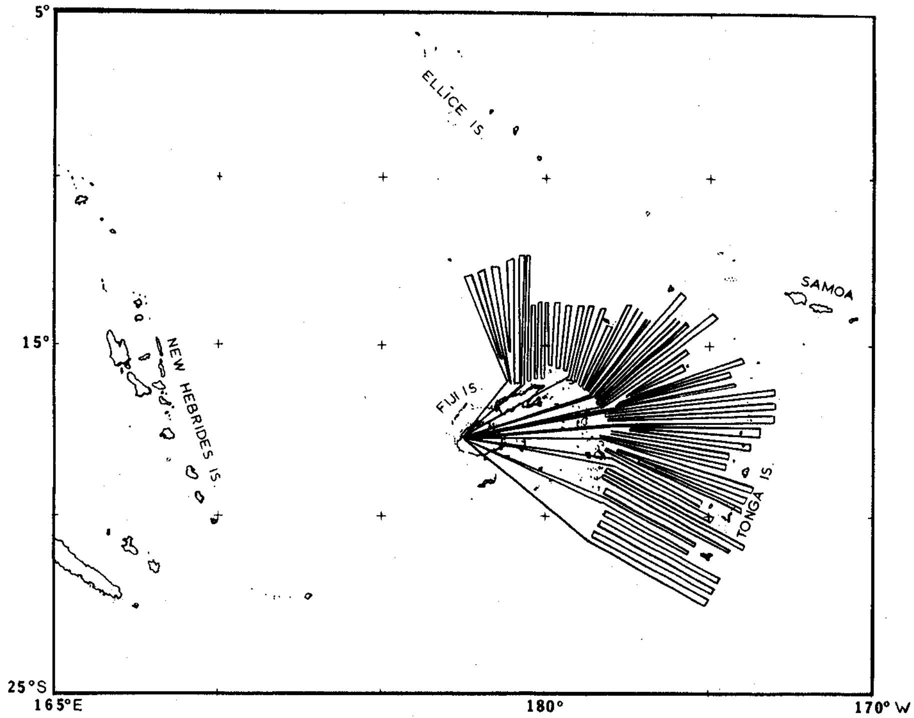

Figure 1. Flight tracks of 1979 aeromagnetic investigation over the northern Fiji Plateau and Lau Basin.

the plateau is directly related to the Lau Basin in origin (A. Malahoff 1979, pers. comm.).

The age of the basins is relatively young. Karig (1971) postulated a late Miocene date for the opening of the Lau Basin. However, geomagnetic data in the western and northern Fiji Plateau (Cherkis *et al.* 1978) revealed the age of the Fiji Plateau to be approximately 4½ million years old (early Pliocene), corresponding to anomalies 1 through 3 (LaBrecque, Kent and Cande 1977). Although basement was not attained at DSDP drill site 203 (Burns and Andrews 1973b; 22°09.22′S, 177°32.77′W) in the Lau Basin, early middle Pliocene strata were identified before the termination of the hole at 409 m corresponding to anomaly 2′ (3.3 mybp.). Extrapolating our aeromagnetic lineations southward, anomaly 3 would be encountered near the drill site indicating that agreement is possible between the drill site and our aeromagnetic age determinations. Although nearly 1 s (2-way travel time) of sediment was seismically encountered and over 400 m were penetrated at the drill site (Burns and Andrews 1973a, p. 20) it would be unwise to determine basin age through downward extrapolation of the sedimentation rates since sedimentation rates reported in the area appear to be highly variable (Karig 1970, Chase 1971, Griffin *et al.* 1972). The relatively small amounts of sediment generally found throughout the axis of the basin (Chase 1971, Hawkins 1974) further attest to the youth of the area.

DISCUSSION

The Fiji Plateau appears to have been created by extensional centers to the west and north of Viti Levu and Vanua Levu (Figs. 3 and 4). Characteristic residual magnetic anomalies obtained on an earlier experiment in 1977, flown to the west and northwest of Fiji (Fig. 3), are in the order of 500 gammas between 18°S and 14°S (Cherkis *et al.* 1978). The spreading center in this area is represented by a linear topographic high that is offset by northwest-trending fracture zones. At 15°30′S, 174°30′E, the spreading center turns almost

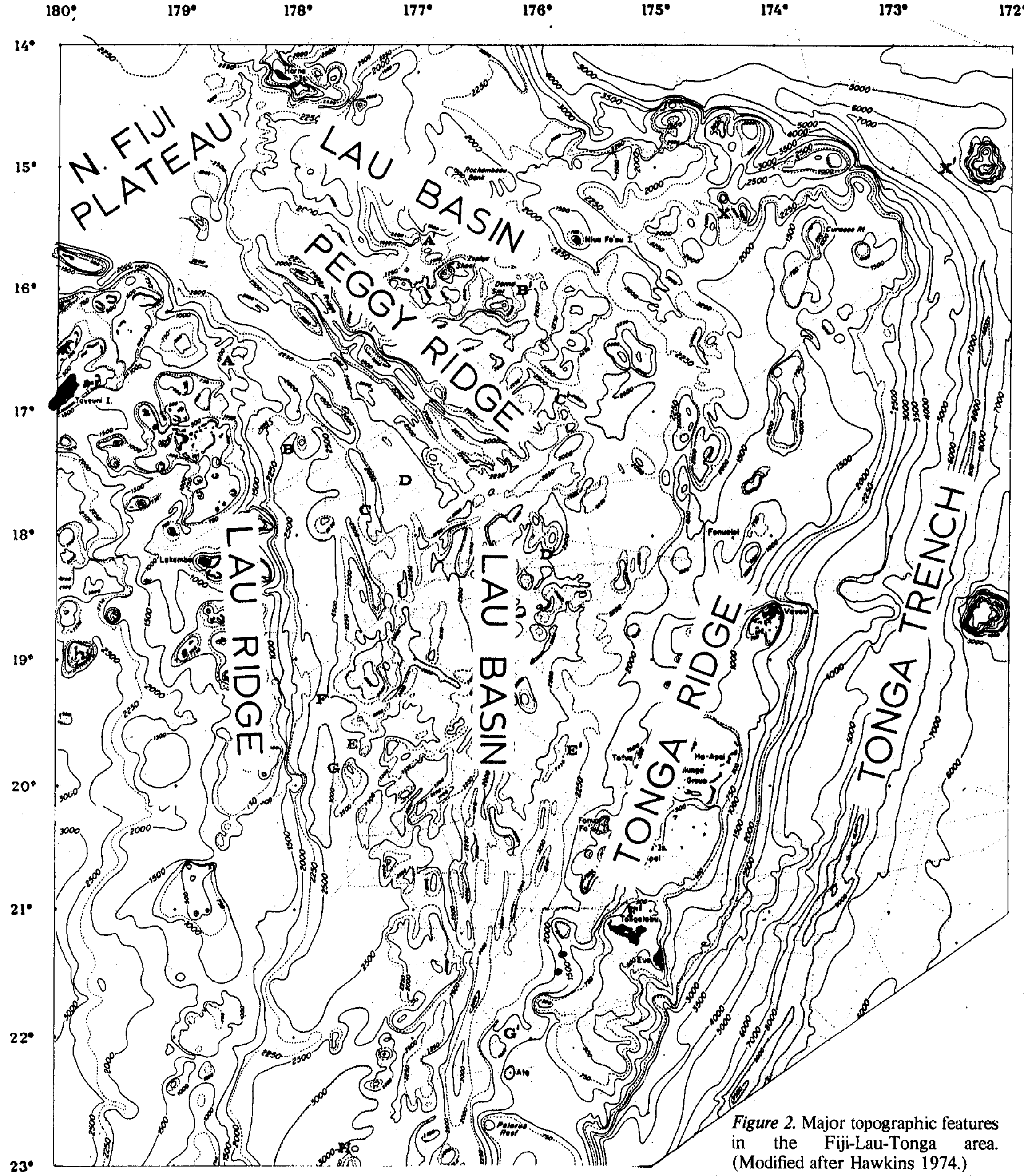

Figure 2. Major topographic features in the Fiji-Lau-Tonga area. (Modified after Hawkins 1974.)

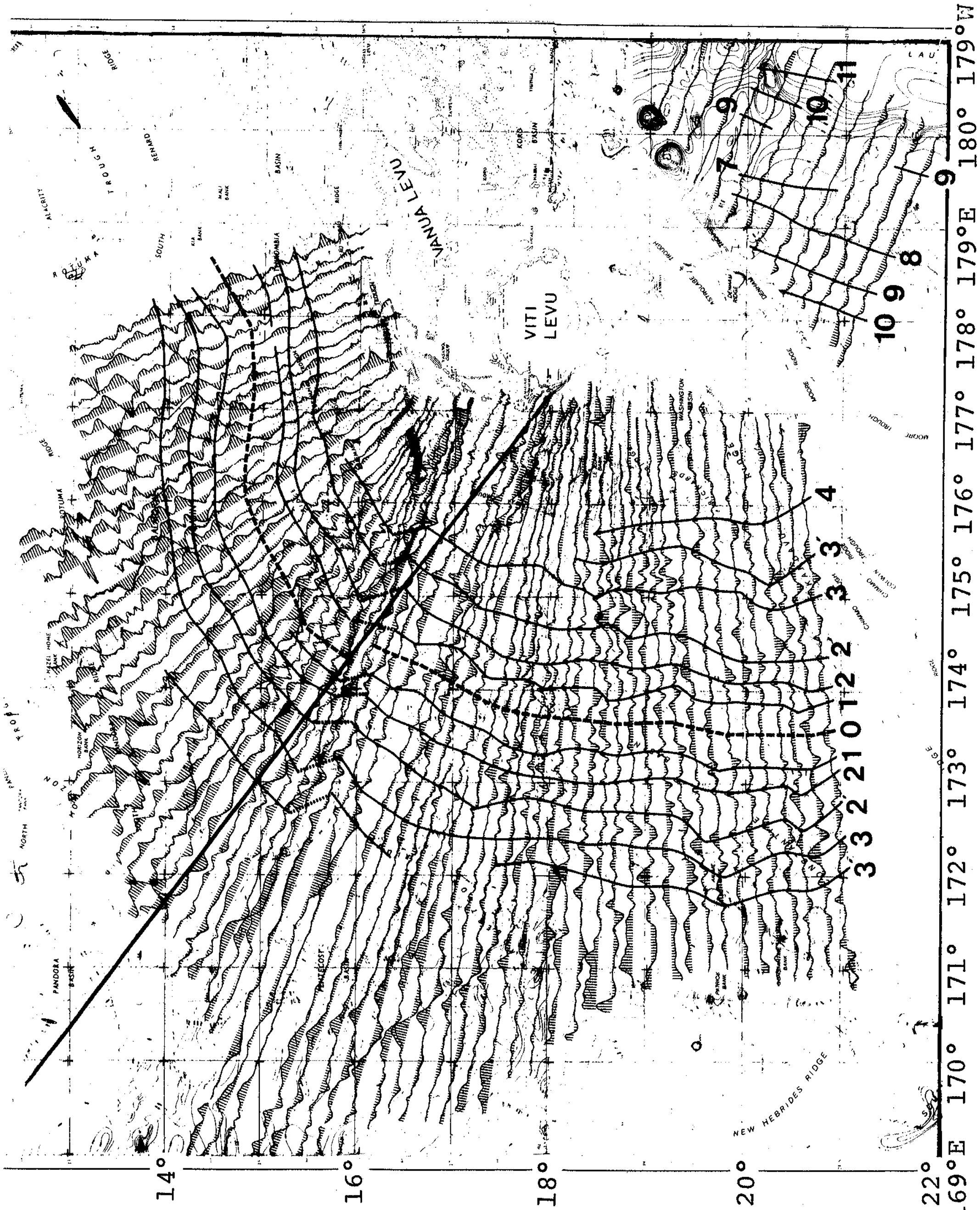

Figure 3. Magnetic anomaly lineations in the western and northern Fiji Plateau. Northwest-trending dark line indicates change of Matthews' sound-velocity correction zones.

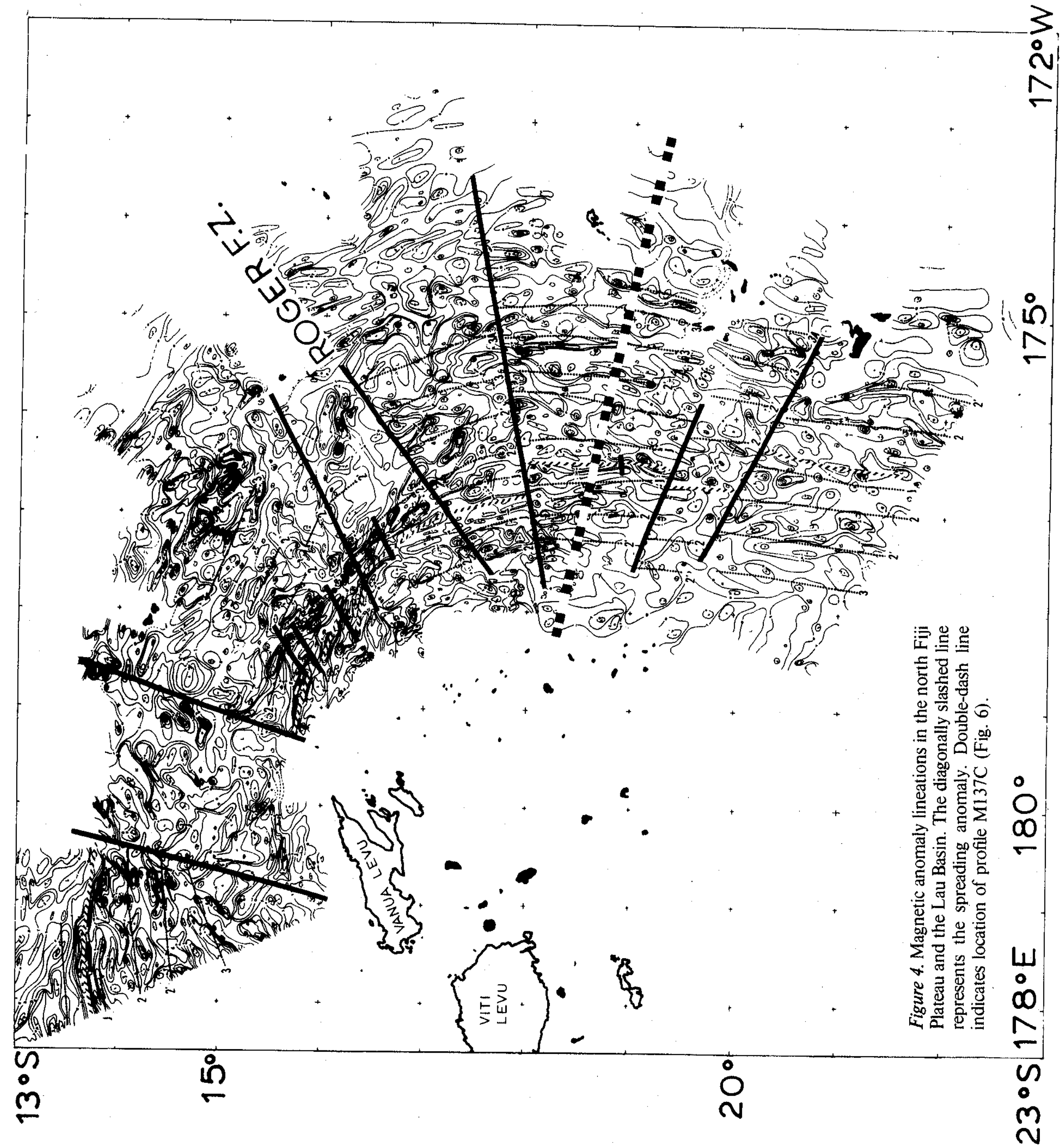

Figure 4. Magnetic anomaly lineations in the north Fiji Plateau and the Lau Basin. The diagonally slashed line represents the spreading anomaly. Double-dash line indicates location of profile M137C (Fig. 6).

abruptly eastward allowing for dilation in a N–S direction. The anomalies have been positively identified through anomaly 2′. Anomalies 3 and 3′ are questionable, and anomaly 4 is speculative. Anomalies in the area north of Viti Levu and Vanua Levu and south of the plate boundary that separates the Pacific and Austral-Indian Plates display residual magnetic anomalies of up to 1000 gammas (peak to trough). Since this area is a zone of high seismicity (Nagumo and Kasahara 1976), it is suspected that many high-amplitude anomalies are attributable to recent submarine volcanic activity.

Extending further to the east, the Fiji Plateau joins the Lau Basin (Fig. 2). This joining occurs at the point where the trend of the spreading center changes to a northwest-southeast trend, and then to a north-south trend to the east of Vanua Levu. The turning points are marked by fracture zones along which considerable offsets of the spreading ridge have taken place.

The Lau Basin was described as an interarc extensional feature by Karig (1970) and Lawver *et al.* (1976). It appears to be situated near the eastern edge of a slab exhibiting crustal growth from a spreading center located above a downgoing slab of Pacific Plate (Sclater *et al.,* 1972). Charts produced by Scripps Institution and the New Zealand Oceanographic Institute indicate that the Lau Basin contains a great number of minor bathymetric highs such as knolls and small seamounts. Contour mapping in the Lau Basin is very poor, however, because of a paucity of bathymetric soundings. The chartered topographic highs indicate the locations of most of the high-amplitude magnetic anomalies. Ridge-like features are also present, the most prominent of which is Peggy Ridge (Fig. 2).

Peggy Ridge is a linear topographic high that is clearly defined by shallow-focus earthquakes and high-amplitude magnetic anomalies. The ridge was first thought to be a spreading center (Chase 1971) and later a transform fault (Sclater *et al.* 1972). It has been the site of several scientific expeditions, primarily by Scripps Institution. When aeromagnetic lines that crossed Peggy Ridge were contoured, several fracture zones were revealed that terminate the ridge at both ends and intersect the ridge crest several times at angles oblique to the strike of the ridge (Figs 4 and 5). From these data and the epicenter locations it appears that Peggy Ridge is an active spreading center, as Chase (1971) first proposed.

The average residual magnetic anomalies in the Lau Basin are of the order of 200 to 300 gammas, excepting the high-amplitude magnetic anomalies associated with Peggy Ridge and isolated highs that are probably associated with submarine volcanoes. They exhibit a roughly N–S trend that is offset in several places by fracture zones normal to the direction of spreading. Minor offsets, as well as some large transform faults, were seen in the lineations.

Magnetic-anomaly lineations associated with the process of sea-floor spreading that have been recognized and dated along the northern Fiji Plateau (Fig. 4) are anomalies 1 through 2′ (0 to 3.3. mybp); anomalies 3 and 3A are questionable. The lineations trend east–west and are frequently offset. Displacement can be seen on the eastern side of a probable fracture zone that trends about N10°E from about 16°S, 179°E. To the east, the central anomaly lineation does not become apparent to about 15°50′S, 179°10′W, where it is flanked by very-high-amplitude negative anomalies of up to −500 gammas in amplitude. A probable fracture zone trending about N15°E cuts across the axis of the spreading anomaly near 16°S, 179°30′W. This fracture also marks the northern terminus of Peggy Ridge and marks an area where a change in the direction of the spreading ridge between the Fiji Plateau and the Lau Basin is apparent. Indeed, the whole length of the currently active spreading center appears to contain offsets, giving credence to Weissel's (1977) theory of the growth of the Lau Basin floor by small plates.

Weissel (1977) suggested that after the initial opening of the Lau Basin by the bifurcation of an incipient Tonga ridge at anomaly 3A time into the present day Lau and Tonga Ridges, the newly formed basin experienced growth at a half rate of 3.8 cm/y for about 1.5 million years. At about anomaly 2′ time, the accretionary plate boundary shifted from the Lau Ridge margin to the central part of the Lau Basin where it is presently active. Further, Weissel suggested the presence of

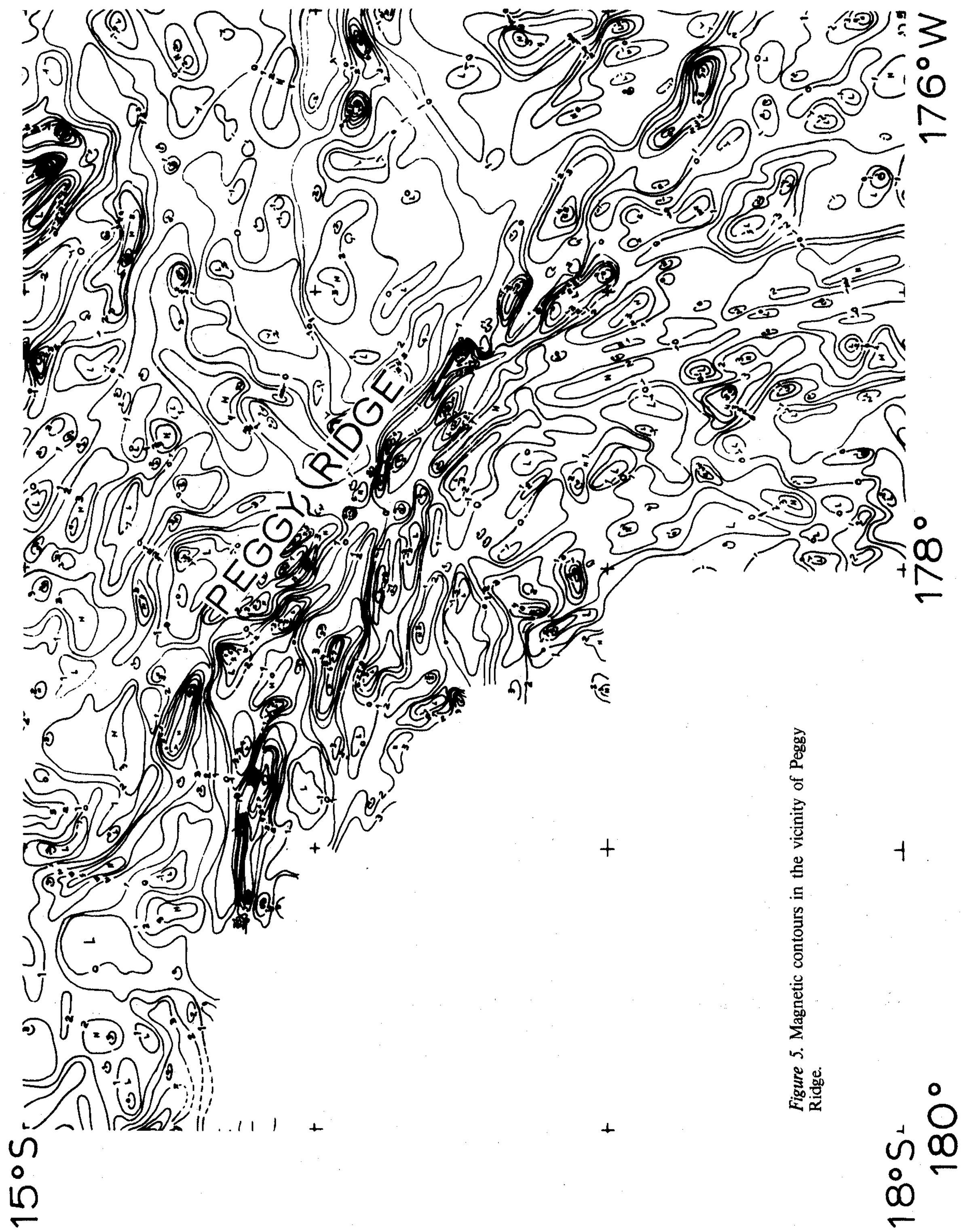

Figure 5. Magnetic contours in the vicinity of Peggy Ridge.

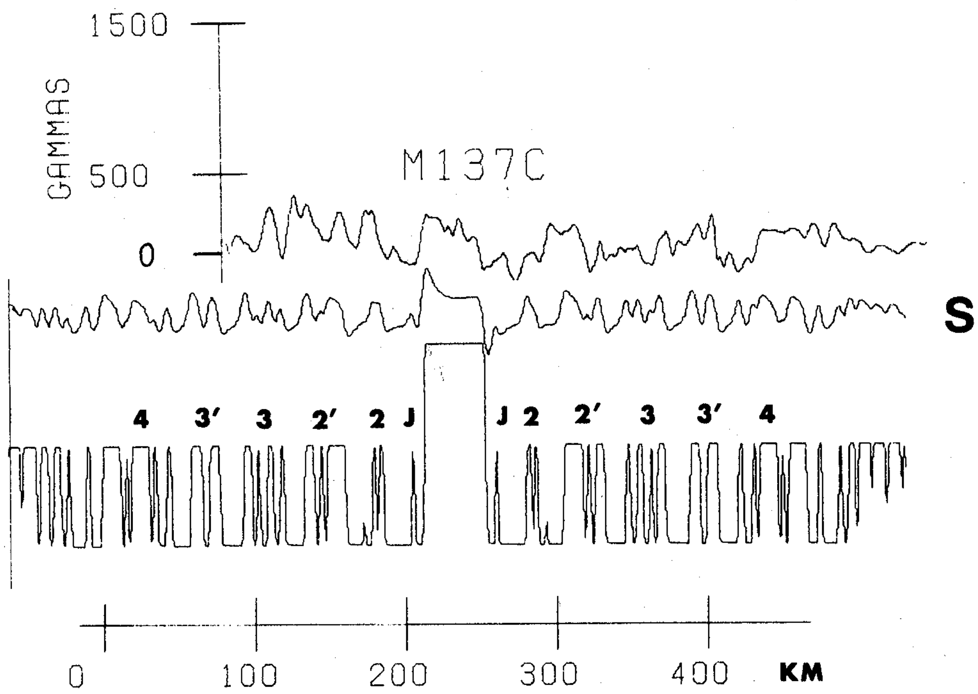

Figure 6. Representative magnetic profile in the Lau Basin. S = Synthetic profile, based on Schouten and McCamy (1972) model.

Roger Fracture Zone, which formed the third arm of an RFF (ridge-fault-fault) triple junction with Peggy Ridge and the spreading ridge about 3.5 million years ago. Roger Fracture Zone, according to Weissel, struck about N40°E from the southern terminus of Peggy Ridge. However, little bathymetric or seismic expression of this fracture zone exists, and magnetic interpretation is tenuous with respect to the fracture zone being a part of a triple junction. Instead, the presence of a transform fault is here proposed, which terminates Peggy Ridge and truncates the magnetic lineations on the north side of the transform. Roger Fracture Zone exists, therefore, not as an arm of a triple junction, but rather as a transform fault that offset the spreading ridge to a minor degree. The timing for the start of this activity probably coincided with the onset of the major rotation of the Fiji platform (about 4 mybp) (A Malahoff, pers. comm.), which would allow for the directional change of the spreading ridge from the north to the northwest, and at the same time allows for leakage at the northern end of the transform. A gradual widening of the distance between the anomalies can be seen on the south side of Peggy Ridge where the Lau Basin begins to exhibit normal growth once the curvature caused by the rotation of the insular platform loses its influence on the shape of the basin.

The magnetic contour chart provides a useful base to use in following the anomalies. Using the residual profiles as a reference, the spreading anomaly was identified (Fig. 4, diagonally slashed line). Other lineations are also seen as well as the major offsets in the anomaly trends. A plot of the residual anomaly profiles on a mercator projection shows a great degree of continuity of the anomalies, as well as confirming the existence of the offsets.

Using the standard Schouten and McCamy (1972) model for the synthetic comparisons,

Fig. 6 is given as an example of one track in the southern Lau Basin. The correlation is quite clear, and corresponding age data are easily applicable. The present spreading half-rate based on these data is between 2.5 and 2.8 cm/y.

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SEISMIC REFRACTION STUDIES IN THE NEW HEBRIDES AND TONGA AREA

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ABSTRACT

Seismic refraction surveys have been conducted over the convergent plate boundaries of the New Hebrides and Tonga, in a joint programme by the Office de la Recherche Scientifique et Technique Outre-Mer (Centre de Nouméa) and the University of Texas (Institute of Marine Science). From 24 profiles taken during the EVA II, EVA IV and EVA VII cruises, it appears that

1. In spite of overall similarities, differences in the shallow structure of the two arcs can be seen. In the New Hebrides the low-velocity layers are much thicker than in Tonga, especially in the lower part of the inner slope of the trench. This difference in thickness can be correlated with the differences in thickness of the transition and oceanic layers of the crust of the dipping plate.
2. Refraction leaves uncertainty as to the structure at depth, particularly as to the joining of the arc and the back-arc basins. The existence of 7.6–7.7 km/s velocity layers complicates the interpretation in classical terms of crust and mantle. One possible interpretation of the evolution of the crust under the island arcs could be a thinning-down in time.

INTRODUCTION

Since 1976 a joint research programme has been operating in the Southwest Pacific, with the participation of the Office de la Recherche Scientifique et Technique Outre-Mer (Centre ORSTOM de Nouméa), Cornell University, the University of Texas (Marine Science Institute) and the National Ocean Survey of NOAA; the naval facilities used were provided by the Centre National pour l'Exploitation des Océans. Part of this programme involved seismic-refraction measurements across the convergent plate boundaries of the New Hebrides and Tonga (Fig. 1).

Together the two arcs form a double zone of convergence with to the west a dip eastwards and to the east a dip westwards under the Tonga arc. On the edge of the Australo-Indian plate, at the level of the area under study, lies a marginal basin, generally called the North Loyalty Plateau or Basin. Its age was determined from core samples JOIDES 286 (Andrews *et al.* 1975) and confirmed by the existence of magnetic anomalies (Lapouille 1978) as being Middle Eocene. The Pacific plate, which dips under the Tonga arc, is covered by an oceanic crust dating from Early Cretaceous (Burns *et al.* 1973). The two arcs also differ in age: it has been estimated that in the New Hebrides subduction began, in the present position, 5 m.y. ago (Carney and Macfarlane 1977), but in Tonga 45 m.y. ago (Gill 1976).

Earlier refraction measurements were taken during the expeditions *Capricorn* (1952) and *Nova* (1966-1967) of the Scripps Institution, the results of which were detailed by Raitt *et al.* (1955), Raitt (1956), and Shor *et al.* (1971). The measurements, taken during large-scale exploratory expeditions, dealt with greatly varying structures. It was therefore of interest to study the structures in more detail.

FIELD OBSERVATIONS AND DATA ANALYSIS

Twenty-four seismic refraction profiles were taken during the cruises EVA II, EVA IV and EVA VII from 1976 to 1978, 18 in the New Hebrides (Fig. 2) and six in the Tonga area (Fig. 3). The profiles were planned to parallel the structural units in such a way as to provide two cross sections of the New Hebrides trench system and one of the Tonga-arc trench system.

Three different sources were used: airguns (5- and 15-litre capacity), explosives, and Flexichoc. With the airguns, distances were obtained of 15–20 km (5-litre capacity airgun) and 40 km (15-litre capacity airgun). The

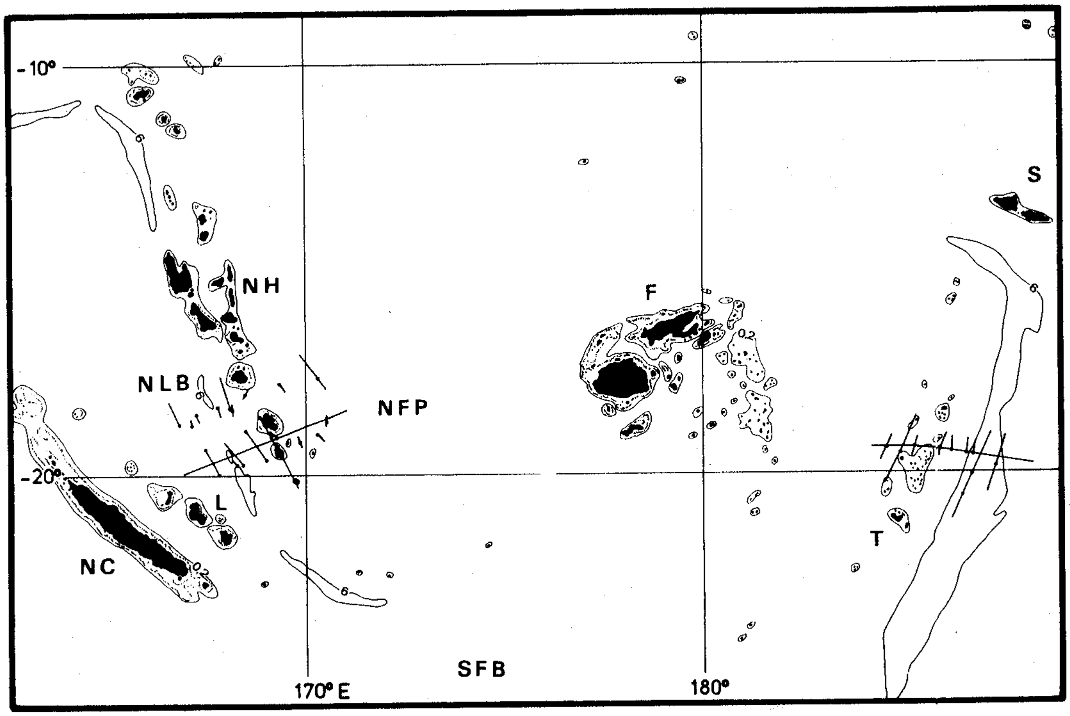

Figure 1. Location of refraction profiles. Bathymetric contours of 0.2 and 6 km. NC, New Caledonia; NLB, North Loyalty Basin; NH, New Hebrides; NFP, North Fiji Plateau; SFB, South Fiji Basin; F, Fiji; T, Tonga; S, Samoa.

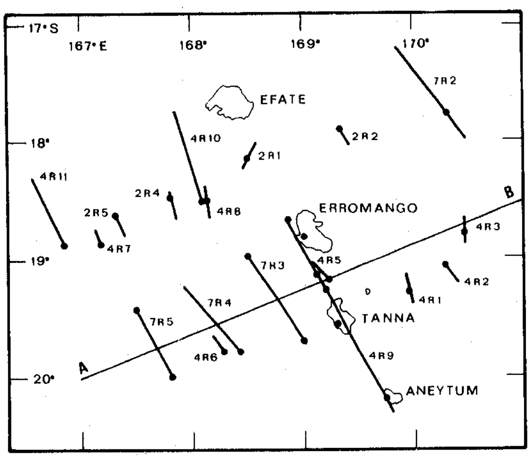

Figure 2. Location of the profiles across the New Hebrides subduction zone. All the profiles are projected on AB cross section (Fig. 5).

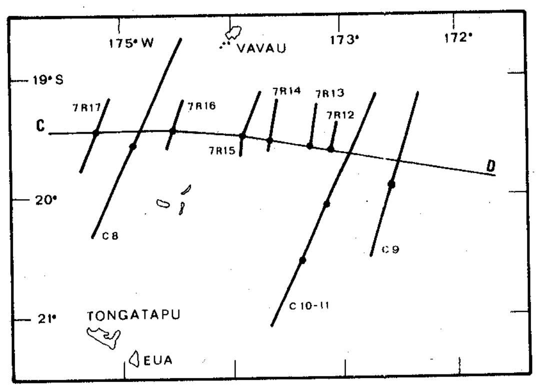

Figure 3. Location of the profiles across the Tonga subduction zone. All the profiles are projected on CD cross section (Fig. 6). C8, C9 and C10-11 profiles were shot on the Scripps Institution 1952 *Capricorn* Expedition.

Flexichoc, a high resolution implosion seismic source, developed by the Institut Français du Pétrole, is perfectly adapted to seismic reflection because of its bubble-effect free signal; however, the technique was found to be inefficient at the recording stations used and its range was only 8–9 km. For the explosives (gomme F 15), the charges varied between 1 kg and 200 kg according to the distance from the station. Calculations were made using the empirical experimental formula:

$$d = 22.1 \sqrt{P}$$

where P is the charge in kg and d the distance in km.

The stations used were the Ocean Bottom Seismographs (OBS) built by the University of Texas (Institute of Marine Science) and described by Latham *et al.* (1978) and Ibrahim and Latham (1978).

Four types of profiles were taken (Fig. 4):

Type (1): single profile, which has the advantage of being done quickly and with only one OBS. However, it gives no indication of the dip of the layers.

Type (2): split profile, which takes longer to do, but has the advantage of giving true velocities and of giving the dip of the layers assuming uniform velocity layers with constant dip.

Type (3): reversed profile, which is done with two OBS. It gives true velocities and the dip for the deep layers. If the velocity in the shallow layers varies from one OBS to another, it will be defined for plane horizontal layers only.

Type (4): compound profile, which gives the inclined shallow structure under each OBS.

The characteristics of the different profiles and the results obtained are shown in Table I.

Data analysis techniques used were the classic ones: the sequences were played back on paper and collated according to a time fixed by the firing time, and for a distance either from the navigation if that was precise enough (radar and bearing) or from theoretical hodochrone of sound propagation in the water. The arrival times of correlatable phases were corrected for topographic effects by bringing the penetration points of the rays to the same depth as the OBS, using the classic formulae of plateau correction.

The arrival times observed were then linked by segments, the parameters of which were obtained by fitting to least squares. A model assuming uniform velocity in each layer was then constructed in accordance with the observations. In fact, the hodochrones obtained almost never had the curve characteristics of the hodochrones corresponding to layers with velocity gradient.

RESULTS AND DISCUSSION

The results obtained are shown on the two cross sections AB and CD (Figs. 2, 3, 5, 6). In fact, although in the New Hebrides the profiles were done on two cross sections, the results were plotted on a single profile. For ease of comparison, the Tonga cross section was reversed (east on the left). Below, the different

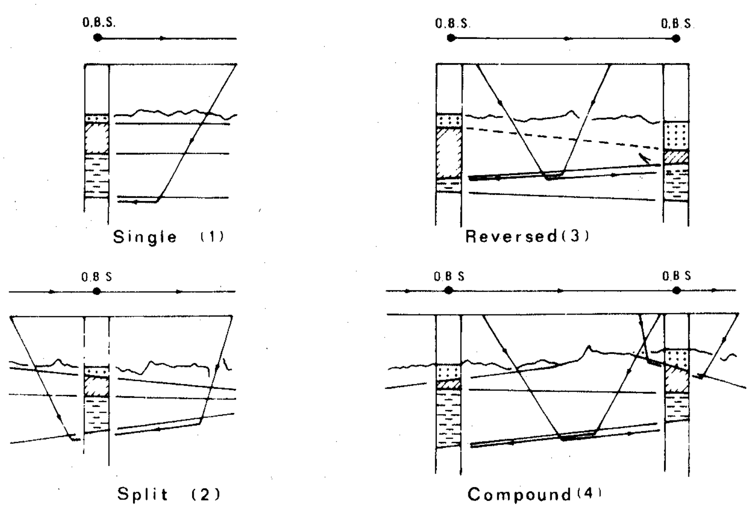

Figure 4. Combinations of linear refraction profiles.

TABLE 1

Seismic Refraction Profiles

Note. For each group of refractors, V = apparent velocities, T = thickness, t = intercept; velocities marked * are defined either from second arrivals or on the corresponding multiple, those marked † are assumed.

Profile No. and Type ( )	Position of Recording Station	Source	Water Depth km	Az.	L km	Sediment (1) V km/s	Sediment (1) T km	Sediment (1) t s	Transition (2) V km/s	Transition (2) T km	Transition (2) t s	Oceanic (3) V km/s	Oceanic (3) T km	Oceanic (3) t s	Mantle (4) V km/s	Mantle (4) T km
					NEW HEBRIDES (EFATE–ERROMANGO)											
2R1(2)	18°07.8 S	AG 5	0.81		14	1.8*	0.03	0.30	5.5		1.84					
	168°31.0 E					2.4	0.18	0.47								
						2.6	1.09	0.54								
						4.3	1.29	1.39								
2R2(1)	18°02.0 S	AG 5	2.61		13	2.2*	0.12	1.28				6.0		2.71		
	169°26.0 E					2.7	0.43	1.52								
						3.5	1.31	1.87								
2R4(1)	18°27.0 S	AG 5	4.32		14	1.9*	0.02	1.77	4.7		3.97					
	167°50.0 E					2.1*	0.60	1.98								
						4.1	3.00	3.20								
2R5(1)	18°36.0 S	AG 5	4.83		13	2.0*	0.30	2.08	5.4	0.48	4.37					
	167°20.8 E					2.4*	1.35	2.72	5.6		4.40					
4R7(2)	18°51.8 S	Flex.	4.31		12	2.0*	0.20	1.90	5.3		3.37					
	167°14.0 E					3.6	1.08	2.78								
4R8(2)	18°28.7 S	AG 5	2.80	N	17	2.9	0.60	1.59	5.2		2.62					
	168°09.8 E					4.1	1.60	2.04								
				S	18	3.2	1.39	1.67	5.3		2.52					
					100				5.1	3.91	2.95	6.1	8.06	3.91	8.1	6.05
4R10(1)	18°30.8 S	Dyn.	2.86													
	168°08.0 E															
4R11(1)	18°51.9 S	Dyn.	4.16		75	4	2.73	2.82				7.0	7.76	9.1	8.1	5.
	166°53.4 E															
7R2(2)	17°46.0 S	Dyn.	2.56	N	83	2.0*	0.55	1.13	5.3	0.85	2.30	6.7	5.75	2.57	7.81	3.58
	170°20.0 E	AG+15				4.2*	0.50	2.05								
				S	30	2.0*	0.55	1.13	5.2	1.55	2.30	6.3		2.70		
						3.9*	0.43	2.05								

NEW HEBRIDES (ERROMANGO–TANNA)

4R1(2)	19°17.0 S	AG 5	2.39	N	20	2.5	0.73	1.23	5.9		3.13					
	169°59.0 E					3.4	2.27	1.77								
				S	16	2.1*	0.33	1.10	5.9		3.13					
						3.6	3.00	1.65								
4R2(1)	19°05.5 S	IG 5	2.12		17	1.8*	0.32	0.72	5.5		2.93					
	170°17.0 E					2.4*	0.44	1.23								
						3.1	1.96	1.64								
4R3(2)	18°47.0 S	AG 5	3.01	N	18	2.0*	0.48	1.21	5.3	3.64?	2.89				7.6?	4.00?
	170°30.0 E					3.2	1.22	2.05								
				S	6	2.0	1.48	1.26	5.0		3.1					
4R5(1)	19°00.0 S	AG 5	0.97		18	2.0	0.60	0.44	5.0		2.28					
	169°10.0 E					3.4	2.66	1.06								
4R6(2)	19°47.5 S	AG 5	5.12	N	12	2.1*	0.21	2.26								
	168°18.6 E					2.8	0.66	2.96								
						4.1		3.60								
					18	2.1*	0.27	2.30								
						2.6	1.00	2.85								
						3.8		3.83								
4R9	(see Fig. 1)	Dyn.			170				5.0	4.11	2.28	6.6	18.5	3.48	7.9	7.08
7R3(4)	(N)18°57.2 S	Dyn. +	1.00	N	10	2.2*	0.2	0.52								
	168°30.0 E	AG 15				4.5		0.80								
				S	95	2.4*	0.2	0.52	5.3*	0.90	0.93	6.2	2.18			
						4.9	1.50	0.70	5.6	5.60	1.08					
	(S)19°42.0 S	Dyn. +	1.00	N	95	2.1	1.25	0.46	5.5	1.95	2.01	6.9	8.15	3.45	7.7	4.40
	169°03.0 E	AG 15				4.5*	0.62	1.65	5.8	3.82	2.30					
						4.9*	0.86	1.80								
				S	30	2.1*	1.22	0.46	5.2	1.30	2.01					
						4.4	1.12	1.75	5.6		2.60					
						4.7*	0.55	1.80								
7R4(1)	19°47.0 S	Dyn. +	5.30	N	82	2.5*	0.35	2.90				7.2	6.70	7.12	8.1	8.47
	168°27.0 E	AG 15				3.7*	0.72	3.54								
						4.7	8.75	3.96								
7R5(3)	(N)19°24.8 S	Dyn.	4.80	S	72	2.0	0.55	2.08	5.2	1.50	3.69	7.0	5.5	4.88	8.3	6.03
	167°30.9 E					3.7*	0.65	3.37	5.9	3.35	4.05					
	(S)19°59.5 S	Dyn.	4.50	N	70	2.8	0.54	2.54	5.4	2.08	3.75	7.0	7.63	4.46	8.3	5.81
	167°50.0 E					4.2	1.80	3.09								

TABLE 1 *(Contd.)*

Profile No. and Type ( )	Position of Recording Station	Source	Water Depth km	Az.	L km	Sediment (1) V km/s	Sediment (1) T km	Sediment (1) t s	Transition (2) V km/s	Transition (2) T km	Transition (2) t s	Oceanic (3) V km/s	Oceanic (3) T km	Oceanic (3) t s	Mantle (4) V km/s	Mantle (4) T km
							TONGA									
7R12	19°37.3 S	AG 15	6.55		27	2.54	0.76	3.52	4.17	2.55	4.92					
	173°08.8 W								4.82	5.7						
7R13	19°35.9 S	AG 15	4.91		38	2.43	0.91	2.56	4.74	1.30	3.75	7.04		5.26		
	173°21.0 W								5.71	4.02	4.20					
7R14	19°35.2 S	AG 15	3.65		53	1.99	0.3	1.60	6.0	1.20	3.20	6.6	7.39	4.13		
	173°41.8 W					2.68	0.7	2.22				7.6		5.29		
						3.84	1.74	2.87								
7R15	19°27.8 S	AG 15	2.01	S	15	2.87	1.25		4.39	1.28	2.58					
	173°54.9 W					3.38	0.99		5.99		3.20					
			2.01	N	46	2.1	0.72	0.90	4.55	1.51	2.58	6.72	2.84	3.35		
						2.76	1.23	1.57	6.33	0.88	3.21	7.5		3.84		
7R16	19°23.7 S	AG 15	0.51	S	17	2 †	0.35	0.25	5.01		1.51					
	174°31.1 W					2.87	0.83	0.54								
						3.87	1.19	1.00								
			0.51	N	27	2 †	0.39	0.25	4.28	2.33	1.50	7.20		3.20		
						2.67	0.79	0.54	5.97	2.61	2.52					
						3.53	1.19	1.01								
7R17	19°25.5 S	AG 15	2.19	S	50	2.15†	1.46		5.83	1.88	2.68	6.9	2.57	3.08		
	175°12.9 W											7.6		3.50		
			2.19	N	35	2.15†	1.24		5.20	1.73	2.45	6.2	1.04	2.86		
												6.8	1.20	3.08		

structural units are examined by comparing the results obtained on each of the arcs.

Outer oceanic basins (dipping plates)

Figure 7 presents the results for the New Hebrides on profiles 4R 7, 4R 11 and 7R 5, and for Tonga on profile C9 (*Capricorn* Expedition); also included is the structure of the standard oceanic crust of Ludwig *et al.* (1970). While the structure of the oceanic crust of the Pacific Basin near Tonga (T) is very similar to that of the standard oceanic crust (O), the crust of the North Loyalty Basin (NH) is considerably thicker, as the depth of the Moho there, in relation to sea level, can exceed 16 km (as against 12 km for the standard crust).

Comparing crust thickness in the North Loyalty Basin with that in the marginal basins of the West Pacific, it can be seen that some of the basins have quite thin crust, for example (as shown in Fig. 8) the Parece Vela Basin (PV) and the Philippine Basin (PH B) (Murauchi *et al.* 1968), in which the crust is thinner than the standard oceanic crust (O) and the depth of the Moho less than 10 km. In other basins, such as the South Fiji Basin (SFB), crust thickness exceeds 15 km. Further, the structure of the North Loyalty Basin crust is very similar to that of the South Fiji Basin, which supports the suggestion of Lapouille (1978) that there is just one basin. Another feature to be noted is that whereas the thickness of crust is greater near the New Hebrides than at Tonga (Fig. 7), the reverse is true for the lithosphere calculated by Dubois *et al.* (1977) from bulge parameters, i.e., 24 km in the North Loyalty Basin, but 34 km in the Pacific Basin near the Tonga Trench.

Inner wall of the trench

In the inner wall of each trench (Figs. 5 and 6) is a layer with velocity ranging from 4.7 to 5.3 km/s. The maximum thickness of this layer in the New Hebrides is 9 km, but only 3 km in Tonga. The difference in thickness parallels that of the two transition layers of the oceanic crust of the dipping plates: in the North Loyalty Basin the 5.3 km/s velocity layer is about 3 km thick, whereas the equivalent layer (5.1 km/s velocity) of the Pacific crust is only 0.5 km thick. This lends support to the hypothesis of accretion of material from the dipping plate on the inner wall of the trench, to a varying degree depending on the crustal structure.

The arcs themselves

In each arc there is a conspicuous rise in the deep layers (Figs. 5 and 6), 75 km from the trench axis in the case of the New Hebrides arc and 100 km away in Tonga. It could be said that the rise is in each case the limit of the arc, and it seems that the two morphologies are different. In the New Hebrides the rise can be seen even in the uppermost layers (4.9 km/s velocity); near Erromango there is even a frontal horst (the fore-horst of Dugas *et al.* 1977). In contrast, the uppermost layers in the Tonga arc are relatively regular; in particular, the rise of the basement is found well before the top of the arc (Fig. 6). It would therefore seem that in the New Hebrides vertical movements are more active.

The structure is also different under the arcs. Under the New Hebrides arc is an extensive layer, more than 15 km thick, of 6.6 km/s velocity. Under the Tonga arc the same layer is nowhere thicker than 10 km. In the New Hebrides on profiles 4R 9 and 4R 10 the velocity reaches 7.9 (4R 9) or 8 km/s (4R 10) under this layer. In Tonga the velocities are only 7.6 to 7.7 km/s. Reinterpreting the refraction measurements of the *Capricorn* Expedition (Raitt 1956) and taking into account gravimetric measurements, Talwani *et al.* (1961) evaluated the 7.6 km/s layer as having a thickness of 23 km, under which is found a 'normal' mantle. Our measurements did not reach the mantle, but it should be noted that all the profiles were shot with an airgun and therefore probably lacked power. According to this model the crust would be 36 km thick under the Tonga arc but 26 km, only, under the New Hebrides arc. As we may not completely exclude, from under the New Hebrides arc, the existence of a greater than 6.6 km/s layer (this layer could have been missed because of a velocity gradient), the two crustal structures could be comparable, the older arc (Tonga) having a thicker crust than the younger one (New Hebrides).

However, the existence of low velocities in the uppermost part of the mantle under the

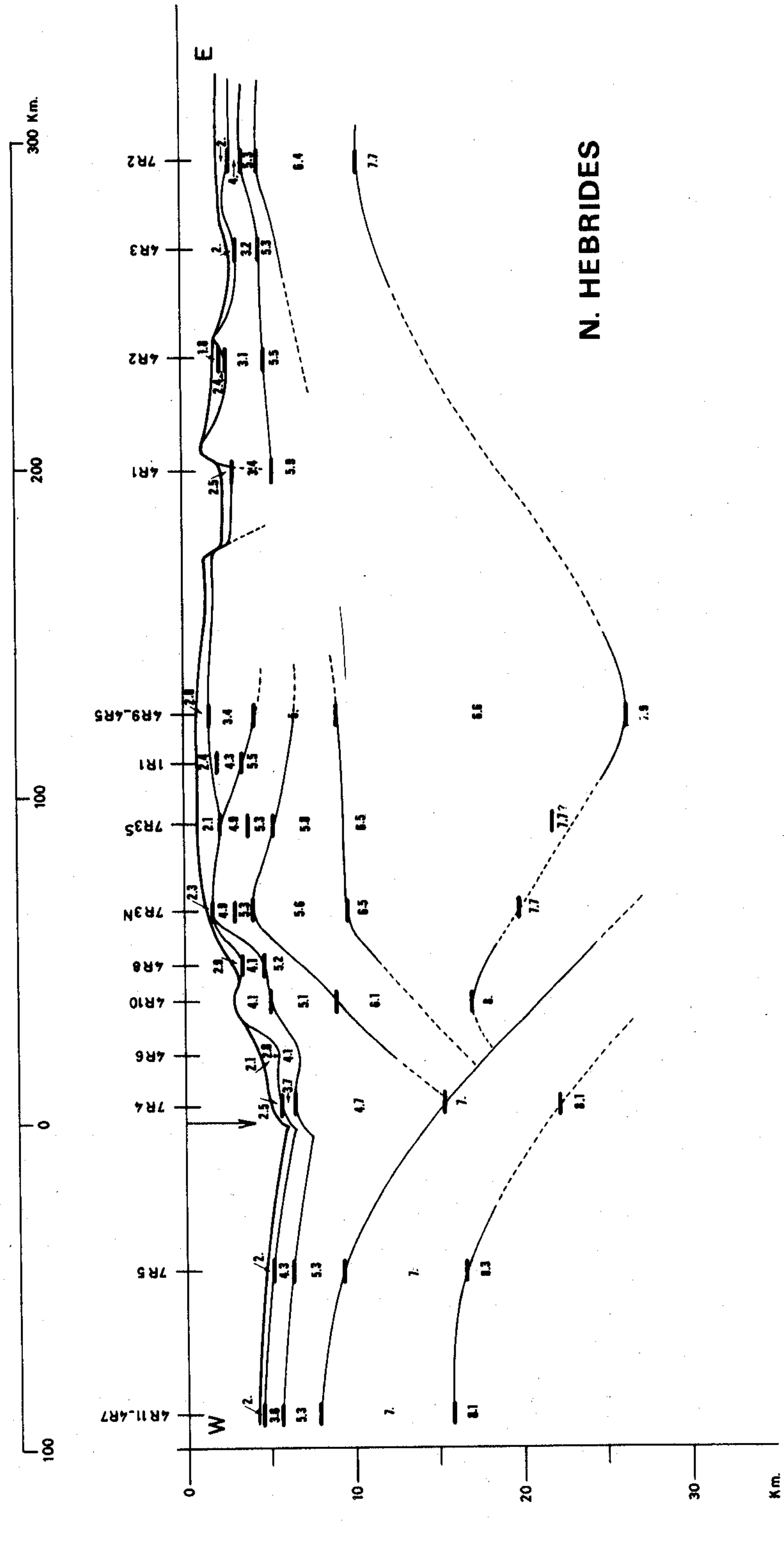

Figure 5. Structure section AB across the New Hebrides subduction zone.

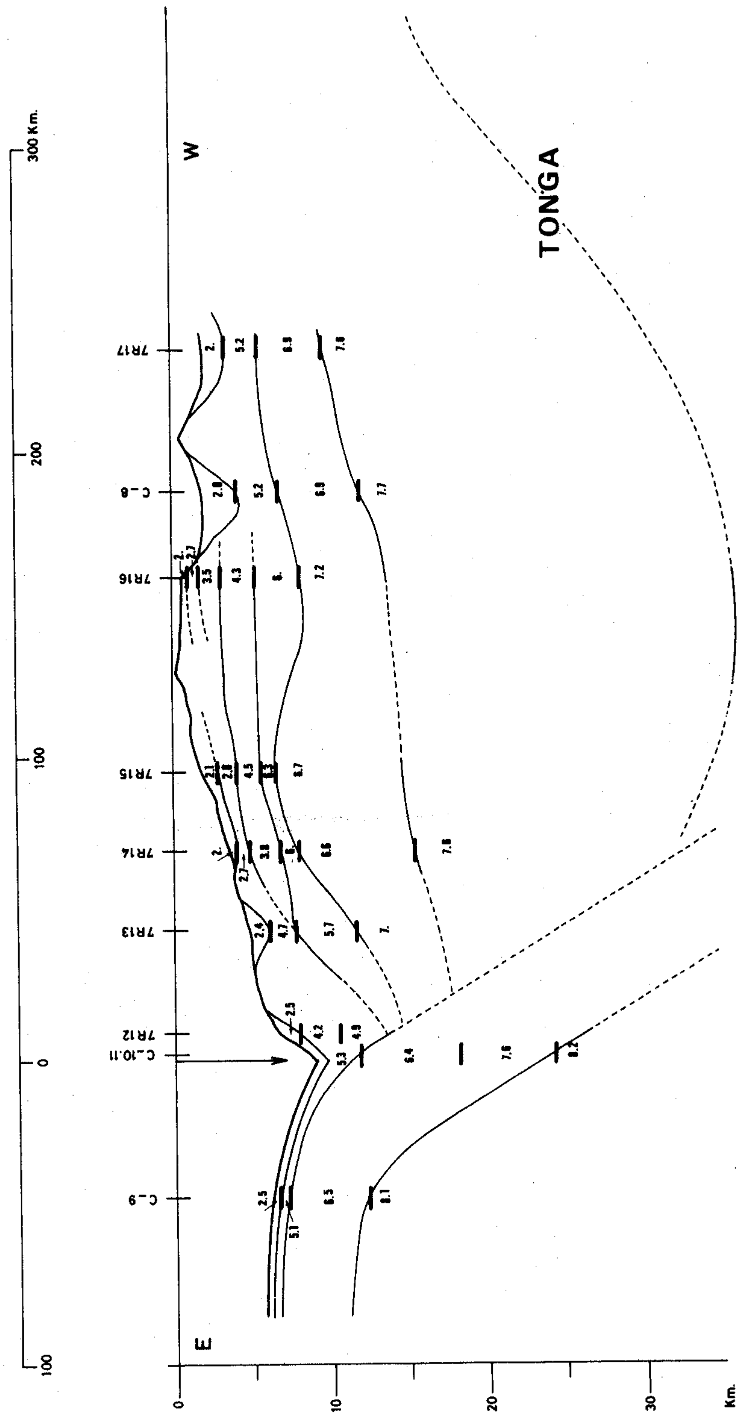

Figure 6. Structure section DC across the Tonga subduction zone.

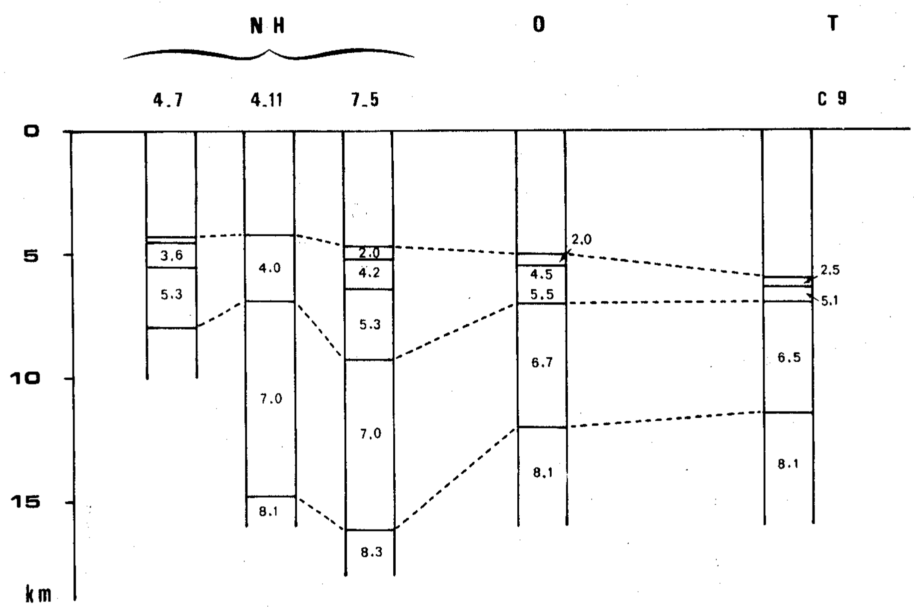

Figure 7. Structure sections of the crust of the dipping plate on the North Loyalty Basin (NH) and Pacific Basin (T) compared with standard oceanic crust (O) (from Ludwig *et al.* 1970).

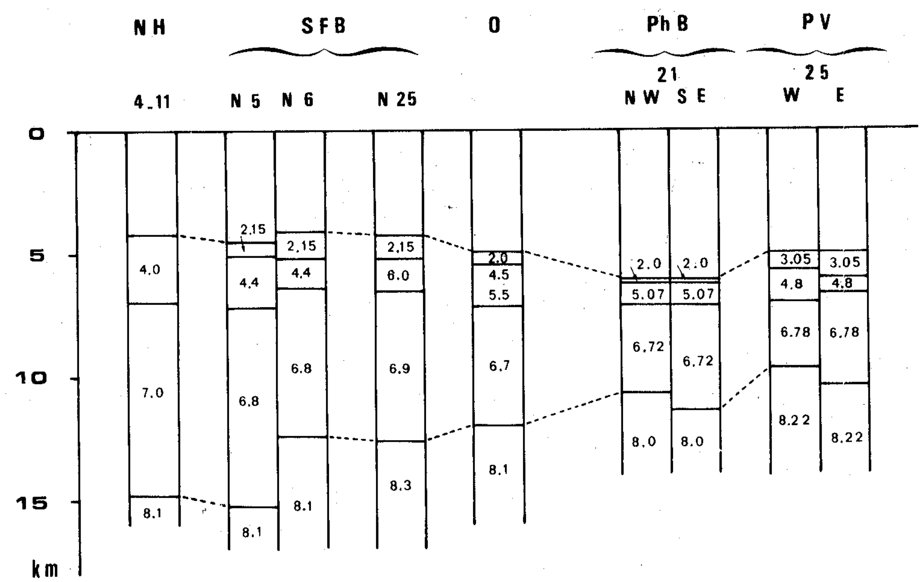

Figure 8. Structure sections of the crust of marginal basins compared with standard oceanic crust. NH, North Loyalty Basin on the dipping plate of the New Hebrides subduction zone; SFB, South Fiji Basin (Shor *et al.* 1971); O, standard oceanic crust (Ludwig *et al.* 1970); Ph B, Philippine Basin; PV, Parece Vela Basin (from Murauchi *et al.* 1968).

island arcs as well as under oceanic ridges is well known. Uyeda (1974) suggests that low velocity 'may be a characteristic of the mantle under island arc volcanic zones'. The 7.6 km/s layer under the Tonga arc could therefore be considered as uppermost mantle. The observed gravity anomaly should then be explained by a density variation with respect to the velocity-density curve of Ludwig *et al.* (1970). In this case, the crusts of the New Hebrides and Tonga arcs would be quite different: the crust of the older arc (Tonga) would be thinner (16 km) than that (26 km) of the younger one (New Hebrides), and the uppermost mantle velocity lower at Tonga (7.6 km/s) than in the New Hebrides (7.9 to 8 km/s).

Another problem is the structure of the crust in the troughs at the rear of the New Hebrides arc. These troughs were considered either as initial stages of marginal basins (where one could expect to find a much thinner crust than under the arc itself) or as extensional troughs (Dubois *et al.* 1975) associated in some way with intermediate and deep seismicity.

Seismic refraction gives no evidence of oceanic crust under the troughs (profile 4R 1, Fig. 5), and the structure of the upper layers is quite comparable to that on the arc. Furthermore, the observed gravity anomaly (J Y Collot, personal communication) leads to the supposition of a progressive joining between the arc (profile 4R 9) and the North Fiji Plateau (profile 7R 2). The possible existence of a 7.6 km/s layer above the Moho under the arc poses the problem of a joining, at the mantle level, with the North Fiji Plateau, where low-velocity uppermost mantle (7.6 km/s), characteristic of mid-oceanic ridges, can be observed.

CONCLUSION

The results obtained from seismic refraction on the New Hebrides and Tonga island arcs show that:

1. Although there are broad similarities in the two arcs, differences can be seen in the shallow structure. In the New Hebrides arc low-velocity layers are much thicker than in the Tonga arc, especially in the lower part of the inner slope of the trench. Further, differences in thickness can be correlated with differences in thickness of the transition and oceanic layers of the crust of the dipping plate.
2. Refraction leaves uncertainty as to the structure at depth, particularly as to the joining of the arc and the back-arc basins (North Fiji Plateau and the Lau Basin). The existence of 7.6–7.7 km/s velocity layers complicates the interpretation in classical terms of crust and mantle. One possible interpretation of the evolution of the crust under the island arcs could be a thinning-down in time.

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BASIN DEVELOPMENT IN THE SOLOMON ISLANDS AND THEIR PETROLEUM POTENTIAL

H R KATZ

ABSTRACT

Two major tectonic and sedimentary provinces are distinguished. In the oceanic-pelagic Malaita (Ontong Java) province an Early Cretaceous tholeiitic basalt basement is conformably overlain by 1500 m of a monotonous and continuous, pelagic carbonate mud sequence of Late Cretaceous to Pliocene age. In the main Solomons province, which is adjacent to the Malaita province along a tectonic boundary, the basalt basement is partly metamorphosed to greenschist and amphibolite facies and was intensely sheared, thrust and faulted in Eocene times. Various gabbro (including ultramafics) and diorite-tonalite intrusions occurred in the Late Cretaceous to Early Tertiary. This basement was deeply eroded in times up to Oligocene. Overlying sediments are unconformable and consist of Late Oligocene to Mid Miocene shallow-water carbonate bioherms (limestone coralgal reefs) and Miocene to Pliocene deep-water turbidites, sandstones and silt-mudstones mainly of volcanoclastic composition; these are over 4000 m thick in Guadalcanal. Mio-Pliocene andesitic volcanics are intercalated locally. The basin generally became shallower in the Pliocene, probably both through infilling and tectonic uplift. Major block faulting occurred in the Late Pliocene in the areas of present-day islands, resulting in widespread erosion and unconformable deposition of Pleistocene reef limestones, as well as thick debris fans. Extensional tectonics became more pronounced and resulted in regional collapse, i.e. taphrogenic breakdown of the Central Solomons Trough since the Pleistocene, accompanied by further strong uplift in the adjacent islands. Recent andesitic-dacitic volcanism is concentrated within and along the southern flank of the Central Solomons Trough.

Structurally this trough is subdivided into two segments arranged en echelon. Both trend WNW–ESE and are further affected by NE-trending cross faults. Overall length is about 500 km, and approximate width 30–70 km. Water depth to the generally very flat sea floor is between 800 m and 1800 m. Sediments underneath this trough are thickest in the west and east, and measure (from refraction velocities) some 2.5–4.5 km; seismic sections do not show basement in the deeper part of the basin, where good reflection events are obtained to 2.8 s. Apart from a shallow (probably Pleistocene) unconformity in some sections across the central part of the trough, little deformation affects the sediment pile except a sharp downwarp accompanied by faulting along the trough margins. However, some fold structures of probably Late Pliocene age, unconformably capped by younger sediments, are in evidence in some areas along its northern margin.

Regarding petroleum potential in the main Solomons province, porosity and permeability in the sediments generally is low but good reservoir properties may exist in Miocene reefal limestones and calcarenites, and probably also in some of the sandstones of Mio-Pliocene turbidites. The large amount of mudstones, and of clay matrix provided by the breakdown of plagioclase and pyroxene in the finer volcanoclastics would amply provide for good cap rocks. Both structural and stratigraphic traps may exist, particularly along some parts of the basin margins, although the very young faulting up to Pleistocene and later may be a deterrent. Little evidence exists for potential hydrocarbon source rocks and the degree of maturation, but favourable conditions may have existed for both, at least locally. Potential targets in general, however, would be in comparatively deep water except for some limited areas on or near the island of Guadalcanal.

INTRODUCTION

The Solomon Islands (for location see inset, Fig. 1) have by various authors been called an island arc, a double arc, an island arc with flipped subduction zones, a zone of convergent double subduction, a fractured arc, also a non-arc. Whatever they are, they certainly do not fit the classical model of island arcs, as has clearly been demonstrated by Coleman (1975, 1976).

The present paper does not deal with these problems, but gives a brief account of the geological history and stratigraphy so far as is relevant to the petroleum geologist. However, the formation of sedimentary basins and the structure in the region, as conceived in this study, is relevant to future discussions of the tectonic relationships in the Solomon Islands

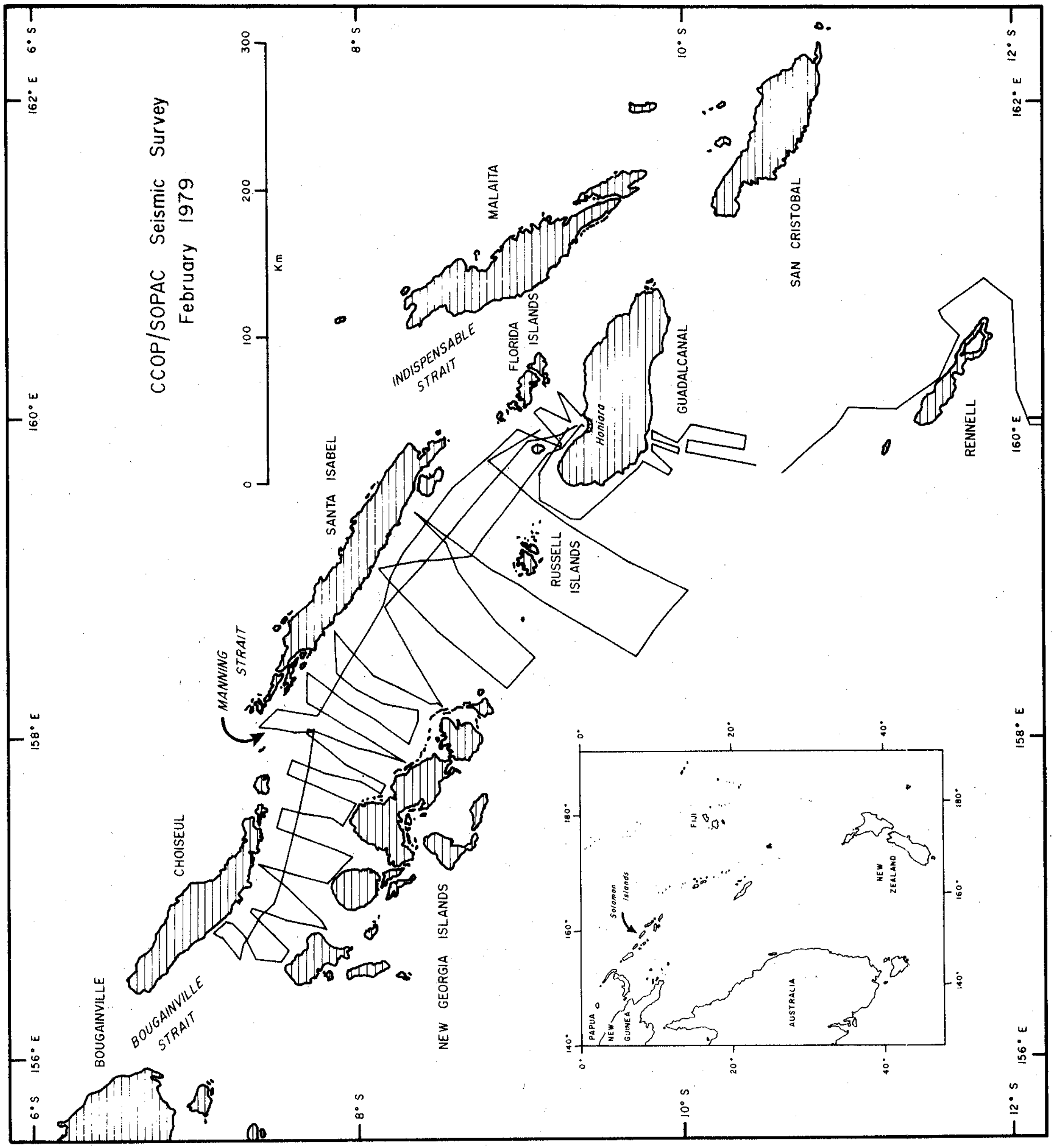


Figure 1. Location of Solomon Islands and track lines of CCOP/SOPAC Seismic Survey of February, 1979.

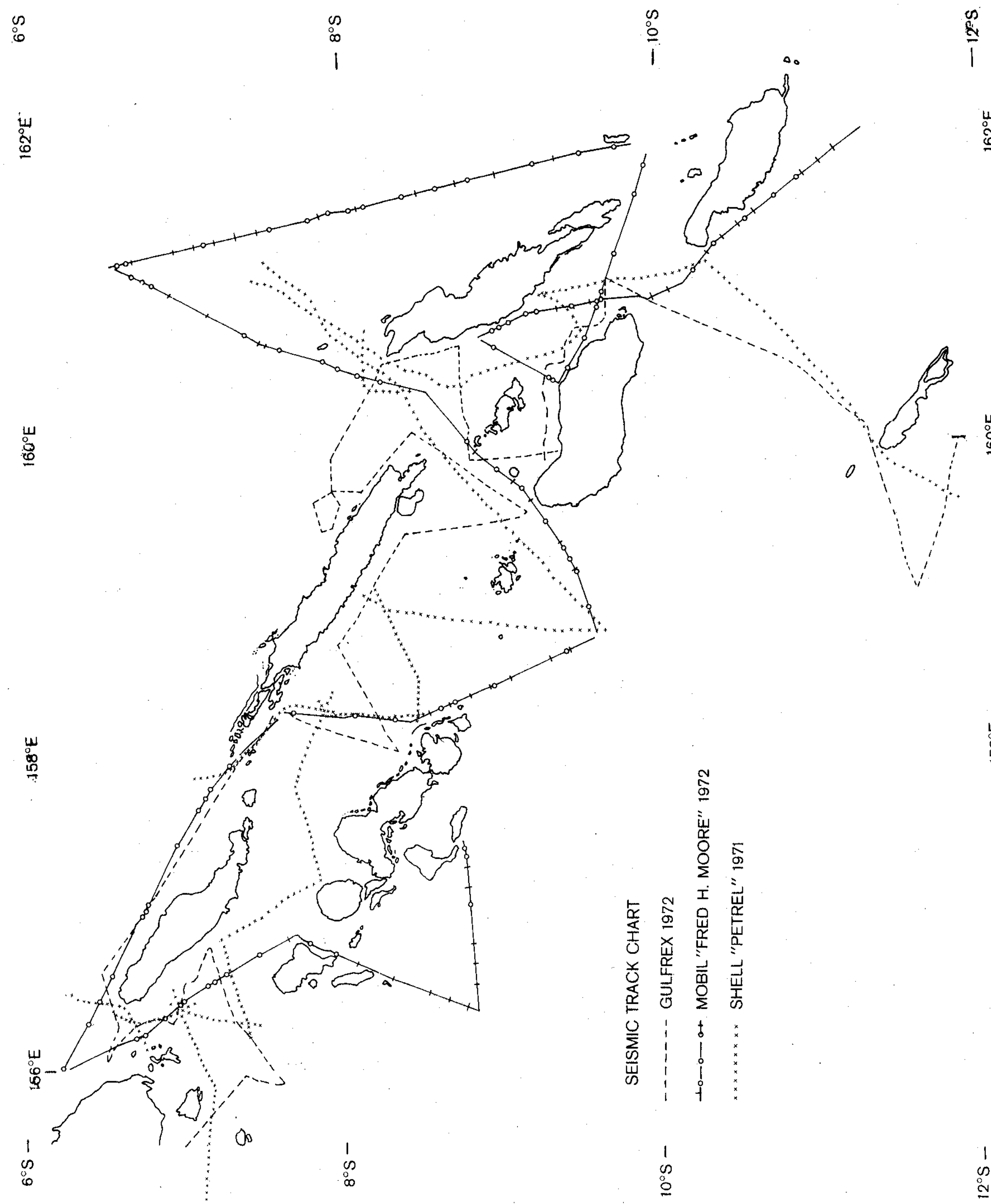


Figure 2. Seismic track lines of Shell, Mobil Oil and Gulf, 1971-72.

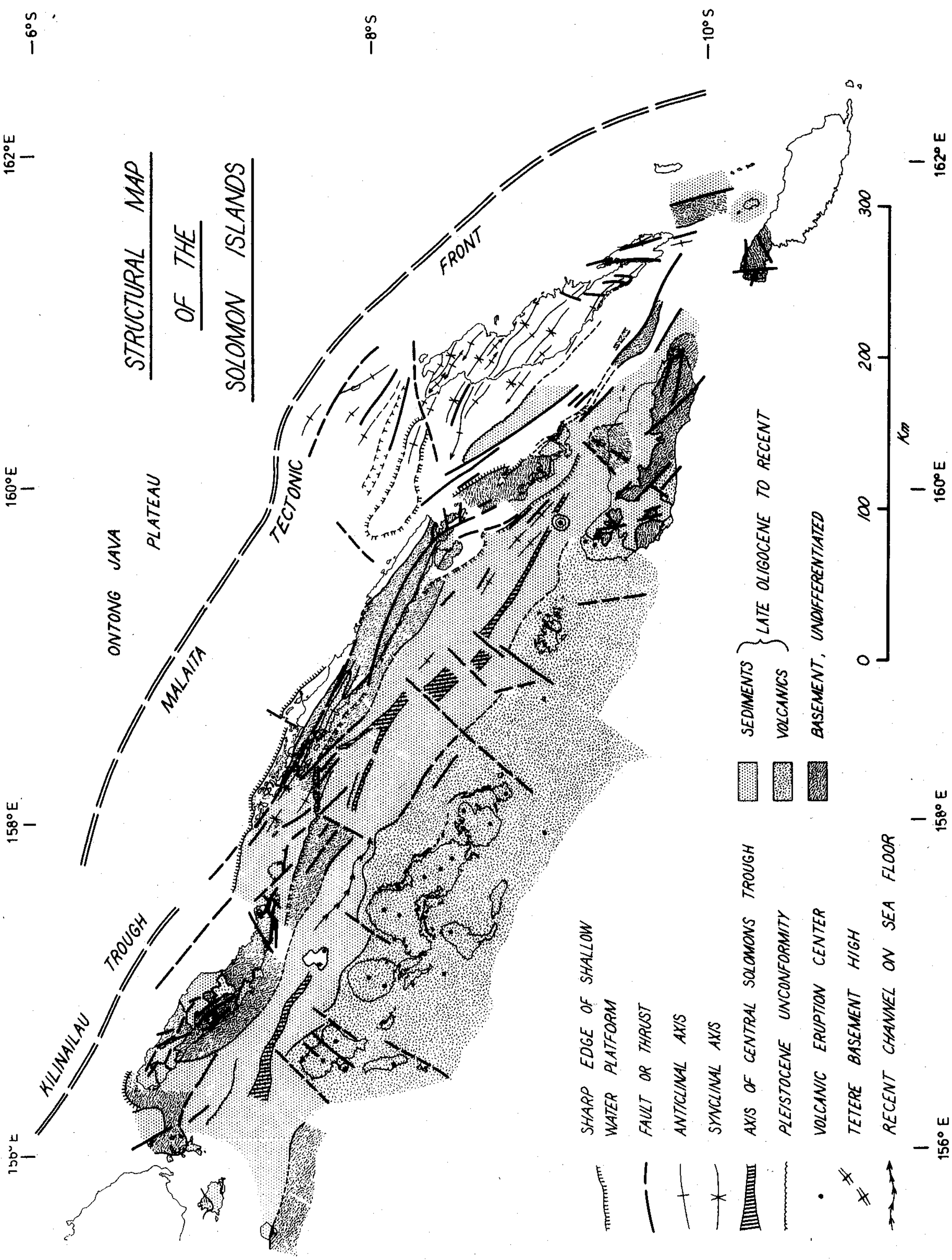


Figure 3. Structural map of the Solomon Islands.

, and to their position and evolution with respect to the Indian-Pacific plate boundary.

This study is to a large extent based on offshore seismic data from oil companies, held on Open File at the Geology Division of the Ministry of Natural Resources in Honiara. More precisely, the multi-channel records from Shell, Mobil and Gulf (Fig. 2) were used mainly, as well as some sections from the French Austradec survey and a very small programme in Manning Strait by Southern Pacific Petroleum Company. A single-channel survey by CCOP/SOPAC (Fig. 1) gave valuable additional information, which allowed correlation of structural features observed in the higher-quality oil-company lines, and provided for more detailed and continuous mapping coverage of the entire offshore area (Fig. 3).

The geological information was gathered — apart from a study of published literature — from extended discussions with members of the British Mapping Team (Institute of Geological Sciences) working at the time in the Solomon Islands, supplemented by personal observations in the field during excursions to the interior of Guadalcanal (Lungga River), to the Florida Islands, and to Malaita.

The study was undertaken while the author was employed as a UNDP Consultant in Petroleum Geology.

GEOLOGICAL SETTING

For the purpose of this study the region of the Solomon Islands is divided into two main provinces (Fig. 4). To the northeast lies the oceanic-pelagic Malaita (Ontong Java) province, which corresponds to the Pacific Province of previous authors (e.g. Hackman 1973); their Central and Volcanic Provinces are here grouped informally as the 'main Solomons province'.

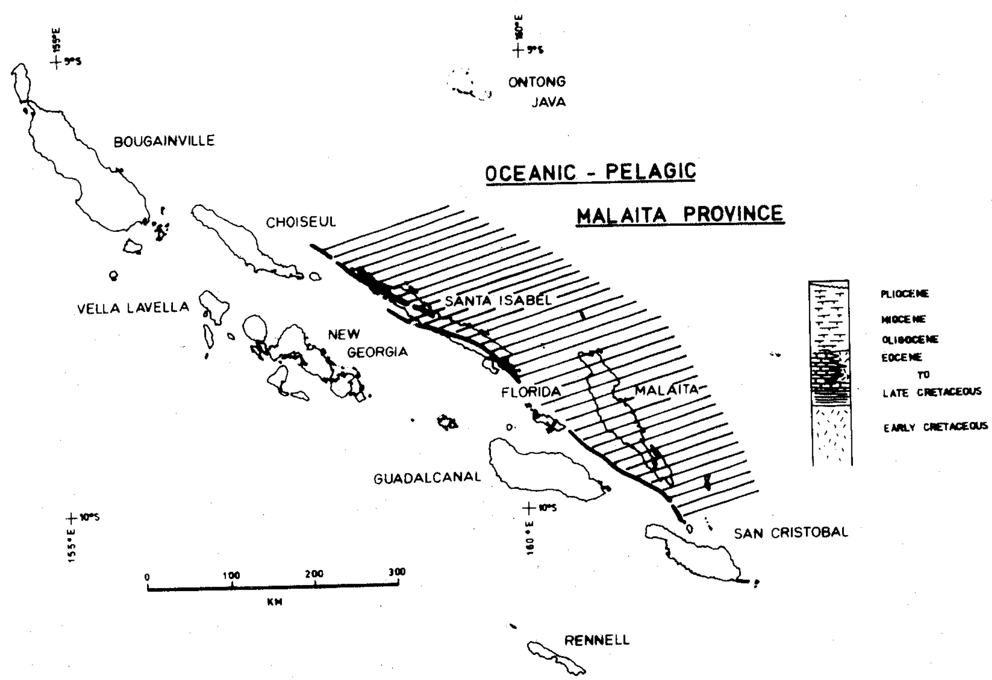


Figure 4. The two main provinces in the Solomon Islands region: the Malaita (Ontong Java) province in the northeast, and the main Solomons province covering all the rest of the region to the west and south. Composite stratigraphic section of the Malaita province, above Early Cretaceous basement basalt, measures about 1500 m.

The boundary between the two provinces appears to be tectonically controlled. It is marked by an extensive zone of shearing, thrusting and faulting all the way from Manning Strait northwest of Santa Isabel to the area between eastern Guadalcanal, southern Malaita and San Cristobal (Figs. 3 and 4). Locally, ultramafics (mainly serpentinized harzburgites) have been pushed up along this major tectonic zone.

MALAITA (ONTONG JAVA) PROVINCE

An oceanic, tholeiitic basalt basement of Early Cretaceous age (Hughes and Turner 1977) is overlain conformably and without intervening erosion by a rather monotonous, pelagic sediment cover ranging from Late Cretaceous to Pliocene (van Deventer and Postuma 1973). The sediments form a continuous, mainly carbonate mud sequence including siliceous marl with chert, white aphanitic limestone with brown chert nodules, nannofossil chalk and blue-gray and brown calcisiltites. A closely similar sequence was found in DSDP drillholes on the Ontong Java Plateau (Andrews and Packham *et al.* 1975). The total thickness across both areas ranges from 1300 m to 1750 m.

Rarity of sedimentary structures in the earlier part of the sequence indicates a quiet environment of deposition, but extrusion of alkalic basalt and the formation of peperites also occurred during this period, i.e. until the Eocene (Younger Basalts in Malaita, Sigana Volcanics of Santa Isabel). An influx of volcanogenic silt-sand detritus begins to appear in Oligocene time on Santa Isabel, and later on Malaita also (Coleman *et al.* 1978); in the younger part of the sequence, distal turbidites and slump deposits become an important feature in the otherwise predominantly pelagic limestone environment (Hughes and Turner 1977). Bioturbation is very common, indicating well oxygenated conditions; the seas are rapidly shallowing during the Late Miocene and Pliocene.

In the Late Pliocene compressional tectonics created an impressive series of fault anticlines, cascading down to the northeast, all along from the northwest of Santa Isabel to east of southern Malaita and Ulawa. Their tectonic front lies along a line of an echelon troughs which mark the boundary of the fold belt against the undeformed, gently down-bending Ontong Java Plateau to the north (Kroenke 1972).

MAIN SOLOMONS PROVINCE

Geology exposed on the islands

The older history in this area is somewhat obscured. The basement of oceanic, submarine basalt flows (locally with pelagic limestone lenses intercalated) passes at depth into dynamo-metamorphically altered metabasalts of greenschist to amphibolite facies (Voza basalts and Choiseul schist in Choiseul, Mbirao volcanics and metabasalts in Guadalcanal), which exhibit intense shearing resulting in three different sets of schistosity in Guadalcanal (Hackman, in press), and are affected by thrusting and faulting followed by deep erosion; in many places schists are now exposed in juxtaposition with unmetamorphosed basalt, in fault or thrust contact. The climax of tectonism (and metamorphism) seems to have occurred in the Eocene to Early Oligocene (one K-Ar age of Choiseul schist is 44 ± 18 m.y.; Arthurs 1978). Locally there are intrusions of gabbro and diorite to tonalite of Late Cretaceous and possibly Early Tertiary age; ultramafics in Santa Isabel and other areas may be Cretaceous too (Stanton and Ramsay 1975).

Overlying sediments, of Late Oligocene to Mio-Pliocene age, are unconformable and to a large extent derived from erosion of the older basement. Local depressions are filled with thick, monomict orthobreccias composed of large, angular basement clasts enclosed within a matrix of basaltic grit (Koloe Breccia in Choiseul, 250 m thick — Hughes 1979a). In shallow areas around still emergent islands of basement volcanics, metabasalts and gabbro and diorite intrusions, extensive marine-carbonate platforms were formed at the base of the sedimentary sequence; these include coralgal reef limestones and reef slope deposits.

The Mount Vuasa Limestone in *Choiseul* (Arthurs 1978), containing *Lepidocyclina* *martini* and lenticular bands of basaltic conglomerate up to 5 m thick (Strange and Danitofea 1979), represents a Lower to Mid Miocene fringing reef established on the flanks of an emergent ridge of basement schist and basalt; it is about 50 m thick. Elsewhere in Choiseul the irregular, eroded surface of basement basalt and schist is overlain by deep-marine non-calcareous grit, sand and shale, often as turbidites, of the Mole formation; this is generally of Miocene age, but on the central north coast of Choiseul, where *L. (Eulepidina) ephippioides* Jones and Chapman was found (Hughes 1979b), deposition of Mole formation sediments may have begun in Late Oligocene. The thickness in that area, where an unknown upper portion, however, is missing because of erosion, measures 2500 m. In northwest Choiseul, deposition of the Mole formation did not begin till Mid to Late Miocene, and the total thickness is only a few hundred metres (Strange 1979a). In the Shortland Islands further west and southwest, the oldest sediments above basalt basement are as young as Late Miocene to Pliocene (Turner 1978). Transgression thus appears to have advanced in a generally western direction in this northernmost part of the Solomon Islands.

In northwest Choiseul the Mole formation passes upwards into the calcarenites and marls of the Pemba formation, 400 m thick and of Early to Late Pliocene age — indicating an open marine carbonate shelf environment, thus a shallowing of the sea but also slower deposition. The original total thickness of the Pemba formation, however, is not known because of uplift, tilting and erosion after the Pliocene.

Both Mole and Pemba formations are absent from southeast Choiseul, where the Early Pliocene Vaghena formation (Hughes 1979c) probably rests directly on basement, as is suggested also from offshore seismic information. The Vaghena formation (from Vaghena Island east of Choiseul) consists of sand-siltstones and various calcareous units, its benthonic microfauna suggesting deposition on a shelf.

Block-faulting and uplift, and regional tilting to the northwest, west and southwest occurred in Choiseul by the end of Pliocene time. The thick Pleistocene Nukiki reef limestones were deposited unconformably across Mole and Pemba formations, as well as over basalt and schist basement. Uplift continued during its deposition as shown by the reef growth in successive concentric ring patterns around the emerging island of Mt Talaevondo, southeast of Choiseul Bay (Strange 1979a). Further southeast along the central south coast of Choiseul, Nukiki reef limestone in a large block 150 m thick is uplifted to 800 m only 7 km inland from the coast, but dips 5–15° to the southwest and to below sea level just off the present coast (Strange 1979b). Also, Recent coral reefs are uplifted and tilted seawards, passing from dead into living reefs. The extensive drowned coastline in southeastern Choiseul, downwarped to the south, is indeed a most impressive feature particularly as viewed from the air.

In *Guadalcanal* a very similar history is recognized (Hackman, in press). At the base of the sedimentary sequence — very widespread though clearly restricted to the basin margins and around shallow or emergent basements highs of schist and diorite intrusions — are the Miocene Mbetilonga and equivalent limestones forming bioherms 200–400 m thick; they are overlain by the Tina calcarenites (100–300 m thick), the Lake Lee calcarenites and related deposits. The somewhat younger Valasi limestones in eastern Guadalcanal consist of 500 m of predominantly off-white, massive coralgal reef limestones directly overlying basement.

In the deeper parts of the basin in western Guadalcanal, the Lungga Beds — greywacke-type sandstones and silt-mudstones — are lateral correlatives of the biogenic limestones; a reducing environment is indicated in their basal member, the Kavo greywackes (Wyn Hughes, pers. comm.). This may be at least partly due to the extensive volcanic activity manifested by the Suta volcanics, which consist of thick basalts and basaltic andesites, both flows and pyroclastics, which interfinger with the Kavo greywackes.

Similarly along the northern slope of the uprising ridge of Guadalcanal, deeper marine clastic turbiditic sequences may have formed (now buried) perhaps along with deep marine carbonates. Certainly in the Late Miocene to

Pliocene, accelerated subsidence of the basin to the north took place, accompanied by increased faulting, upheaval and erosion in adjacent land areas. Such increased tectonic instability with rapid uplift and subsidence made reef building virtually impossible. At the slope edge large chunks of older reef limestones were broken off and fell into the deep (van Deventer 1971), where huge masses of erosion products, possibly including contemporaneous volcanic material, were deposited. The 4000 m thick Mbokokimbo formation with its various, laterally and vertically intertonguing members (Hackman, in press) is to a considerable extent characterized by bathyal turbiditic and locally chaotic slump sedimentation. Muddy silt-sandstones, partly calcareous and mainly of volcanic composition, are the principal lithologies.

Upwards coarsening, partly conglomerate debris fans of great thickness continued well into the Plio-Pleistocene. The pronounced uplift of the hinterland caused the basin to fill in more and more rapidly; in southern Guadalcanal Pliocene lignite beds are found in the upper part of the Lungga Beds. But tectonic uplift now began speading further north, too. While locally, river and related submarine canyon systems still provided for the deposition of deep-marine (bathyal) fluxo-turbidites (van Deventer 1971) in the Late Pliocene and Pleistocene, the thick conglomerates of the Pliocene Vatumbulu beds in northeastern Guadalcanal are shallow marine and include lenses of coralgal reef limestones. More recent uplift is demonstrated by the spectacular series of terraces up to 200 m above sea level, immediately behind Honiara, which consist of raised Pleistocene coral reefs and reef slope deposits. According to Hackman (1973), Pleistocene surfaces of marine erosion further inland on Guadalcanal are found up to 800 m a.s.l.

The island of *Santa Isabel* has unfortunately not been covered yet by modern mapping, although it is one of the most critical areas geologically, particularly for an understanding of the relationship between the Malaita and main Solomons provinces. From data presented mainly by Stanton (1961), Stanton and Ramsay (1975) and Coleman *et al.* (1978), it appears that basement consists of an ophiolite complex, including ultramafics and layered microgabbros, which in places attain amphibolite-grade metamorphism. This basement probably is of Cretaceous age and is unconformably overlain by a 3 km thick pile of basaltic pillow basalts (Sigana Volcanics), which at least in their upper part are Paleocene and Eocene and may compare with the Younger Basalts of Malaita (Hughes and Turner 1977). Closely associated sediments, partly intercalated with pillow lavas, are hard, cream-coloured, carbonate pelagites often with pinkish chert bands and nodules. Tuffaceous sandstone units, breccias and marls occur also, the last-named of consistent cream to pale buff colour, rather chalky in texture and grain size, and containing pelagic foraminifera and radiolaria (Stanton 1961). Deep-sea pelagite limestones on the northeast coast of Santa Isabel occur from Late Paleocene through to Oligocene, but locally also younger ones are found. In general, however, Mio-Pliocene deposits are rather more shallow-water type volcanic wackes and sandstones carrying larger foraminifera and coralline algae (Coleman *et al.* 1978).

Stanton (1961) has drawn attention to the geographical distribution of rock types on Santa Isabel. Pelagite limestone (mostly massive but sometimes also bedded, sparsely fossiliferous (radiolaria), silicified and often recrystallized, and generally close to Sigana Volcanics) is a nearly ubiquitous, main constituent of sediments all along the north coast. In Manning Strait massive limestone hills extend continuously along the north side of Dart Sound: on the south side of the Sound no limestone is found, but only interbedded clastics (silt-mudstone, sandstone, greywacke, tuffs and occasional conglomerate). There is also a pronounced structural distinction, the limestones to the north (and in most places along the north coast of Santa Isabel) being strongly folded and contorted, with dips mainly steep to perpendicular, whereas the rocks to the south, from Thousand Ships Bay (San Jorge Island) to Dart Sound, are only very gently folded with dips generally of 5–15°. The Barora syncline immediately south of Dart Sound has dips of < 40° only. The postulated

Kia Thrust in Dart Sound would mark the tectonic boundary between the two facies as well as between the two structural provinces. It is here suggested that the Kia Thrust is of similar significance to the Korigole Thrust further to the southeast in Santa Isabel, of which it may be an en echelon displacement, i.e. continuation.

From all the available evidence, it seems probable that Santa Isabel combines elements of both the Malaita and main Solomons provinces, possibly in a horizontal as well as vertical sense. If so, future mapping will be important not only to give detailed stratigraphic and structural information, but possibly to clarify the original relationship between the two provinces and the process of the tectonic amalgamation.

In this context it should be noted also that Stanton's (1961) informal grouping of all sediments in his Tanakau Group lacks any stratigraphic and facies connotation, but was meant only to outline the various lithologies included in the Tertiary sediment cover found above basement. In particular, he has never established a sequential stratigraphic order, as was erroneously implied by Landmesser (1977) who gave a quite misleading picture in his seismic correlation and interpretation.

Geology of the Central Solomons Trough

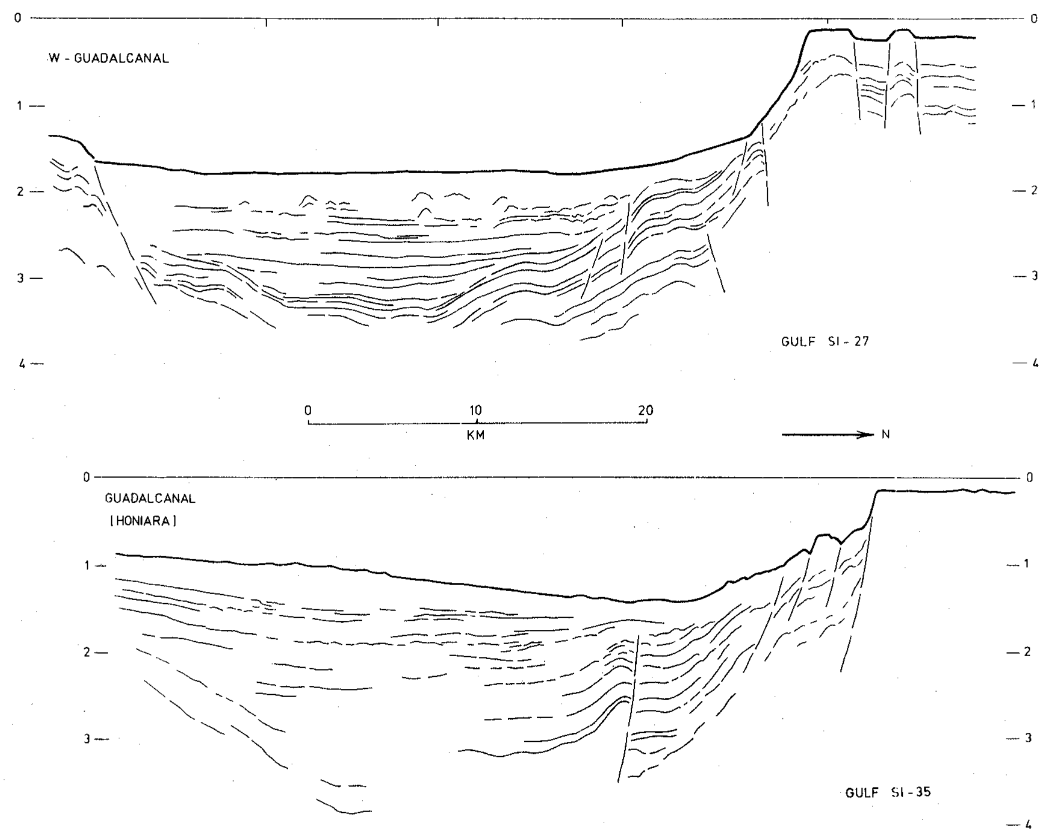
Seismic investigation in the New Georgia Sound has shown a continuous, deep sedimentary basin between Choiseul – Santa Isabel – Florida on the one side and the Guadalcanal – New Georgia trend on the other (the 'Solomon Basin' of de Brion *et al.* 1977). There is little doubt that the sediments in this basin correlate in principle with those found on the islands of Choiseul and Guadalcanal, of Late Oligocene to Plio-Pleistocene age. This implies that the idea of Coleman (1975) and others, of a Neogene displacement of an originally single chain resulting in the present double chain character of the Solomon Islands, is untenable.

However, the basin does show some internal structural complexities, perhaps fragmentation. While its overall length is 500 km and approximate width 30–70 km, it is markedly narrow and shallow between Kolombangara (157°E, 8°S) and Manning Strait (Figs. 1 and 3). There is a strong suggestion of a broad fault zone in Manning Strait having the effect of an en echelon arrangement of two partly separate basins. A younger fragmentation along a set of northeast-trending faults is of minor implication yet certainly quite real, although the particular location of individual faults is inferred mainly from circumstantial evidence. Their direction closely follows the line-up of young volcanic centres and fault lines observed on Vella Lavella, including the geothermal graben of Paraso Bay on the same island.

Present water depth of the trough is 700–800 m in the shallowest part north of Kolombangara, deepening to 1500 m in the northwest (south of Bougainville Strait), and to 1800 m in the southeast (northwest of Guadalcanal). This conforms with an increase in sediment thickness in the same directions. The lesser thickness in the central part, however, may be apparent only and due to a masking effect of a more coherent, volcanoclastic overburden derived from the young volcanoes in the south.

From refraction measurements the sediments underneath this trough are some 2.5–4.5 km thick. Good reflection events are obtained to 2.8 seconds two-way travel time; basement, however, is never seen in the central part of the basin. There is mostly very little deformation of the sediment pile, except along the basin margins, particularly in the north, where a sharp upturn of sediments occurs, generally accompanied by faults (Fig. 5a/b); locally, anticlinal fold structures have also been formed in association with upfaulting along the northern margin, e.g. north and northwest of western Guadalcanal (Fig. 6a/b). However, although such fold structures are unconformably capped by younger rocks and therefore manifest themselves as probably of Late Pliocene age, most normal faulting towards the uplifted basin margin (which also must have begun in the Late Pliocene) has obviously continued until Recent time as evidenced not only by the broken sea-bottom relief but also by the geologic history as seen on land. The total amount of vertical displacement since the Pliocene is undoubtedly considerable, probably amounting to several kilometres in most places. Within the basin an unconformable relationship of younger, probably Pleistocene

Figure 5. Sections across Central Solomons Trough north of western Guadalcanal. (a) upper, (b) lower; for location see Fig. 2.



to Recent sediments is clearly seen in many sections across the trough (Fig. 3 and profiles Figs. 5a, 6a); their thickness distribution indicates very recent, differential subsidence.

On or above this unconformity, some seismic sections show features interpreted as buried reefs. Between Vella Lavella and western Choiseul (Fig. 7), such 'reef structures' are developed very extensively across the central part of the basin, where water depth is about 1300 m. They are 0.2–0.4 s thick, with a karst-like surface relief of 0.1–0.2 s, and are buried below 0.2–0.4 s of well layered sediments. Assuming average velocities for limestones and unconsolidated silt-mudstones above, the limestones would be about 400–500 m thick, and their overburden 200–250 m. Considering that the erosional remnants of Pleistocene Nukiki reefs in Choiseul immediately north of here still cover hundreds of square kilometres and are some 150 m thick, it is suggested that the buried reefs in the basin are lateral equivalents downfaulted since the Pleistocene. Taken at the base of the reef formation, subsidence over the last 1 million years, approximately, would therefore have been 2000 m, or about 1600 m since the top of the reef formation was formed. Both the thickness of limestone plus overburden, and the amount of subsidence, could have been achieved within the Pleistocene to Recent period (subsidence of about 2000B; see Katz 1979, for an analysis of comparable movements elsewhere). Indeed, the tilt established by vertical uplift in Choiseul as measured between 800 m high Nukiki limestone 7 km inland and its zero position at the present coast, is virtually the same as an average tilt from zero to 2000 m below sea level, measured between the coast and roughly the location of the reef occurrence on the seismic section about 20 km seaward; instead of an even tilt, however, vertical displacement of the same amount is established by uneven warping and step-wise up- and downfaulting.

Pronounced subsidence in more recent times is also indicated by extensive drowned reef occurrences about 1 s below sea level — with no overburden whatever — on seismic sections between Mono and Alu islands (Shortland Islands southwest of Bougainville Strait, Fig. 1).

Volcanism

The bulk of sediments in the main Solomons province are of volcanic composition and generally classed as volcanoclastics. Without further discussion or evidence, this is often taken as proof for widespread contemporaneous volcanic activity, on the basis of which the proto-Solomons have (like other similar regions) been interpreted as a volcanic island arc.

From the previous discussions of the geology exposed on the islands, however, it is clear that locally very deep and widespread erosion of volcanic basement rocks has occurred; this erosion must have supplied repeatedly very large volumes of detritus into the basin. On the strength of this, it is clear that volcanoclastic sediments *per se* do not necessarily indicate contemporaneous volcanism. It is of interest, therefore, to investigate and discuss the extent of direct and unequivocal evidence for volcanic activity.

At the very beginning of the basin formation is the thick sequence of basalt flows, basaltic andesites and associated pyroclastics of the Suta Volcanics, which cover about 150 km^2 of

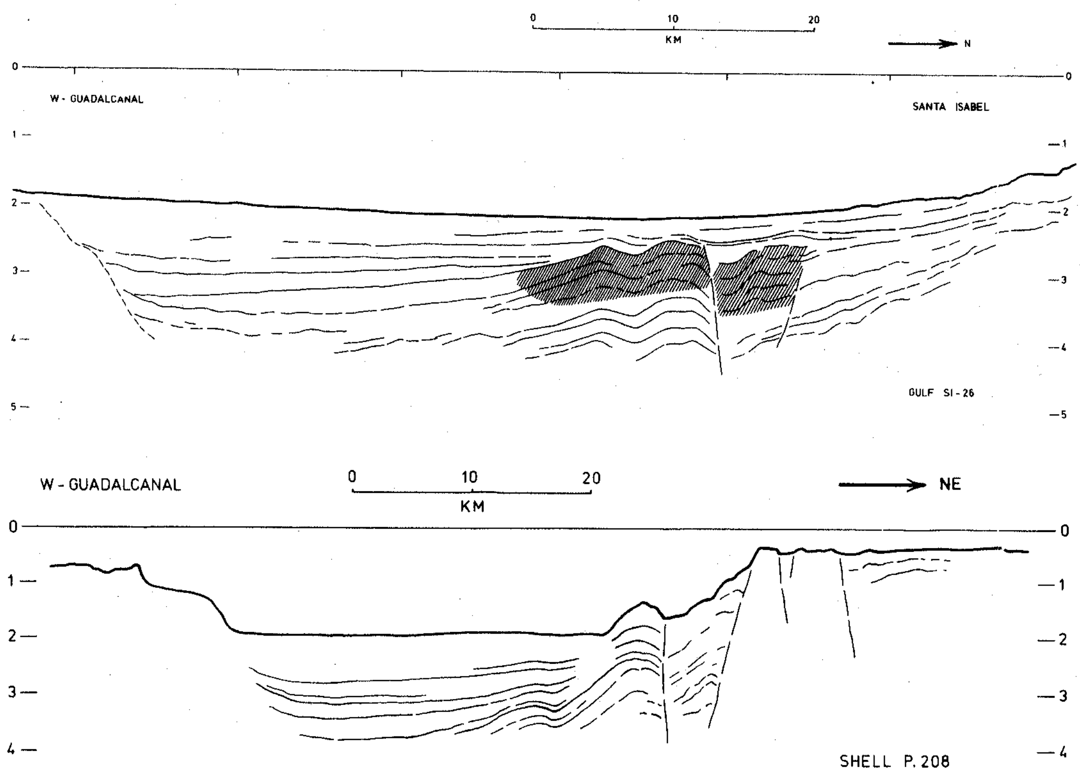


Figure 6. Fold structures in the Central Solomons Trough, northwest of Guadalcanal. (a) upper, (b) lower; for location see Fig. 2.

south-central Guadalcanal (Hackman in press). From benthonic forams in subsidiary limestone intercalations, the age is Early Miocene. Basaltic andesite flows also interfinger with Lower Miocene Kavo greywackes of the Lungga Beds, while to the north intercalated calcarenites indicate a transition to the Lower Miocene Mbetilonga limestones. However, considering the 2500 m thick pile of volcanics (lava flows and intercalated pyroclastics) below these sediments, it seems possible that the Suta Volcanics reach down into the Oligocene.

Compositionally they include a great variety of volcanic rock types, which are characterized as island arc tholeiites (Hackman 1973 and in press).

In central Choiseul the Maetambe Volcanics are andesitic tuffs forming discrete horizons up to 50 m thick, intercalated in the lower part of the Mole formation. In the central northeast coast they are Early to Mid Miocene (Hughes 1979b), but further to the west such tuffs reach up into the Early to Middle Pliocene, while hot springs associated with Maetambe Volcanics suggest that volcanic activity ceased relatively recently (Strange 1979b). The Kömboro volcano in easternmost Choiseul, surrounded by the Early Pliocene Vaghena formation, has supplied ash fall crystal tuffs (mainly hornblende and plagioclase crystals) into the Vaghena formation (Hughes 1979c).

Detrital sandstone-tuffs in the younger sedimentary formations of Santa Isabel — but of undefined age — are not derived from underlying basement according to Stanton (1961), but are of subaerial origin and distributed probably from volcanoes to the west or southwest (Choiseul or New Georgia). Their principal constituent is fresh and zoned plagioclase, but they also contain rock fragments of hypersthene-andesite. These tuffs contain isolated vegetal remains, carbonaceous lenses up to an inch long and occasionally

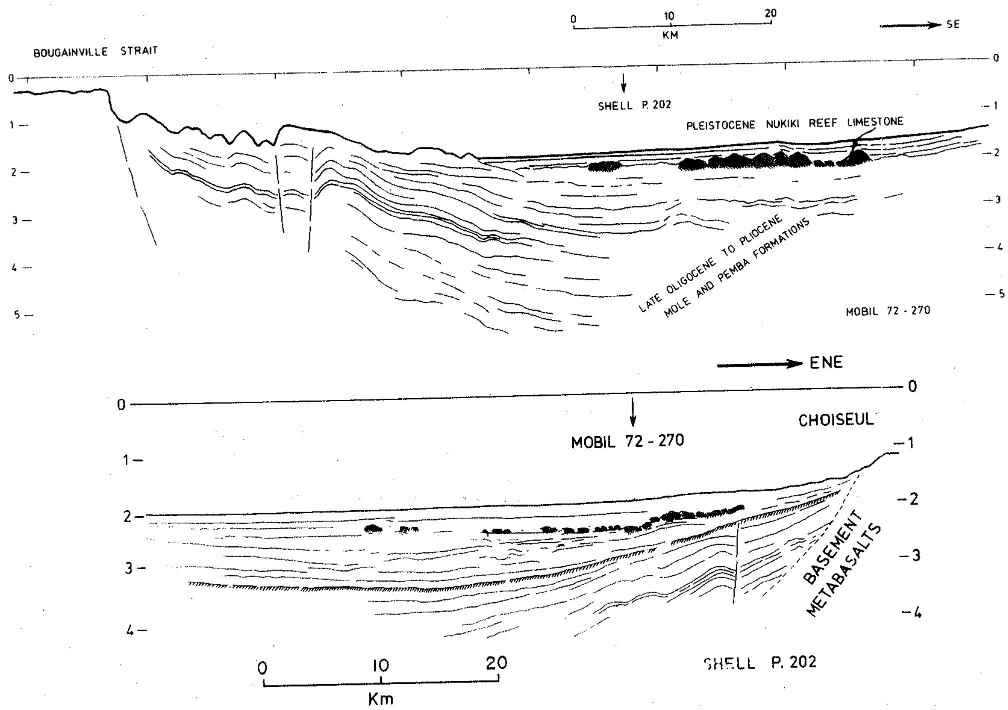


Figure 7. Buried reefs of inferred Pleistocene age in northwestern part of Central Solomons Trough, between the islands of Vella Lavella (northwesternmost of New Georgia Islands) and Choiseul. For location see Fig. 2.

pieces of poorly carbonized wood up to a foot long — apart from remains of foraminifera, radiolaria, bryozoa and corals.

On Fauro Island near the southeastern tip of Bougainville, the shallow-marine, 1400 m thick Koria Sandstone consists predominantly of andesitic-dacitic tuffaceous material, partly also primary tuffs and is disconformably overlain by 650 m of Togha pyroclastics, mainly agglomerates and tuffs and some intrusive andesites and basalts (Turner 1978). While their age is unknown, these volcanic formations most likely are young and probably belong to the Pliocene to Recent andesitic-dacitic volcanic domain of Bougainville Island. Oema Island immediately to the north of Fauro consists of andesite of probably Quaternary age.

In the New Georgia Islands, which are the main centre of young volcanism in the Solomons, activity must have commenced in the Pliocene if not earlier, as is demonstrated by overlap of Late Pliocene calcarenites on older volcanics in Vella Lavella (Wyn Hughes, pers. comm.). Volcanic activity in the New Georgia Islands continued through Pleistocene to Recent times. Of Late Pliocene to Pleistocene age, and probably ranging even younger, are the andesitic-dacitic Gallego Volcanics of northwestern Guadalcanal. Related to these extrusives is the shallow-intrusive Plio-Pleistocene Koloula granodiorite in southern Guadalcanal (Chivas 1975). Within the Central Solomons Trough are the volcano of Savo Island and another volcanic centre discovered on seismic evidence and closely outlined on detailed bathymetry, which to the north of Kolombangara Island (see map Fig. 3) intersects the youngest sediments on the sea bottom and sharply protrudes about 100–200 m above its flat surroundings. Both these volcanic occurrences apparently are of Recent age, and are lined up along the very axis of the Central Solomons Trough.

In summary, unquestionably, volcanic activity occurred throughout the history of the main Solomons sedimentary basin. In the Oligocene-Early Miocene, island-arc tholeiites were extruded in Guadalcanal, but at the same time or soon thereafter a calcalkaline trend, producing hornblende andesites and dacites was established in Choiseul where it continued until at least the Pliocene. Calcalkaline, Pliocene to Quaternary volcanism is manifest in the Shortland Islands (and Bougainville), the New Georgia Islands and Guadalcanal. It is interesting to note that, apart from the Bougainville-Shortland islands, younger volcanism has come to be concentrated entirely to the south of, and within, the Central Solomons Trough. Volcanism was of limited extent, however, through most of Oligo-Miocene and Pliocene time, and rather isolated and randomly distributed. Thus it can have supplied only a small part of the sediments in the basin; there appears to be no question that the bulk of the so-called volcanoclastics is not derived from contemporaneous volcanism, but from the erosion of an older, volcanic basement. To what extent, and in which geographic location, a volcanic island arc may thus be postulated remains an open question beyond the scope of this paper.

Summary of basin evolution in the main Solomons province (Fig. 8)

After intense tectonism in Late Cretaceous to Early Tertiary time, resulting in shearing, thrusting and metamorphism of ocean floor basic extrusives and oceanic sediments, and intrusion of gabbro and diorite, uplift, probably along regional fault lines, occurred during the Oligocene, accompanied by volcanism. Together with deep erosion of these land areas, marine transgression commenced in the Late Oligocene to Miocene. Shallow seas near and around still emergent basement highs gave rise to the development of reef limestones and open marine carbonate shelf deposits, while in deeper marine areas clastic turbiditic sequences were deposited.

Rapid deepening of the basin occurred in the Late Miocene-Pliocene, accompanied by fault-controlled upheaval and erosion in adjacent areas, and sporadic volcanism. Bathyal turbidites and locally chaotic slump deposits are characteristic. The intensive erosion in uprising land areas supplied large volumes of sediment into the basin; deposition was rapid and sediments many thousands of metres thick were deposited.

While in the central part of the basin deep-marine conditions continued throughout the Pliocene, marginal areas experienced a pronounced shallowing as a result of both infilling and tectonic uplift. In the areas of present-day islands such uplift was accentuated in the Late Pliocene, with block-faulting and tilting of blocks accompanied by strong erosion. Huge, coarse marine debris fans developed in the Pleistocene to the north of Guadalcanal, while extensive, unconformable reef limestones formed from Choiseul south into the now shallow basin.

The prevailing extensional tectonics that were dominant in the region from the very beginning of basin formation gradually intensified to a first climax in the Late Pliocene and culminated in the Pleistocene to Recent period of major diastrophism. Regional taphrogenic breakdown resulted in the formation of the Central Solomons Trough, 500 km long and downthrown about 2000 m perhaps within the last 1 million years. This was accompanied by further strong uplift in the adjacent islands.

Volcanism — tholeiitic in the early stages of basin formation but mainly calcalkaline thereafter — appears to have been of limited importance only, but occurred sporadically through most of Miocene and Pliocene time in one place or another. In the course of recent tectonic climax, volcanism has become more intensive and continuous, but restricted to the area immediately south of, and within, the Central Solomons Trough.

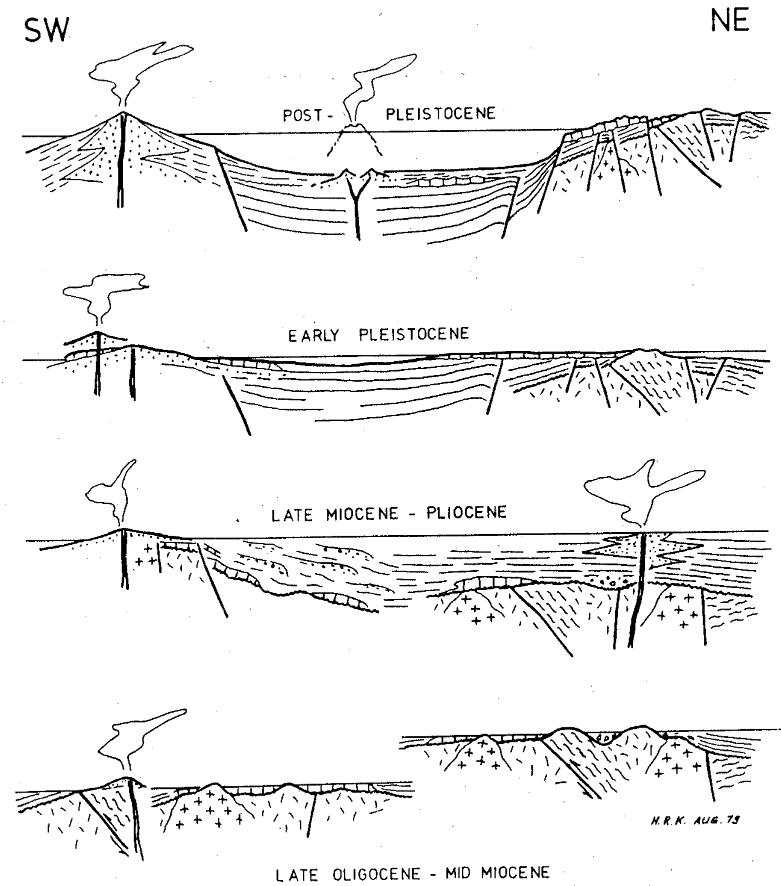


Figure 8. Diagrammatic sections across main Solomons province, showing evolution of sedimentary basin and Central Solomons Trough. Not to scale.

PETROLEUM POTENTIAL

Little if any potential seems to exist in the Malaita province. Although no source rock analyses have been carried out, the sediments of mainly pelagic character have unquestionably very low organic contents. Slow rate of deposition under conditions of open oceanic circulation has lead to well oxygenated environments (widespread bioturbation through much of the sequence), while the limited thickness would hardly be sufficient for thermal maturation. Rocks are generally highly indurated, dense and tight, and no potential reservoir types have been seen or reported.

In the main Solomons province sediments are mainly terrigenous and many thousands of metres thick. High rate of deposition in rapidly subsiding basins may have created favourable environments for the accumulation and preservation of source material. However, no analyses are known, and there are no seepages of hydrocarbons reported from the region. Vitrinite reflectance measurements on three Miocene and Pliocene samples from Guadalcanal (van Deventer 1971) gave DOM values of 45–51, i.e. well within the range of immaturity. This may markedly improve offshore, however, where sediments are much thicker.

Porosity and permeability in the predominantly clastic rock sequence generally is low. Overall quartz content is low, although some metabasalts on Guadalcanal from where sediments were derived contain up to 40% by volume of quartz (van Deventer 1971); sandstones consist mainly of plagioclase and pyroxene. Flaky minerals such as diorite and clay minerals tend to negatively influence permeability, while the response to burial is an overall deterioration of reservoir qualities.

Nevertheless, there are fine- to medium-grained, well sorted volcanoclastic arenites which should have a fair potential as reservoir rocks. Similar rock types are good oil producers in the Niigata basin of Japan. Also, the potential of turbiditic sequences such as occur in the Mio-Pliocene of the Solomons is well established in many places, and should not be overlooked here.

Coralgal reef limestones, which mainly occur in the Miocene sequences, but locally also in the Late Pliocene to Pleistocene, are probably the most promising targets as potential reservoirs. Although they are often recrystallized and dense, there are zones of good porosity, often vugular. They are widespread and thick on Guadalcanal and probably also occur offshore to the north of it, and are known from Choiseul too. Reef-slope deposits and extensive calcarenites associated with the reefal limestones may also be of interest.

The considerable clay content of fine-grained volcaniclastic mudstones, which are abundant throughout the sequence, should amply provide for good cap rocks.

Since erosion has deeply cut into the sedimentary sequence on all the islands, there is little prospect of any accumulation on land, except perhaps underneath the northern part of Guadalcanal. There is no anticlinal folding there, but the nature of the block-faulting (Hackman in press) may provide for closed structures of reasonable size in subsurface. Besides, there is a good potential for stratigraphic traps, or combined stratigraphic-structural ones formed by the biohermal reef limestones. However, the Tetere shallow basement high (Coleman and Day 1965, Fig. 3) reduces the prospective area considerably. This high projects north-northeastwards offshore towards the little Nughu Island and adjacent reefs and has been recorded on seismic profiles in the locality.

A thick and complete sedimentary sequence is preserved offshore underneath the Central Solomons Trough, where the best prospects for hydrocarbon generation would be expected. Prospective areas of trapping and accumulation may be found along its northern margin, where a number of Late Pliocene anticlines, unconformably capped by Pleistocene, could form prospective structural targets particularly in the area between the Florida Islands and south of Santa Isabel (Figs. 3, 6). Fault blocks along this steeply upturned basin flank may form additional targets in this eastern area and also in the west, i.e. south of western Choiseul; however, active faulting has continued throughout the Quaternary and may have impeded the trapping and accumulation of much of the hydrocarbons in at least some of the fault blocks.

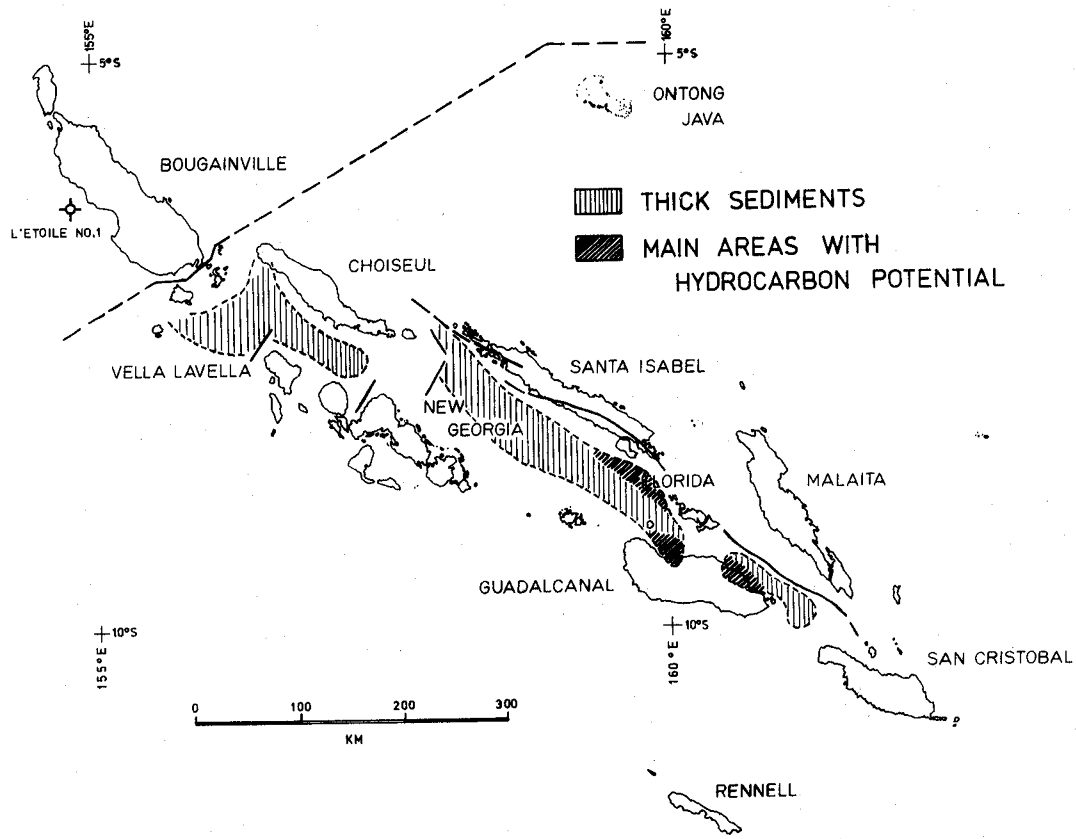


Figure 9. Areas with hydrocarbon potential in the Solomon Islands.

While probably the best and certainly most extensive areas with hydrocarbon potential lie offshore (Fig. 9), it should be noted that most of them are in fairly deep to very deep water, i.e. down to about 1000 m. Shallow-water prospects could only exist close to the northern coast of Guadalcanal, and possibly in Manning and Bougainville Straits. However, seismic coverage at present is not sufficient in these shallow and reef-infested seaways, while existing lines often show basement very close to the surface. In most parts of Manning Strait in particular, sediments are only about 1 s thick and probably consist exclusively of the Early Pliocene Vaghena formation and younger beds. Only along its eastern border are thicker sediments apparent on seismic profiles, but track lines mainly follow an important fault zone there and are in no way conclusive. Thus Shell line P.205A depicted and discussed by Landmesser (1977) is nearly parallel to the regional structural grain, wherefore a correlation with fold elements as shown is very tenuous to say the least (compare his Figs. 2 and 3 which are quite inconsistent).

In summary, and as conceived by this study, there is a fair potential for hydrocarbon generation and accumulation in the main Solomons province, with possible target areas mainly offshore along the northern flank of the Central Solomons Trough, but also on its southern flank in Guadalcanal both on and offshore.

ACKNOWLEDGEMENTS

The author is grateful for the assistance he received from the Solomon Islands Geological Survey, for both accommodation and facilities in the Survey offices during his stay in Honiara, and the organization of several trips into the field. Many fruitful discussions were held with geologists of the British Mapping

Team (I.G.S.), in particular Wyn Hughes, who liberally shared his wide knowledge of Solomons geology and generously extended what other help he could.

The author's wife Charlotte draughted most of the profiles and maps at the Geological Survey office in Honiara, and reductions were arranged by Jack Lai at the Lands and Survey Division, Ministry of Agriculture and Lands, which also produced slides for the oral presentation of the paper at the Suva Symposium. The final draughting of the Structural Map (Fig. 3) was done by Tony Jaegers, and the final typing of the manuscript by Robin Millow, both of New Zealand Geological Survey at Lower Hutt.

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A SEDIMENTARY STRUCTURE SOUTHWEST OF VITI LEVU, FIJI: THE BARAVI BASIN

B M LARUE, J Y COLLOT, AND A MALAHOFF

ABSTRACT

Geophysical data collected during the cruise EVA VI, jointly organized by ORSTOM and NOAA/NOS, identified a basin 2 to 2.5 km deep, oriented NNW–SSE, 20 km southwest of Viti Levu Island, Fiji. Its main features are: (1) Length 80 km; width 30 km (from a preliminary bathymetric map); (2) Sediment thickness equivalent to 2 s (two-way travel time) on a single-channel seismic-reflection profiler; (3) a −70 mgal gravity anomaly indicating a sedimentary section 3 km thick; (4) commencement of filling in Early Miocene times. The hypothesis that the basin was formerly connected with the Aoba Basin, New Hebrides, is discussed.

INTRODUCTION

During EVA VI cruise (organized by l'Office de la Recherche Scientifique et Technique Outre-Mer (ORSTOM) Noumea and the National Oceanic and Atmospheric Administration / National Ocean Survey (NOAA/NOS) Washington) in June 1978, on Research Vessel *Coriolis,* two profiles EVA 648 and 649 were made offshore to the southwest of Viti Levu Fiji (Fig. 1). The seismic reflection, gravity and magnetic data of EVA 648 show a sedimentary basin previously unknown. It was named Baravi Basin by the Fiji Government during the time of the Symposium.

The purposes of this paper are to describe this structure, to identify some of its essential characteristics (depth, sedimentary thickness, age, limits) and to try to fit it into the structural pattern of the southwest Pacific.

STRUCTURE AND AGE

The single-channel seismic reflection, of which an interpreted cross section is given in Fig. 3, suggests that there are two sedimentary basins defined by three ridges (A, B, C). However, the preliminary bathymetric map drawn by the Mineral Resources Department of Fiji (MRD 1979) and used as background to Fig. 1, shows that the central ridge B is simply an undersea advance heading out of Viti Levu Island and made visible by the proximity of the coast. The singleness of the basin is indicated by the long wavelength of the magnetic anomaly, probably produced by a volcanic basement only slightly perturbated by the central ridge B (Fig. 2a).

The sedimentary filling is thicker to the east of the Central Ridge (in the B-C area of Fig. 3) than to the west, probably owing to terrigenous drifts from Viti Levu. The acoustic penetration, slightly less than 2 stwtt (seconds of 2-way travel time) increases as the profile edges away from the coast, i.e. draws closer to the centre of the basin. The basement on which the sediments lie has not been reached. Three distinct sedimentary series can be described:

(1) The youngest, 0.9 stwtt thick, is strongly disturbed by terrigenous drift, coming most probably from the Singatoka River (Fig. 1). A bathymetric bulge and disorder in the sedimentation may be due to the presence of an undersea fan at the river mouth. On both sides of this fan, the layering is more regular.

(2) The second layer, very thin (0.2 stwtt) and comparatively well bedded, is transgressive on the deepest one.

(3) The third layer — base not reached — is not less than 0.6 stwtt thick.

Eastwards, the lowest layer laps and thins on to ridge C. Thus the ridge and the sedimentary layer are probably coeval, i.e. Plio-Pleistocene (Geological Survey of Fiji, 1965), confirmed as 4.6 m.y. by K-Ar dating of volcanics of Vatulele by Whelan and Gill (1979).

Westwards the lowest layer abuts ridge B. A comparison with land geology (see Fig. 1,

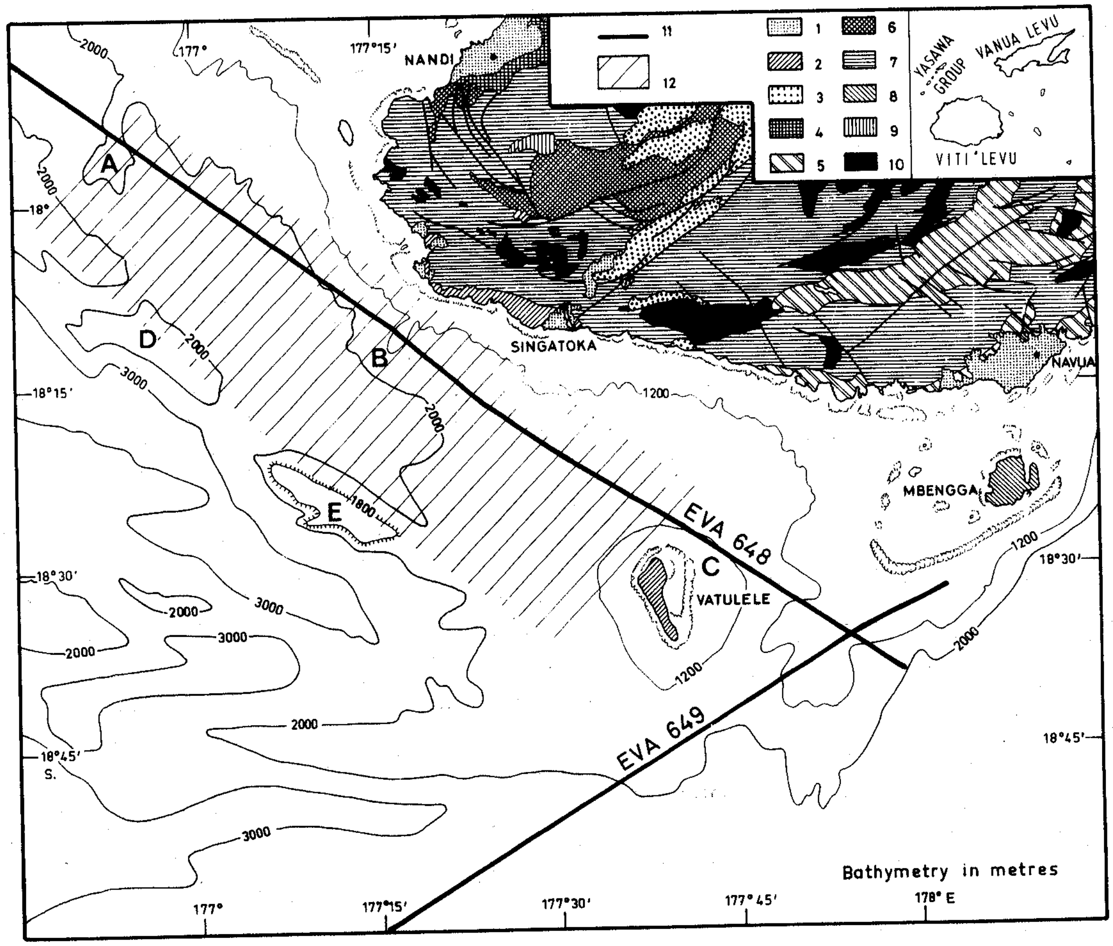

Figure 1. Baravi Basin: location and ship tracks; bathymetry (after MRD 1979); geology (after Rodda 1967).

KEY TO LEGEND

Plio-Pleistocene and Recent
1 Alluvium
2 Thuvu Sedimentary Group and Vatulele

Mio-Pliocene
3 Navosa Sedimentary Group
4 Nadi Sedimentary Group
5 Mendrausuthu Andesitic Group

Eocene Miocene
6 Singatoka Sedimentary Group
7 Wainimala Group
8 Undifferentiated Cainozoic volcanics

Intrusives
9 Upper Tertiary
10 Lower Tertiary

11 Ship track
12 Baravi Basin

based on Rodda 1967, Rodda *et al.*, 1967, and Geological Survey of Fiji, 1965, who assign Vatulele to the Thuvau sedimentary group) leads us to identify ridge B with the Wainimala group, aged Eocene–Miocene. Thus we see the same structure on land as at sea: Eocene–Upper Miocene basement, ridge B, and Wainimala group, on which lie sediments, at least as old as Pliocene, covered by recent terrigenous drifts and alluvium. To the west of the central ridge (section AB of Figs. 1, 2 and 3) only two separable sedimentary layers are seen. An irregular sedimentation is peculiar to the first one (1′), the thickness of which can reach 1 stwtt. A network of relatively shallow normal faults indicates tensional movements, parallel to the shore. Layer 1′ is discordant on a strongly tectonized layer 2′ + 3′, which is thought to be the equivalent of layers 2 and 3 of section B-C. Northwestwards, away from

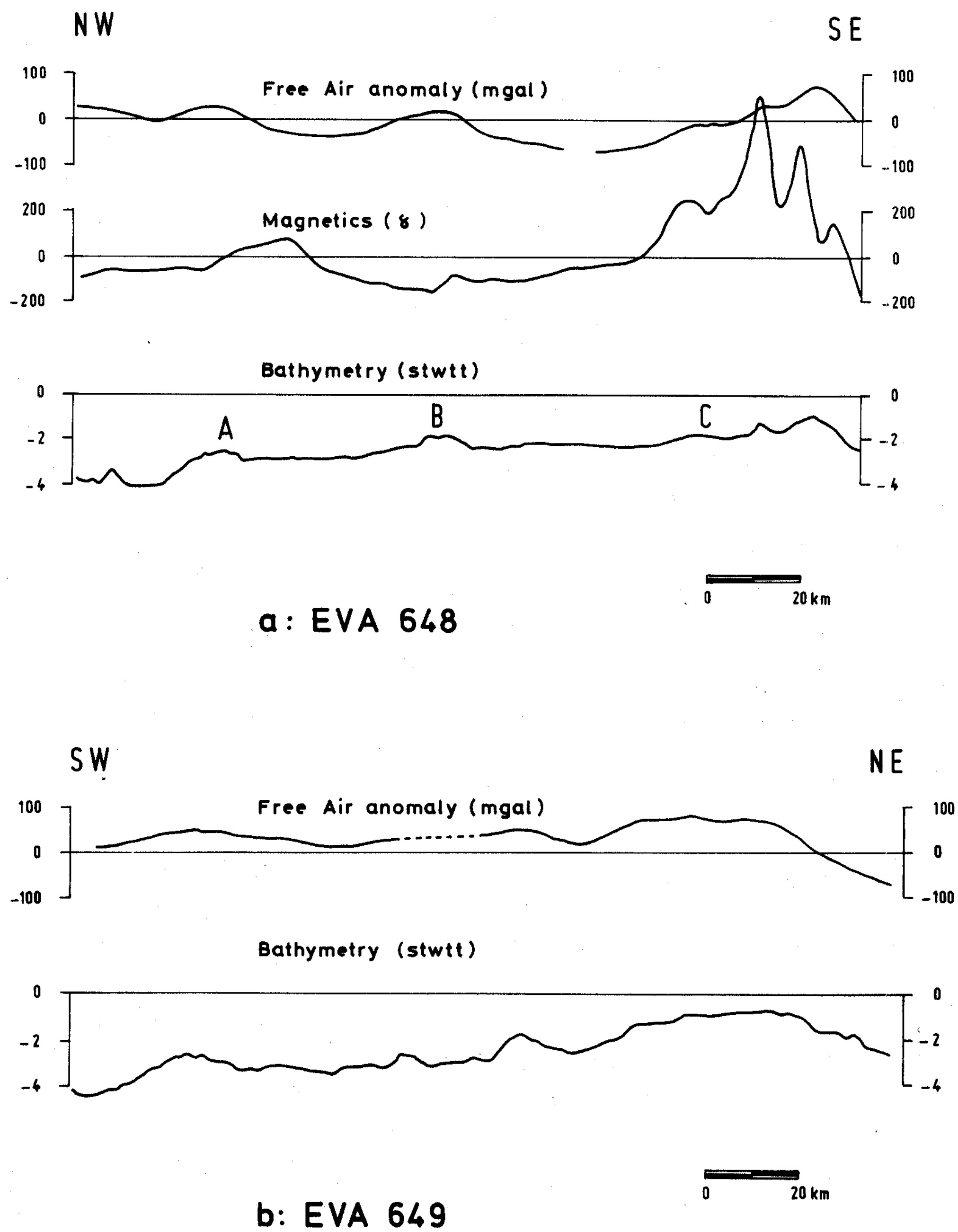

Figure 2. Profiles EVA 648 and 649 — bathymetry, magnetics, gravity.

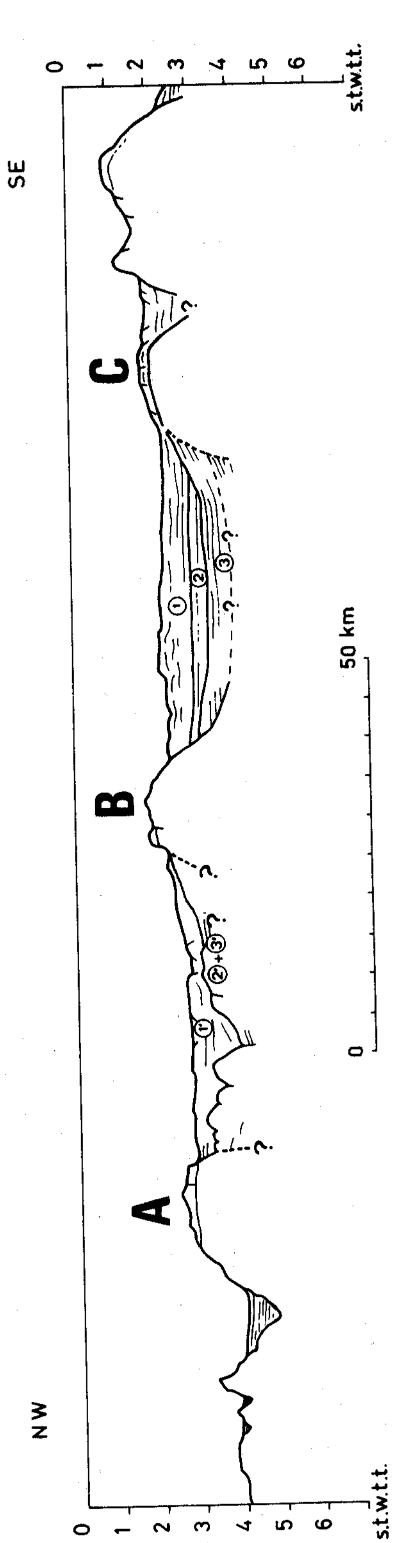

Figure 3. Interpreted seismic reflection EVA 648.

ridge A, deepening of the sea floor, thinning of the sedimentary blanket and fracturing of basement characterize the transition zone between the Fiji Platform and the North Fiji Basin.

SEDIMENT THICKNESS

The total thickness of the sedimentary filling can be estimated by gravity. The free-air anomaly indicates a −0.4 mgal/km regional gradient in contrast to the bottom slope, which reaches −0.9 mgal/km after reduction of the water influence. This is due to the deepening of the Moho, whose depth reaches 24 km under Viti Levu as calculated by Robertson (1967) from the −30 mgal Bouguer anomaly. Two negative anomalies, respectively −70 and −90 mgal, correlated with the sedimentary filling, are superimposed on this gradient. The model has been calculated in two dimensions, made necessary by the disposition of the single gravity profile. In this case, in consequence of the basin configuration (Fig. 1) the sediment thickness is underestimated. Estimations of densities are needed to evaluate thickness. Nettleton's (1940) method gave a value of 2.6 for the ridges, which is close to that found by Robertson (1967) for the rocks of the Wainimala group, namely 2.6 ± 0.13 for the 'intermediate and acidic igneous rocks' and 2.76 ± 0.1 for the 'basic igneous rocks'. A density of 2.7 has been adopted for the base of the basin. As far as the sediments are concerned, a comparison with other sedimentary basins (Collot and Missegue 1977, Pontoise *et al.* 1980) leads us to estimate the density of layers 1 and 1′ at 2.0 and that of layers 2, 2′ and 3 at 2.2. These figures, introduced with a geometrical feature compatible with the seismic reflection in a 2-D programme (Talwani *et al.* 1959) adapted for a HP 9845 computer (Missegue 1979), produced the model shown in Fig. 4. It should be noted that a higher density, 2.9, must be attributed to Mbengga in order to match the observed anomaly: the volcanic character and strong magnetism of Mbengga seem to justify this. The maximum thickness of sediments is 3 km close to ridge B.

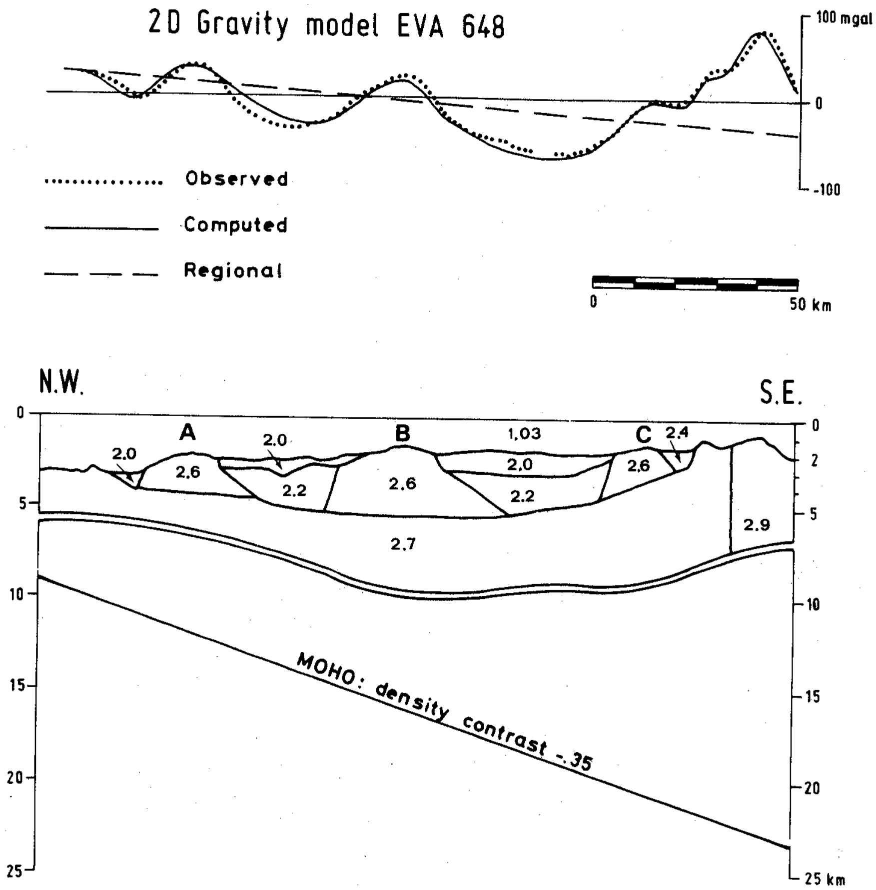

Figure 4. 2 D gravity model, EVA 648.

DIMENSIONS AND EXTENT

The Baravi Basin is estimated to be 80 km long and 30 km wide. As shown by Fig. 2, no structures similar to those seen on EVA 648 are visible on profile 649. From this observation and the bathymetric chart, the south-eastern limit of the basin is undoubtedly the Vatulele and Mbengga headings. On the north-west, the bathymetric bulge characteristic of a ridge (Fig. 1) is aligned with the Yasawa Group, which forms the ultimate limit of the Fiji Platform.

Crosswise, only bathymetric information is available: two rises are present — D and E (Fig. 1) — a few hundreds of metres high and about 5 km wide. They must have acted as a trap in the sedimentary process, in which the importance of terrigenous drift has been previously noted.

The Baravi Basin is the most southerly structure trending parallel to the edge of the Fiji Platform. (Fig. 1).

STRUCTURAL CONTEXT

Geodynamic reconstructions of the South-west Pacific based on petrology (Carney and Macfarlane 1978), paleomagnetics (Falvey 1978), or magnetic anomalies (Malahoff *et al.*, 1979) suggest that the New Hebrides and Fiji were linked in a single island arc in Miocene times. In such reconstructions one can note the alignment of the Baravi Basin and the North and South Aoba basin (excluding Aoba, Ambrym, other recent islands and the southern part of the arc).

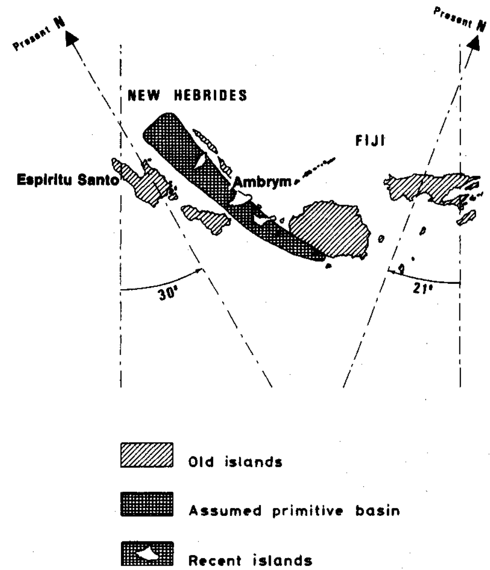

Figure 5. Reconstruction of the Melanesian Island Arc in Miocene times.

The similarities between the two basins are not limited to their tectonic position, but extend to their crosswise dimensions (Luyendyk *et al.* 1974, Ravenne *et al.* 1977), the sedimentary fill (Ravenne *et al.* 1977, Dugas *et al.* 1977), gravity anomaly (Luyendyk *et al.* 1974) and even the date of commencement of sedimentary infilling (Coleman 1969, Carney and Macfarlane 1980).

It should be noted that they predate the opening of the North Fiji basin (Chase 1971; Falvey 1978; Malahoff *et al.* 1979; Halunen 1979). Thus a continuous structure (Fig. 5) could have been split by expansion of the marginal basin. Since, according to Wood (1980), the Yasawa Group marks the transform fault which gives evidence of the early opening of the North Fiji Basin, the existence of, and similarities between, the two basins impose contraints to the Melenasian arc reconstruction (Fig. 5) and lead to a consideration of the southern islands of the New Hebrides as postdating the opening of the North Fiji Basin, which is in accordance with their age.

CONCLUSION

Though the discovery of fairly deep basins does not provide any short-term interest in terms of petroleum potential, it is essential to make a systematic inventory of such sedimentary structures. This paper indicates that a number of them might still be discovered in the Southwest Pacific.

ACKNOWLEDGEMENTS

We extend our thanks to the Mineral Resources Division, Fiji for permission to publish extracts from their preliminary bathymetric map, and to C. Jouannic, CCOP/SOPAC, for his valuable help.

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MARIANA FOREARC TECTONICS

CARY L MROZOWSKI AND DENNIS E HAYES

INTRODUCTION

As known petroleum reservoirs become depleted and substantial new deposits more difficult to locate, the petroleum industry has directed considerable attention towards the less well known, geologically more complicated areas of the earth. For example, the complex regions of crustal convergence in the southwestern Pacific and in Southeast Asia have come under close scrutiny. These regions have also been the sites of intense investigation by various academic and governmental agencies.

Consequently, many geophysical and geological data have recently been collected from the forearc region of the Mariana arc-trench system. These data were obtained during pre-drilling site surveys for the Deep Sea Drilling Project (DSDP) Leg 60, aboard the D/V *Glomar Challenger* during the actual drilling leg, and as part of the International Decade of Ocean Exploration - Southeast Asia Tectonics and Resources (IDOE-SEATAR) program. In this paper we summarize our preliminary results from seismic-reflection mapping of shallow structures in the Mariana forearc region.*

THE DATA SET

Although the data set for this study included all available geophysical and geological data from the Mariana forearc region, the study was based primarily on single-channel seismic reflection data, 24-fold multichannel (MCS) reflection data and 3.5 kHz echograms. The ship track distribution is shown in Fig. 1. These data were supplied by the Lamont-Doherty Geological Observatory, the Hawaii Institute of Geophysics, the Scripps Institution of Oceanography, and the Deep Sea Drilling Project of the National Science Foundation.

THE TECTONIC SETTING OF THE STUDY AREA

The Mariana arc-trench system is a manifestation of the continued west-directed underthrusting of the Pacific Plate beneath the Philippine-Mariana 'Plate'. The forearc region of this system is bordered to the west by the volcanically active Mariana Ridge (arc) complex and to the east by the trench-slope break (the upper culmination of the inner wall of the Mariana Trench) (see Fig. 2). The Mariana Ridge complex is composed of island-arc type volcanic and intrusive rocks, and of volcaniclastic and calcareous sediments. On the southern end of the ridge at Guam, calcareous sediments date back to Eocene (Tracey *et al.* 1964).

RESULTS

In this study we have mapped the locations, trends and surface offsets of shallow faults, and the positions of 'basement' outcrops and other relevant structures in the Mariana forearc region (Fig. 1). Within the mapped area (145°30'E to 148°00'E, 17°25'N to 18°15'N) the seafloor is thickly sedimented and generally smooth. Water depths increase from 2400 m near the foot of the Mariana Ridge to 4200 m at the trench-slope break. Many fault scarps offset the seafloor, more than were expected from our review of studies in other forearc basins. Two seamounts are

* The complete background information and supporting data for this study and a detailed discussion of the results are contained in a paper titled A Seismic Reflection Study of Block Faulting in the Mariana Forearc Region by C L Mrozowski and D E Hayes. This paper has been submitted for publication in a special American Geophysical Union monograph on 'The Tectonic/Geologic Evolution of Southeast Asia', edited by D E Hayes.

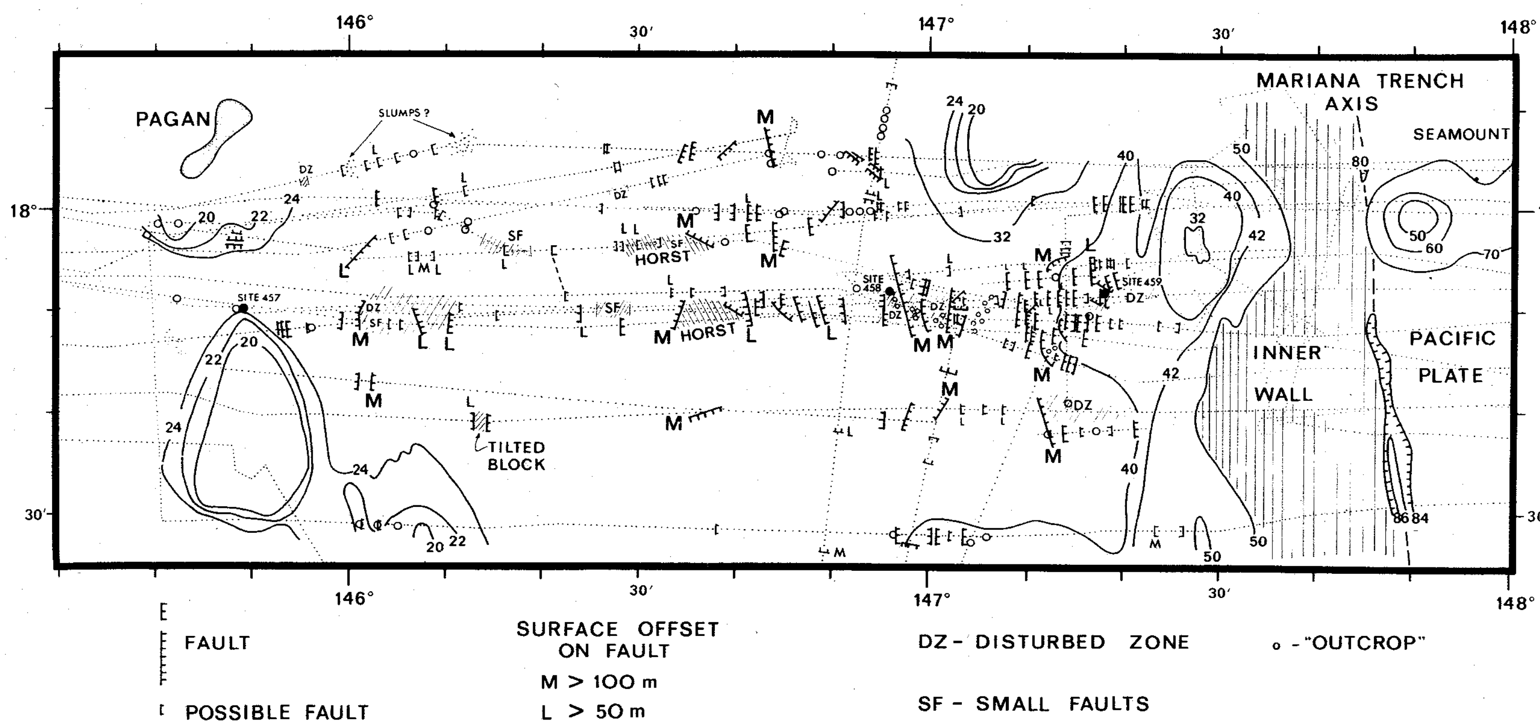

Figure 1. Structural map of the Mariana forearc region. Pagan Island is on the north-south trending Mariana Ridge (arc) complex. 'Outcrop' is defined as the intersection of acoustic basement with seafloor; some outcrops may not be igneous. Disturbed zones may be areas of high density of minor faults. Dotted lines are tracks from which reflection data were used; minor data gaps exist on some tracks. Numbered sites are DSDP sites. Preliminary bathymetry courtesy of D Hussong. Contours in 100's of meters.

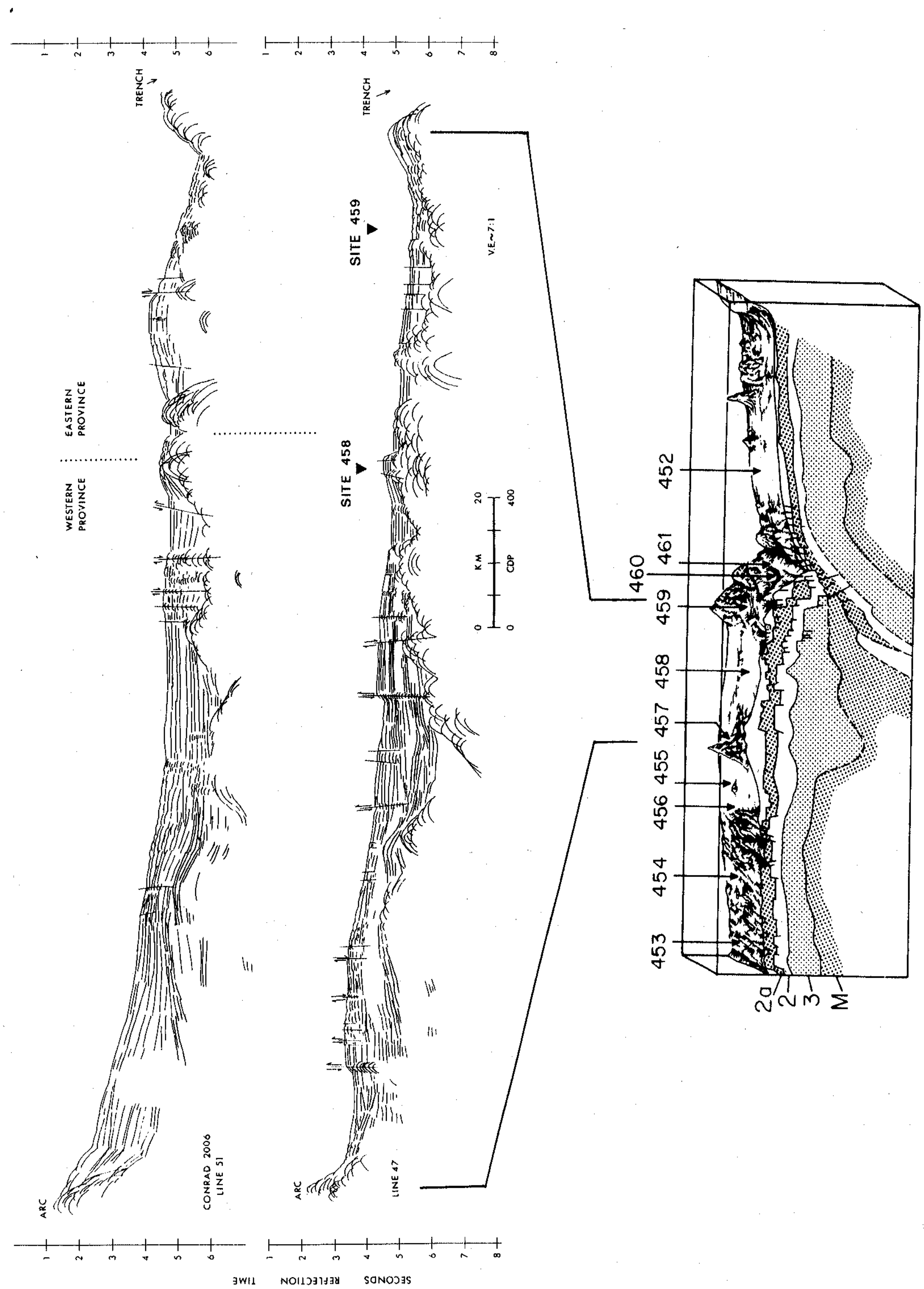

Figure 2. Line drawings of two stacked, E–W oriented MCS profiles from the study area and a physiographic diagram of the Mariana arc-trench system (redrawn from Hussong, Uyeda *et al.* 1978). On the MCS profiles the eastern and western structural provinces are indicated. Note the thick, stratified sediments and the hyperbolated nature of basement in the western province. On the physiographic diagram, 2a, 2, and 3 are crustal layers; M indicates mantle. MCS profiles were collected by the R/V *Conrad* of the Lamont-Doherty Geological Observatory. DSDP sites 452-461 are indicated.

located in the study area at approximately 147°10′E and 147°30′E.

From the seismic reflection data, we have interpreted the faults to be dominantly very high angle normal faults. Reflection records show offsets of (1) seafloor, (2) reflectors within the sedimentary section and occasionally, (3) acoustic basement across the faults (Fig. 2). Seafloor offsets range from 10 m to 250 m. We have no evidence to indicate that strike-slip motion is also occurring on these faults, but the possibility cannot be excluded, since it is difficult to detect with our data set. The well defined fault scarps and the presence of diffuse shallow seismic activity in the forearc suggest the faults are currently active. Some faults may have been active since the formation of the arc-trench system. (Hussong, Uyeda *et al.* (1978) contend that DSDP Sites 458 and 459 have undergone continued uplift and subsidence, respectively, throughout their history. This motion was most likely accommodated on fault planes.) Most fault trends parallel those of the ridge complex and trench, but transverse trends are also noted. The data distribution favors the identification of N–S trending faults.

The forearc region can be divided into two structural provinces which parallel the arc. The western province occupies about two-thirds of the forearc region (Fig. 2). In this province substantial sequences of flat-lying, well stratified sediment rest above a distinct acoustic basement. These sediments thicken toward the west. The acoustic basement is characterized by strong, overlapping diffraction hyperbola on unmigrated MCS profiles. Basement shoals to the east and crops out at approximately 147°E. This outcrop zone defines the eastern boundary of the western structural province. Drilling at Site 458 revealed that the basement is igneous rock of Early Oligocene (?) age (Hussong, Uyeda *et al.* 1978). The western province is extensively faulted. Some horst blocks have boundary faults trending transverse to the trend of the ridge complex.

The eastern structural province rests on the crest of a volcanic ridge (the western flank of which forms part of the basement of the western structural province). The eastern province is characterized by a complex pattern of many normal faults and a relatively thin cover of stratified sediments. The down-dropped blocks of the faults are commonly located towards the trench.

DISCUSSION

Our study reveals extensive high-angle faulting throughout the Mariana forearc region. In the eastern structural province the local pattern of faulting indicates that trenchward subsidence may be occurring. Subsidence in the eastern forearc region, also, has been suggested by Hussong, Uyeda *et al.* (1978) on the basis of drilling results from DSDP Site 459. Models of typical arc-trench systems predict uplift, not subsidence, for this portion of the forearc (e.g. see Karig 1974). This uplift would be evidenced by convergent, arcward-tilted sediment reflectors in the distal regions of the forearc basin, (see discussion on p. 16, Dickinson and Seely 1979), but none is observed. The forearc volcanic ridge, on which the eastern structural province rests, appears to have formed early in the history of the arc-trench system and since that time has served both as the seaward dam for arc-derived sediments and as the trench-slope break. Faulting in the western portion of the western structural province may be related to faulting or igneous activity on the arc massif itself; faulting in the eastern portion of the western province to motions in the forearc volcanic ridge. The origin of faults in the central portion of the province is unknown.

ACKNOWLEDGMENTS

We would like to thank Donald Hussong for allowing us access to unpublished data from the Hawaii Institute of Geophysics, and Daniel Karig for his helpful discussion. The manuscript was reviewed by John E Damuth and Jeffrey Weissel. Mary Ann Garland and Hester Haring Cason assisted in preparation of the manuscript. This work was supported by National Science Foundation grants UC/NSF C482-2, OCE 76-23382, OCE 75-16027 and OCE 78-17021.

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LATE QUATERNARY UPLIFT HISTORY FROM EMERGED REEF TERRACES ON SANTO AND MALEKULA ISLANDS, CENTRAL NEW HEBRIDES ISLAND ARC

C JOUANNIC, F W TAYLOR, A L BLOOM, AND M BERNAT

Currently uplifting reef terraces on Santo and Malekula islands, in the central New Hebrides arc, offer a detailed record of vertical tectonics related to the subduction of the Australian plate beneath the Pacific plate.

Inferred uplift rates on Santo range vary widely from 1 to 7 mm/y, generally increasing from east to west. They define an elliptical zone of maximum uplift, which corresponds to the western mountain belt of Santo. This must be related to the location of western Santo very close to the axis of the New Hebrides trench, which is interrupted in front of Santo. The uplift of southern Santo may be influenced particularly by subduction of the d'Entrecasteaux fracture zone, a major bathymetric feature on the underthrusting Australian plate. The uplifted reef terraces which form the Eastern Plateau Limestones, overlying the eastern half of Santo, are tilted E to ENE in the northeastern and central parts of the island, E to ESE in southeastern Santo and Malo. This is consistent with the apparently elliptical shape of the zone of maximum uplift rate.

Such high uplift rates are not observed on Malekula, which is not on the trench axis, although still abnormally close to the thrust zone. Northern and southern parts of the island behave differently because they are separated by a tilt discontinuity across central Malekula. In northern Malekula, inferred uplift rates range from 0.5 to 4.3 mm/y, with a zone of maximum uplift in the southwestern part of north Malekula and a general tilt to the NE in its main northeastern part. The southern Malekula inferred uplift rates range from a few tenths of a mm to about 1 mm/y, with a zone of maximum uplift near the tilt discontinuity of central Malekula and a general tilt to the SE. Such an uplift pattern on Malekula island is likely to be related to the presence of the southern margin of the d'Entrecasteaux fracture zone which is being underthrust beneath northern Malekula.

The main surface of the Eastern Plateau Limestones in Santo and the main terrace of northern Malekula are believed to correspond to the 125 000 years old paleosea level by comparing the terrace levels of both islands with the paleosea levels as estimated for the last 140 000 years in the Huon peninsula area (New Guinea). This correspondence of terraces with paleosea levels is supported by uranium-series dates for a few of the lower terraces on Malekula and Santo. If the inferred ages of the higher terraces are correct, then the uplift rates of both Santo and Malekula may have increased significantly some time between about 40 000 y.b.p. and the Holocene epoch. In any case, it is clear that much of Santo and north Malekula emerged quite recently.

INTRODUCTION

The New Hebrides island arc extends NNW–SSE between latitudes 11° and 22°S, from the Solomons island arc to the Hunter fracture zone (Fig. 1). It is related to the subduction of the Australian plate beneath the Pacific plate. The Australian plate is thrusting steeply in a N75± 11°E direction (Pascal *et al.* 1978) at about 12 cm/y (Dubois *et al.* 1977).

The corresponding New Hebrides trench is interrupted between latitudes 14° 30′ and 17° 30′S, where the rugged relief of the d'Entrecasteaux fracture zone on the Australian plate is being thrust beneath the islands of Santo and Malekula (Fig. 2). These two islands would occupy the position of the inner trench slope if the New Hebrides trench were continuous: this unusual relationship places Santo and Malekula on the very thin western edge of the Pacific plate and therefore very close to the thrust zone. Despite the lack of a trench west of Santo and Malekula, seismicity associated with the Benioff zone is comparable to that to the north and south where a trench exists (Pascal *et al.* 1978).

North of Efate, the New Hebrides can be divided into three belts, corresponding to three main phases of volcanism (Carney and Macfarlane 1976): (1) a western belt, containing the oldest volcanic rocks (Oligocene to Middle Miocene), includes Santo, Malekula and the Torres islands; (2) an eastern belt, related to an Upper Miocene to Lower Pliocene volcanism, is represented by Maewo and Pentecost; and (3) between lies a modern volcanic chain consisting of Lower Pliocene to modern volcanic islands. The western belt is the focus of this report because of its unusual position close to the trench and away from the modern volcanic chain and therefore from possible interference by volcanic activity.

Coral-reef terraces have developed widely upon a substrate of volcanic and sedimentary rocks during the Quaternary and have been uplifted throughout much of the New Hebrides. These reef terraces represent paleosea levels (Mesolella *et al.* 1969, Bloom *et al.* 1974), and they record an absolute chronology of uplift history and deformation patterns, because the fossil corals they contain can be radiometrically dated by uranium-series methods and radiocarbon.

Two surveys were done in 1976 and 1977 on Efate, Malekula, Santo and the Torres islands. Data concerning Efate are published elsewhere (Bloom *et al.* 1978), and the Torres islands study is still in progress. In this paper, geomorphological observations and age determinations on Santo and Malekula are first presented and then interpreted.

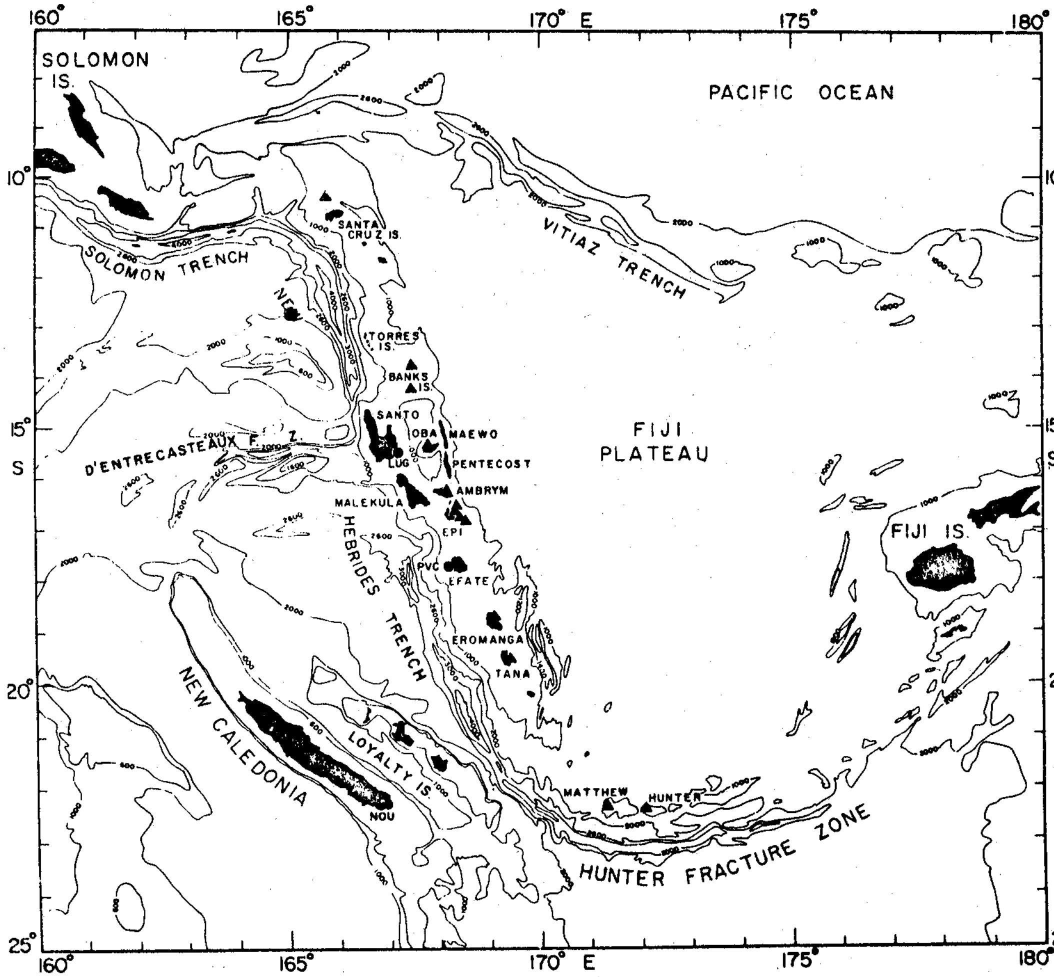

Figure 1. The New Hebrides island arc, bathymetric setting.

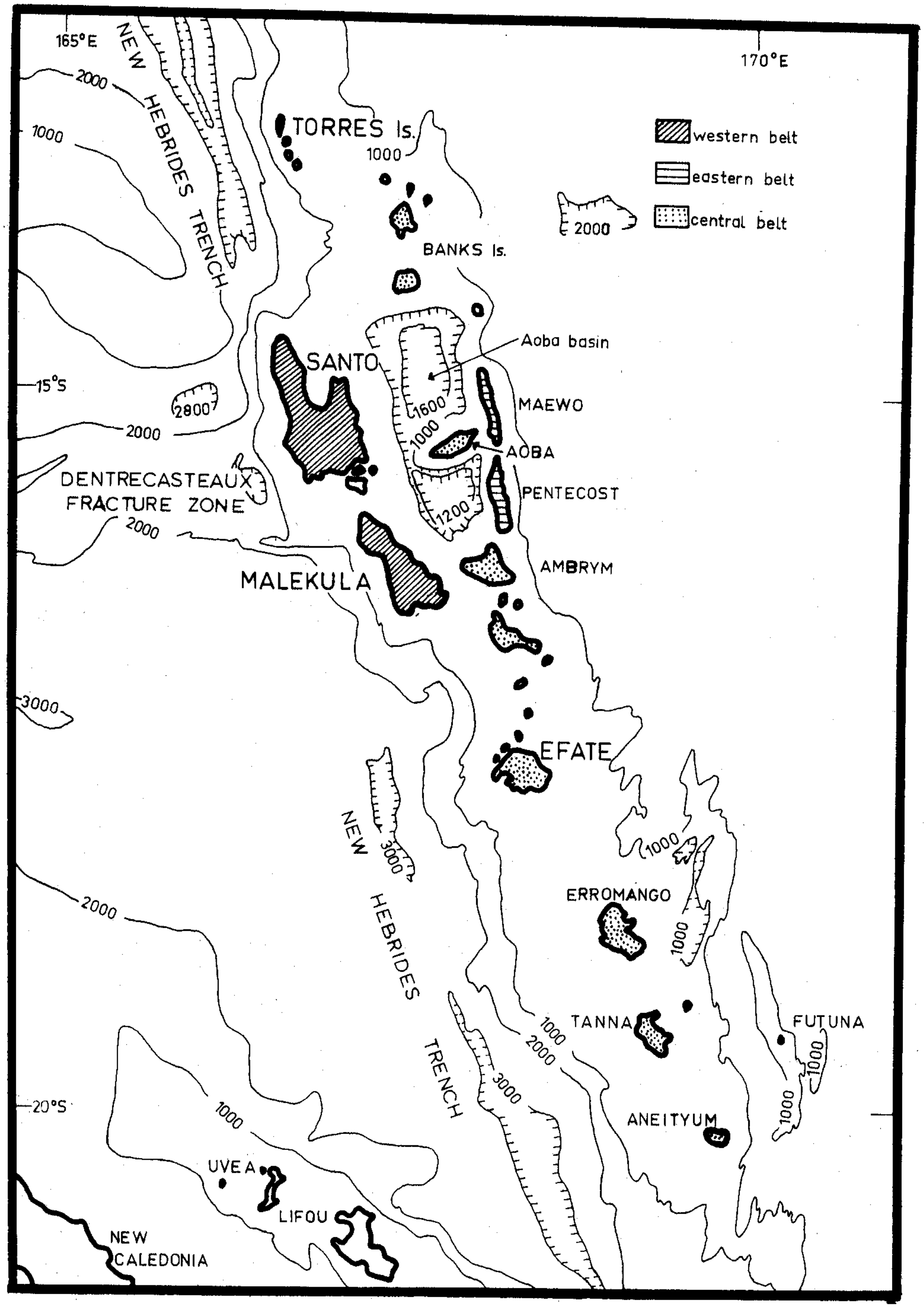

Figure 2. The New Hebrides archipelago.

METHOD

An important part of this paper is represented by the calculations of uplift rates based on the altitudes of emerged reef terraces and the radiometric ages of fossil corals from the terraces. A third factor is the paleosea level at the time each dated coral lived. The higher paleosea levels of the past 140 000 years are reasonably well known (Mesolella *et al.* 1969; Bloom *et al.* 1974; Bender *et al.* 1979). For the older terraces, slight errors in age, altitude and paleosea level are not very important because the large amount of uplift of Santo and Malekula overwhelms other errors. But, for the very young Holocene terraces, close to present sea level, small errors in age, height and paleosea level are more important. For comparison, both a generalized Holocene sea-level-rise curve (Clark *et al.* 1978) and specific Holocene sealevel-rise curves (Bloom 1977) are used to correct the amount of uplift undergone by a coral of a given age. For example, a coral at 14 m altitude is 7000 years old: it must have undergone at least 20 m of uplift because sea level 7000 years ago was about 6 m lower than present. Because this correction assumes that the coral was growing at low-tide level, it gives a minimum amount of uplift.

When two corals from a Holocene terrace give different uplift rates, the higher rate is probably correct; and the coral giving the lower rate lived well beneath the surface.

SANTO ISLAND

The two fundamental geomorphological divisions of Santo are a western chain of volcanic and sedimentary rocks and an eastern limestone plateau zone (Robinson 1969; Mallick and Greenbaum 1977). The western belt is made up of semi-independent uplifted or depressed blocks, in which rocks are steeply tilted and intensely faulted: major faults are mostly parallel to the western Santo coastline or trend NE–SW. In the east, extensive, massive reef limestones rest almost undisturbed over nearly the entire eastern half of Santo: overlying Pliocene sediments and probably Oligo-Miocene volcanic rocks, they form a series of step-edged plateaux which have been uplifted during the late Quaternary. Photogeological interpretation indicates that these plateaux are tilted down slightly E to ENE in the northeastern and central parts of the island, E to ESE in southeastern Santo and in Malo (Mallick and Greenbaum 1977): the angle of tilt is in the order of 1° but reaches 5° (in the higher and older Boutmas plateau, for instance). Thus, there appears to be a sudden change in tectonic style from eastern to western Santo. However, this may be due in part to the lack of a coral limestone carapace on western Santo combined with a gradual increase in uplift rate and faulting from east to west.

The coastal reef-terrace complex of Santo

A continuous coastal fringe of slightly uplifted reef terraces can be observed all along the eastern coast of Santo, much of the southern coast (except where terrigenous sediments are dominant), and the southern islands (Toutouba, Aore, Malo, etc.). Except in the southeast corner of Santo, including islands such as Aore or Toutouba, where it occupies a few square kilometers, the low coastal platform is rather narrow relative to the extent of the high plateaux; its width ranges from a few tens of meters up to a kilometer, generally in the order of a few hundred meters.

The coastal platform, when narrow, consists of uplifted fringing reefs, with a subhorizontal surface inland and a gentle slope seawards, usually edged by a small modern sea cliff and a fringing reef. Where the coastal terrace is wider, its consists mostly of lagoon-type deposits (friable, coarse calcarenite, with many sticks of *Acropora* sp. and numerous *Porites* coral heads).

Queiros peninsula

An uplifted fringing reef surrounds the Queiros peninsula. It is usually edged on its seaward margin by a high modern sea cliff, 3 to 6 m high (Lotoror, Ekar, Bilion, Dolphin island, Areb bay, etc.). In some places, the sea cliff is replaced by a low, sandy shore (Dr Keller's estate, Port Olry mission, Hog Harbour bay). One or two subsidiary terraces sometimes occur between the main terrace and the sea:

(1) The lowest terrace, not well developed, is seen in Dr Keller's estate at 3 m above low-tide level (ALT).

(2) The second one, more obvious, though narrow, is seen between 5 and 8 m ALT in Dr Keller's estate, near Mar hill and at Hog Harbour bay.

Two radiometric dates were obtained from the Queiros peninsula:

(1) A coral head (S-N-1, see Fig. 3), found in growth position close to the surface of the main terrace (13 m ALT in that place), gave a date of 7460 ± 230 y b.p. (Table 1b). If the 7500 y.b.p. sea level was 9 m lower than present (Bloom 1977, Clark *et al.* 1978), a total uplift of 22 m and an uplift rate of 3 mm/y are inferred.

(2) A 2650 ± 100 y b.p. date has been reported previously by Launay and Recy (1972), collected 0.5 m from the top of a 3 m high sea cliff near Hog Harbour: an uplift rate of about 1.1 mm/y is inferred.

SE Santo

The southeastern corner of Santo is characterized by a wide coastal platform and a number of reef-terraced islands offshore. The backreef calcarenite facies dominates the coastal terrace surface. This is illustrated by the terraces of Turtle bay (8–11 m ALT) and Matewulu airfield (9 m ALT). One sample (S-AC-1) from Matewulu terrace is dated at 5940 ± 190 y b.p.: at that time, sea level had not quite reached its present height (Bloom 1977; Clark *et al.* 1978) and a total uplift of 11.5 m is inferred, yielding an uplift rate of 1.9 mm/y.

For a coral from the lowest terrace on western Malo, at 4.5 m ALT, Neef and Veeh (1977) reported a date of 4000 ± 500 y b.p. An uplift rate of 1.1 mm/y is inferred. It is suspected that at least part of a broad terrace levelled at 8–18 m ALT on western Malo may also be Holocene in age.

Southern Santo

From Rose point westwards to Tangoa mission, along the southern coast of Santo, are many emerged paleo-patchreefs on the Holocene terrace. Such a morphology of modern patchreefs exists immediately offshore.

Two of these paleo-patchreefs have been dated (sites S-A and S-B close to Navota Farm School). The top of the first one is 18 m ALT, and dates are 5000 ± 600 and 6500 ± 700 y b.p.: uplift rates of 3.6 and 3.3 mm/y are inferred, after correction of 4 m for 6500 y paleosea level (Bloom 1977; Clark *et al.* 1978). The second fossil patchreef culminates at 13 m ALT: a date of 4180 ± 130 y b.p. (S-B-1) leads to an uplift rate of 3.1 mm/y.

Two other corals from a lower terrace, 2 m ALT, below Navota Farm School (site S-C) have been dated at 1055 ± 80 and 1415 ± 100 y b.p.: the inferred uplift rates, 1.4 and 1.9 mm/y, are not used further in regard to the higher 3.1 and 3.6 mm/y from former sites S-A and S-B, quite close to S-C (see method).

In the vicinity of South Santo lies the small island of Araki, entirely capped by reef terraces. Two samples from the first main terrace, 27 m ALT, are dated consistently at 5430 ± 200 and 5470 ± 160 y b.p.: a high uplift rate of 5.2 mm/y is therefore inferred.

SW and NW Santo

Uplifted fringing reefs form coastal terraces at 6 and 14 m ALT near Tasmaloum, 6 and 10 m ALT at Cape Sinotarip, and 3 and 8 m ALT near Wounpouko, in the area of Cape Cumberland. Older Holocene reefs probably occur at higher altitudes in each of these areas. Instead of being limited by small sea cliffs on their outer edges, the lower terraces are usually bounded by fairly steep slopes without well formed notches. Furthermore, evidence of relatively recent emergence of unknown age occurs along the coast of SW and NW Santo, similar to the 1965 uplifted reef platform visible in NW Malekula (Taylor *et al.* in press).

A coral sample from the 10 m high terrace of Cape Sinotarip (S-Z-5) gives an age of 3200 ± 350 y b.p.: a minimum uplift rate of 3 mm/y is inferred, which is not higher than in the Queiros peninsula or close to Navota Farm School. However, it is possible that this coral was living at some meters depth, or has been downthrown along one of the local faults cutting the coast near Cape Sinotarip.

Six dates come so far from NW Santo. Ranging from 545 ± 90 to 6700 ± 150 y b.p. (Tables 1a and 1b), they indicate uplift rates from 2.1 to 5.5 mm/y. As previously noted (see method), the rate of 5.5 mm/y is selected and is used further in the interpretation.

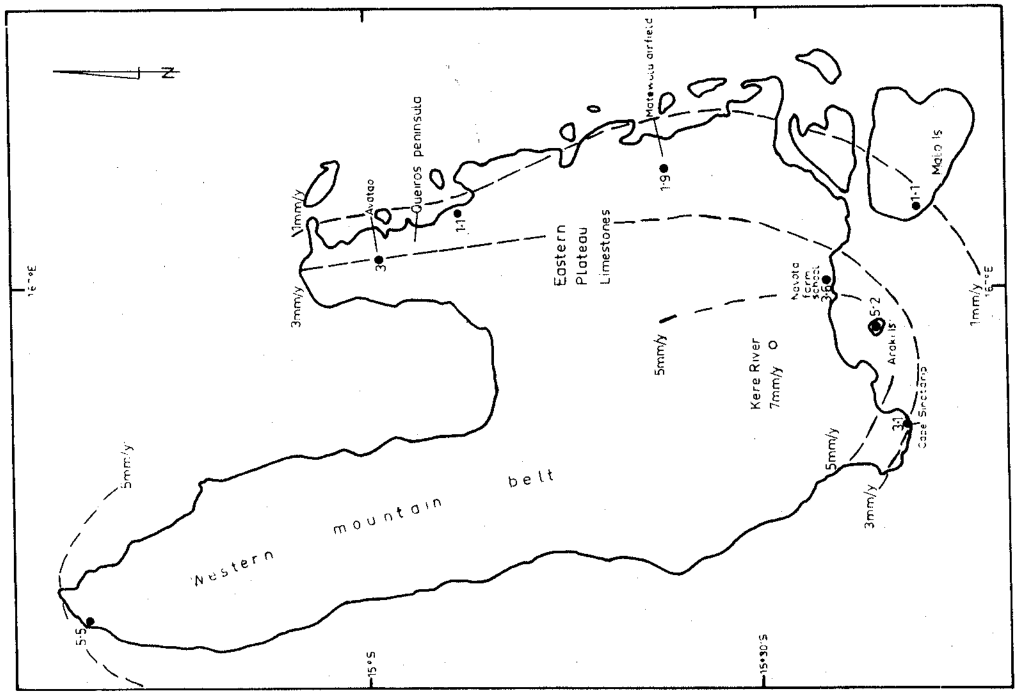

Figure 4. Holocene uplift rates on Santo island.

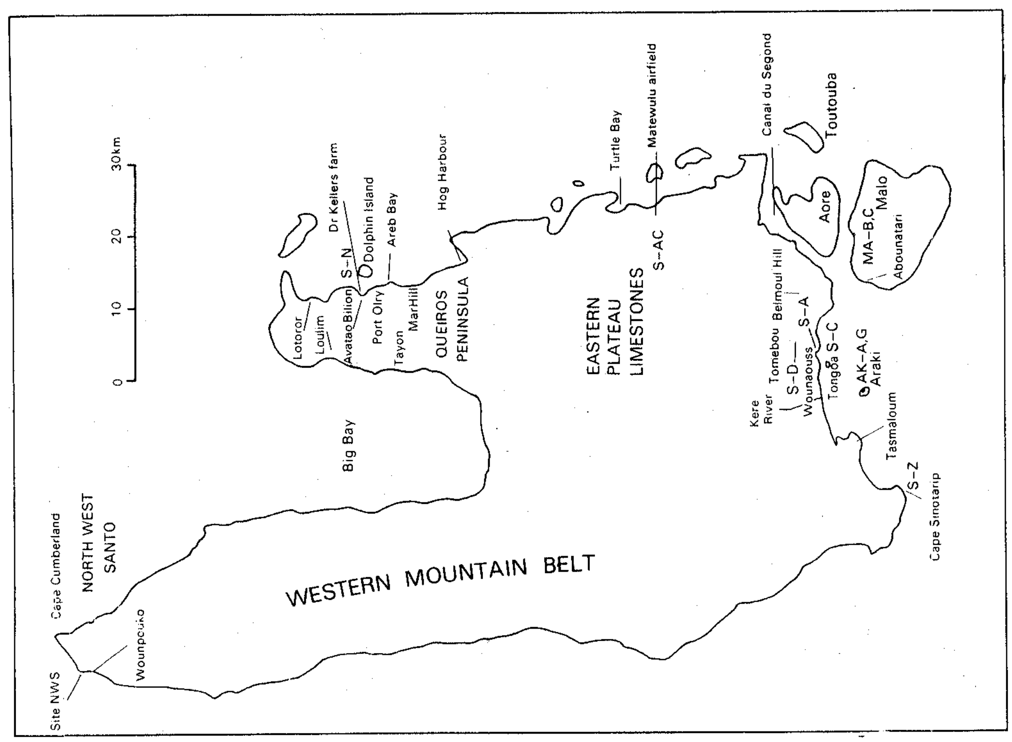

Figure 3. Santo island toponymy and sample locations.

Interpretation

An emerged Holocene reef platform occurs all along the east coast, most of the south coast and on the end of the northwest point of Santo. It seems to consist of as many as three terraces, the uppermost one usually very broad and sloping gently down towards the sea. The variation in maximum height of the Holocene terrace is probably due, at least in part, to differences in distance from the tilt axis, which is thought to trend N–S or NNW–SSE.

The Holocene uplift rates range from 1 to 5.5 mm/y. The rates of 3 to 4 mm/y on the Queiros peninsula and in southern Santo, and of 5.2 mm/y for Araki island, are high. However, on a map of Santo (Fig. 4), the Holocene uplift rates tend to define an elliptical zone of maximum uplift rates, which corresponds topographically to the high western part of Santo. This is consistent with the structural pattern of Santo, if one assumes that the high topography of western Santo occurs because that area has been uplifted most.

The high reef terraces of Santo

Geomorphology

The higher reef terraces are well developed along the northeast, east and southeast coasts of Santo. They occur as steps on the edges of the high wide plateaux that form the extensive Eastern Limestone Plateau Zone of Santo (Robinson 1969, Mallick and Greenbaum 1977).

Most altitudes of the terraces were measured with two Wallace and Tiernon altimeters (one of them recording changes in barometric pressure at the shore) and are compiled in Table 2. The traverses back to the base station show that the instrumental shift can sometimes reach the equivalent of 3 m: but this drift in most cases is not higher than the altitude inaccuracy in estimating the right place to be measured on the terrace. A mean accuracy of ± 3 m is adopted in this report for altitudes of high-reef terraces.

Nine altimetry traverses were made on Queiros peninsula, including one previously reported by Mallick (1970) on Walraoul plateau. The principal traverse is near Avatao (on the 1 : 100 000 IGN map of Santo), where the terraces are visible under the coconut trees of the plantation. In most other cases, dense vegetation partly conceals the morphology, leading to confusion, as between a simple slope break and a sloping, but definite, terrace.

As many as eight different terrace surfaces above the Holocene platform were seen on the Queiros peninsula. Among these high levels, the second one from sea level (39–46 m), the seventh one (182–190 m) and the upper one (218–253 m) are morphologically the most definite terraces. Particularly, the upper one forms most of the top surface of the Queiros plateaux.

In southern Santo, two altimetry traverses have been made. On Tomebou hill, a former island that is now part of the emerged mainland, the altitudes presented here are quite consistent with those previously reported by Mallick (1970) on the same hill. No correlations are obvious among the southern Santo altimetry traverses and another one previously reported on Malo island by Neef and Veeh (1977). This must be due to the fact that the hill behind Belmoul plantation, Tomebou hill and Malo island have been uplifted at different rates. However, the Tomebou terraces correspond well to the main terraces of the Queiros peninsula, particularly with the terraces observed at Avatao: there appears to be a close relation between the uplift rates of both areas.

Radiometric dates

Four new Th^{230}/U^{234} dates come from the high reef terraces of Santo and Malo (Tables 1a and 1b). Two of them date the 41 m high terrace of Tomebou and are quite consistent with one another: they are 38 ± 2 and 37± 2 ka.

Two more dates of 55 ± 4 ka and 223 ± 44 ka have been obtained from corals in terraces on Malo Island at 43 and 53 m respectively, although the higher terrace seems much younger than the lower one. The older terrace may correspond to an older reef exposed by erosion. These two dates must be added to two other dates previously reported by Neef and Veeh (1977) on the same part of Malo: 60 ± 4 and 130 ± 10 ka, dating two terraces, at 49–55 m and 94–98 m, respectively.

TABLE 1a

U/Th radiometric datings and inferred uplift rates on Santo

Sample No. and Locality (see Fig. 3)	Coral Species	Aragonite (%)	U (p.p.m.)	$^{234}U/^{238}U$	$^{230}Th/^{234}U$	Age (× 1000y)	Terrace Altitude (m ALT)	Correction Altitude (m) for Paleosea Level	Inferred Uplift Rate (mm/y)
S-A-1*	*Oulophyllia crispa*	100	2.52±0.07	1.12±0.03	0.045±0.005	5.0±0.6	18	..	3.6
S-A-3†	*Diploastrea heliopora*	99	..	1.14	0.04	6.5±0.7	18	+4	3.3
S-D-1*	*Acropora humilis*	100	3.44±0.07	1.13±0.02	0.30±0.01	38 ±2	41	+38	2.0
S-D-2*	*Montipora* sp.	100	3.58±0.09	1.13±0.02	0.29±0.01	37 ±2	41	+38	2.0
S-N-1*	*Porites lutea*	100	2.75±0.06	1.15±0.02	0.063±0.003	7.1±0.4	13	+6	see text see ^{14}C
S-Z-5†	*P. lutea*	100	..	1.17	0.03	3.2±0.3	10	..	3.1
S-AC-1*	*P. lutea*	100	2.73±0.05	1.14±0.02	0.051±0.004	5.7±0.5	9	+2	see ^{14}C
NWS-C-1†	*Porites* sp.	99	2.78±0.05	1.10±0.02	0.04	4.5±0.3	10	..	2.2
NWS-D-1†	*Porites* sp.	100	2.52±0.05	1.13±0.03	0.03	3.5±0.3	16	..	4.6
MA-B-2†	*P. lutea*	100	2.72±0.05	1.07±0.02	0.89	223±44	53	?	?
MA-C-3†	*P. lutea*	100	2.85±0.05	1.15±0.02	0.40	55 ±4	43	+28	1.2

* Dates from Broecker W S and Goddard J G, Lamont-Doherty Geol. Obs., N.Y., USA
† Dates from Bernat M and Gaven C, Géol. Struct., Univ. Nice, France

TABLE 1b

Radiocarbon datings on Santo, with inferred uplift rates

Sample No. and Locality (see Fig. 3)	Coral Species	Aragonite %	Age	Terrace Altitude (m ALT)	Correction Altitude for Paleosea Level (m)	Inferred Uplift Rate (mm/y)
S-B-1*	*Diploastrea heliopora*	100	4180±130	13	..	3.1
S-C-2†	*Platygyra sinensis*	100	1415±100	2	..	1.4
S-C-4†	*Favites virens*	100	1055±80	2	...	1.9
S-N-1*	*Porites lutea*	100	7460±230	13	+9	3.0
S-AC-1*	*P. lutea*	100	5940±190	9	+2.5	1.9
NWS-A-1†	*P. lutea*	100	6700±150	10	+5	2.2
NWS-A-5†	*P. lutea*	100	5745±200	10	+2	2.1
NWS-B-1†	*Pavona clavus*	100	545±90	3	..	5.5
NWS-B-2†	*Acropora* sp.	100	650±95	3	..	4.6
AK-A-1†	*Acropora* sp.	99	5470±160	27	+1	5.2
AK-G-2†	*Acropora* sp.	100	5430±200	27	+1	5.2

* Dates from Broecker W S, and Goddard J G, Lamont-Doherty Geol. Obs. N.Y., USA
† Dates from Fontes J-Ch, Hydrol. et Géoch. Isot., Univ. Paris Sud, France

TABLE 2

High-terrace altitudes measured on Santo

	Altitude in metres ALT*																														
Traverse Locations	0	10	20	30	40	50	60	70	80	90	100	110	120	130	140	150	160	170	180	190	200	210	220	230	240	250	260	270	280	290	300	310
NE SANTO																																
Lotoror		11											127	133			161							232								
Bilion	9	18		32	40																											
Dr Keller's farm		14	24		39	(53)																										
Avatao		13			42		(60)			(92)			121		145				185			218			240		260					
Walraoul plateau	7			30				70									160			190						253						
(Mallick 1970)																																317
W. Queiros (Loulim)		19			46		63									154			189													
W. Queiros (Tayon)		15												136					182													
Dolphin island		10	(21)															173														
Mar hill		13			42								128																			
SOUTH SANTO																																
Belmoul hill			(29)								106			136			164															
Tomebou		14			41		(64)			90						154		170							240							
Tomebou (Mallick 1970)			27	35					85							153									240							

*ALT = above mean low tide. Altitudes in ( ) mark a less distinct terrace, rather corresponding to a break in slope.

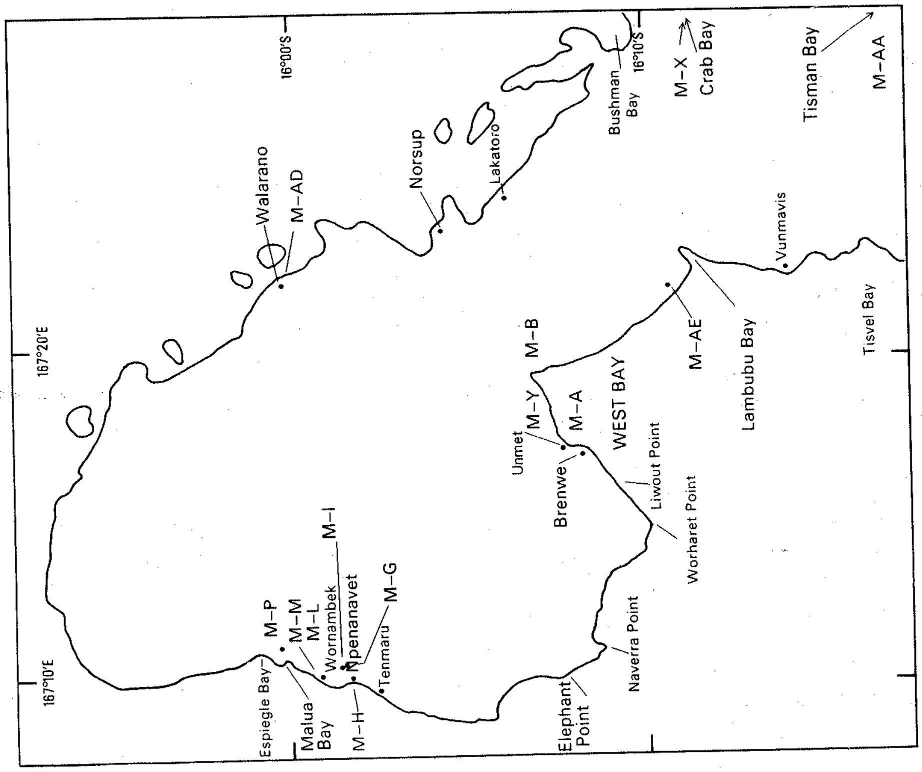

Figure 5. Malekula island: toponymy, sample locations.

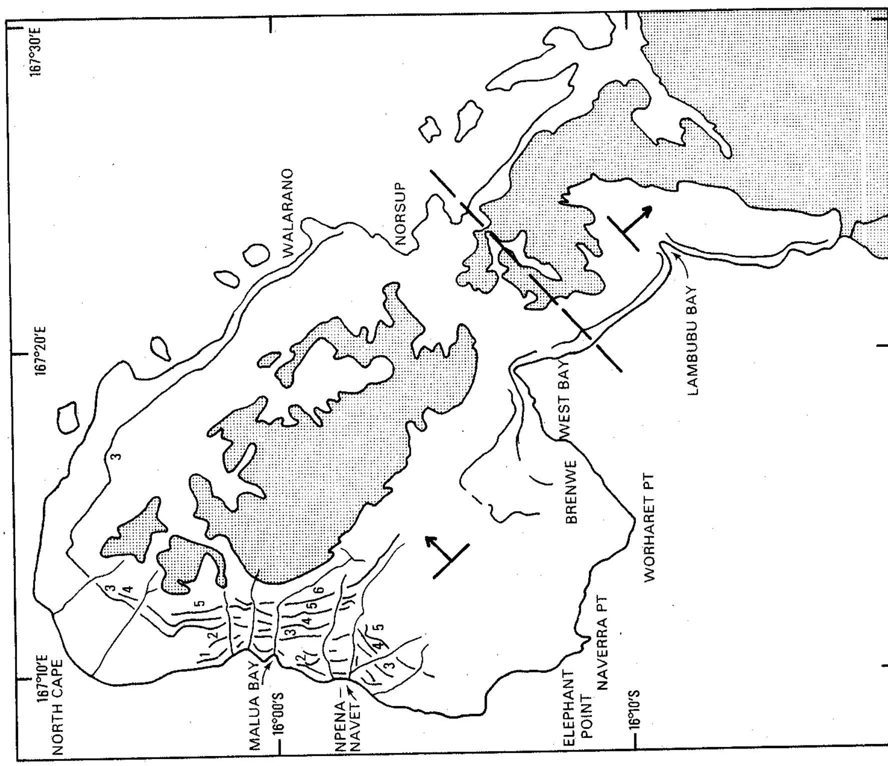

Figure 6. Malekula island: location of uplifted reef terraces (after Mitchell 1971) and tilt features. Stippled areas pre-Pliocene igneous rocks; clear areas Plio-Quaternary.

Finally, one well preserved solitary coral collected in a richly fossiliferous deposit located about 70 m above sea level on the Kere river (south Santo), about 5 km inland, is reported by Ladd (1976) to give a radiocarbon age of 25 280 ± 460y. From faunal analyses (especially from molluscs), the deposits appear to represent an offshore bed laid down at a depth in excess of 50 m.

Inferred uplift rates

It has been clearly demonstrated (Mesolella *et al.* 1969, Bloom *et al.* 1974) that uplifted reef terraces record a succession of high sea levels and represent an interaction between glacio-eustatic sea-level oscillations and regional vertical movements. A sea-level curve for the last 140 000 years has been established from uplifted reef terraces of the Huon peninsula, New Guinea (Bloom *et al.* 1974), and of Barbados, West Indies (Mesolella *et al.* 1969; Bender *et al.* 1979).

Uplift rates of the high reef terraces of Santo can thus be inferred on the basis of the radiometric dates, terrace altitudes and sea-level history:

(1) The 41 m high terrace of Tomebou, dated at 37 and 38 ±2 ka, is likely to correspond to the 40 ka Huon paleosea level, which is assumed to have been 38 m below present sea level (Bloom *et al.* 1974). Thus, a total uplift of 79 m and an uplift rate of about 2 mm/y are inferred for this Tomebou terrace.

(2) In Malo, the 43 m high terrace and Neef and Veeh's (1977) 49–55 m high terrace, respectively dated at 55 ± 4 and 60 ± 4 ka, can be considered equivalent to the 60 ka Huon terrace. The 60 ka paleosea level is supposed to have been 28 m below present sea level (Bloom *et al.* 1974). A total uplift of 71 m and therefore an uplift rate of about 1.2 mm/y are inferred for this terrace. The older dates (223 ± 44 ka and 134 ± 10 ka) are not consistent with one another and are to be confirmed by additional work on older terraces of Malo.

(3) If the 25 280 y coral from the Kere river is contemporaneous with the sediments in which it was collected, then it was deposited at a depth of more than 50 m when sea level is estimated to have been around 50 m below present sea level (Bloom *et al.* 1974). Given the present altitude of the fossiliferous outcrop (70 m above sea level), a minimum uplift of 170 m and a very high uplift rate of 7 mm/y are inferred for this area.

MALEKULA

Malekula consists of a mountainous, deeply dissected core of Miocene volcaniclastic sediments and igneous rocks surrounded mostly in its northern part by a cap of terraced coral limestones, the Tenmaru Reef Limestones (Mitchell 1969, 1971), which reaches a maximum altitude of 614 m. As in Santo, there is a nearly continuous low terrace along most of the north Malekula coast. At the inland border of the modern reef, much of north Malekula is fringed by dead corals killed by earthquake-related emergence in 1965. A detailed comparison of the 1965 uplift pattern and that recorded by the older reef terraces is presented elsewhere (Taylor *et al.* in press).

The coastal terrace complex of northern Malekula

From Vunmavis (south of Lambubu bay) around north Malekula to Tisman bay, on the eastern coast, the low coastal reef terrace disappears for only about 3 km between Naverra and Elephant points (western coast). In the west, from Lambubu bay to Malua bay, it usually does not exceed a few hundred meters wide. It widens to a maximum of about 2 km around the northern coast. On the eastern coast, its width is more variable, ranging from a few meters up to several hundred meters.

The reef terraces edging this low platform along the western coast are uplifted fringing reefs. However, lagoonal deposits are present along the northeastern coast, where former embayments have emerged (e.g. Norsup, Sarmet).

Geomorphologically, in the West bay area the low coastal platform is formed by

(1) the 1965 reef platform, up to about the present high tide level (e.g. Lambubu bay, site M-B, Brenwe, Liwout point, Worharet point); (2) the remains of an apparently slightly older reef, 1 m higher than the 1965 platform (e.g. Lambubu bay, Brenwe, Liwout point);

TABLE 3a

U/Th radiometric datings and inferred uplift rates on Malekula

Sample No. and Locality (see Fig. 3)	Coral Species	Aragonite %	U (p.p.m.)	$^{234}U/^{238}U$	$^{230}U/^{234}U$	Age (× 1000 y)	Terrace Altitude (m ALT)	Correction Altitude for Paleosea Level (m)	Inferred Uplift Rate (mm/y)
M-G-3*	*Platygyra lamellina*	100	2.76±0.08	1.12±0.03	0.40±0.02	55±4	68	+28	1.6 (see text)
M-G-4*	*Leptoria phrygia*	100	2.73	1.11	0.41	57±2	68	+28	1.6 (see text)
M-I-2*	*Platygyra sinensis*	100	2.60±0.06	1.12±0.02	0.016±0.004	1.7±0.5	57	aberrant	aberrant
M-I-3*	*Acropora* sp.	100	3.35	1.13	0.35	47±2	57	see text	see text
M-L-1*	*Plesiastrea curta*	99	2.89±0.07	1.14±0.03	0.035±0.004	3.9±0.5	11	..	2.8
M-L-2†	*Favia stelligera*	99	..	1.14	0.03	5.8±0.3	11	+2	2.2
M-M-1†	*Acropora humilis*	100	..	1.13	0.03	9.5±0.6	11	see text	see text
M-M-2*	*Favia stelligera*	100	2.34	1.13	0.036	4.0±0.25			
							11	..	2.75
M-P-3*	*F. stelligera*	100	2.37±0.05	1.13±0.02	0.43±0.02	60±4	37	+28	1.1 (see text)
M-Y-2*	*F. stelligera*	100	2.64±0.07	1.13±0.02	0.050±0.004	5.6±0.5	19.5	+1	3.6
M-AA-2†	*Acropora* sp.	100	3.19±0.09	1.08±0.03	0.60	98±14	30	+15	0.45 (see text)
M-AD-4†	*Acropora* sp.	100	3.16±0.05	1.13±0.02	0.49	72.5±5	30	+13	see text
M-AE-1*	*Porites lutea*	100	2.97±0.09	1.11±0.02	0.50±0.015	75±4	60	+13	see text

* Dates from Broecker W S and Goddard J G, Lamont-Doherty Geol. Obs., N.Y., USA
† Dates from Bernat M and Gaven C, Géol. Struct., Univ. Nice, France

TABLE 3b

Radiocarbon datings on Malekula, with inferred uplift rates

Sample No. and Locality (see Fig. 5)	Coral Species	Aragonite %	Age	Terrace Altitude (m ALT)	Correction Altitude for Paleosea Level (m)	Inferred Uplift Rate (mm/y)
M-A-6†	*Favites pallida*	100	1250±80	5	..	4
M-A-7†	*Favia stelligera*	100	1155±90	5	..	4.3
M-A-8†	*Porites lutea*	100	2530±100	5	..	2
M-H-1†	*P. lutea*	100	6900±180	3	+5	1.2
M-H-4†	*Platygyra lamellina*	100	2475±100	3	..	1.2
M-H-5†	*Porites lutea*	100	2865±130	7.5	..	2.6
M-M-2*	*Favia stelligera*	100	3810±140	11	..	2.9
M-X-2†	*Platygyra sinensis*	100	2970±200	1	..	0.3

* Dates from Broecker W S and Goddard J G, Lamont-Doherty Geol. Obs., N.Y., USA
† Dates from Fontes J-Ch, Hydrol. et Géoch. Isot., Univ. Paris Sud, France

(3) the edge of next terrace, 4–6 m ALT, which is continuous in the West bay area;
(4) the edge of a yet higher terrace, 12 m ALT (site M-B, Brenwe);
(5) a wide terrace, 19.5 m ALT at Unmet.

Along the western coast, between Tenmaru and Espiegle bay, are observed

(1) the 1965 reef, about 0.8 m ALT (Taylor *et al.* in press);
(2) a reef, about 3 m ALT (e.g. Npenanavet, Wornambek, southern bank of Malua bay);
(3) a main terrace, the seaward edge of which is 7.5 m ALT at Npenanavet, 8 m at Malua bay, perhaps 6 m at Wornambek (not obvious in this locality); this terrace rises gently inland up to 12 m at Npenanavet, 11 m at Wornambek and Malua bay;
(4) the remains of a terrace (or huge tumbled blocks from the next high terrace), 23 m ALT near Wornambek.

On the northeastern coast, between Norsup and Lakatoro, a 2–3 m ALT terrace is formed mainly of lagoonal calcarenites. It is bounded landwards by an ancient sea cliff about 10 m ALT corresponding to the outer edge of a higher terrace. At Walarano mission (north of Norsup), two terraces occur at 8 and 30 m ALT.

Thirteen dates have been obtained thus far (see Table 3) for the low platform of north Malekula (they do not include the last recent uplifts: for these, see Taylor *et al.* in press). In West bay area, samples M-A-6, 7 and 8 (Brenwe) represent the lower part of the Holocene terrace from 2.5 m to 4 m ALT seaward of, and within, a sea cliff that reaches 5 m ALT (Table 3b; Fig. 5). Nearby, at Unmet, a coral from the uppermost scarp of the 19.5 m ALT Holocene terrace (M-Y-2) is 5.6 ± 0.5 ka. Across north Malekula, at Npenanavet and Wornambek (sites M-H, M-L and M-M), corals from 3 to 11 m ALT give ages ranging from 2.5 to 9.5 ka.

Maximum uplift rates on the basis of these dates are 4.3 mm/y at M-A, 3.6 mm/y at M-Y, and 2.9 mm/y at M-H, M-L and M-M. The date of 9.5 ka (M-M-1) is considered suspect because it disagrees with the age of M-M-2 and M-L-1 and 2 on corals from a few meters away. Possibly, this older age is from a coral that was transported from a presently submerged part of the Holocene reef as sea level rose from about −30 m to about present sea level between 9.5 ka and 5.8 ka.

Sample M-X-2 is from 1 m ALT at Crab bay on the east coast of Malekula and gives an age of 2970 ± 200 y b.p. This coral was flat-topped and presumably had grown to about low tide level when it died. It indicates a low uplift rate of 0.3 mm/y.

The high reef terraces

The high reef terraces of north Malekula have been described by Mitchell (1969, 1971). They consist of a flight of six major reef-capped terraces, with up to four additional terraces locally (Fig. 6). These terraces extend well to the north and around the northeast coast in the case of the terrace No. 3 of Mitchell. They are untraceable 10 km south of Tenmaru, where they are intersected by a series of northwest-trending faults (Elephant point area). The terraces can be seen again in West bay area and southwards further than Lambubu bay. The terraces on north Malekula are very consistently tilted to the northeast, with an axis of maximum uplift trending northwest from Tenmaru to Brenwe. South from a tilt discontinuity across central Malekula (which runs from West bay to Norsup area), the terraces on south Malekula tilt slightly to the southeast (Taylor *et al.* in press) (Fig. 6).

Altimetry of the reef terraces from two localities above Npenanavet and Malua bay (north Malekula) is presented in Table 4, with a third traverse previously reported by Mitchell (1969) near Malua bay. The N-S component of tilting is obvious when terrace altitudes between the two new traverses are compared.

The first major terrace is at 37 m ALT at Malua bay (site M-P). A minor, lower terrace occurs on its seaward edge above Npenanavet. Here, the crest of the high cliff above the coastal platform is 57 m ALT, but a few tens of meters landwards is a second small sea cliff, 68 m ALT, forming the outer edge of a quite wide flat terrace (site M-G). Air photograph interpretation indicates that the 37 m terrace at Malua bay and the upper terrace at Npenanavet (68 m ALT) correlate. The 57 m terrace at Npenanavet occurs for only a short interval along the coast.

Several age determinations on corals from this first high terrace were obtained at these two localities:

(1) A coral sample from the 37 m terrace at Malua bay (M-P-3) is dated at 60 ± 4 ka.
(2) Two corals from the upper of the two terraces (68 m ALT) at Npenanavet (M-G-3 and 4) are dated at 55 ± 4 and 57 ± 2 ka respectively.
(3) A coral from the minor terrace on the seaward edge of the wide terrace at Npenanavet (M-1-3) gave an age of 47 ± 2 ka.

Although the ages of 55 and 57 ka are young compared to the 60 ka paleosea level as inferred in the Huon peninsula (Bloom *et al.* 1974), and although the morphology of this high reef terrace appears to be unusually well developed for a 60 ka paleosea level, the two dates seem relatively consistent with one another and strongly suggest that this first high terrace corresponds to the 60 ka paleosea level.

The 60 ka paleosea level is estimated to have been at 28 km below present sea level (Bloom *et al.* 1974). Total uplifts of 65 m above Malua bay and 96 m above Npenanavet are inferred, yielding uplift rates of respectively 1.1 and 1.6 mm/y, which are significantly lower than the Holocene uplift rate of 2.7 mm/y, calculated from coral samples collected nearby at localities M-H, M-L and M-M.

An age determination comes from the locality of Walarano, on the northeastern coast of Malekula: a coral (M-AD-4) sampled in the 30 m high terrace gives an age of 72.5 ± 5 ka. By assuming the terrace corresponds to the 82 ka paleosea level, this leads to an uplift rate of 0.6 mm/y. Such a low uplift rate on the eastern coast of Malekula is in agreement with the general tilting of the northern part of the island towards the northeast.

South from the tilt discontinuity across central Malekula, two dates have been obtained on high terraces:

(1) A coral from the terraces above Lambubu bay (M-AE-1) gives an age of 75 ± 4 ka for a terrace altitude of 60 m. The terrace is

TABLE 4
High-terrace altitudes measured on north Malekula

Terrace Nos (Mitchell 1971)	Mitchell's Traverse (1969) (m)	Malua Bay Traverse (m)	Npenanavet Traverse (m)
6	240	223-240	353
5	215	204	(289)*
4	180	160	231
3	120	122	173
2a	..	(89)*	154
2	75	71	(113)*
1	45	37	57, 68

* Altitudes in parentheses mark a less distinct terrace, usually corresponding to a break in slope

TABLE 5
Comparison of the Tomebou terraces with Huon paleosea-level changes
(assumption, steady uplift rate of 2 mm/y)

Paleosea-Level Ages (ka)	Calculated Uplift (m)	Position of Paleosea Levels from Present Sea Level* (m)	Terrace Heights Calculated (m)	Terrace Heights Measured (m)
28	+56	−41	+15	+14
40	80	−38	42	41
60	120	−28	92	(64), 90
82	164	−13	151	154
103	206	−15	191	170 (to 190)
125	250	+6	256	240

* As determined in Huon peninsula (Bloom *et al.* 1974)

assimilated as in the case of Walarano to the 82 ka paleosea level. An uplift rate of 0.9 mm/y is inferred.

(2) A coral (M-AA-2) from Tisman bay locality (southeast Malekula) gives an age of 98 ± 11 ka for a terrace altitude of 30 m. The terrace is assumed to correspond to the 103 ka paleosea level. An uplift rate of 0.45 mm/y is inferred.

DISCUSSION AND INTERPRETATION

Santo Island

When plotted on the map of Santo, the uplift rates based on the high terraces agree somewhat with the Holocene data. The Kere river uplift rate, particularly, if confirmed, is quite consistent with the presence of a zone of maximum uplift coinciding with the high spine running the length of western Santo (Fig. 4).

Nevertheless, the 2 mm/y uplift rate in Tomebou is lower than the 3.6 mm/y uplift rate of site S-A (very close to Tomebou). This may mean that the uplift rate in this area has increased in the Holocene epoch. Both uplift rates are based on only two samples each. Additional age determinations would be useful to confirm the difference in late Pleistocene and Holocene uplift rates in this area. It is possible that the reef terrace from which corals gave dates of 37 and 38 ka was not deposited during the maximum sea level near 40 ka, but, instead, during a pause during the fall of sea level after the 40 ka high sea stand.

It is interesting to compare the terrace flight of Tomebou with the paleosea levels changes as inferred from Huon peninsula for the last 140 ka (Bloom *et al.* 1974). A similar method was used by Konishi *et al.* (1970) for the Ryu-Kyu islands. The basis of the calculation rests on the assumption of a constant 2 mm/y uplift rate, which has been previously inferred for the supposed 40 ka terrace on Tomebou (see Table 5).

The correspondence between the calculated and measured heights of Tomebou appears to be consistent. A 64 m level is reported in Table 2 but was not reported by Mallick, 1970: it does not correspond to an obvious geomorphic terrace, but rather to a break in the steep slope.

This comparison is based on the assumption of a uniform uplift rate, which is not in full agreement with the former observation about the possible increase of the uplift rate in the Holocene. Yet, it can be suggested that the summit of the Tomebou records the 125 ka paleosea level and that, consequently, the 90, 154 and 170 m terraces represent the 60, 82 and 103 ka paleosea levels.

Two remarks must be made:

(1) In the case of the assumed 82 ka terrace, the observed measure, on the chosen traverse, 170 km, does not quite fit with the height as calculated, 190 m. But this terrace is effectively 190 m high on the northern edge of Tomebou.

(2) The 28 ka terrace, which is predicted to be at 15 m ALT in this sequence, presents an interesting theoretical problem. This altitude corresponds to much of the Holocene platform. Holocene coral reefs may have developed as a veneer upon a 28 ka old reef that was inundated as sea level rose in the Holocene.

Previously, it has been pointed out, from Table 2, that the terraces of Tomebou and Queiros peninsula, particularly the terraces of Avatao, are nearly accordant in altitude. Therefore, the terraces at Avatao can also be considered as consistent with the paleosea level history of the past 125 000 years and a constant uplift rate of 2 mm/y. If one compares the terrace heights as measured at Avatao and as calculated for a constant uplift rate of 2 mm/y, one can observe the correlation set out below.

TABLE 6
Comparison of the Queiros terraces (Avatao) with the Huon paleosea-level changes

Paleosea-Level Ages (ka)	Terrace Heights Calculated (see Table 5) (m)	Measured (m)
28	+15	+12
40	+42	+42
		(60)
60	92	92
		(121)
82	151	145
103	191	185
125	256	218

This observation, however, requires two qualifications:

(1) Two levels, 60 and 121 m high, are observed at Avatao, but do not seem to corres-

pond to any main paleosea level recognized in Huon peninsula. In fact, they are only minor geomorphic terraces at Avatao and they may record minor events. There is, however, a correspondence of the 60 m level of Avatao with the 64 m level of Tomebou. The two may correspond to one of the several terraces in the 40–50 ka time range on the Huon peninsula. (2) The highest terrace altitude, as measured at Avatao (218 m), seems rather low in comparison to its calculated altitude (256 m). But it must be emphasized that the measurement was made on the low eastern edge of the uppermost terrace of the Walraoul plateau. The 1:50 000 IGN map of Santo indicates that near Avatao the terrace is between 240 and 260 m. The eastern edge of Walraoul plateau is at about the correct altitude to have formed during the 125 ka plaeosea level.

It has been stated that the surface of the Walraoul plateau corresponds to the main surface of the other eastern limestone plateaux of Santo: all of these plateaux are therefore likely to be 125 000 years old.

Nevertheless, the calculations of Table 5 rest on the assumption of a constant uplift rate of 2 mm/y for the eastern edge of the Walraoul plateau. This is not consistent with the uplift rate of 3 mm/y inferred from the Holocene platform, at site S-N, very close to Avatao. Calculations for the terrace altitudes of Avatao were made with the assumption of a constant uplift rate of 3 mm/y instead of 2 mm/y: 3 mm/y give calculated terrace heights (43, 82, 152, 233, 294 and 381 m) which cannot be correlated with the measured terrace heights (Table 3). Thus, the uplift rate appears to have increased in the Queiros peninsula in the Holocene, as in the case of the Tomebou area.

North Malekula

Based on the assumption of constant uplift rates as inferred in north Malekula (1.1 mm/y at Malua bay, 1.6 mm/y at Npenanavet), the same calculations as on Santo can be made for the terraces of Malua bay and Npenanavet (Tables 7 and 8).

TABLE 7
Comparison of the Malua Bay terraces with Huon paleosea-level changes
(assumption, constant uplift rate of 1.1 mm/y)

Paleosea-Level Ages (ka)	Calculated Uplift (m)	Position of Paleosea Levels from Present Sea Level (m)	Terrace Heights Calculated (m)	Terrace Heights Measured (m)
28	+31	−41	−10	..
40	44	−38	+6	+3; 8-11
60	66	−28	38	37
82	90	−13	77	71
103	113	−15	98	89
125	137	+6	143	122

TABLE 8
Comparison of the Npenanavet terraces with Huon paleosea-level changes
(assumption, constant uplift rate of 1.6 mm/y)

Paleosea-Level Ages (ka)	Calculated Uplift (m)	Position of Paleosea Levels from Present Sea Level (m)	Terrace Heights Calculated (m)	Terrace Heights Measured (m)
28	+45	−41	+4	3; 7.5-12
40	64	−38	26	(23 at Wornambek)
60	96	−28	68	68
82	131	−13	118	113
103	165	−15	150	154
125	200	+6	206	173

As on Santo, there is a similarity between calculated and measured terraces heights in north Malekula. Thus, terraces No. 2, 2a and 3 (Mitchell's numbering, 1971; see Fig. 6) are assumed to correspond to the 82, 103 and 125 ka paleosea levels. Such a correlation between the terrace No. 3 and the 125 ka paleosea level is consistent with the great extent of this terrace on north Malekula (Fig. 6).

In both traverses, the presumed 125 ka terrace is rather low relative to its calculated position: without neglecting the possible effects of erosion, this may indicate that the uplift rate slightly increased during the last 100 000 years.

As in Santo, calculations show that the Holocene and 28 ka reef may interfere in north Malekula.

CONCLUSION

Uplifted Quaternary reef terraces on Santo and north Malekula give evidence of uplift continuing into the present on both islands. The main surface of the eastern limestone plateaux of Santo is assumed to correspond to the 125 000-year-old paleosea level, by comparing the terraces levels of Santo with the paleosea levels as estimated for the last 140 000 years in the Huon peninsula area (New Guinea). The emergence of a large part of eastern Santo thus appears to be quite recent. A similar conclusion is inferred for north Malekula.

The second clear implication of these early results is that the uplift rates for both Santo and Malekula have increased dramatically in the Holocene. Pre-Holocene paleosea level history is sufficiently well known that if our terrace ages from radiometric dates and inferences are correct our pre-Holocene uplift rates must also be correct. They appear to have been relatively constant until the Holocene. There are certainly errors in our assumed Holocene sea-level history, but the errors must be small compared to the amount of Holocene emergence which was measured, and our Holocene uplift rates must be essentially correct. They are significantly higher than the inferred late-Pleistocene uplift rates.

The cause for the increase in uplift rates, which, if confirmed by further studies, constitutes the main conclusion of this paper, may be a change in convergence rate between the Australian and Pacific plates, or a change in ruggedness of topography of the descending plate (particularly in regard to the d'Entrecasteaux ridge). The latter explanation seems, however, inadequate because the present convergence rate of 1.2 cm/y (Dubois *et al.* 1977) allows subduction of only 1.2 km of Australian plate in 10 000 years. The inferred average late-Pleistocene uplift rates seem constant enough through time for a long-term change in average uplift rates to be unlikely. Possibly, we have merely discovered a short-term increase in uplift rates that will not significantly affect the long-term average rates.

ACKNOWLEDGEMENTS

This research is part of a joint agreement between the Government of the New Hebrides, Cornell University (Ithaca, NY, USA) and the Office de la Recherche Scientifique et Technique Outre-Mer (Noumea, New Caledonia). It was supported by grants from NSF (EAR-77-13685 and EAR-78-15188) and ORSTOM, with the cooperation in field assistance from the New Hebrides Service des Mines. Professor John Wells of Cornell University kindly identified the coral samples for radiometric dating.

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A SEDIMENTARY BASIN IN THE CENTRAL NEW HEBRIDES ARC

J N CARNEY AND A MACFARLANE

ABSTRACT

A sedimentary basin with water depths of some 3000 m lies between the Western Belt and Eastern Belt islands of the central New Hebrides Arc. Extrapolation of the onshore geology indicates that the western margin of this 'Central Basin' is underlain by faulted and flexured island-arc volcanics and derived sediments, Upper Oligocene to end-Middle Miocene in age. These are capped by Upper Miocene to Plio-Pleistocene pelagites and hemipelagites, mainly foraminiferal mudstones and turbidite-calcarenites, over 2000 m thick. The eastern margin of the basin comprises a thick sequence of Middle Miocene volcaniclastics overlain by Upper Miocene to Plio-Pleistocene pelagites with an interbedded, locally voluminous, volcanic sequence Mio-Pliocene in age.

The basin axis was in existence during the Pliocene, but the main subsidence dates from a Plio-Pleistocene taphrogeny during which complementary uplift of the margins led to the raising of the Western Belt and Eastern Belt islands above sea level. Prior to this, basinal development along the western margin had taken place during an unrelated Middle Miocene deformational event and had formed graben structures, now buried beneath the later Central Basin deposits.

Structurally the eastern margin of the Central Basin is the more complex and shows evidence of stepfaulting and flexuring through Late Pliocene to Recent times.

INTRODUCTION

The New Hebrides archipelago is an active island arc underlain by an easterly dipping Benioff Zone and adjoined to the east by a marginal sea, the North Fiji Basin. It has been subdivided into three provinces of island-arc volcanics, namely a *Western Belt* of Upper Oligocene to Middle Miocene age, an *Eastern Belt* of Mio-Pliocene age, and a Pliocene to Recent volcanic line known as the *Central Chain* (Mitchell and Warden 1971). These provinces are well exposed in the central latitudes (Fig. 1) and are probably continuous along the length of the arc both to the north and south.

The three volcanic episodes are a reflection of the arc's complex history, which has involved a reversal of subduction from west- to east-facing in post-Middle Miocene times (e.g. Mitchell, in Colley 1970). Deformation prior to or accompanying reversal is shown by an initial faulting and flexuring of Middle Miocene strata in the Western Belt; subsequent tectonism involved block faulting and took place along the Eastern Belt during and after the Mio-Pliocene volcanism and in the backarc zone and central latitudes from the Lower Pleistocene. As a result, many of the presently emergent parts of the arc are fault bounded, with narrow shelfal areas and intervening deep-water basins.

The distribution of the major sedimentary basins has been summarised by Ravenne *et al.* (1976), who delineated a zone of narrow, backarc rifts (first described by Karig and Mammerickx 1972) along the eastern margin, between the latitudes of Efate and Tanna in the south and of Gaua and the Torres Islands in the north (see Inset, Fig. 1). In addition they postulated a 'mid sedimentary basin' extending along the length of the arc and partly infilled by volcanic deposits from the Central Chain. In the central latitudes the backarc rifts are absent and the 'mid sedimentary basin' becomes a much deeper and wider feature, which they named the Central Basin.

This account describes the morphology of the Central Basin and traces its likely evolution in terms of the geology exposed on the islands along its eastern and western margins (Fig. 1). Extrapolation of the onshore geology into the offshore parts of the basin is based on seismic reflection profiles taken from Oil Company data held on open file at the Geological Survey and incorporated into the diagrammatic profiles of Fig. 2.

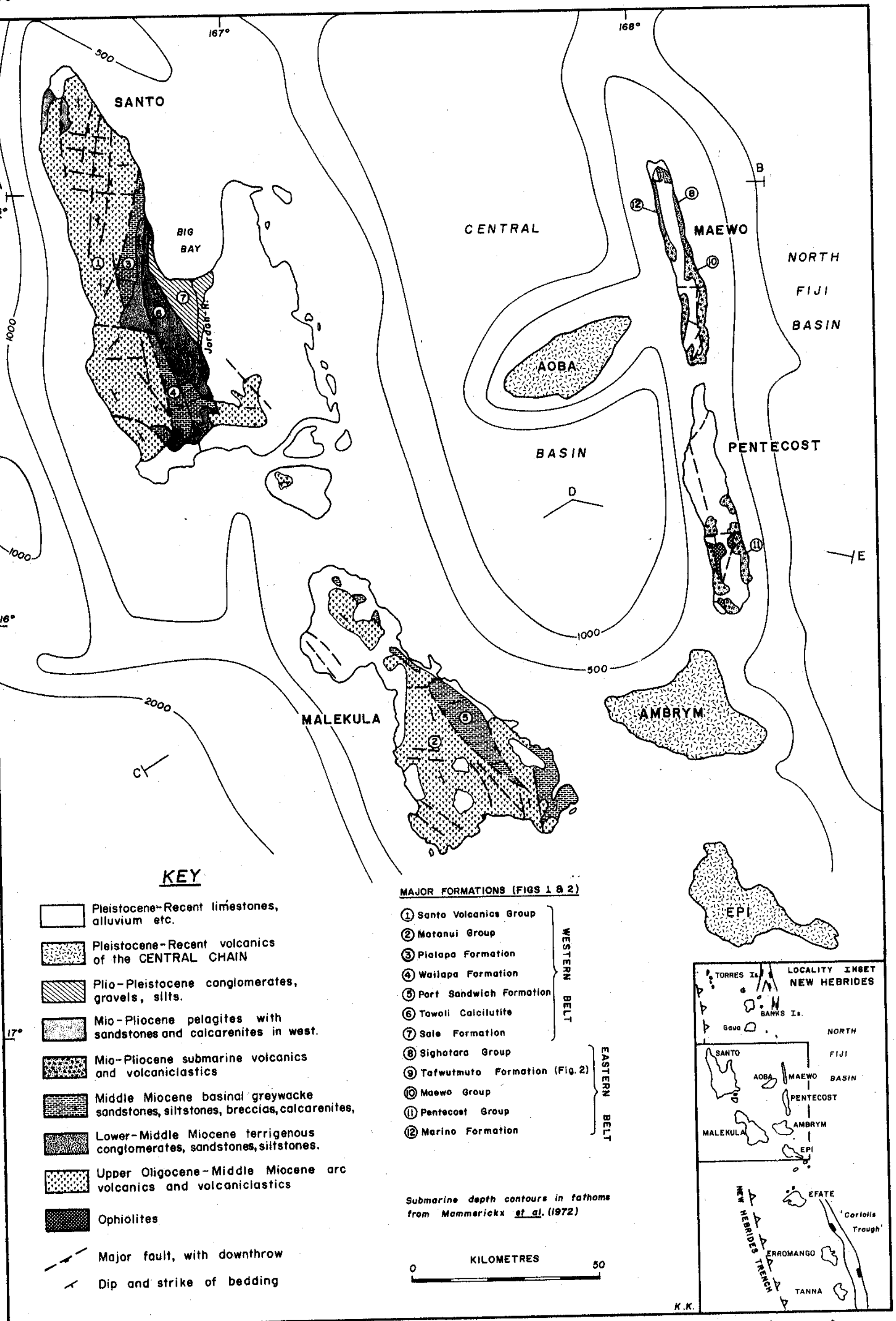


Figure 1. Geology and bathymetry of the Central New Hebrides Arc. Cross sections A-B and C-E are shown in Fig. 2. *Note:* Several formations referred to in the text but precluded in this figure, are too small to be illustrated at this scale.

GEOLOGY AND STRUCTURE OF THE CENTRAL BASIN

The Central Basin, located between the Eastern Belt and Western Belt islands (Fig. 1), is a trough, 200 km long, with an average width of 60–70 km. It is bounded to the north and south by the Central Chain active volcanoes of Gaua and Ambrym and is divided into two compartments by the northeast-trending shield volcano of Aoba, now in the solfataric stage (Warden 1970). The basin floor is smooth and flat-lying, with depths of *ca.* 2300 m and *ca.* 3000 m to the south and north of Aoba, respectively. Within the basin most of the Pliocene-Recent sedimentation has occurred along the present axis, whereas Middle Miocene accumulations took place in an older series of graben along the western flanks and belong to a separate taphrogenic episode (e.g. see Fig. 2).

Western margin — onshore geology

The main geological formations on the islands of Santo and Malekula are, respectively, the Santo Volcanics Group (Mallick and Greenbaum 1977) and the Matanui Group (Mitchell 1966, 1971); estimated minimum thicknesses are in the order of 4000–6000 m. They comprise early Neogene island-arc accumulations of coarse polymict or monomict volcanic breccias with interbedded tuffs, quartz-free greywacke sandstones, turbocarbonates and rare, massive or pillowed lavas; local intrusion by andesite stocks, pyritised and with potassic, phyllic and propylitic alteration is common. On southern Malekula 30 m-thick units of bedded carbonaceous silty mudstones, usually pyritic and with thin laminae of plant remains concentrated into lignite horizons, are also included within the succession (Mitchell 1966): (recent fieldwork by the writers in northern Santo has located similar sediments intercalated within volcanic flow-breccias). Abyssal red mudstones, of uncertain age, exposed along a fault zone on northwest Malekula, may form a floor to these arc deposits (Mitchell 1971); their position as shown on Profile II of Fig. 2 is, however, purely conjectural.

Faulting and flexuring are the dominant tectonic components in the arc volcanics. Strata dips are 20–40° both to the east and to the west, the latter direction of inclination being the more prevalent on Malekula (Mitchell 1971). Deformation was associated with a Middle Miocene taphrogenesis (Robinson 1969), during which grabens were formed on both Santo and Malekula. Subsequently these basins were infilled by rapidly deposited sediments belonging to the Port Sandwich Formation of Malekula (Mitchell 1966, 1971) and to the Pialapa and Wailapa formations of Santo: (the latter two formations have been included by Mallick and Greenbaum (1977) within the Santo Volcanics Group but have been differentiated here (Fig. 1) for descriptive purposes). On Santo these basinal sediments occupy downfaulted strips along the western margin of the Big Bay - Jordan River trough, while on Malekula they are preserved in graben along the island's eastern margin (Fig. 1). On both islands the basinal lithologies are volcanogenic but contain a lower proportion of breccias than the older island-arc deposits. They comprise quartz-free greywacke sands and silts, turbocarbonates, tuffs and volcanic breccias; degree of induration is generally greater, but diagenetic alteration rather less, than that of the earlier volcaniclastic successions. Carbonaceous and lignitic horizons are known in the basinal deposits from both islands, e.g. in south Malekula (Mitchell 1966).

Estimated thicknesses for the basinal lithologies on Santo are 2000–4000 m although faulting and flexuring may have introduced considerable repetition (Mallick and Greenbaum 1977); on Malekula Mitchell (1971) estimates a minimum thickness of 700 m. On Santo the upwards gradation of the Wailapa Formation into foraminiferal mudstones suggests cessation of volcanism in the late Middle Miocene (Mallick and Greenbaum 1977), whereas continued activity on Malekula into the Upper Miocene is demonstrated by a 10.7 m.y. age (Gorton 1974) for an uncorrelated breccia sequence exposed on the east coast.

Overlying the Middle Miocene successions of the western margin are pelagite and hemipelagite sequences of the Tawoli Calcilutite on Santo (Fig. 1 and Robinson 1969) and the Wintua and Malua formations

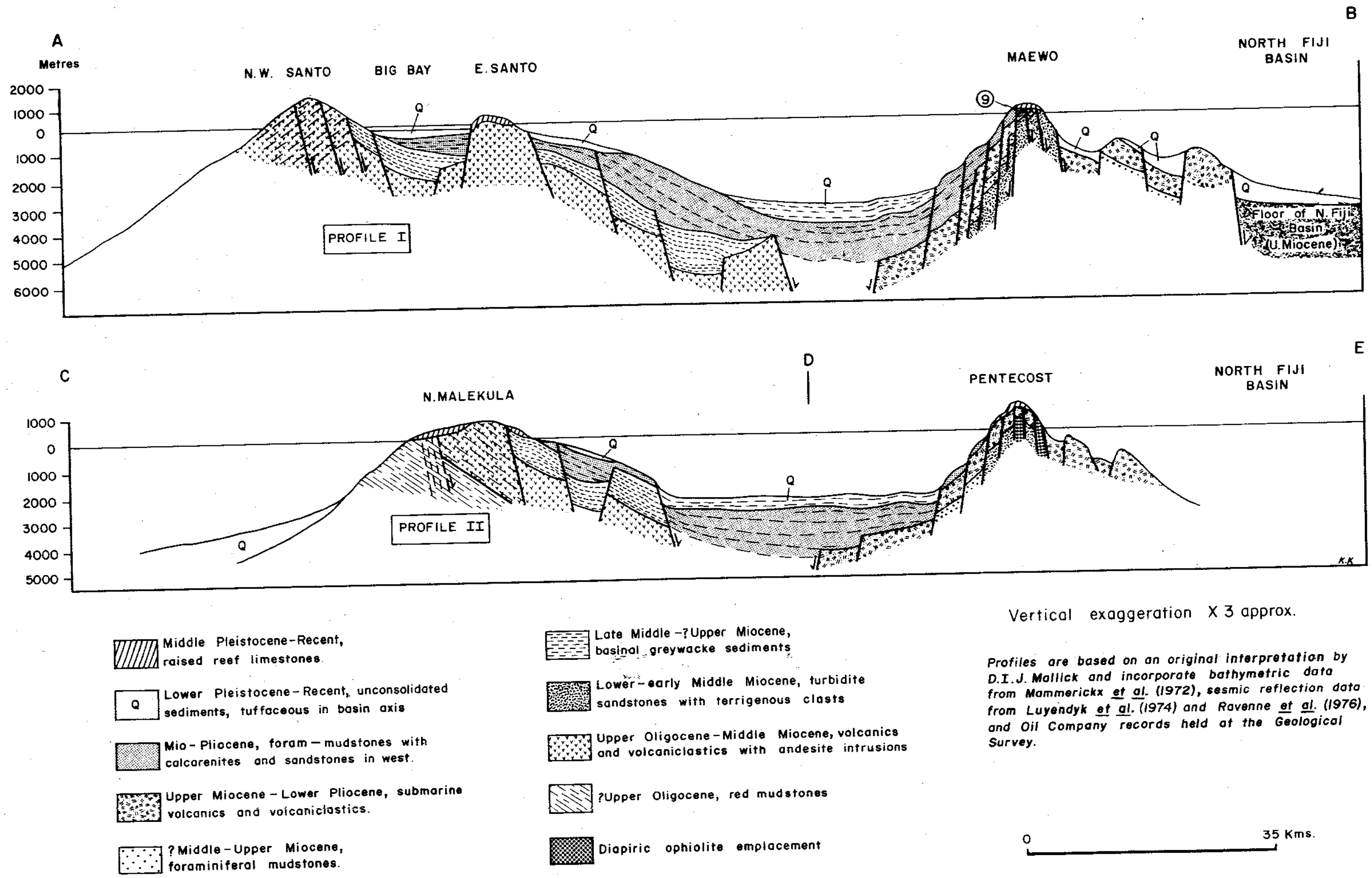


Figure 2. Sections across the Central Basin of the New Hebrides Arc. Location of cross sections A-B and C-E are shown on Fig. 1.

on Malekula (grouped by Mitchell (1966, 1971) within the raised limestones and not, therefore, separately defined on Fig. 1). On north Santo ages for the Tawoli Calcilutite are given as Lower Pliocene to Plio-Pleistocene by Barsdell (1976) and Taylor (1977). However, a recent identification of *Globorotalia plesiotumida* (Ruth Todd, pers. comm. 1979) from a previously unmapped locality places the lower limit of this formation in the Upper Miocene zone N17, or at 5–8.5 m.y.b.p. The junction of these pelagites with the Middle Miocene strata is markedly irregular and is consistent with a major erosional episode within the Upper Miocene. On south Santo the fine-grained components are included as a capping sequence to the Tawoli Calcarenite (Mallick and Greenbaum 1977). This formation (not differentiated from the Tawoli Calcilutite on Fig. 1) has a basal calcarenite component dated at Middle Miocene but in its upper part lacks microfauna appropriate to zones N16 and N17. By analogy, therefore, with relationships in north Santo the Pliocene pelagites constituting the younger strata of this formation may be unconformable upon the older calcarenites. This impression is strengthened by the overstepping relationship of the Tawoli Calcarenite onto the southern and northern flanks of other Middle Miocene formations in the south of the island (Fig. 1). Likewise on Malekula the Wintua and Malua formations rest with angular unconformity on the Lower to Middle Miocene beds and have provisionally been assigned a Pliocene age (Mitchell 1966, 1971).

In eastern north Santo the Tawoli Calcilutite consists mainly of foraminiferal mudstones with (at the base) local development of hard, forereefal calcarenite and calcirudite. By contrast, in the northwest the formation (Wounpouku Calcarenite of Robinson 1969) has an appreciable content of friable, well sorted quartz-free volcaniclastic sandstones with interbedded mudstone and calcarenite. The maximum thickness for the formation, *ca.* 1000 m, is found along the margins of the Big Bay - Jordan River trough (Fig. 1) but the estimate is complicated by faulting and slumping of strata (Mallick and Greenbaum 1977).

A latest Pliocene change in sedimentation on Santo is demonstrated at the southern end of Big Bay by the Sale Formation (Fig. 1) comprising fine- to medium-grained clastics with a basal 250 m-thick accumulation of shallow-water conglomerates. The age of this formation is mainly Plio-Pleistocene (Mallick and Greenbaum 1977).

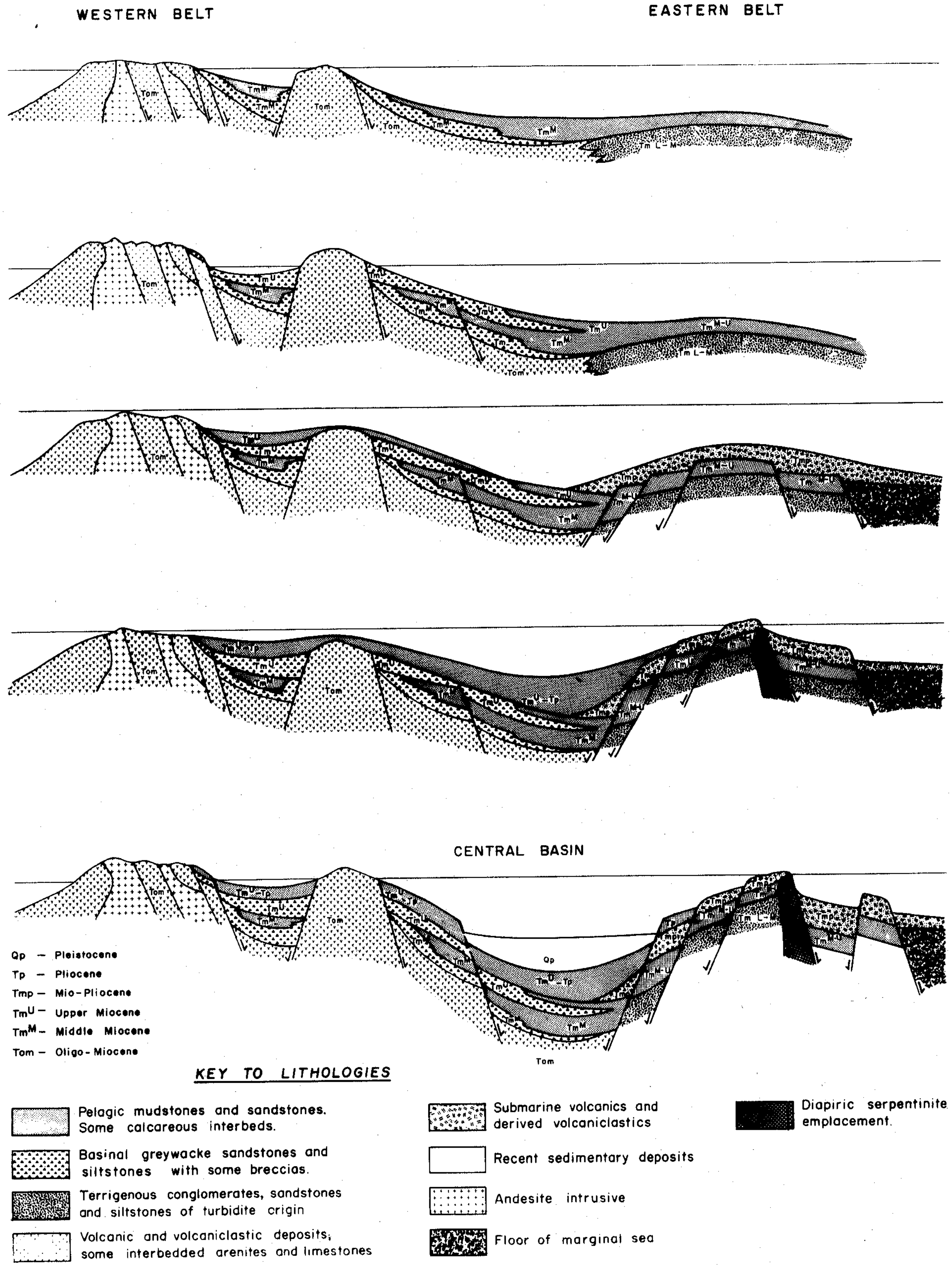
Overlying the older strata on both Santo and Malekula is an unconformable series of Pleistocene to Recent raised reef limestones. In the main these limestones are tilted at angles of 1–2° eastwards towards the Central Basin (e.g. Mitchell 1969; Robinson 1969).

Western margin — offshore geology

An interpretation of the structure and stratigraphy of the offshore slopes of the Central Basin's western flank is given in Fig. 2 and is based on seismic-reflection data and an extrapolation of the onshore geology. A 'lower' series of reflectors, more than 1000 m thick, unconformably overlies and infills a faulted topography to the east of Malekula (Profile II, Fig. 2) and is taken as the offshore continuation of the unconformable Upper Miocene to Pliocene pelagite/hemipelagite sequence. The 'basement' on which this unit rests is thought to be Middle Miocene strata, either the Matanui Group or the Port Sandwich Formation. A major fault, of ? post-Pliocene age, cuts the 'lower' sediment series in Profile II, whereas in Profile I (Fig. 2) the basinwards slope seems to be a fairly uniform downwarp.

Central Basin axis

The axial zone of the Central Basin contains a seismically transparent 'upper' sedimentary series, almost 1000 m thick, overlying the Pliocene or 'lower' reflectors at the margins with angular unconformity (e.g. Ravenne *et al.* 1976) and is interpreted as Pleistocene to Recent unconsolidated ashes derived from the active Central Chain volcanoes of Aoba, Ambrym and Gaua. Seismic penetration in the axis is 2.4 s within the 'lower' series, hence the 'basement' cannot be located precisely. Moreover, refraction data are not available at the Geological Survey to determine the structures in the faulted basement of the lowermost western slopes.


WESTERN BELT
EASTERN BELT
CENTRAL BASIN
Qp — Pleistocene
Tp — Pliocene
Tmp — Mio-Pliocene
TmU — Upper Miocene
TmM — Middle Miocene
Tom — Oligo-Miocene
KEY TO LITHOLOGIES
Pelagic mudstones and sandstones. Some calcareous interbeds.
Basinal greywacke sandstones and siltstones with some breccias.
Terrigenous conglomerates, sandstones and siltstones of turbidite origin
Volcanic and volcaniclastic deposits; some interbedded arenites and limestones
Submarine volcanics and derived volcaniclastics
Recent sedimentary deposits
Andesite intrusive
Floor of marginal sea
Diapiric serpentinite emplacement.

(A)
11–10 m.y. Quiescence of Western Belt arc volcanism. Followed by block faulting, intrusion and deposition of basinal sediments. Termination of terrigenous sedimentation in Eastern Belt.

(B)
10–7 m.y. Uplift and erosion of Western Belt, probably with further deepening of sedimentary graben. Continued pelagic sedimentation to east. Block faulting in Eastern belt at ca. 8 m.y associated with inception of North Fiji Basin (see following section C)

(C)
7–6 m.y. Onset of arc volcanism, faulting and uplift in Eastern Belt. Subsidence of Western Belt leading to pelagic sedimentation may have occurred within the period N17 or 8·5–5 m.y.

(D)
5–2 m.y. Continuation of pelagic sedimentation in Western Belt. In Eastern Belt, uprise of serpentinites, faulting and cessation of volcanism. Uplift to reef limestone zone by end - Pliocene

(E)
2–0 m.y. Uplift of Western Belt and Eastern Belt, block faulting and main collapse of the Central Basin axis and accumulation here of volcaniclastic sediments derived from Pleistocene Central Chain eruptions

Figure 3. Geological evolution of the Central Basin, New Hebrides Arc.

Eastern margin — onshore geology

The oldest exposed *in situ* rocks are a sequence of Lower to Middle Miocene volcanogenic, turbidite sandstones, siltstones, calcarenites and conglomerates, more than 600 m thick, known on Maewo as the Sighotara Group (Fig. 1) and on Pentecost as the Olambe Formation (not shown on Fig. 1). Structures in the Maewo sediments indicate a basinal flysch-type accumulation, and their large content of cobbles and blocks of low-potash, iron-enriched lavas with the composition of island-arc tholeite point to derivation from the frontal part of a former east-facing arc system (Carney and Macfarlane 1978). Conformably overlying the Sighotara Group on Maewo is the Tafwutmuto Formation, a 100–150 m thick sequence of foram-rich mudstones (Liggett 1967; Carney in press), whose youngest beds are of uppermost Miocene age.

Island-arc volcanism on Maewo began in the Upper Miocene at about 7 m.y.b.p. with extrusion of Maewo Group pillow lavas, which thin out northwards and eastwards onto a massif of older Miocene sediments. However, although there is slight angular discordance between the lavas and the sediments, the absence of erosional features and of a microfaunal hiatus at the contact suggests that the thinning-out was against west- and south-facing fault scarps which developed immediately prior to volcanic activity (Carney in press). On north Maewo the Tafwutmuto Formation and Maewo Group are replaced by a sequence of tuffs, mudstones and dark grey-blue marls; the marls are of euxinic facies and contain about 10–15% pyrite.

Further block faulting and graben formation on Maewo about 6 m.y.b.p. was followed by final effusions of palagonitic volcaniclastic breccias, distally resorted as tuffaceous horizons intercalated with mudstones in the north. On Pentecost the Mio-Pliocene volcanics of the Pentecost Group contain breccias bearing clasts of limestone and serpentinite; this suggests that by *ca.* 5 m.y.b.p. islands were then present and the 'Basement Complex' ophiolites, now exposed as horst blocks to the south, had been unroofed (Mallick and Neef 1974).

Cessation of arc volcanism by *ca.*

4–5 m.y.b.p. was reflected on north Maewo by deposition of the Marino Formation, a succession of foraminiferal pelagites over 100 m thick dated at 3–5 m.y.b.p. Increasing unconformity, with westwards tilting, preceded or accompanied deposition of the Nasawa Formation (not shown in Fig. 1), a thin Plio-Pleistocene foram- and pteropod-rich sequence, locally with reefal bioclastic intercalations (Carney, in press). Similar sediments on Pentecost were included as the basal part of an unconformable limestones sequence (Mallick and Neef 1974). Continued tilting, faulting and uplift into Recent times was associated with erosion and reefal growth on the older rocks and the elevation of these limestones above sea level; on Pentecost, faulting and anticlinal warping of the limestones were related to continued uplift of 'Basement Complex' ophiolite horst blocks (Mallick and Neef 1974), whilst on Maewo similar limestones were warped and tilted gently westwards (Carney, in press).

Eastern margin — offshore geology

The profiles of Fig. 2, compiled from both bathymetric and seismic reflection data, show that the Eastern Belt is a narrow, complex horst. Off Maewo (Profile I, Fig. 2) the seismic reflectors both dip steeply and are stepfaulted basinwards. They are correlated with the Pliocene mudstones of the Marino Formation, and are thought to continue westwards into the prominent 'lower' reflectors of the Central Basin axis which have been equated with the Upper Miocene to Pliocene Tawoli Calcilutite on Santo (see above). Profile II is compiled from a reflection traverse between Pentecost and Maewo, although the geological relationships are those observed in south Pentecost; thus the Pliocene mudstones are either missing or little developed until the basin floor is reached.

It is worth noting that on both profiles gentle flexuring of the inferred Pliocene and Pleistocene–Recent reflectors occurs near the structurally complex eastern margin and that the 'igneous basement', as defined by the basal part of the reflecting series, is older in the west (Middle Miocene) than in the east (Mio-Pliocene). Also noticeable on Profile I is the contrast between the smooth slope towards the floor of the North Fiji basin and the adjoining rugged submerged margin of the Eastern Belt.

EVOLUTION OF THE CENTRAL BASIN

In the sequential diagrams of Fig. 3 major volcano-tectonic events are diagrammatically portrayed, and an attempt is made to relate them to sedimentation. A study of the unconformities in this part of the arc shows that block faulting and basinal sedimentation in Middle Miocene times was restricted to the western margin and brought about the early development of the Big Bay - Jordan River trough on Santo, and of narrow graben on Malekula (Fig. 3A). This taphrogenesis was accompanied or followed by cessation of the Western Belt volcanism — presumably consequent upon blocking of subduction at the former trench, located at that time to the east of the New Hebrides rather than to the west as at present (e.g. Gill and Gorton 1973; Falvey 1975). Reversal of the arc's polarity may also have commenced during this episode.

Uplift and erosion of a frontal 'Vitiaz arc' has been postulated to explain the appearance of island-arc tholeiite clasts in contemporaneous sediments of the Sighotara Group on Maewo (Carney and Macfarlane 1978), and a later subsidence of this source area may have led to deposition of thin and slowly accumulating pelagites of the Tafwutmuto Formation. The subsequent sedimentary hiatus on the Western Belt in the Upper Miocene is ascribed to uplift and erosion following the Middle Miocene taphrogeny; however, clastic detritus from this episode (Fig. 3B) may have been deposited offshore in the Big Bay area, and along the western margin of the Central Basin. The Eastern Belt remained submerged, with pelagite sedimentation continuing through to the uppermost Miocene.

The first unconformities signalling formation of what is now the eastern margin of the Central Basin date to the Eastern Belt Mio-Pliocene volcanism (Fig. 3C). In terms of arc development the eruptions are thought to reflect thermal activation of the new eastwards-descending lithosphere following its inception during the Middle Miocene deformation of the Western Belt. The major faulting that preceded, and caused thinning of, the Eastern Belt lavas on Maewo may also be attributable to rifting movements along the arc's margin associated with the initial expansion of the North Fiji Basin; a date of 8 m.y.b.p. for the opening of this basin is suggested by the recognition of Anomaly 4 to the east of Maewo/Pentecost (A Malahoff, pers. comm.). Further block uplift evidently continued until volcanism had ceased about 4–5 m.y.b.p. (Carney, in press). This Upper Miocene episode of arc volcanism and marginal-basin inception apparently correlates with a subsidence of the western margin which led to pelagic and hemipelagic sedimentation on Malekula and Santo (Tawoli Calcilutite). These sediments, together with their equivalents in the Eastern Belt (Marino Formation), are thought to thicken towards the axis of the Central Basin (Fig. 3D), although this feature was probably not well defined at that stage.

The main development of the Central Basin dates from the Plio-Pleistocene on the basis of unconformities and sedimentary-facies changes along the western and eastern margins (Fig. 3E). On Santo, pelagic sedimentation ceased and the Big Bay - Jordan River trough became infilled with terrigenous deposits of the Sale Formation; the associated 250 m-thick shallow-water conglomerates suggest contemporaneous subsidence along this trough (Mallick and Greenbaum 1977), interpreted here as due to the rejuvenation of fault lines first active in Middle Miocene times. Major fault movements at this stage, however, were operative mainly along the submerged central axis of the basin. Basinward dips in the Pliocene 'lower' reflectors (Fig. 2) suggest that block uplift of the western and eastern margins, as shown by the onshore development of Pleistocene to Recent raised limestone terraces, was accompanied by a corresponding 'lag' or downwarping of the axial zone.

Whereas on the western margin basinal subsidence was mainly controlled by monoclinal flexuring and simple downfaulting, the eastern margin is marked by complex stepfaulting, and some 'crumpling' of strata has also taken place (Fig. 2). Young sedimentary infills overlie the Pliocene reflectors with angular conformity in the axial part of the basin, but towards the margins the relationship, owing to faulting and flexuring there, becomes an increasingly unconformable one (Ravenne *et al.* 1976). Warping of these younger deposits (Fig. 2) and the presence of Recent fault scarps displacing the raised limestones of both the Western Belt and Eastern Belt islands, provide complementary evidence that the Central Basin is still developing.

ORIGIN OF THE CENTRAL BASIN

The first accounts of the tectonic framework of the New Hebrides by Karig and Mammerickx (1972) placed the Central Basin as part of a system of extensional rifts located mainly to the rear of the arc (Inset, Fig. 1). This broad correlation is supported by the regional geology, which shows that initial subsidence of the axial zone of the Central Basin in Plio-Pleistocene times (as inferred above from unconformities along both its margins) was synchronous with early development of the backarc rifts. There are, however, important morphological differences between the backarc rifts and the Central Basin; as pointed out by Karig and Mammerickx (1972) the Central Basin has a considerably greater sediment infill, although the bulk of this probably pre-dates the most recent phase of subsidence (see above). Luyendyk *et al.* (1974) noted that the Central Basin lacks the steep walls and outward external slopes of the backarc rifts and is more in the form of a downflexure or 'collapse' feature. The absence of a central magnetic anomaly in the Central Basin further distinguishes it from the backarc rifts (e.g. Luyendyk *et al.* 1974; Dubois *et al.* 1975), which are considered as embryonic interarc basins (Karig and Mammerickx 1972).

The unique morphology of the Central Basin cannot be considered in isolation from other fundamental differences between the central sector of the New Hebrides archipelago and the northern and southern segments of this island arc. Perhaps the most notable of these differences is the absence of both the trench and the backarc rifts in the central sector between the islands of Efate and Gaua (Inset, Fig. 1). This morphological break coincides with the presence on the Australasian plate, to the west of Santo and Malekula, of an aseismic ridge of thickened crust known as the D'Entrecasteaux Fracture Zone. Deformation associated with the subduction of this feature at the New Hebrides Trench has been linked with the formation of the Central Basin either as a rift caused by collapse of an upwarped crust (Luyendyk *et al.* 1974) or as a synclinal downflexure of the crust which behaved essentially as a thin, elastic plate (Chung and Kanamori 1978). The writers prefer Chung and Kanamori's interpretation of the Central Basin with the proviso that there may be a greater crustal heterogeneity, due to differential uplift of fault-bounded blocks, than they envisage; in addition their model does not adequately explain the extreme elevation of the Eastern Belt nor the apparent relationship of this to the Pliocene-Recent uplift of ophiolite horsts on south Pentecost (e.g. Mallick and Neef 1974). Profiles demonstrate that the arc is slightly uparched, in its northern and southern sectors, and narrow backarc rifts are developed on, or just to the rear of the 'anticlinal' crest (Dubois *et al.* 1975).

While these differences in deformational style along the New Hebrides arc can be attributed to local complexities imposed by subduction of the thickened crust in the central sector, the synchronous formation of the backarc rifts and the Central Basin must also imply underlying stress-regime of regional importance, operative since the Plio-Pleistocene. In this wider context there are, for example, marked similarities between the Central Basin and the Central Solomons Trough (Katz, this volume), namely in structural style, Plio-Pleistocene age of latest development and position within the tectonic framework of the respective arcs. If the comparison is valid then the factors responsible for the formation of collapsed basins, and of the related backarc rifts along some 2000 km of the Pacific/Australasian plate boundary, may lie in kinematic changes in geologically young plates in this region, as proposed by Chase (1971) and Falvey (1978). Confirmation of this may be sought in the change in orientation of the spreading centres in the North Fiji Basin at the time of Anomaly 2, *ca.* 1.8 m.y.b.p. (A Malahoff, pers. comm.).

HYDROCARBON POTENTIAL

The hydrocarbon potential of a sedimentary succession is generally evaluated in terms of parameters that include source rock and reservoir potential, thermal regime and structure. In these respects the Upper Oligocene–Middle Miocene successions along the western margin of the Central Basin and their continuation north to the Torres Islands and south towards Tanna offer some hope for hydrocarbon potential in that appreciable oil and gas fields have recently been found in island-arc rocks of similar lithology and early Neogene age (e.g. Miyazaki *et al.* this volume). No extensive source rock evaluations have so far been made, however, although four of 30 surface samples analysed from the Middle Miocene and Plio-Pleistocene successions of Santo, bore gas-type generating kerogens and contained organic matter in the 'Early – Early Peak' or 'Early Peak – Peak' generation states (Mallick and Greenbaum 1977).

Furthermore, the pyritiferous carbonaceous sandstones described from Malekula by Mitchell (1966) and also known on Santo attest to the presence of euxinic facies and locally to appreciable amounts of organic material trapped within rapidly deposited greywacke-type sediments. Induration is generally high, however, and thus adverse to hydrocarbon migration.

Structural and stratigraphic conditions along the western margin of the Central Basin could be favourable for hydrocarbons in that the Middle Miocene successions show evidence of contemporaneous block-faulting, with local basinal thickening in the graben so formed. The Upper Miocene erosional unconformity on Santo is adverse to hydrocrbon entrapment there, but in the offshore areas east of Santo and Malekula, and in the northern continuation to the Big Bay - Jordan River graben, sedimentation may have been more continuous, with Upper Miocene clastics perhaps forming prograding wedges enclosed by pelagic sediments, and the whole series overlain by the pelagite 'lower' reflectors of the Central Basin (see Fig. 2). These sediments are mainly impermeable and therefore good cap rocks; reservoir potential may also exist in the extreme northwest of Santo where there is a facies change to friable, well sorted, medium-grained volcaniclastic sandstones and calcarenites (the Wounpouko Calcarenite of Robinson 1969).

The eastern margin of the Central Basin has less obvious hydrocarbon potential; it contains a thick, basal Middle Miocene sedimentary sequence, but this appears to be in the form of a rapidly deposited, flysch-type apron with very little included carbonaceous material. The Upper Miocene mudstones overlying these sediments are very fossiliferous, with 40–90% forams (Carney, in press) although they evidently accumulated very slowly, are only 100–150 m thick and thus may not represent good source rock. The Mio-Pliocene volcanics are largely submarine and may be ruled out as good source rock; in northern Maewo they are replaced by a tuff/mudstone sequence which locally contains pyritiferous marls of euxinic facies. However, a variety of structural traps is present along the eastern margin of the basin, where sediments and volcanics have been stepfaulted and flexured (Fig. 2).

The axis of the Central Basin, as a zone of extreme subsidence, may have developed only since Plio-Pleistocene times (Fig. 3) and is thus not regarded as an area of important early Neogene sedimentation. However, Lower to Middle Miocene sediments thought to extend beneath the basin must have been subjected to two phases of heating related to arc volcanism — one during the Mio-Pliocene (Eastern Belt) and one during the Pleistocene (Central Chain). If this has enhanced maturation of enclosed hydrocarbons the area may hold promise as an oil or gas generator; migration paths would be affected by thermal effects and by pressure differentials prevailing during the Lower Pleistocene to Recent downwarping and faulting event.

ACKNOWLEDGMENTS

We are grateful to Dr H R Katz for helpful criticism of an early draft of this paper.

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THE STRUCTURE OF THE YASAWA ISLANDS AND ITS PLATE TECTONIC SIGNIFICANCE

B L WOOD

ABSTRACT

Two major units, Loto Basaltic Group (probably Lower Miocene) and Nalauwaki Andesitic Group (probably Lower Pliocene) occupy a large syncline on Waya Island trending NNE for at least 50 km and are present also on other islands of the Yasawa and Mamanutha groups. Nalauwaki sediments 1.3 km thick represent subaerial proximal and marine distal fan deposits emplaced during and after andesitic eruptions in an elongate depocentre which rapidly opened by at least 5 km and subsequently closed at least 2 km. They are correlated with movements on a 200 km epicontinental transform fault (Yasawa Zone) resulting from changes in adjacent plate trajectories. The changes resulted from translation (or jump) of the spreading ridge in the adjacent North Fiji Basin following arc-arc collision between the Fiji and Lau islands in the Late Miocene, and included anticlockwise rotation of the Fiji Group and opening of the Lau Basin by seafloor spreading.

INTRODUCTION

The Melanesian region from the Tonga Group to the New Hebrides has been the subject of several tectonic studies and reconstructions in recent years (Chase 1971, Packham 1973, Falvey 1975, Hawkins 1976, Katz 1976, Weissel 1977, Lawrence and Wood (in press)). Many aspects still remain obscure or uncertain, and for many large areas where there are few or no islands critical data have yet to be obtained by marine or airborne surveys.

The present paper describes the local geology of part of the outlying Yasawa Islands in the western part of the Fiji group, and suggests a possible plate-tectonic relationship with the nearby North Fiji Basin. This work is part of a continuing program of regional structural study in the west Fiji region, for the purposes of obtaining significant data critical to certain of the reconstructions cited above and of contributing to the regional geological information of the area. Field work carried out in October, 1978 was greatly facilitated by prior regional mapping by Mr P Rodda of the Mineral Resources Department, Suva; approval to use this unpublished information was kindly granted by the Director of Mineral Development, Mr R Richmond.

GENERAL GEOLOGY

The Yasawa Islands form a rather regularly spaced linear chain trending 045°, near to but not along the western edge of the Fiji platform (Fig. 1). Further to the south in the Mamanutha Group the linear trend curves south and southeast to 150°. A few of the Mamanutha Islands include extensions of the Yasawa structural trends, others are small volcanic or sand islands. To the north the Yasawa Islands decrease successively in relief and the most northerly is dominated by sand and coral with limited volcanic outcrops. The linear trend of the group terminates northward at Round Island Passage, but the structural trend is continued by the outer part of the Passage, with its small presumably volcanic seamount at about 1 km depth, and by the platform edge beyond (Fig. 1).

The strike of folded rocks in most islands of the Yasawa group parallels the general linear trend of 045°. However, at Waya Island, the most southerly, the local structural trend is significantly different at 010°. Southwest from Waya Island bathymetric expression of the 045° trend is apparent in the deeper part of the shelf slope. There, narrow well defined ridges are separated by a downslope groove or trough, which merges beyond the base of the slope into

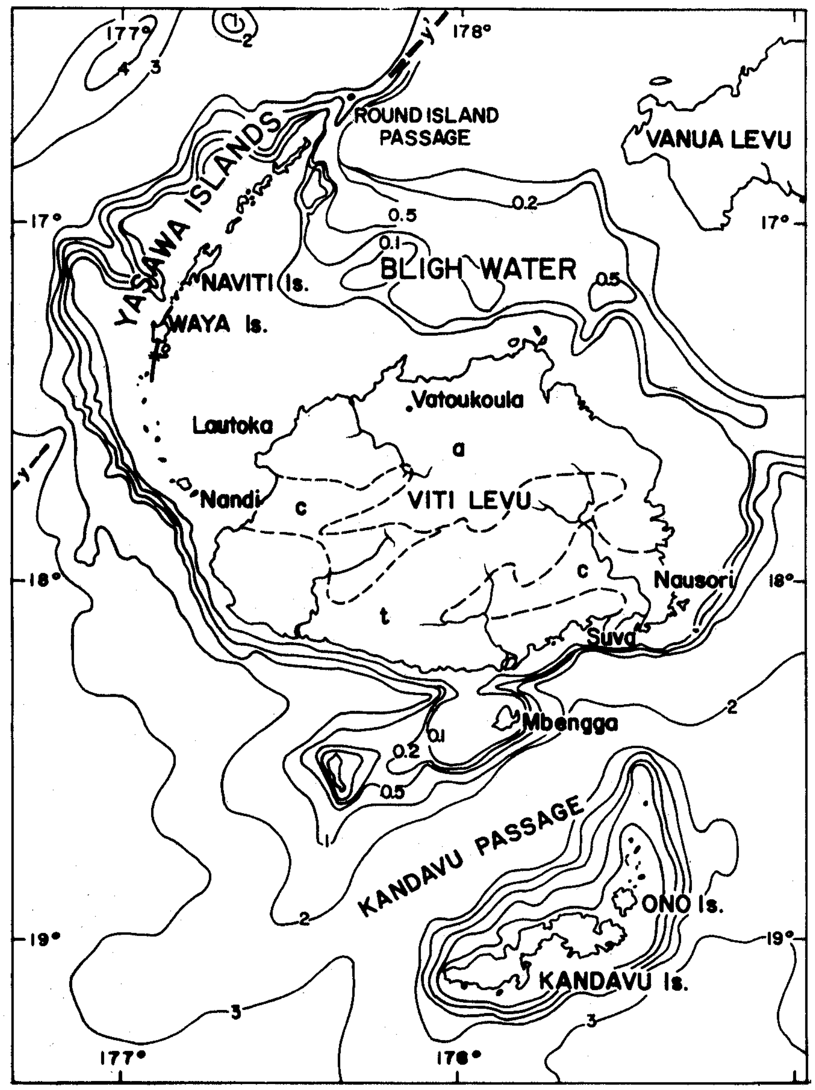

Figure 1. Islands of the Fiji group, and surrounding bathymetry (depths in km). The position and trend of the Waya Syncline is shown; the trend of the inferred Yasawa Zone is indicated between y-y'. On Viti Levu, t = tholeiitic rocks, c = calc-alkalic rocks, a = alkalic rocks, after Gill 1970. Bathymetry from World Map 1:2 500 000 Sheet 162, Administration of Cartography and Geodesy, Moscow, 1968: modifications in detail near Yasawa Islands from unpublished bathymetric compilations by the Mineral Resources Department on open file in CCOP/SOPAC Project Office, Suva, included by permission of the Director, MRD.

a +3000 m depression in the sea floor (Fig. 1).

Waya Island thus appears to lie near a critical intersection of a local structural trend (010°) and a regional trend (045°) of a larger entity herein referred to as the Yasawa Zone. The island clearly reveals details of the local structure, in contrast to others further north in which younger eruptives and sediments obscure large parts.

WAYA ISLAND

The Island and its smaller neighbour Wayasewa are hilly and steep (maximum height 550 m) with prominent crags and strike ridges of steeply dipping well bedded strata. Two major lithostratigraphic units are recognised, and parts of both are exposed in the east and west limbs of a large syncline — the Waya Syncline — that occupies two-thirds of the island (Fig. 2).

In addition small intrusive plugs of dacitic composition (map symbol D) are present west of Wayalevu village and on the east side of Wayasewa Island. Their affinities are uncertain but they are provisionally grouped as late members of the Loto Group.

The Loto Group is well exposed along the east coast and in the nearby hills, with a total thickness of approximately 800 m. On the west at Loloto Point, black basaltic breccia is gradationally intermixed with Nalauwaki andesitic breccia and is associated with a few flows of basalt. The lower undifferentiated sequence some 500 m thick is exposed in coastal sections at the east side of Yalombi Bay and at Vatukavika Point. Lava flows have well developed pahoehoe surfaces and are intercalated with red oxidised interflow and carapace breccias, all suggestive of subaerial conditions. Minor tuff-breccia is present and appears to be mainly a primary airfall deposit, with little or no resorting.

The dominant rock type is dark clinkery flow breccia and debris. The main flow rock is a dark porphyritic augite basalt commonly with plagioclase phenocrysts and well crystallised fluxional or trachytic groundmass. Alteration effects are common, many phenocrysts are embayed, and clouded, and some ferro-magnesians are completely obscured and chloritised. Zeolites, laumontite and stilbite are common in the matrix, in amygdales and breccia vugs.

The Motukuro Pillow Basalts comprise the upper 300 m of the Loto Group, crop out to each side of Motukuro Point, and are best exposed at and northwest of Naikawakawa Point. Many flows with distinct pillows 0.7 m to 1 m in diameter alternate with beds of breccia and agglomerate and dykes of similar basalt. Interpillow fillings of spheroidal fibrous zeolite and of dense microfossiliferous calcite are common. The assemblage appears to represent shallow marine conditions. No ages can be inferred from the microfossils. The Nalauwaki Group (1.3 km thick) lies in fault contact on the east side with Loto Basalts, but on the southwest side is probably in transitional contact with several beds of black basaltic breccia and one flow at Loloto Point.

The west limb and centre of the Waya Syncline consist of Nathilau and Vatunaremba Breccia, and the subvertical east limb is mainly Koromasoli Sandstone. The laterally transitional relationship between them is shown in the geologic section of Fig. 2. Nathilau Breccia (500 m thick) includes a small proportion (10-15%) of rounded weathered basaltic pebbles and coarse xenolithic clasts of hornblende diorite, occasionally also embedded in fine-grained andesite. Vatunaremba Breccia-Conglomerate (800 m thick) lacks these varieties but includes more rounded clasts of andesite. Both units are dominated by pale yellow-grey hornblende andesite clasts up to 10 cm diameter with lesser dacitic pumiceous clasts. Even where clasts are very large, bedding up to 4 m thick is well developed, regular and planar, with a dense non-porous matrix of cream-grey altered andesitic sand. Finer sandy interbeds are regular, occasionally laminated, up to 1 m thick, indurated and non-porous. No fossils, coral or other carbonate remains could be found, although small isolated clasts of granular calcite are present in Nathilau Breccia. All rocks are pervasively altered and cemented with chlorite, laumontite and silica.

Koromasoli Sandstone (1000 m thick) occupies most of the east limb of the Waya Syncline in steeply dipping to overturned beds. No second order folds were seen, and, apart from a weak set of calcite-filled fractures, there

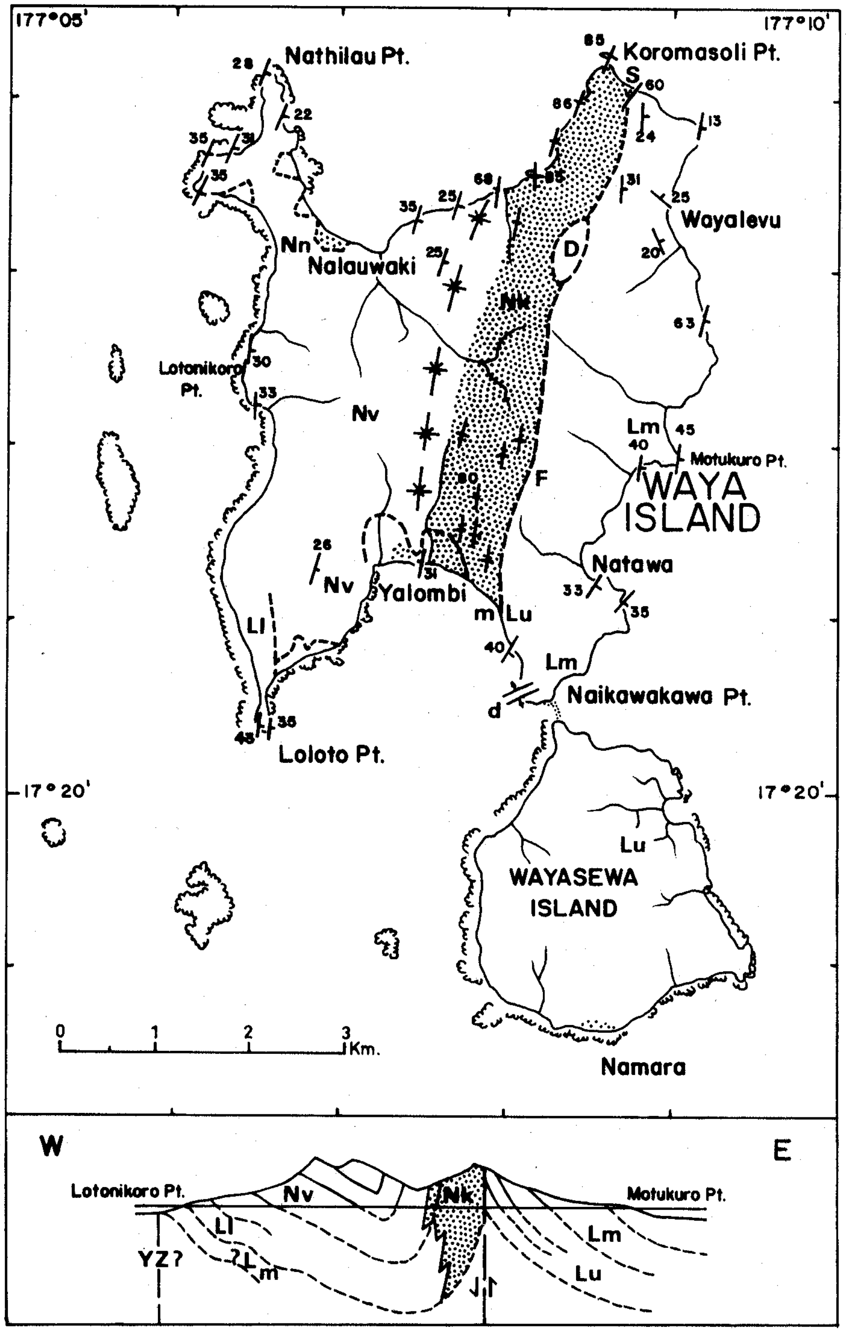

Figure 2. Geological map and cross section of Waya Island, Yasawa Group. Formational symbols as in Table 1. s = sulphidic quartz vein near Koromasoli Pt; m = manganiferous tuff in Yalombi Bay; d = subvertical dykes at Naikawakawa Pt. In section, YZ represents inferred Yasawa Zone.

TABLE 1

Lithostratigraphic units on Waya Island

	West Limb	East Limb
Nalauwaki Andesitic Group	Vatunaremba Breccia-Conglomerate (Nv) (800 m) Nathilau Breccia (Nn) and minor flows (500 m)	Koromasoli Sandstone (Nk) (1000 m)
Loto Basaltic Group (800 m)	Lololo Basalt (Ll) and intermixed basaltic and andesitic breccia (100 m+)	Motukuro Pillow Basalt (Lm) (300 m) Undifferentiated pahoehoe flows, breccia and tuff (Lu) (500 m+)

is little evidence of internal deformation such as slaty cleavage or schistosity. The fault contact with Loto Basalt is exposed southeast of Koromasoli Point, where intensely altered chloritic andesitic flow rocks and volcanic breccia at the base of the sandstone unit occupy a partly mineralised crush zone 50 m wide. A crushed chloritic band containing an 8 cm thick quartz-sulphide vein dipping 70°NW gave on analysis 2.61% Cu. Other sulphidic quartz veins with malachite stains are present in the fault zone and in adjacent Loto Basalt.

Bedding in the sandstone is simple, planar and for the most part indistinct and of low grainsize contrast. Repetitious bedding resembles turbiditic flysch, although well graded beds are rare. A typical thin section reveals an andesitic crystal-lithic tuffaceous sandstone, equigranular with approximately 10% clastic matrix, but with little or no porosity because of extensive secondary chloritic matrix. Some specimens include many calcareous microfossils* and could conceivably be petroleum source rocks. Isolated angular clasts of fine granular calcite are present in some beds but nothing of organic origin could be recognised in them. Most beds between 0.5 m and 1 m thick tend to lack internal detail, but thinner layers down to 2 cm are commonly fine-grained mudstone or argillite with current ripples, load pocketing, rip clasts and minor slump effects. The latter are well developed along the east side of Yalombi Bay where there are also slabby layers of manganese nodules associated with specific tuff beds. All sandstones contain dispersed chlorite, laumontite and other zeolites; some include calcite.

The component units of the Nalauwaki Group are interpreted as (a) proximal volcaniclastic fanglomerates (Nathilau and Vatunaremba) of mainly subaerial emplacement, and (b) distal marine flyschoid basin-fill (Koromasoli). The source would appear to have been a nearby andesitic volcano or volcanic chain lying west of Waya Island. Similar rocks are present on Naviti Island to the north, and bedded sandstones are known on some of the Mamanutha islands to the south (P Rodda, personal communication). The area of deposition was thus meridionally elongate, which together with the narrowness of the E-W transition zone, the proximity of coarse fanglomerates to a contemporaneous marine flyschoid basin, and the rapidity with which these conditions were developed, require fault control of basin formation.

The ages of the rocks are known only within wide limits as yet, and then only by lithologic correlations with units on Viti Levu (Rodda, 1967). Further study of foraminiferal samples of Koromasoli Sandstone, and of interpillow limestone from Motukuro Basalt is expected to provide palaeontologic ages. The Loto Basaltic Group is tentatively correlated with the younger part of the Wainimala Group and considered to be lower to middle Miocene. The fanglomerates and few flows of the Nalauwaki Andesitic Group are correlated with the Mendrausuthu Andesitic Group (Rodda

* Preliminary identifications in thin section by Dr A N Carter, School of Applied Geology, University of New South Wales, indicate probable late Miocene age.

1967) and the Koromasoli Sandstone with the Nandi Sedimentary Group; their age is provisionally late Miocene to early Pliocene.

The strongly folded Waya Syncline resulted from crustal compression which had been preceded by an episode of crustal extension of a fault-controlled elongate depocentre. The transverse dimensions of the syncline and the stratigraphy indicate extension over at least 5 km accompanied by subsidence, transgression and deposition, followed by shortening of at least 2 km accompanied by faulting, uplift and emergence. Since the Waya structural trend curves into the Yasawa Zone trend, the two must be dynamically associated probably in the manner of oblique congruent folds to a major wrench fault (Bishop 1968). Such folds converge at an acute angle to the fault in a direction *opposite* to that of movement of the fault block in which they are located (Fig. 3). It follows from this relationship that the Yasawa Zone underwent an unknown amount of right-lateral movement, during the Pliocene folding of the Waya Syncline, but it also seems probable that prior left-lateral movement had occurred during the preceding extensional-deposition phase. The significance of these events and changes of movement are considered in a wider and more speculative context below.

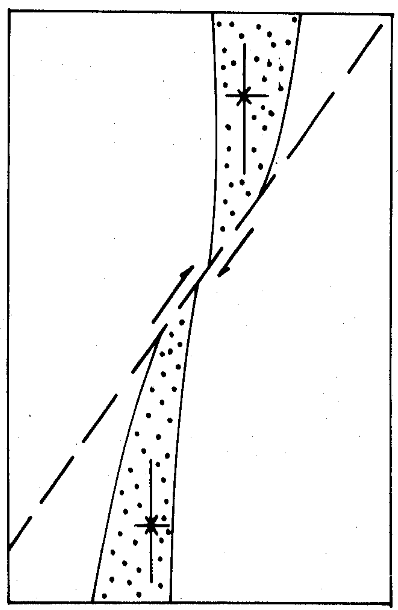

Figure 3. Diagram showing relationship between congruent regional folds and large wrench fault. A reversal of fault movement would be accompanied by meridional extensional rifting.

PLATE-TECTONIC SCHEME

Of the tectonic reconstructions cited earlier, several begin with a colinear reassembly of the New Hebrides-Fiji-Lau Tonga groups along the edge of the Australian-Indian plate. Serious objections to this can be raised on several points. For example, the Vitiaz-Samoan part of the plate edge differs in form and activity now and probably differed then from the Tongan segment, the directions of relative plate movements must have been different at each segment, the N-S petrochemical 'polarity' of Viti Levu (Gill 1970) cannot be reconciled with the position on the plate edge, and the geology of Viti Levu correlates with that of the New Hebrides but not with that of the Lau or Tonga groups.

A reassembly of only the New Hebrides and Fiji groups near the present Vitiaz-Samoan Lineament (or New Hebrides - Samoan Lineament of Hawkins 1976), is proposed on the following grounds: (a) suggested correlations between the Wainimala Group of Fiji and the Western Belt succession of Malekula, New Hebrides (Carney and Macfarlane 1976), (b) a N-S petrochemical 'polarity' in each group (Gill 1970, Gorton 1974) from tholeiitic through calc-alkali to alkalic rocks, related to their former positions on a south-facing subduction zone, (c) the probability of clockwise arc rotation of the New Hebrides of approximately 30° from a former position near the Vitiaz Trench, (confirmed by palaeomagnetic analyses by Falvey 1978).

This position for the initial New Hebrides arc was earlier suggested by Packham (1973) and later incorporated in a scheme of mineralisation, petrogenesis and tectonic evolution by Lawrence and Wood (in press).

The scheme is summarised in serial diagrams (Figs. 4, 5 and 6) and developed as follows.

During the Eocene and Oligocene, Fiji and the New Hebrides were parts of a continuous island-arc system developing near a major plate boundary — the Vitiaz-Samoan Lineament. Following arc-polarity reversal a trench and subduction zone formed along the south side of the arc with a north-dipping Benioff zone (Fig. 4). The Lau and Tonga ridges then lay close together or formed a single ridge extending south from near the east end of the trench. The nature of the plate boundary east of the Lau-Tonga ridge at that time is not known; it may not have been convergent as at present, and could have been a major transform connecting with the Alpine Fault of New Zealand.

During the early Miocene, seafloor spreading began 'behind' the Fiji - New Hebrides arc (Malahoff *et al.* in press) similar to that presently occurring in the Lau Basin 'behind' the Tonga arc. Spreading caused the arc to migrate southward and brought the Fiji segment into collision with the Lau-Tonga ridge, resulting in closure and failure of the entrapped part of the subduction zone. Continued spreading progressively displaced the New Hebrides arc southwesterly with clockwise rotation about a pole located approximately at 8°S, 167°E. The New Hebrides segment, separated from the Fiji segment by transform faults — the Yasawa Zone and the Hunter Fracture Zone — subsequently migrated southwestward to its present position.

Clockwise rotation of the New Hebrides arc by approximately 30°, and anticlockwise rotation of the Fiji segment by almost the same amount are indicated by palaeomagnetic data (Falvey 1978; James and Falvey 1978). The former is the result of seafloor spreading in the North Fiji Basin, the latter is here suggested to have resulted from oblique collision of the Fiji segment with the Lau-Tonga ridge.

Shortly after collision of the Fiji-Lau segments (Middle Miocene), the Tonga arc began eastward migration, and at about 4 m.y.b.p. (Pliocene), the Lau Basin began spreading, (Weissel 1977, Hawkins 1976). This spreading in the Lau Basin is considered to be an extension of spreading in the North Fiji Basin and a

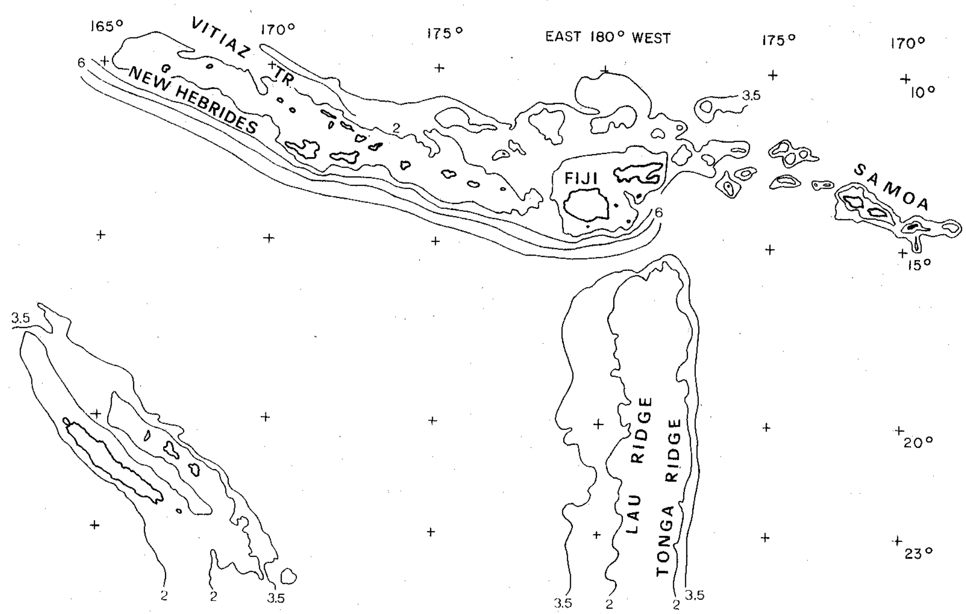

Figure 4. Suggested configuration of eastern Melanesia about early Miocene. The New Hebrides and Fiji islands formed one chain with a north-dipping subduction zone beneath and a trench to the south. They may have been part of the Pacific plate or may have been separated from it by the formerly active Vitiaz Trench. The Lau and Tonga islands occupied a single ridge.

probable result of changes in the system caused by the Fiji-Lau collision.

Several features and consequences of this scheme can be tested by field observation. For example, the closure of a subduction zone south of Viti Levu, and the Fiji-Lau oblique collision, would have resulted in a wide east- to northeast-trending zone of deformation, volcanic activity and alignments, and possibly of cataclastic or high-pressure metamorphism. A preliminary search on Kandavu Island failed to reveal the presence of either cataclasites or blueschists in situ, in xenoliths or in alluvium and other detritus. However, the presence on Kandavu of more than 12 andesitic volcanoes which erupted successively from northeast to southwest during the Pliocene and Pleistocene (Woodrow, in press) is regarded as a consequence of collision. The presence of prominent northeast-trending zones of faulting, young volcanic alignments, and earthquake epicentres in southeastern Fiji, is also significant. Related northeasterly basinal alignments occur on the seafloor along the Hunter Fracture Zone south of Kandavu, in Kandavu Passage north of the island, and in Natewa Bay and Somosomo Straits of Vanua Levu (Ibbotson and Coulson 1967).

Other consequences of Fiji-Lau collision should be observable where the New Hebrides segment parted from the Fiji segment, that is, in the islands of the western shelf, the Yasawa and Mamanutha groups.

The foregoing plate-tectonic scheme necessitates two major northeast-trending transform faults, one along the northwest side of the Fiji segment (Yasawa Zone), the other along the southeast side (Hunter Fracture Zone).

The former functioned probably for only part of the time after collision, allowing spreading of the North Fiji Basin to continue while the New Hebrides arc moved away to the southwest. The latter probably began as an oblique wrench fault system in the collision

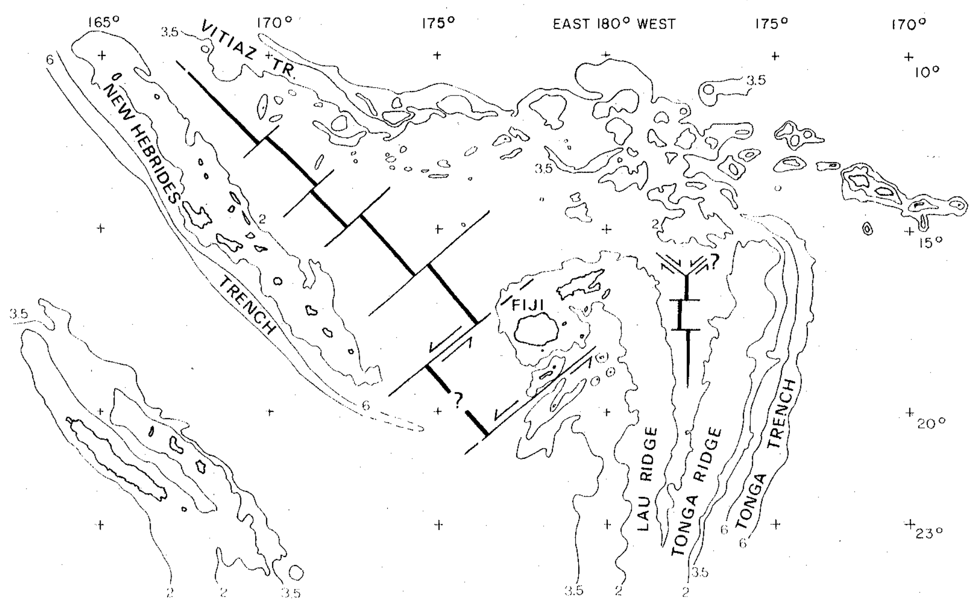

Figure 5. Suggested configuration of eastern Melanesia about 5 m.y. ago (early Pliocene). The Fiji and Lau islands formed one group following collision and closure of part of the trench between them. The New Hebrides continued moving southwest with seafloor spreading in the North Fiji Basin, and the Tonga Ridge separated from the Lau Ridge also by spreading. The spreading ridge in the North Fiji Basin had passed the Fiji Platform, and an extended segment (or possibly a newly jumped segment) connected with the Hunter Fracture Zone, a former wrench fault in the collision zone.

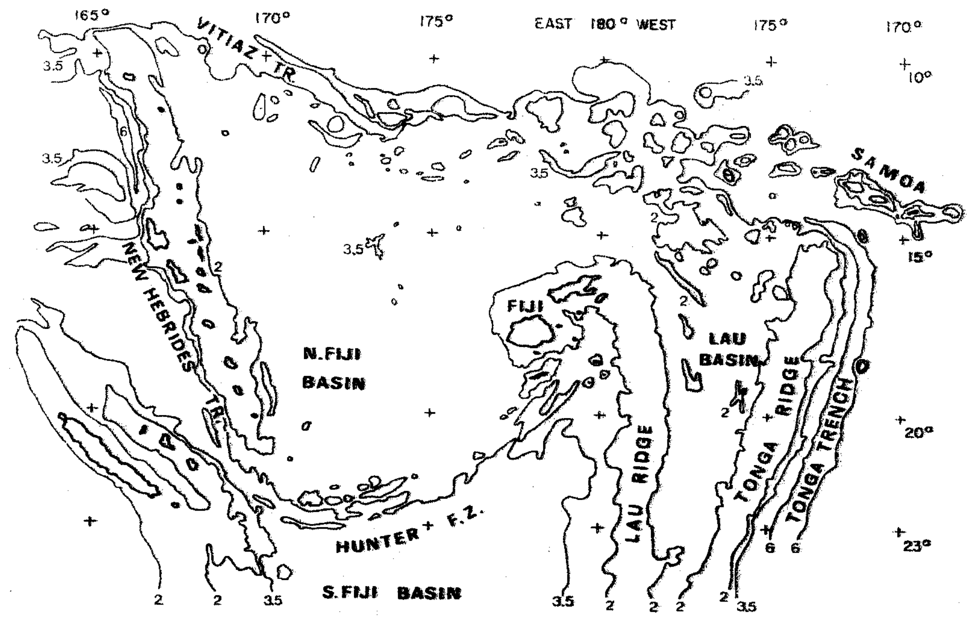

Figure 6. Island groups and regional morphology of eastern Melanesia. Water depths are indicated by 2 km, 3.5 km and 6 km isobaths (after Hawkins 1976).

zone, then, as further spreading ensued in the North Fiji Basin, developed into the major transform boundary (the Hunter Fault Zone) separating the still active spreading system of the North Fiji Basin from older crust to the south.

CONCLUSION

The Yasawa Zone probably began as a simple left-lateral wrench fault across the New Hebrides-Fiji arc during the Fiji-Lau collision, and for some time thereafter left-lateral movement continued as the New Hebrides arc moved away. During this late Miocene phase, north-trending extensional rifting along the Zone, augmented by anticlockwise rotation of the Fiji Platform, led to andesitic eruptions and deposition of Nalauwaki epiclastic sediments in an elongate fault-controlled local basin (or basins). However, at some later stage, seafloor spreading in the adjacent North Fiji Basin would have translated the principal spreading centre past the Fiji Platform, or as suggested by Malahoff *et al.* (in press) the spreading zone jumped to a new position southwest of the Platform. The time at which such a jump occurred is indicated by the seafloor magnetic-anomaly pattern as being just prior to anomaly No. 3, at the commencement of the Pliocene (LaBrecque *et al.* 1977).

The shift of the spreading zone past the Yasawas would have resulted in a change of movement on the Yasawa Zone from left lateral to right lateral, and on the nearby platform area from extensional rifting to compressional folding. The changes recorded in the seafloor magnetic sequence thus correlate closely in time with the tectonic changes recorded in the stratigraphy and structure of Waya Island.

Generalisations arising from these data may be relevant to the search for economic resources in the western platform area. For example, obliquely convergent folds like the Waya Syncline probably occur along the Yasawa Zone and widen out with decreasing amplitude into the adjacent platform. The more distant stratigraphy may include petroleum source rocks and greater porosity, whereas near the zone local cataclastic greenschist metamorphism provides favourable conditions for sulphide mineralisation.

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POTENTIAL DELINEATION OF ISLAND ARC TECTONICS BY FORAMINIFERAL DISTRIBUTION PATTERNS*

IAN DEIGHTON AND DAVID TAYLOR

ABSTRACT

Conflicting data provided by paleontologists has resulted in widely varying interpretations of the geological history of island arcs and adjoining marginal seas. This variance is seen to be primarily the result of a misunderstanding of the (usually biased) paleontological studies, during geological synthesis.

Such conflict of information is in fact a reflection of the sedimentological and geotectonic processes operating in an active island-arc/marginal-basin environment. These processes include relatively slow horizontal movements over large distances (as shown by plate tectonics), relatively rapid vertical tectonic movements on both a localised and widespread scale, rapid lateral environmental change over small distances and intermixing of sediments from widely disparate environments. With these processes in mind:

Delayed arrival of biostratigraphically important planktonic microfossils in Southwest Pacific marginal basins (relative to the Pacific Ocean on the other side of the arc) is explained by vertical tectonic displacement of the Melanesian Arc. Staggered delays of various species leads to a refinement of the vertical scale of such movement since planktonic foraminifera are and were stratified in the oceans. Bottom-living, benthonic foraminiferal assemblages provide further data for relative paleobathymetric scales.

Interbedding of reefal or shoal benthonic assemblages within pelagic oozes (with deep-water benthonic assemblages) is explained by turbiditic displacement of shallow-water sediments into the adjacent deep-water basin or trench. Quantification of the proportions and frequency of these allodapic limestones can lead to the construction of tectonic instability scales for the arc, plotted against time.

The Papuan late Tertiary 'stages' based on benthonic foraminifera, have little or no absolute chronological validity. These benthonic assemblages varied in age (relative to the planktonic zonation) from one section to another. Comparison with modern benthonic foraminiferal depth ranges allows production of well-site paleobathymetric curves. Comparisons of curves from many wells indicates basin-filling trends and vertical tectonics. On the basin-wide scale, such data could lead to prediction of possible oil-migration paths in underlying reservoir rocks.

Thus analysis of foraminiferal distribution patterns leads to delineation of island-arc movements, both regionally and, more importantly for petroleum exploration, within individual sedimentary basins.

INTRODUCTION

Biostratigraphy provides the scale against which geologic development is measured. Early biostratigraphy in the Southwest Pacific was based on evolution of the larger benthonic foraminifera.

A major advance in the Tertiary biostratigraphy of the Southwest Pacific region (locations Fig. 1) occurred in 1967 when new planktonic foraminiferal zonal schemes developed in Trinidad were applied to the younger Tertiary sediments of Papua (Lloyd 1974). Against the planktonic foraminiferal schemes the older benthonic foraminiferal schemes were seen to be grossly inaccurate for the late Tertiary (Lloyd loc. cit.). Since then, with the advent of the Deep Sea Drilling Project and similar marine geological, geophysical and oceanographic research cruises, major advances in understanding of oceanographic and sedimentologic processes of island-arc regions have taken place.

Reworking is now recognised as an important factor in previously recorded zonations, especially those based on 'larger benthonic foraminifera', and even recent statements such as those of Lloyd (1974) are now recognised as deficient. Lloyd (loc. cit. p. 19) and recently McGowran (1979 p. 244) make blanket assertions that there is a regional hiatus in the Late Miocene of New Guinea. Our re-examination of 'type' sections, such as

* *Paltech* Pty. Ltd., Report 1979/12.

Wana-1 (location, Fig. 4) reveals there is no such regional hiatus, nor is the sequence 'strongly regressive' as suggested by McGowran (loc. cit.). Regional overviews are self defeating when analysing geological processes of active margins.

Techniques are now available which make use of these advances in paleontologic dating and paleoenvironmental analysis and, when used judiciously, can accurately quantify island-arc tectonic and sedimentologic processes.

On the basis of our own studies, we present new and revised data, showing three methods of increasing understanding of past events in the Southwest Pacific island arcs.

PLANKTONIC FORAMINIFERAL RANGE ANALYSIS

Figure 2 shows the Pliocene distribution ranges of 15 important planktonic foraminifera from sections on the eastern and western sides of the Melanesian Arc (location, Fig. 1). Of the 15 taxa plotted, 11 species have identical overall ranges. These will be referred to as Group A. The other four, Group B, either did not appear on the western side of the arc *(Sphaeroidinella dehiscens immatura)* or appeared later on the western side of the arc than to the east. These latter three species, in order of first appearance on the western side of the arc are *Globigerinoides rubra, Pulleniatina obliquiloculata,* and *Sphaeroidinella dehiscens dehiscens.*

There are, theoretically, several possible explanations for this delayed distribution pattern, relating to the influence of temperature, latitude, plate movement or water stratification. Temperature alone cannot be the controlling factor, since today many of the forms in Group B have mean and cut-off temperatures similar to those of Group A (Bé 1977, p. 35, fig. 8, tables 3 and 5), which were present on both sides of the arc in N.19 (refer Fig. 2). Nor can latitude be the controlling factor, since the species of Group B were all present approximately 4 m.y. ago at Lat. 16°S

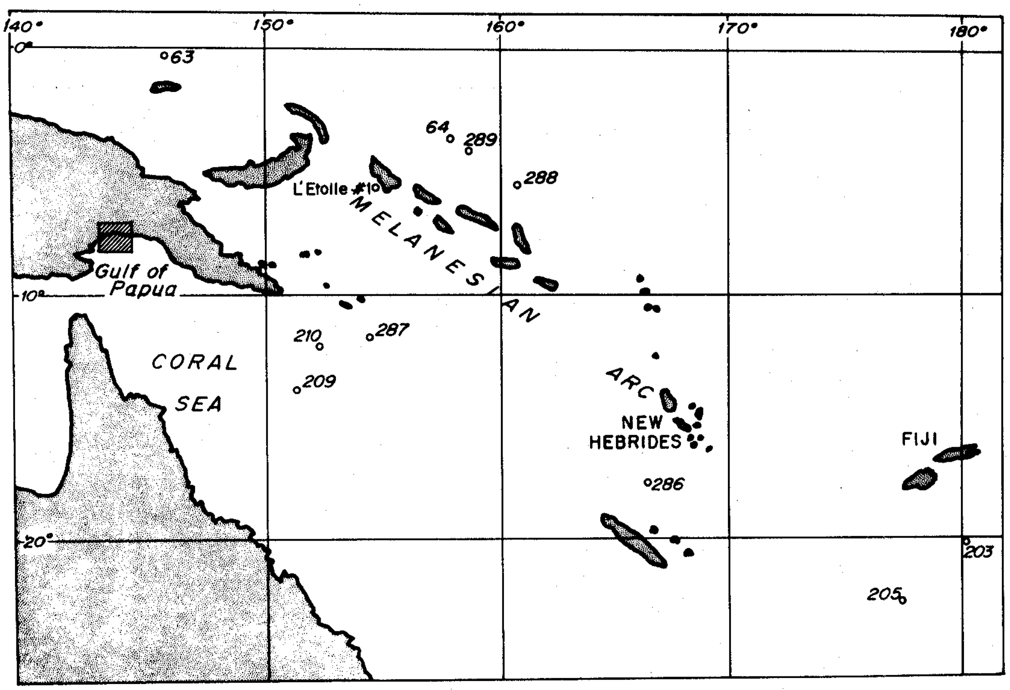

Figure 1. Location Map, showing wells and areas referred to in text. Small box in the Gulf of Papua indicates location of Fig. 4.

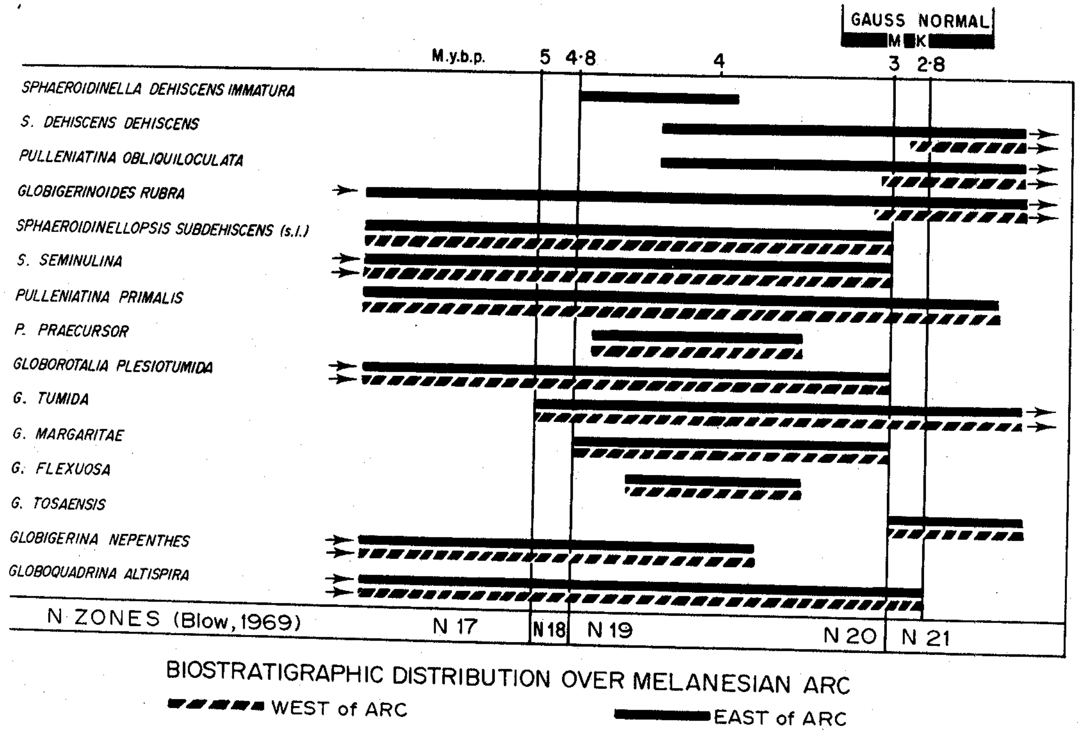

Figure 2. Planktonic foraminiferal range diagram for sites east and west of the Melanesian Arc. Sites presently located on the arc, (e.g. Espiritu Santo) are part of the east group. Sites 'east' of the arc used in compiling the diagram were: Fiji (Blow 1969), Espiritu Santo (our examination), Ontong Java Plateau, DSDP Site 64 and East Caroline Basin, DSDP Site 63 (Bronnimann and Resig 1972). Sites 'west' of the Arc: Coral Sea, DSDP Site 210 (Kennett 1973), Gulf of Papua (our examination) and L'Etoile-1 off Bougainville (Taylor 1975).

(Fiji) but were delayed till 3 m.y. ago in latitudes up to 10° further north, such as in L'Etoile-1 near Bougainville at Lat. 6°S (Taylor 1975). Recently developed plate-tectonic reconstructions, based on magnetic lineations and paleomagnetism, certainly do not involve any large-scale relative N–S displacements of the Indian and Pacific plates at this time (e.g. Falvey 1978, fig. 5), and in any case such an explanation would rely on temperature and latitude as controlling factors as well.

Planktonic foraminifera are stratified in the water column (Bé 1977 and Van Donk 1977). Many of the taxa of Groups A and B are extant in the modern Pacific Ocean, and depth preferences for these species have been tabulated by Bé (loc. cit.) and Van Donk (loc. cit.). Using these depth preferences we have constructed Fig. 3, which shows a model of the planktonic foraminiferal stratification at 4 m.y.b.p. on both sides of the Melanesian Arc. The stratification on the right hand side of Fig. 3 is nearly identical to the stratification in the present-day Pacific Ocean. Reference then to the range diagram (Fig. 2) shows a direct relationship between presumed depth habitat and the order of appearance of Group B taxa west of the arc; *Globigerinoides rubra* appears before *Pulleniatina obliquiloculata,* which appears just before *Sphaeroidinella dehiscens dehiscens*. This ordering implies that surface water from the Central Pacific penetrated into the Coral Sea earlier than the deeper-water layers. An important consideration here is that while individual specimens of the named planktonic foraminifera may have filtered across the arc *prior* to the suggested breaching, it was not until water masses (ecologic conditions) suitable for that species' propagation were established on the west side of the arc that it could exist in any great numbers there, and hence appear regularly in sampled sections. Figure 3 has not been vertically scaled but the ***Sphaeroidinella dehiscens* / *Globorotalia tumida*** association was probably well below 100 m (Bé 1977, table 4 and text).

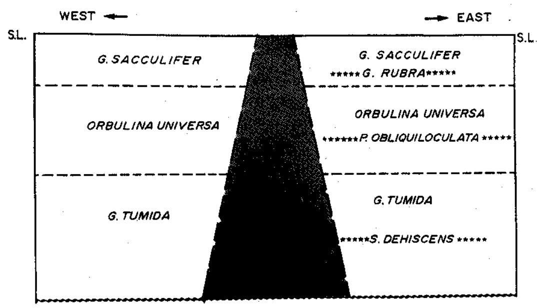

Figure 3. Planktonic foraminiferal depth stratification — east and west of the Melanesian Arc — at 4 m.y.b.p.

Thus, matching planktonic foraminiferal biostratigraphic ranges with their known water-depth habitats, we have shown that changes in the topographic configuration of the Melanesian Arc took place at about 3 to 2.8 m.y.b.p., allowing increasing penetration of Central Pacific water masses over the arc and into the Coral Sea.

The appearance of *Sphaeroidinella dehiscens dehiscens* behind the Melanesian Arc is here dated at 2.9 m.y.b.p. Van Hinte (1978) has shown, from evidence collected in Gulf of Mexico well sections, that a glacio-eustatic drop in sea level of about 200 m took place at this time, associated with the onset of the Nebraskan glacial (Stainforth *et al.* 1975, fig. 22). The change in topographic configuration of the Melanesian Arc must have involved breaching (due to subsidence or transform faulting) of at least part of the arc, and the rate of 'subsidence' of the breach must have been greater than the suggested contemporary glacio-eustatic drop in sea level.

Indeed, delays in the first appearance of *Sphaeroidinella dehiscens* behind island arcs have been recorded elsewhere: for the Gulf of Mexico relative to the central oceanic Atlantic sequences on Trinidad (compare the zonations of Stainforth *et al.* (1975) with Blow (1969)), and for the Andaman Sea relative to the Indian Ocean (M. S. Srinivasan, pers. comm. 1978). The first appearance of *S. dehiscens* in the Gulf of Mexico occurs at 3.0 m.y.b.p. (Stainforth *et al.* 1975, fig. 19). A chronologic date for the first appearance of *S. dehiscens* behind the Andaman Arc is not available. If it were similar to the two other cases (i.e., at or around 3.0–2.8 m.y.) this could indicate synchronous movement on widely separated island arcs. An alternative possibility is that in one or both of these cases (*viz* Andaman Sea and Gulf of Mexico) gross oceanic circulation changes associated with glaciation could be responsible for the sudden appearance of *S. dehiscens.* Staggered delays, as evidenced for the Melanesian Arc, are not apparent in the Gulf of Mexico.

Late Pliocene or Pleistocene tectonism of the Melanesian Arc has been confirmed, at least on Espiritu Santo, where surface exposures of substantially deformed carbonate turbidites contain N.19 and N.20 faunas. Associated benthonic faunas indicate deposition in at least 500 m water depth, thus uplift of these turbidites must have occurred later than N.20 in the period 2.8 m.y. to present.

Depth-related staggered appearances of planktonic faunas west of the Melanesian Arc are here considered evidence of deformation of the arc and illustrate a valuable general method for evaluation of tectonic island-arc movement.

ALLODAPIC LIMESTONE ANALYSIS

Because of the high relief of island-arc regions, it is hardly surprising that sediments composed of shallow-water carbonate detritus often are found in very deep water sequences.

These carbonate turbidites or allodapic limestones have been recognised in many areas, both on and offshore. The previously mentioned carbonate turbidites on Espiritu Santo are examples. In Papua, the term 'Kereruan' was introduced by de Verteuil and McWhae (1948) for the Lower Miocene Limestones in the Kereruan Range which comprised

> 'about 400 feet of well bedded and sometimes massive grey orbitoidal limestones and cream dense medium grained limestones containing *Lepidocyclina, Miogypsina, Miogypsinoides* and abundant smaller

foraminifera. Subordinate layers of fine grained *Globigerina* limestone also occur . . . it is relatively easy to fix the lower age limit of the limestone by the presence of *Spiroclypeus* and *Eulepidina*' . . . however . . . 'the age determination of its upper boundary is difficult as the Kereru limestone passes conformably and imperceptibly upwards into the Puri-type limestone.'

The 'Puri-type limestone' is now recognised as a globigerinid ooze. The above passage is obviously a description of an allodapic limestone and 'Kereruan' is an environmental or facies term used erroneously in a chronostratigraphic sense. Surprisingly the age of the 'Kereruan' has withstood the test of time. This is perhaps understandable for the base of the unit, as *Eulepidina* is a useful biostratigraphic marker. The top of the 'Kereruan' in sampled sections and wells usually turns out to be roughly the top of the Early Miocene as measured by planktonic foraminiferal zonal schemes (Lloyd 1974). Younger Papuan 'stages' of the basin-fill regime such as the Muruan and Ivorian are markedly diachronous, as explained in the next section.

Occasional bands of derived calcarenities occur within the 'Puri' deep-water sequences (e.g. Omati-2). The major difference between the 'Kereruan' and 'Puri' facies is one of relative proportions of allochthonous (derived) material to autochthonous (in situ) material. The sudden decrease in proportion and frequency of these derived carbonate turbidite flows which defines the top of the 'Kereruan' seems therefore to be related to a decrease in tectonic disturbance in the Papuan region. It is not purely a function of the increasing water depth at the time (see next section), because the top of the 'Kereruan' is a relatively synchronous event.

This example indicates the possibility of establishing 'tectonic instability scales' for island-arc areas by measuring the relative proportions and frequencies of derived bands in these allodapic limestone sequences. It is obvious that the chronostratigraphy for these scales must be based on the planktonic foraminifera within the autochthonous bands. However, careful analysis of the components of the allochthonous bands would indicate whether the derived material was the result of contemporaneous slumping or subaerial (or submarine) erosion or a combination of both these possibilities.

GEOHISTORY ANALYSIS

Geohistory analysis, a new quantitative approach to the application of micro-paleontology to geologic history, was developed by J E van Hinte during the years prior to 1975 and published in 1978 (van Hinte 1978). Basically, geohistory analysis relies on recent advances in high-resolution biostratigraphic and paleobathymetric determination. Currently attainable chronostratigraphic resolution using planktonic foraminiferal biostratigraphy is about 1–2 m.y. for Tertiary tropical sequences, while bathymetric ranges for extant benthonic microfossils are continually being documented and refined. Many of the bathymetrically restricted extant benthonic foraminifera are found in sequences throughout the Tertiary, so their known present-day depth preferences can be used with considerable reliability to determine Tertiary paleobathymetry. We have already mentioned the water-depth stratification of planktonic foraminifera; such data can be used to supplement depth information from the benthonic microfossils, although it is generally not as accurate. Other depth-related phenomena such as calcium-carbonate dissolution provide increased resolution in the abyssal and deeper-water sequences.

Great care is needed to interpret these phenomena, because changing paleocurrents due to changing basin configuration, which is the norm for island-arc regions, influence the depth ranges of planktonic and benthonic microfossils as well as calcium-carbonate dissolution depths. For instance, in areas of oceanic upwelling the calcite compensation depth (C.C.D.) is known to rise; a relative rise of 650 m has been observed in Panama Basin waters over upwelled areas (Moore *et al.* 1973). For comparisons of sequences separated by short distances within discrete basins such considerations are generally not necessary.

To illustrate the general method of geohistory analysis we have constructed Figs. 5, 6 and 7 as tentative geohistory diagrams for three wells from the Gulf of Papua (based on core samples examined by us and Australasian Petroleum Company well reports on open file). The top curve in the diagram shows the water-depth history for each well site interpreted from the benthonic microfossils (heavy line) and, where this is deficient, from general basin development (dashed line). It is important to note that the water-depth curve is plotted relative to a static sea level. While sea level is known to have fluctuated in the past, for instance during the Oligocene sea-level fall, it is not necessary to remove this component when making comparisons between wells, such as is done on Fig. 8. It should be noted that, for simplicity, neither the Late Miocene nor Pleistocene sea-level fluctuations (both of which were quite significant) has been shown in these curves. Eustatic sea-level fluctuations affect all sites equally. However, in order to determine the true nature of the tectonic mechanism responsible for the basement-depth curve, eustatic sea-level effects must be removed.

Figures 5, 6, 7 and 8 are plotted assuming zero compaction of sediments. Porosity factors obviously need to be taken into consideration to refine the curves, particularly the basement-depth curve (here using the top of the Jurassic as 'basement' in Omati-1 and Iviri-1, and top Cretaceous for Wana-1).

The most reliable parts of the curves shown are the 'basin-fill' Late Neogene sections for Wana-1 and Iviri-1. For these sections paleo-water-depths have been established by reference to the tops of the depth ranges of extant benthonic foraminifera (data contained in Phleger 1960) in most cases. The key is:

A–A′ : Shelf edge break, top of *Cassidulina carinata* range, 200 m water depth.

B–B′ : Top of *Loxostomum*, usually *L. karrierianum*, 500 m water depth.

C–C′ : Top of *Eggerella bradyi*, 1000 m water depth (limits not as reliable as for *P. wuellerstorffi*).

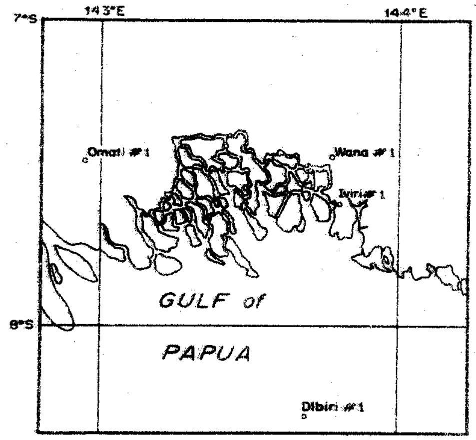

Figure 4. Location map for Papuan Basin wells mentioned in text.

D–D′ : Top of *Planulina wuellerstorffi*, 1000 m water depth.

The curve A–A′ indicates the path traced through time by that particular paleo-bathymetric level; thus A indicates the depth of the shelf break in later N.19, A′ indicates the depth *in the well* of the N.19 shelf-edge indicators.

DISCUSSION

The most immediate feature of the diagrams is that they are similar — reflecting sequences of basin development and eustatic cycles. In detail, however, a number of important differences can be noted. In particular, basin filling during the Neogene commenced at different times in different places. At Omati-1 it appears to have commenced about 14 m.y.b.p. (N.9), at Wana-1 about 7 m.y. (mid N.17), at Iviri-1 about 6 m.y. (late N.17) while further offshore at Dibiri-1 it commenced at 3 m.y.b.p. (N.20). Thus the age of the top of the 'Puri' facies decreased offshore, reflecting shelf progradation. Facies diachronism needs to be considered when basin analysis is dependent on seismic techniques. Seismic reflections depend on lithologies (i.e. facies) and, as shown, these can be markedly time-transgressive.

Comparison between the base Eocene depth curves (Fig. 8) shows that tectonic influences are not synchronous either. Subsidence during

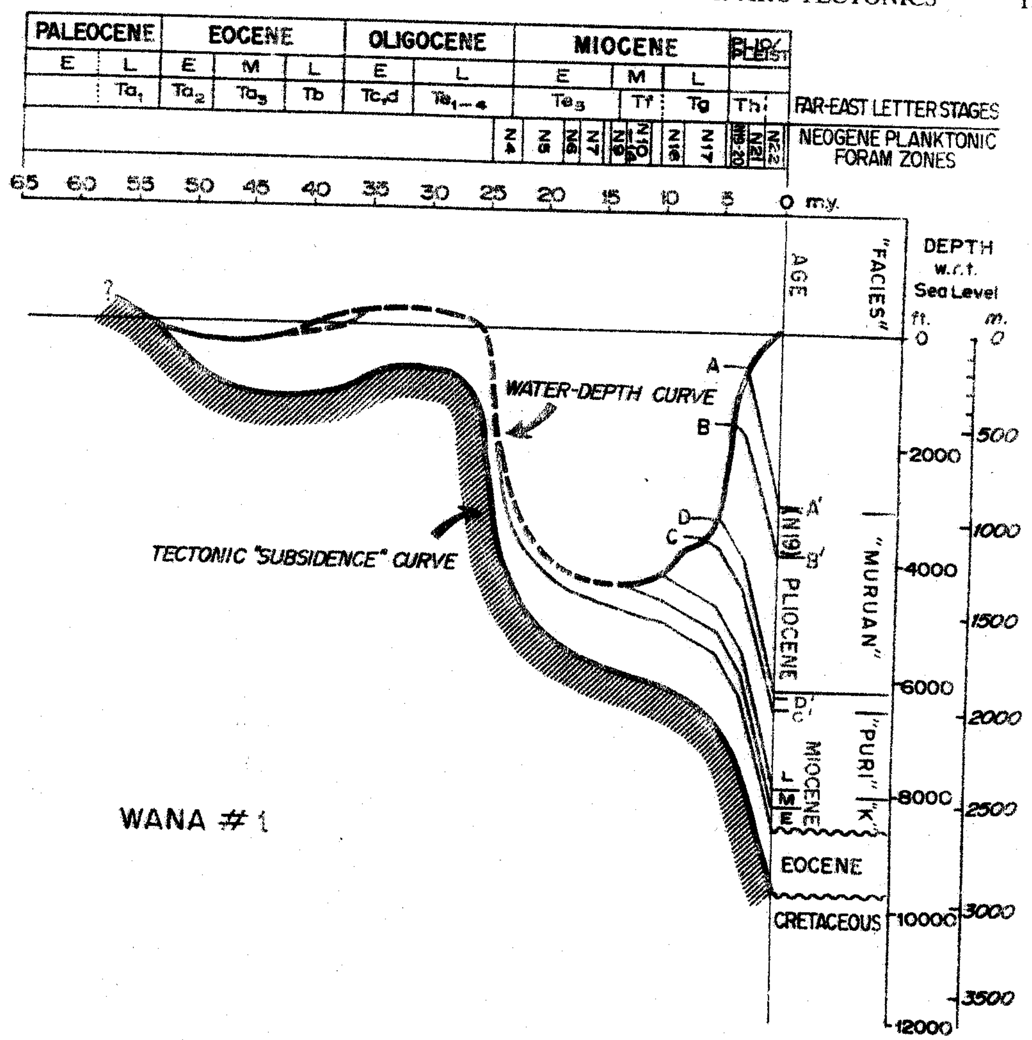

Figure 5. Tentative geohistory diagram for Wana-1 (location Fig. 4). Uncorrected for compaction and eustatic effects. Symbols: K, Kereruan; A-A′ etc., see text. Correlations after Van Couvering and Berggren (1977), Berggren (1972) and Lloyd (1974).

the Neogene commenced earlier at Wana-1 than at Iviri-1. It should be noted here, that the *relative* amounts of tectonic subsidence shown are probably not accurate, as (and it is stressed again) porosity changes were not taken into account for these tentative curves. Similarly, depth curves for any unit within a well can be prepared and compared with other wells: with sufficient data this would allow tectonic paleo-dip curves to be produced and would quantify and date movements along faults. Thus greater precision in establishing past oil-migration paths is attainable, given sufficient well density.

The precision of the geohistory-analysis technique is dependent primarily on sample quality and frequency. For some parts of the curves shown, for instance the 'Muruan' section, a combination of detailed ditch-cutting study with relatively sparse sidewall core sampling may be sufficient to produce an accurate curve. For the 'Puri' facies, on the other hand, only detailed sidewall coring can provide the accuracy required. The same is true for a rapidly subsiding sequence which

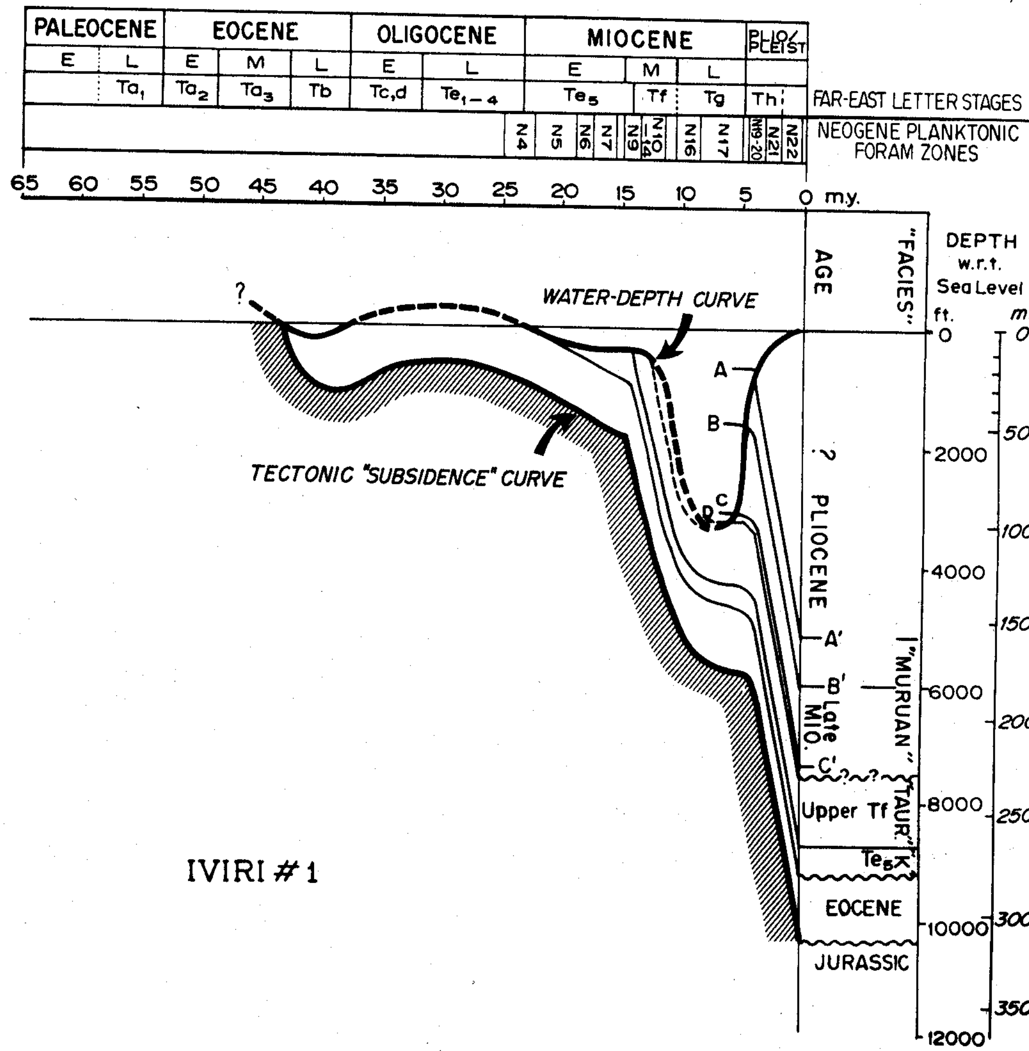

Figure 6. Tentative geohistory diagram for Iviri-1 (location Fig. 4). See caption, Fig. 5, for symbols. It is considered here that reworking of 'Taurian' indicators into younger deep-water sediments (pre N.17) was not previously recognised. Thus the unconformity at the top of the 'Taurian' is an artifact.

shoals later. In this case, downhole contamination of shoaling benthonic microfossils can mask the *in situ* microfossils of the subsiding phase. Over small parts of the section, conventional coring should be considered to resolve such events as unconformities, especially those due to subaerial exposure. Preliminary geohistory diagrams can be prepared before drilling commences, and these can be used to indicate sections for detailed sampling. While drilling is in progress, first downhole appearances of deep-water forms also indicate that the depositional rate should decrease below that level, and sidewall-coring frequency below such a level can be increased to compensate for slower deposition.

Van Hinte (1978) listed many of the advantages of geohistory diagrams. Among these are maturation, migration and porosity prediction. The diagrams form a convenient visual linear time-depth frame against which may be plotted such parameters as heat-flow data and mineralizations. Considerations of such factors are vital in the search for

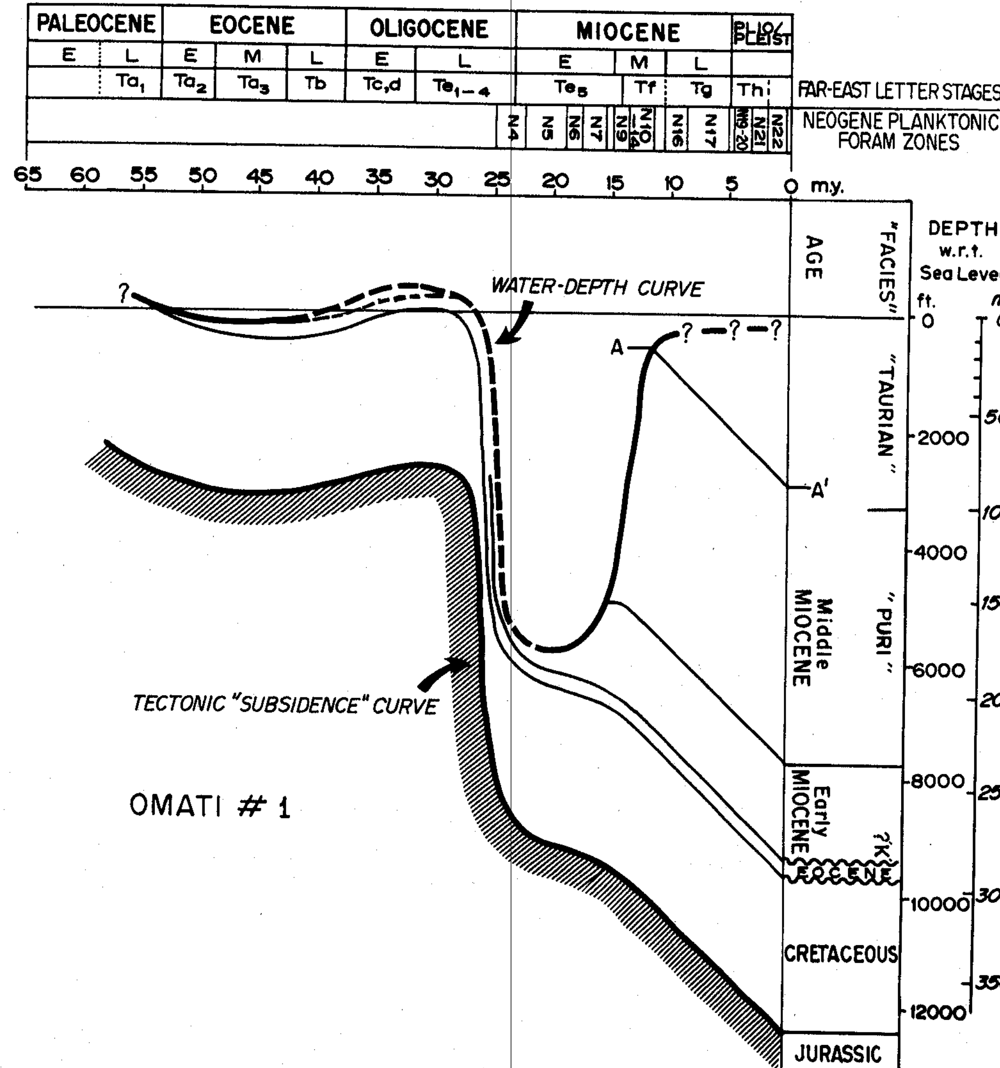

Figure 7. Tentative geohistory diagram for Omati-1 (location Fig. 4). See caption, Fig. 5, for details.

hydrocarbon accumulations. The reader is referred to Van Hinte (1978) for further detail.

CONCLUSIONS

Recent advances in biostratigraphic and paleontologic techniques provide extremely valuable information in the tectonic and sedimentologic processes which have operated in the past in island-arc areas. Such information is invaluable in the search for hydrocarbon accumulations. The technique of geohistory analysis in particular, offers much needed resolution to the petroleum exploration industry, though the technique relies for its success on detailed well sampling, including sidewall and conventional coring.

Petroleum exploration in CCOP/SOPAC countries is about to enter the drilling phase. It is recommended that detailed sampling be undertaken to allow full utilization of these newly developed techniques.

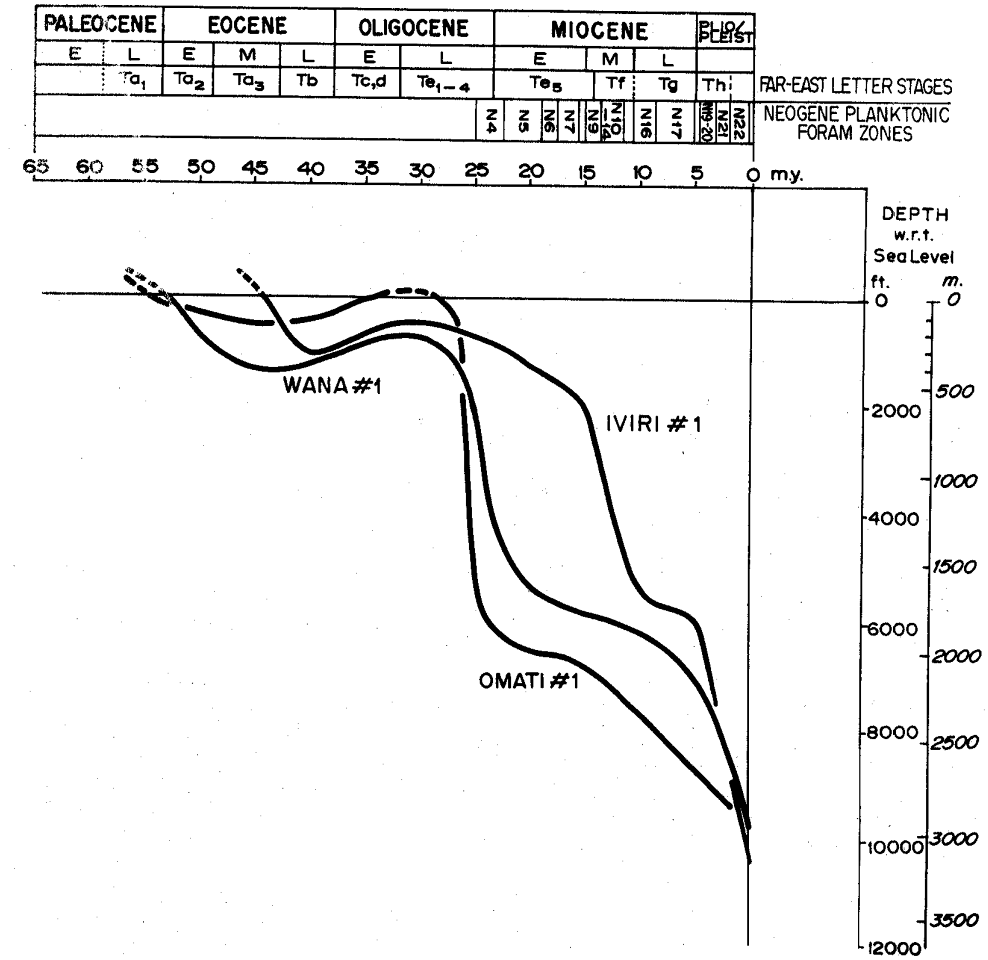

Figure 8. Comparison of subsidence curves for the base of the Eocene for Wana-1, Iviri-1 and Omati-1. Uncorrected for compaction or eustatic effects.

ACKNOWLEDGEMENTS

Reappraisal of the Wana-1 and other Papuan Basin Neogene sections, which led to some of the ideas expressed in this paper, was conducted for ESSO Australia Pty. Ltd. in 1975/76. Assistance during sampling for these studies was provided by Allan Partridge (ESSO) and Ron Walters (B.P. Petroleum Development Australia Pty. Ltd.).

The examination of L'Etoile-1 was conducted for Shell Development (Australia) Pty. Ltd. in 1975. Permission to publish results from the well was granted by Shell Development Australia Pty. Ltd., Oceanic Exploration Company, Mono Power Company (Papua New Guinea), Hematite Petroleum Pty. Ltd., Ted Weiner Oil Properties, Gulf Oil Exploration and Production Company, Bougainville Texas Gulf Inc., Amoco International Oil Company.

During 1978, David Taylor visited the New Hebrides and sampled sequences around Big Bay, Espiritu Santo. Field observations and the microfaunas in these sediments were vital in the formulation of several of the ideas in this paper, and members of the New Hebrides Geological Survey, especially Sandy Macfarlane, are thanked for logistic support, discussion and samples.

Our thoughts on some of these matters benefited considerably through discussion with David Falvey and Gordon Packham at the Department of Geology and Geophysics, University of Sydney.

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PETROLEUM POTENTIAL OF JAMAICA: A CASE STUDY OF PART OF AN ANCIENT ISLAND ARC

ANTHONY N EVA

ABSTRACT

Jamaica's Cretaceous geological history has been widely interpreted as part of an active island arc system. Forearc basin and trench environments were present in what is now eastern Jamaica, and backarc and shelf environments were present in western and central Jamaica. Most volcanism ceased towards the end of the Cretaceous, but during the early Tertiary the arc split to produce active and inactive marginal basins. These initially accumulated thick sequences of arc-derived clastic sediments, but from Middle Eocene to Middle Miocene times thick carbonate sequences were deposited. The frontal and remnant arcs were submerged during this time, and they themselves accumulated thick shelf carbonates. Uplift in Middle Miocene times essentially produced the island's present shape.

Ancient sedimentary environments on Jamaica therefore include fossil examples of forearc basins, trench, backarc basins, active and inactive marginal basins and shelf environments. By examining each of these palaeoenvironments in terms of their petroleum potential it appears that in Jamaica the inactive marginal basins and backarc basins are the most promising exploration targets.

INTRODUCTION

This paper attempts to interpret the petroleum geology of Jamaica in the light of work on active or potentially active island arcs. In other words, the stratigraphy and events of part of an island arc which has long been inactive will be correlated with lithologies and processes which can be observed on present-day arcs. This approach provides a realistic way in which to evaluate the petroleum potential of Jamaica's morphotectonic units.

Any assessment of a region's petroleum potential must rely heavily upon a knowledge of the local geology. A detailed review of Jamaican geology is outside the scope of this paper and for present purposes a summary is given, only, of points considered by the author to be both important and relevant. For an in-depth review of the island's stratigraphy readers are referred to articles by Zans *et al.* (1962) and Meyerhoff and Kreig (1977a).

This paper considers only Jamaica's onshore (11 420 km^2) area, though some experts (Greiner 1965; Meyerhoff and Kreig 1977a, b) favour Jamaica's offshore (9512 km^2) region. Meaningful discussion of the offshore region would require an interpretation of geophysical data obtained recently (some 4065 line km of seismic, gravimetric and magnetic observations (Anonymous 1979)) but not yet available.

The development of tectonic events on Jamaica in terms of plate tectonics was outlined by Horsfield and Roobol (1974) and has since been revised by Jackson (1977) and by Draper (1979). Although several details remain unresolved, the overall development is fairly clear and is summarized in Fig. 1, which relates the positions of the major sedimentary basins and the time of their development to the sites of subduction and active volcanism. Though grossly simplified, it provides a framework for the discussion that follows.

CRETACEOUS SEDIMENTARY BASINS

Backarc basin and shelf

Description and Environmental Interpretation: The Hanover Block of western Jamaica (Fig. 2) represents, at least in part, Santonian-Campanian backarc basin and shelf environments. These have recently been the subject for a study by Grippi (1978), and this section is based mainly on his work.

The backarc basin contains some 5+ km of mudstones, siltstones, sandstones, conglomerates and limestones; most of which were derived from the active arc. Lithological associations include alternating beds of sandstone and shale, which probably represent turbidite facies forming the middle and outer

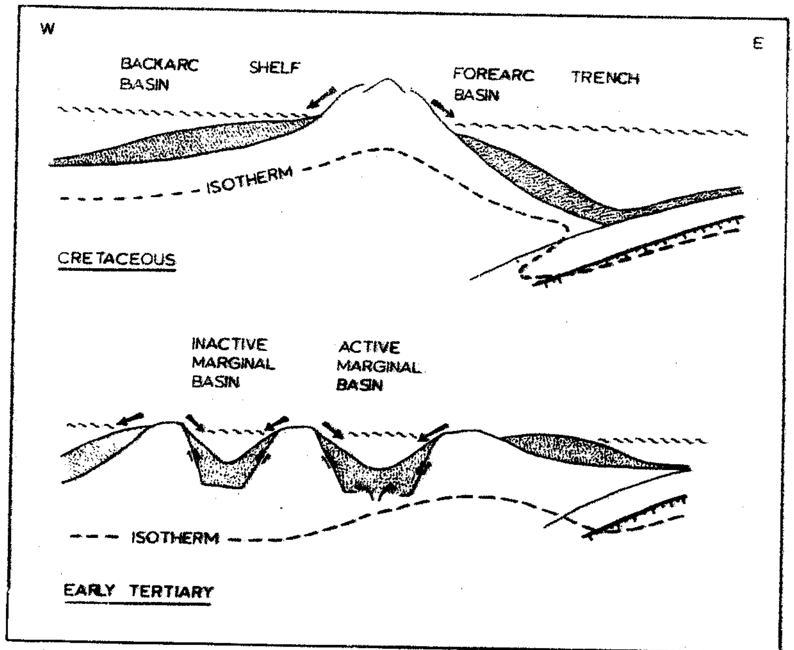

Figure 1. Diagrams showing the positions of the major sedimentary basins in relation to the sites of active volcanism and subduction.

parts of a submarine fan complex. Massive sandstones and local conglomerates cut through these turbidite facies in places and are interpreted by Grippi (1978) as abandoned and infilled distributary channels on the middle part of a fan. Grippi (1978) also found a predominantly conglomeratic sequence, forming a body 4.5 km long by 2.4 km wide, cutting across the regional trend. This he interprets as an infilled submarine canyon, which was originally a feature of the outer shelf and inner fan. Small lenses of fossiliferous limestone (e.g. Clifton Limestone), oolitic in places, represent a small part of the shelf sediments, which appear to be dominated by sandstones and conglomerates. A diagrammatic representation of the backarc basin and shelf sequence as they appeared at the end of the Campanian period is shown in Fig. 3.

Petroleum Potential: As yet there is no well documented case of an ancient submarine fan, now exposed on land, being a major petroleum producer. Much of the following discussion tries to draw comparisons with present-day submarine fans, where the possibility of finding petroleum deposits has been advanced by several authors (Moore 1969, Moore and Fullam 1973, Caughey and Stuart 1976, Wilde *et al.* 1978).

The thick sequences of marine shales on the middle and outer parts of the Cretaceous fan of western Jamaica could have acted as source beds, but unfortunately no measurements have yet been made as to the nature or abundance of their organic content. According to Connan's (1974) data Santonian-Campanian sediments would require a threshold temperature for petroleum generation of less than 80°C. With a palaeotemperature gradient in the range 20-40°C/km, they would need only to have been buried to depths of 2-4 km and this would almost certainly have been the case until Middle Miocene times. The necessary conditions for the conversion of organic material into petroleum were, therefore, almost certainly satisfied.

The geometry of the coarse sandstones and conglomerates on the upper and middle parts of submarine fans and their potential for forming stratigraphical traps have been discussed by Normark (1978), Walker (1978), and Wilde *et al.* (1978). It seems likely that similar stratigraphical traps exist in the

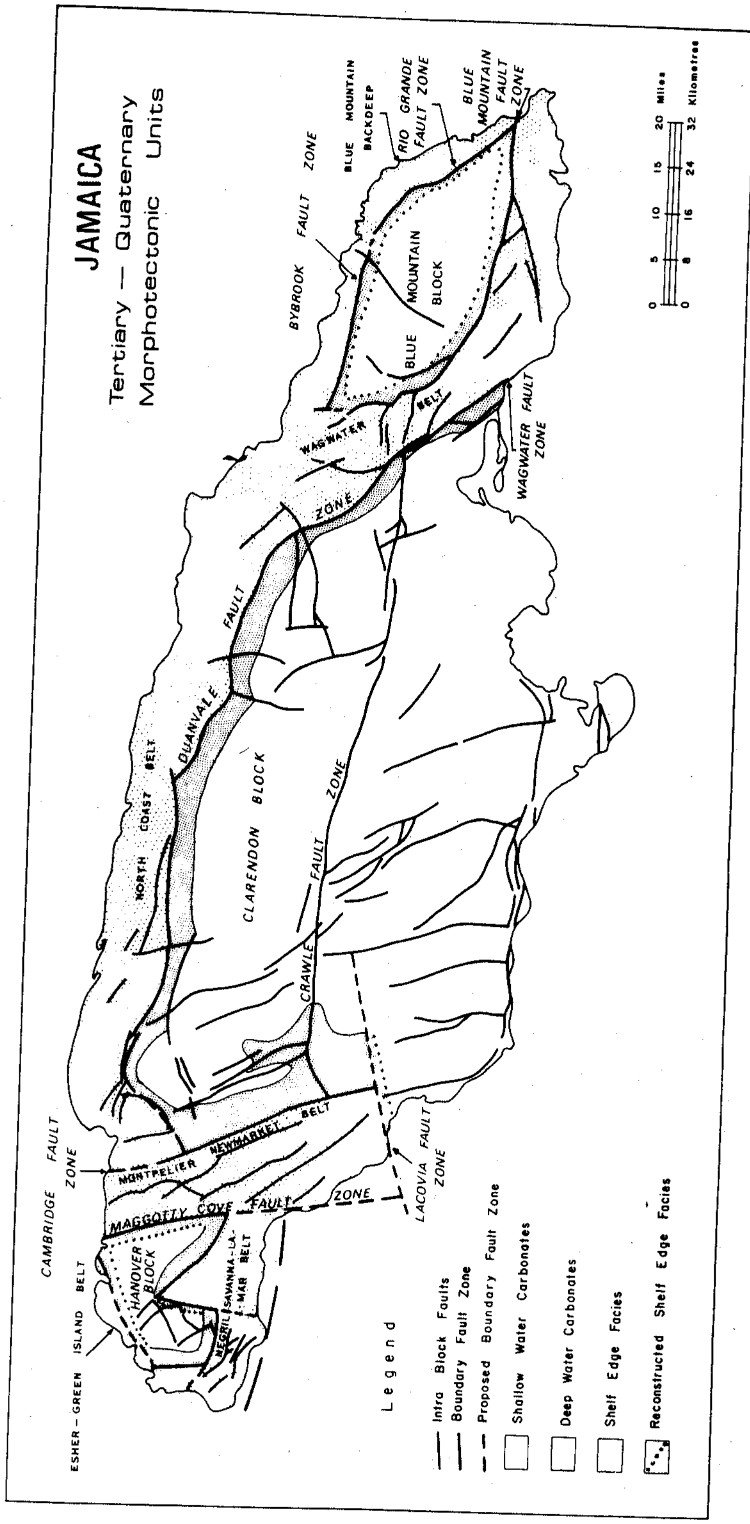

Figure 2. Morphotectonic units of Jamaica.

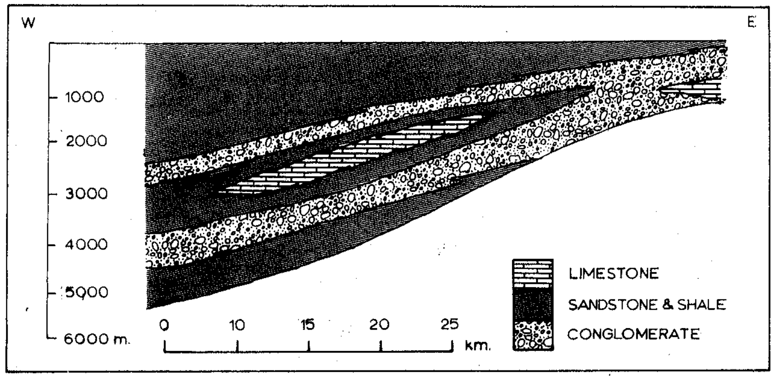

Figure 3. Diagrammatic section through the Cretaceous backarc basin and shelf sediments at the end of the Campanian.

Cretaceous submarine fan system of western Jamaica, but their reservoir potential is little known. Whilst good reservoir rocks have been described from many submarine fans, both modern and ancient, almost all of these have been situated offshore of continental areas. The precise nature of submarine fan sediments offshore from an island-arc system is still little known, and unfortunately Grippi (1978) gave no detailed grain-size data in his report.

The sediments of the Cretaceous backarc basin are not highly deformed, and the folding and faulting are more likely to have generated structural traps than to have created major escape routes for petroleum. The present erosion level, however, is such that a large part of the submarine fan system has been destroyed, so that even if significant petroleum deposits were formed, many of these must by now have been destroyed also. Any future exploration associated with the fan complex should, therefore, concentrate on those areas to the south, and possibly to the east, of the Hanover Block. There it can be expected that reasonably thick sections of the fan are preserved below a cover of Tertiary limestones.

Forearc basin and trench

Description: Sediments and volcanic rocks which originally accumulated in forearc basin and trench environments are now present in the Blue Mountain Block of eastern Jamaica (Fig. 2), the largest and least understood of the Cretaceous inliers on the island. This account is based on the recent works of Krijnen and Lee Chin (1978) and Wadge and Draper (1978).

The oldest exposed rocks are of Campanian age and consist of a series of volcaniclastic, arc-derived, sediments overlain by a thin shallow-water limestone. These are overlain (?) by a thick Campanian/Maastrichtian succession of immature conglomerates, sandstones, and shales, together with volcanic rocks and minor cherts and limestones. Altered carbonaceous material has been found in the sandstones and shales, but only in small amounts. Most of the sandstones and shales are arranged in a flysch-like sequence, but their sedimentology has not yet been studied in any detail. The entire sequence is probably 5+ km thick, and is intruded by granodiorite/tonalite bodies, and broken up by block faulting on a large scale.

Petroleum Potential: Readers are referred to Dickinson and Seely (1979) for a review of forearc regions and a detailed evaluation of the petroleum potential of forearc basins and trenches. Whilst they are able to cite examples of fossil forearc basins which produce petroleum, it seems unlikely that such will be found in Jamaica. Owing to the low heatflow near to the trench, overburdens of perhaps 3–6 km would be needed for petroleum generation to begin and this would result in a great reduction of the porosity of the immature sediments. In any case, the granodiorite intrusions, the large number of major faults, and the present deep erosion level are all likely to have destroyed any petroleum that might have been generated and trapped.

LOWER TERTIARY SEDIMENTARY BASINS

Active marginal basin

Description: The interpretation of the Wagwater Belt (Fig. 2) as an interarc basin (= active marginal basin) was first made by Jackson (1977), and has since been expanded upon by Jackson and Smith (1979). The description which follows is based mainly on the work of Green (1977), Jackson (1977), and Wadge and Eva (1979).

The Middle Eocene to Middle Miocene rocks in the Wagwater Belt are the deep-water facies of the Yellow and White Limestone Groups, which are also seen in the North Coast and Newmarket-Montpelier Belts. The characteristic feature of the Wagwater Belt, however, is the Lower Eocene and possibly Palaeocene section, which consists of clastics, volcanics, evaporites, and minor limestones. These have been given the name 'Wagwater Group' (Green 1977).

The clastic sediments are of two main types: crudely bedded red conglomerates and sandstones forming the Wagwater Formation, and a mainly well bedded flysch-like sequence of sandstones and shales, with minor conglomerates, forming the Richmond Formation. Green (1977) proposed a general stratigraphy with the lower part of the Wagwater Group consisting mainly of Wagwater Formation approximately 1200–1600 m thick. The middle 'subgroup', Green suggested, consists of alternations of Wagwater and Richmond Formations, 450–3000 m thick, and is overlain by an upper subdivision, 650–1000 m thick, consisting mainly of Richmond Formation (Fig. 4). Intercalated at several horizons are basalt flows which vary in thickness from 25–600 m. Locally present are lenses of limestone (e.g. Woodford Limestone, Fig. 4) which reach a maximum thickness of approximately 400 m, and gypsum/anhydrite deposits up to 60 m thick.

The palaeogeography of the Wagwater Belt during the Lower to Middle Eocene was complex. There was essentially a narrow, deep sea in a fault-bounded trough, with mountains shedding detritus into the trough from both the east and west. The Richmond-Wagwater sequence probably represents a series of complex delta/submarine fan systems which often overlapped and interfingered with one another. At least parts of the sea were probably landlocked on several occasions, but the exact details are not yet known.

Petroleum Potential: The Richmond Formation contains a reasonably large proportion of marine mudstones and shales, often containing thin horizons of lignite (Robinson 1976),

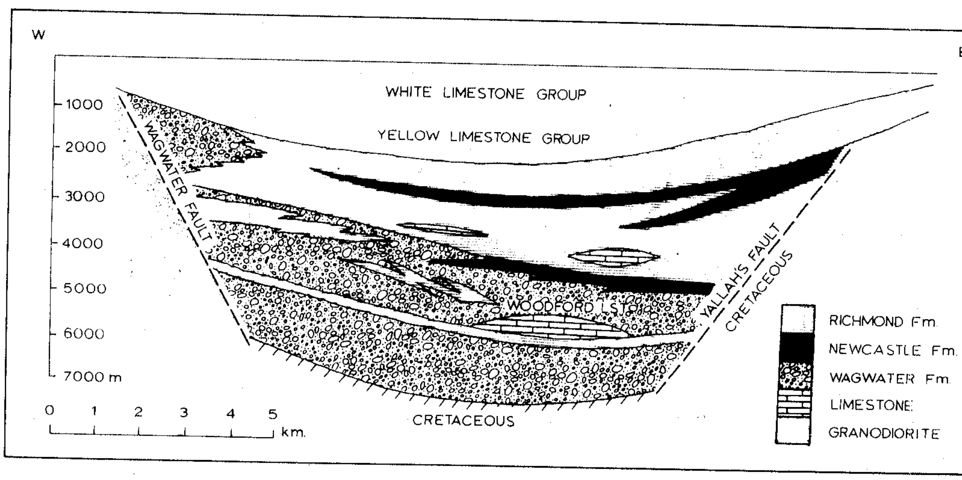

Figure 4. Diagrammatic section through the Tertiary Wagwater Belt — an infilled active marginal basin — just prior to the Middle Miocene uplift.

which locally could have acted as source beds and/or seals. In general the sandstones and conglomerates of the Wagwater Formation are very immature, poorly sorted, and probably of low reservoir potential. Locally, however, coarse sandstones and conglomerates, probably representing channel infillings may provide small stratigraphical traps of reasonable porosity. Similarly, the little-studied limestone members, some of which have been described as 'reef-like' (Meyerhoff and Kreig 1977), could serve as small stratigraphically trapped reservoirs, whose subsurface positions, however, would be difficult to locate.

Many of the sediments in the Wagwater Belt are likely to have been heated to considerable temperatures. Heat flow is relatively high in active marginal basins (Fig. 1), and may have reached 50°C per km. Therefore, if in pre-Middle Miocene times the Richmond Formation was buried to depths of 1–6 km (Fig. 4) it is likely to have been heated to temperatures of 50–300°C. Petroleum might, therefore, only be expected in the shallower parts of the basin — much of which has been lost by erosion since Middle Miocene times. The structure of most of the Wagwater Belt is complicated by both strong folding and faulting, which make detailed structural interpretations difficult and may have provided escape routes for any generated hydrocarbons. Overall the Wagwater Belt would seem to have a very low petroleum potential.

One possible exception is the south-eastern part of the belt, which is not as deeply exposed as the central and northern portions. It does not appear to contain great thicknesses of volcanic rocks, and the structure of the area is less complex. This sector of the belt may, therefore, be regarded as having slight petroleum potential (Meyerhoff and Kreigh 1977a, b).

Inactive marginal basins

Description: The North Coast Belt, Montpelier - Newmarket Belt, Esher - Green Island Belt, and the Negril Savannah-la-Mar Belt (Fig. 2) are all examples of infilled inactive marginal basins, which developed during the early Tertiary. They are of special interest to petroleum geologists because one of them, the North Coast Belt, contains the only known petroleum seep in Jamaica. The seep consists of natural gas containing approximately 84% methane (Greiner 1965): the source of the gas has not yet been positively identified.

Although there are differences in detail between the history of each of these basins, in general terms their Tertiary development is remarkably similar. The Montpelier - Newmarket Belt can be considered as typical and is the one chosen for discussion because its

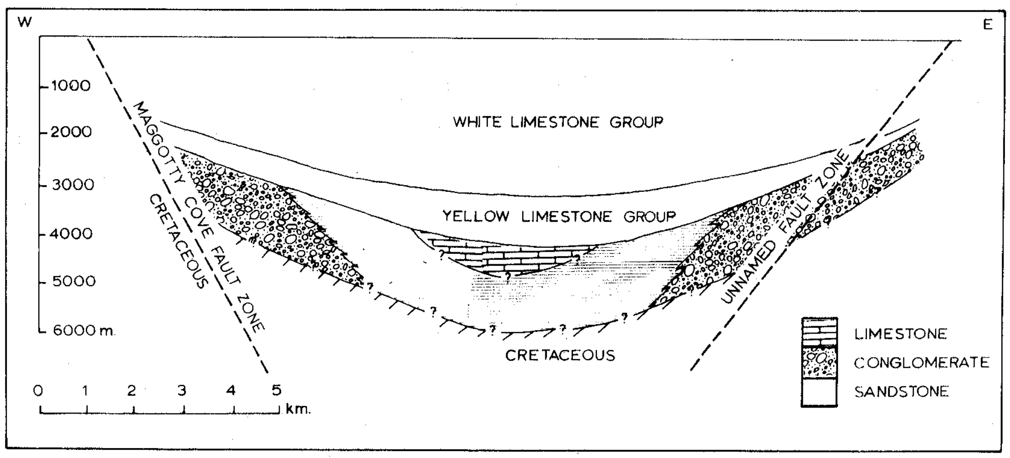

Figure 5. Diagrammatic section through the Newmarket-Montpelier Belt — an infilled inactive marginal basin — just prior to the Middle Miocene uplift.

geology has been studied more closely than the others.

The pre Yellow Limestone (= pre *Globorotalia lehneri* zone) Tertiary rocks in this basin are at present known only from the Content 1 exploration well. The well penetrated 500 m of sandstones, conglomerates, and mudstones but did not penetrate the Cretaceous sequence, which was its original objective. The clastic rocks are thought to represent alluvial and possibly deltaic deposits (Eva 1976), and they were deposited at a time when most of Jamaica was above sea level. Almost all the clasts are volcanically derived, often poorly sorted, and now tightly cemented. No marine fauna has yet been found in them, and there is very little plant material present.

The overlying Yellow Limestone Group begins with a series of more than 30 lignite horizons, ranging from 2 cm to 30 cm in thickness, and passes upwards into impure limestones and minor mudstones and shales with a fully marine fauna. Lateral equivalents of the lower part of the Yellow Limestone Group on the Clarendon Block (Fig. 2) include well sorted quartzo-feldspathic sands with lenses of limestone (Eva and McFarlane in press), and these might also be present within the basin itself. A diagrammatic interpretation of the Tertiary facies in the Newmarket-Montpelier Belt, prior to the Middle Miocene, is shown in Fig. 5.

Petroleum Potential: The source rock potential of this basin appears to be much better than any of those previously described. There is a fully marine sequence including limestones and shales, approximately 4 km in thickness (Fig. 5), which may have acted as source rocks together with the lignite beds at the base of the Yellow Limestone Group. To act as source beds the Lower and Middle Eocene sediments would have needed to attain a threshold temperature of approximately 90°C for petroleum generation to have taken place (Connan 1974). This would necessitate an overburden of only 3–4 km, which is geologically reasonable (Fig. 5). The report by the operators that fresh gas was smelt when many of the Content 1 cores were broken is therefore consistent with the known geology.

Most of the sediments in the Content 1 well are very well cemented, and with the pressure of overburden many of the impure limestones have developed pressure solution contacts. Porosity values are therefore rather low, and future explorationists can only hope that better reservoir rocks are present elsewhere in the basin.

The source of detrital sediment from both

TABLE 1

Tentative assessment of the petroleum potential of different palaeoenvironments found in Jamaica. The table is based on data discussed in the text.

	Forearc Basin	Trench	Backarc Basin	Active Marginal Basin	Inactive Marginal Basin	Shelf
Seals	Fair	Fair	Fair	Fair	Fair	Fair
Structural Traps	Poor to Fair	Poor to Fair	Good	Poor to Good	Good	Poor to Fair
Stratigraphical Traps	Fair	Fair to Good	Good	Fair	Fair to Good	Fair to Good
Reservoir Potential	Poor	Bad to Poor	Poor	Poor to Good	Poor to Fair	Fair to Good
P–T Conditions	Poor to Fair	Poor to Fair	Fair	Bad to Good	Good	Bad to Poor
Source Rock	Poor to Fair	Bad	Bad to Poor	Poor to Good	Fair to Good	Good to Excellent

sides of the basin (Fig. 1), and the gradual transition into a marine sequence, mean that stratigraphical traps can be expected. The structure of the basin is fairly simple, not greatly complicated by major faults, and with several large anticlines, such as the one on which Content 1 was located.

In summary this basin appears to have a good petroleum potential, if suitable reservoir rocks are present. An added advantage of drilling in this basin is that the Cretaceous rocks probably correspond to the backarc submarine fan complex, which itself has some potential (see discussion on 'backarc basin' above).

CONCLUSIONS AND SPECULATIONS

This paper has examined the major sedimentary basins in Jamaica in terms of their position in an ancient island arc setting. Although many of the palaeoenvironments have some positive features as regards their petroleum potential, they all appear to have one or more negative features, which would seem to seriously limit their potential. The overall analysis suggests that the inactive marginal basins are the most attractive exploration targets, followed by the backarc basin and shelf environments (Table 1).

It is tempting, but clearly not yet meaningful, to try to extrapolate these conclusions, based on the study of one small island, to inactive island arcs in general. It is hoped that as more data are collected from other regions such extrapolations will be possible, so that the petroleum potential of other ancient island arc systems can be quickly assessed, at least in a general way.

ACKNOWLEDGEMENTS

I would like to thank Dr L Montadert who presented this paper at the CCOP/SOPAC conference, and Ms I J Luke for constructive editorial comments on the manuscript.

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REVIEW OF HEAT FLOW STUDIES IN THE EASTERN ASIA AND WESTERN PACIFIC REGION

SEIYA UYEDA

ABSTRACT

Heat-flow data in eastern Asia and the western margin are reviewed. The deep western Pacific region shows uniform sub-normal heat flow agreeing with the prediction from plate-cooling models. Trench-arc gaps are generally characterized by lower heat flow, back-arc basins by high heat flow. The exact position of the transition between the low and high heat flow zones is still uncertain. Actively spreading back-arc basins, such as the Mariana Trough, have highly variable heat flow, indicating hydro-thermal circulation in the new crust. Vast shallow-water areas in the region have not yet been explored, except for geothermal gradient in offshore oil wells. Several new techniques for heat-flow measurement in shallow seas are being developed in Japan. Heat flow in the land areas in the region has not yet been studied except in Japan, Sumatra, Thailand, Korea and some parts of far-east USSR and eastern central China. A CCOP-IODE Project to vitalize heat-flow studies in the region is underway.

INTRODUCTION

Terrestrial heat flow is one of the few observable geophysical quantities pertaining to the thermal state of the earth's interior. It provides information important for understanding the earth's interior, global heat budget and regional tectonics. Heat-flow data are also useful for various practical purposes such as the assessment and exploration of geothermal-energy, metal and petroleum resources.

A great many heat-flow measurements have been made during the last two decades. The number of published measurements was only about 30 for the entire earth in 1952 (Birch 1954), but at present it is several thousands. Heat-flow data are compiled by the International Heat Flow Committee (IHFC) of the International Association of Seismology and Physics of the Earth's Interior (IASPEI). IHFC's recent global compilation of heat-flow data (Jessop *et al.* 1976) in the form of a list and map is available from the World Data Center A for Solid-Earth Geophysics, Environmental Data and Information Service, NOAA, Boulder Colorado, USA 80303.

Heat flow in oceanic areas is generally high and variable in the crestal regions of active mid-oceanic ridges (spreading centers) and low and uniform in oceanic basins away from the ridges (Lee and Uyeda 1965; Langseth and von Herzen 1970). This general tendency is interpreted in terms of the cooling of spreading oceanic lithosphere (McKenzie 1967, Turcotte and Oxburgh 1967, Sclater and Francheteau 1970, Parker and Oldenburg 1973, Yoshii *et al.* 1976). Although this line of interpretation predicts that oceanic heat flow, Q, should decrease with increasing age, t, of the lithosphere as $Q \propto 1/\sqrt{t}$, actual data are often widely at variance with this relation. As shown in Fig. 1 observed values are usually much less than the theoretical values near the ridge crest. Observed and theoretical values come to agree only at distance from the ridge crest — namely for older sea floor. This discrepancy between observation and theory has long been an enigma, but is now explained as due to hydrothermal convective heat transfer, in young oceanic crust (e.g. Lister 1972, Sclater *et al.* 1976, Anderson in press, a).

A similar relation has long been noted for the heat flow and the age of the crust, or the time since the last thermal event, in the continental areas (Fig. 2) (e.g. Polyak and Smirnov 1968, Hamza and Verma 1969, Sclater and Francheteau 1970). The time scale concerned here, however, is of an order of magnitude greater than for the oceanic heat flow, suggesting that the factors controlling the heat flow in the continental and oceanic areas are different. These relatively simple

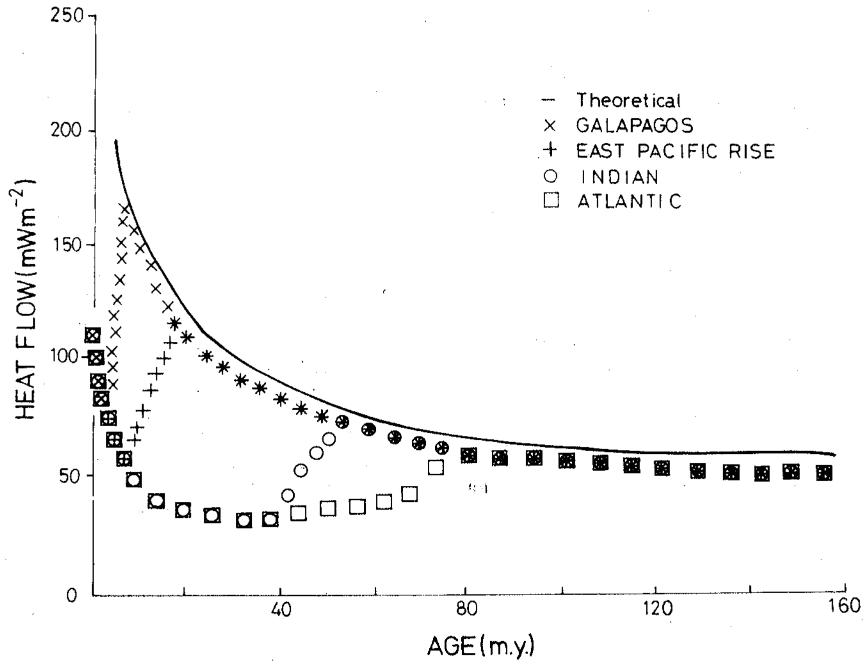


Figure 1. Variation in mean heat flow versus age in each of the major mid-ocean ridge segments (Anderson *et al.* 1979). (1 HFU = 1 μcal/cm^2 s = 42 mW/m^2).

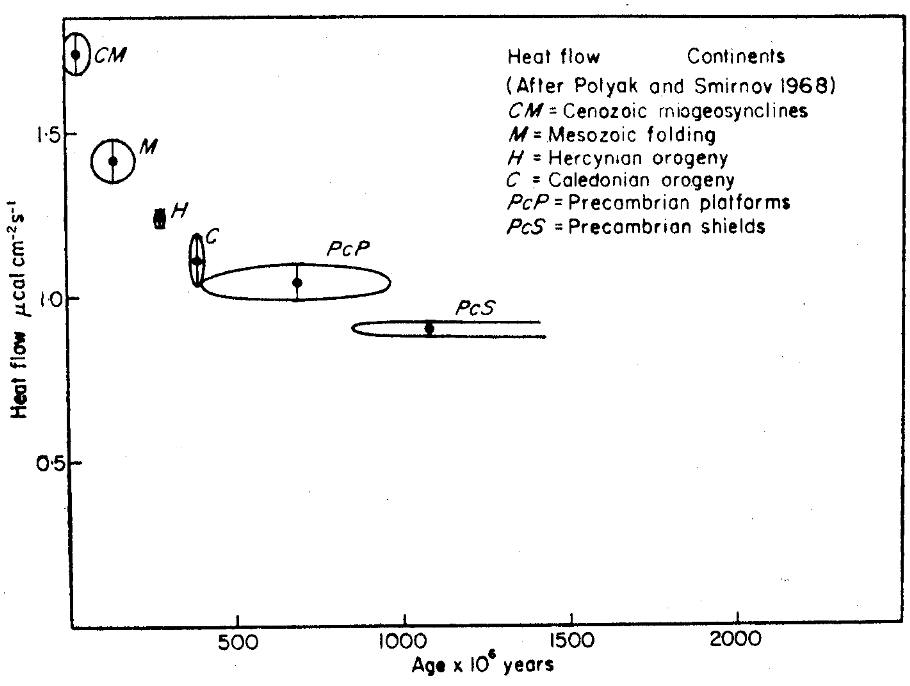


Figure 2. Mean heat flow against age of orogenic province for continents (Polyak and Smirnov 1968, Smirnov 1968).

relationships have been found for the more typical oceans and continents.

The western Pacific and eastern Asia constitute an area of extremely complex and active tectonics, and the elucidation of the processes involved is one of the most challenging targets of earth science today. However, the heat flow in most of this interesting area has not been sufficiently investigated (Fig. 3). Except for the regions in and around the Japanese, Kurile, Mariana and Ryukyu islands (Uyeda and Horai 1964, Yasui *et al.* 1967, 1968; Uyeda and Vacquier 1968, Yasui *et al.* 1970, Watanabe *et al.* 1977, Lubimova and Nikitina 1978, Smirnov and Sugrobov 1980) and several marginal seas (Nagasaka *et al.* 1970) and deep ocean basins, most of the area forms a large gap in the world's heat flow data distribution (see also 'A Geophysical Atlas of the East and Southeast Asian Seas' (Hayes 1978)). Notably, most of the land areas and shallow seas (such as the Sunda shelf) lack heat-flow data.

To remedy this situation, an attempt to foster heat-flow studies in the southeast Asian region was initiated in 1975 under the CCOP/IDOE program. In this project, a group of Japanese experts, under the sponsorship of the Japan International Cooperation Agency (JICA), visits various southeastern Asian countries with a set of down-hole temperature-logging apparatus and conducts initial measurements of geothermal gradient jointly with local scientists. The down-hole temperature-logging apparatus is donated to the appropriate agency of each nation visited, for later use by local scientists. The work is still in its preliminary stage and the results are still to be fully worked out.

DEEP-SEA REGIONS

Deep-sea regions, in this paper, are defined as the seas where the water depth is great enough for the bottom water temperature to be sufficiently stable to permit the use of conventional, short, heat-flow probes. Depending on the conditions the water depth that has to be exceeded appears to range from 1000 to 2000 m. For the areas dealt with in this paper, deep-sea regions include the western Pacific basin, trenches and deep back-arc basins. Fairly extensive marine heat-flow measurements have been conducted in these deep-sea regions by several institutions, including the Japan Meteorological Agency, the Earthquake Research Institute of the University of Tokyo, Soviet Academy of Sciences, Scripps Institution of Oceanography of the University of California and the Lamont-Doherty Geological Observatory of Columbia University.

Heat flow in the old western Pacific basins is generally uniform and low (1.0 – 1.2 HFU). This is in agreement with the theoretical prediction, based on models of a cooling rigid plate (e.g. McKenzie 1967, Sclater and Francheteau 1970) or of cooling half-space with a growing plate (e.g. Parker and Oldenburg 1973, Yoshii *et al.* 1976). Some details, however, would still have to be tested by further observation to determine which types of the above two models might be better (Parsons and Sclater 1977). The criteria for choosing one type of model from another would be as follows: the rigid-plate model predicts that the heat flow would approach an asymptotic value in the older ocean, whereas the half-space model predicts continuing decrease of heat flow with the age of the sea floor.

Heat flow in the marginal areas is more variable. Most of the published heat-flow data in these areas, as of 1978, are shown in the Geophysical Atlas of the East and Southeast Asian Seas (Hayes 1978). (This Atlas also includes the then existent land data, mainly from Japan, and the geothermal-gradient data from the southeast Asian oilfields (see next section, 'Shallow-water Regions')).

Heat-flow distribution in and around the Japanese island arc as of 1972 (Uyeda 1972) is shown in Fig. 4. As can be seen in this figure, heat flow seems to be somewhat lower in the Japan Trench region than in the adjacent Pacific Basin, whereas it is significantly high in the Japan Sea. The boundary between the outer low-heat-flow region and the inner high-heat-flow region cannot be delineated precisely with the data available. Uyeda and Horai (1964), however, noted that the boundary roughly coincides with the so-called front of active volcanoes, or seaward limit of active volcanoes on the arc. Such a distribution of heat flow was considered to be a characteristic of the active trench-arc back-arc systems. (A notable exception to the foregoing is the

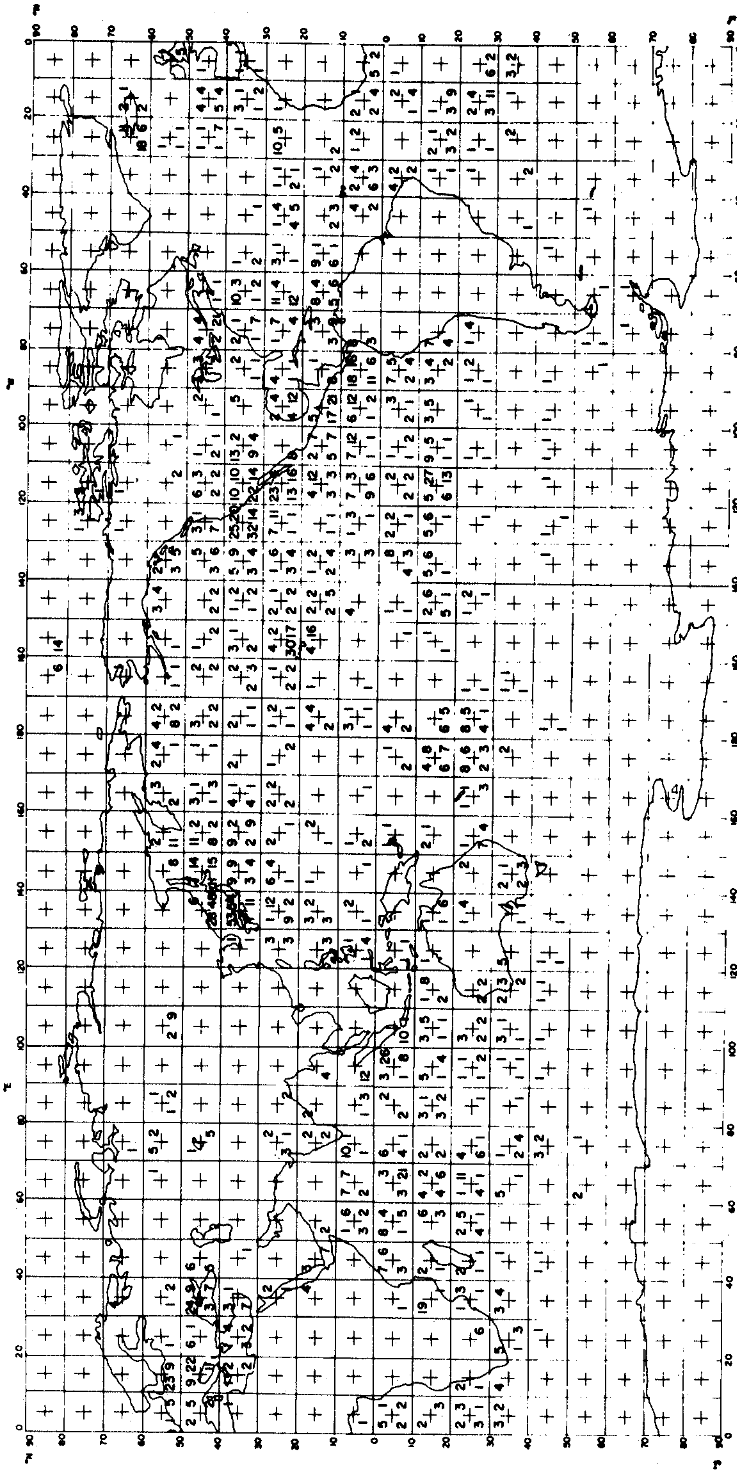


Figure 3. Distribution of heat-flow stations. The number of stations is given for each 5° of latitude by 5° of longitude. (Horai and Simmons 1969.)

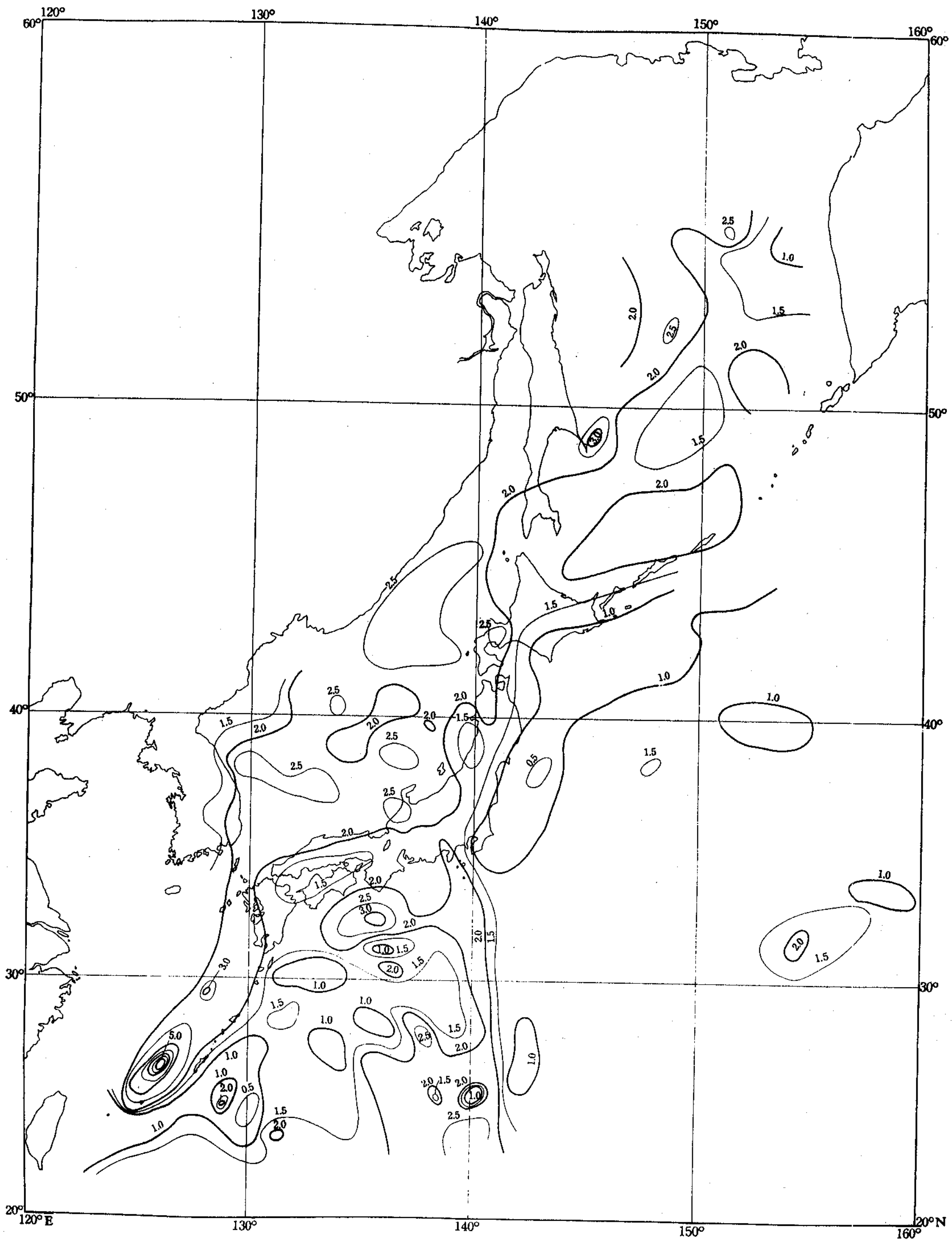


Figure 4. Smoothed contours of heat-flow values in and around Japan (Uyeda 1972).

Wankai Trough which shows a high heat flow, as seen in Fig. 4.) This view was supported by subsequently obtained data from the Kurile (Yasui *et al.* 1967, 1968, Lubimova and Nikitina 1968) and Ryukyu (Yasui *et al.* 1970) systems. Various models have been put forward to explain such a characteristic distribution of heat flow over a subduction zone (e.g., McKenzie and Sclater 1968, Hasebe *et al.* 1970, Oxburgh and Turcotte 1970, Anderson *et al.* 1976). Dehydration of subducted ocean sediments and crust and frictional heat between the slab and the upper mantle above it have been suggested as the main sources of outer negative and inner positive heat-flow anomalies, respectively. Convective flow in the upper mantle wedge above the subducted slab has been also invoked for the high heat flow in the back arc (e.g. Andrews and Sleep 1974). It should be noted, however, that subduction of cold oceanic plate is a very efficient cooling process. Therefore, in order to bring about the observed high heat flow in the back-arc region and arc volcanism, extremely high heat production and extremely efficient upward transfer of heat are needed (Hasebe *et al.* 1970). The latter condition, in particular, requires non-conductive heat transfer in the upper mantle under the active back-arc basins. This point may have an important bearing on the origin of back-arc basins, in that if they were formed by extensional processes (Karig 1971) similar to sea-floor spreading, the mass movement in the upper mantle associated with the process would provide the necessary means for the upward heat transfer (see, for example, Uyeda 1979). If such was the case, the heat flow in back-arc basins should depend on the age of the basin, just as heat flow of the ocean floor depends on the age of the crust.

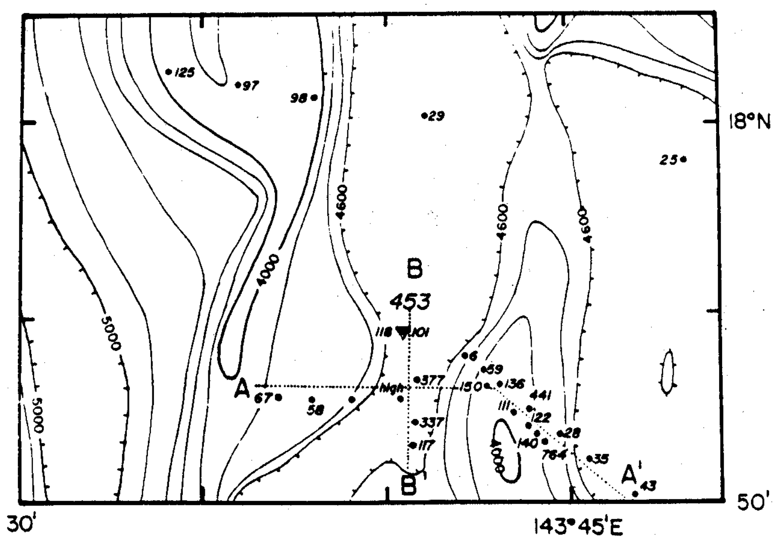
Watanabe *et al.* (1977) made a thorough review on the heat flow in back-arc basins. They postulated, on the basis of data then available that the heat flow / age relationship for back-arc basins is significantly different from that for the mid-ocean ridge: i.e., heat flow in back-arc basins tends to be higher than in the north Pacific basin for the same age. Anderson (in press, b) more recently reviewed the heat flow in the east and southeast Asian seas with more complete information on the ages of the back-arc basins, and noticed that the heat flow / age relationship for back-arc basins may not be different from that of mid-ocean ridges. If this is correct, since the depth / age relationships in the two cases are considered to differ significantly (back-arc basins are deeper for the age), the upper mantle under the back-arc basins should have a higher density than that under oceans, as was suggested by Yoshii (1973). This difference in the upper-mantle density, apparently not affecting the thermal structure, may be of importance in deciphering the possible difference, if any, in the processes of sea-floor spreading in back-arc and mid-ocean ridge situations.

More recently, Smirnov and Sugrobov (in press) have presented a unified heat flow / age relation for the northwestern transition zone between Asia and the Pacific.

Other important discoveries relating to heat flow in the back-arc basins may be the highly variable heat-flow distribution (Uyeda and Horai 1980, Hobart *et al.* 1979) and evidence for the hydrothermal processes (Natland 1980) in the floor of the actively spreading Mariana Trough (Hussong, Uyeda *et al.* 1978). Figure 5 shows some examples of spatially changing heat flow observed in the 5 m.y.-old part of the Mariana Trough near IPOD site 453. The situation is considered to be the same as that reported for the Galapagos spreading center area (Williams *et al.* 1972). The present author believes that vigorous hydrothermal activity as observed by submersibles in the Galapagos and East Pacific Rise (21°N) areas (Nat. Geogr. 1979) is taking place in the crust of the Mariana Trough, and submarine metallogenetic processes similar to those that produced the 'Kuro-ko' type massive sulphide may be operating there also (Uyeda and Nishiwaki 1980). Elucidation of these possibilities will constitute one of the major targets in future study in this area.

The new marine heat-flow probe that made possible such closely spaced measurements as shown in Fig. 5 allows multipenetration of the bottom and retrieval of real-time data. The tool was developed by R P von Herzen at Woods Hole Oceanographic Institution and C B Lister of the University of Washington. Known as 'POGO', it has already been in use for the Galapagos (Williams *et al.* 1972) and other areas (e.g. Anderson *et al.* 1979, Hyndman *et al.* 1979) and proved most useful for detailed heat-flow studies.

An important suggestion was made recently by Anderson (in press, a), namely that the heat flow may not be really low in the trench-arc gap, as commonly assumed. Instead, heat flow may start to rise in the landward half of the trench-arc gap. Such a situation may be suspected from the observation (Yoshii 1979 a, b) that the upper-mantle aseismic front, which presumably is the seaward limit of high-temperature upper mantle under Japan, is significantly displaced seaward of the front of volcanoes (Fig. 6). Recently, discoveries have been made on the existence of arc volcanism in the fore-arc region of the Japan and Mariana Trenches through IPOD drilling (von Heune, Nasu *et al.* 1978, Hussong, Uyeda *et al.* 1978). Volcanism in the area seaward of the present front of volcanoes seems difficult to explain by the current models for generation of arc magmatism. One way would be to call for tectonic erosion of the overriding plate as bringing the site of volcanism towards the trench. But at the same time, it may be worth examining the possibility that at some early stage of the subduction episode the thermal



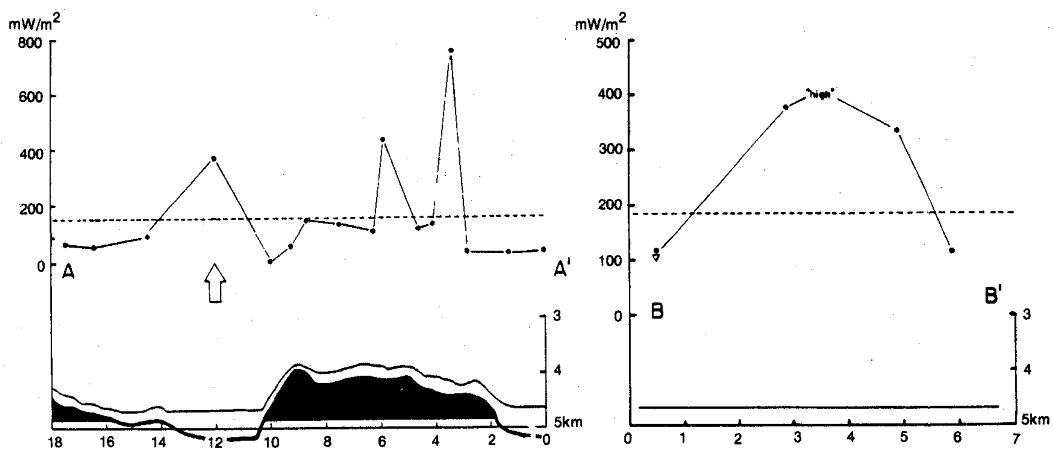


Figure 5. Examples of heat flow profiles in the Mariana Trough (after Hobart *et al.* 1979). A-A′, B-B′ in lower figures correspond to the survey lines in the top figure. Heat flow in mW/m^2. (1 HFU = 42 mW/m^2). The results are based on preliminary shipboard data reduction.

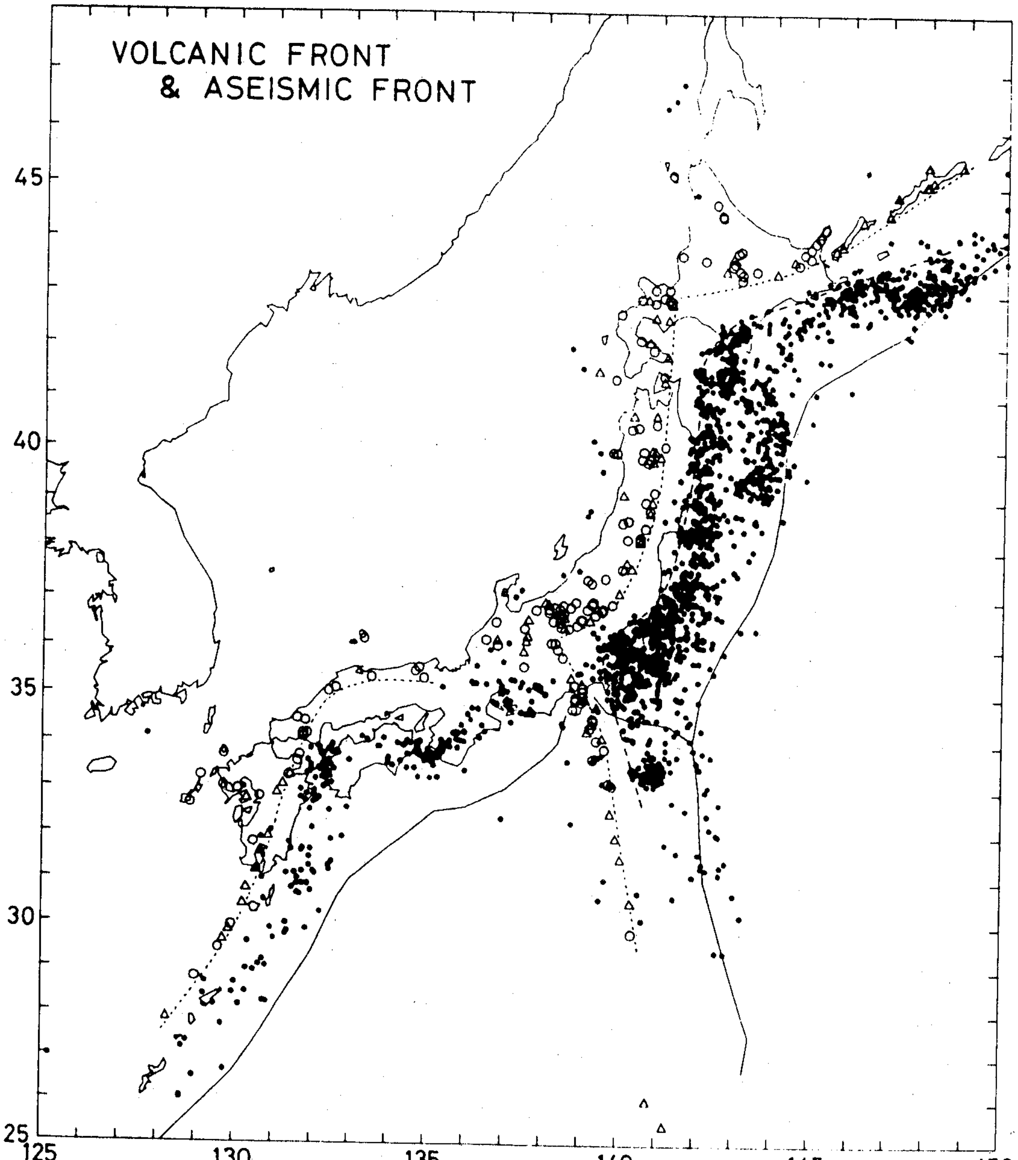


Figure 6. Position of Quaternary volcanoes and earthquakes with depths of 40-60 km in the Japanese area. Dotted and dashed curves indicate the 'volcanic front' and the 'aseismic front' respectively. (From Yoshii 1979b).

state under the fore-arc region might have been favourable for the production of a certain type of magma, such as boninite.

However, the data relevant to this problem are still too few (see Figs. 3 and 4). Precise determination of heat-flow profiles over trench-arc gap areas would be the first step to take. Here, again, the multipenetration heat-flow probe mentioned above would prove useful for the deep-water part of the trench-arc gap.

To make a complete heat-flow profile, both land and shallow-water parts must also be investigated. For the heat-flow measurements in shallow seas, some new techniques are being developed in Japan, as briefly described in the next Section.

SHALLOW-WATER REGIONS

The eastern asia and western Pacific region has vast continental shelf areas (Fig. 7) where the water depth is shallow and conventional deep-sea heat-flow techniques cannot be used because of the unsteady bottom-water temperature. These shelf areas are very important in the tectonic development of continental margins, and the heat flow in these areas deserves much attention. The problem of trench-arc gaps touched upon in the previous section is one example. Heat flow in the shelf areas is especially important to hydrocarbon maturation.

A number of oil wells attest to the importance of the shelves. The Southeast Asian Petroleum Exploration Association (SEAPEX) and Indonesian Petroleum Association (IPA) have compiled geothermal gradient data from many wells and published a Geothermal Gradient Map of Southeast Asia (1977). From the map one can infer, to some extent, the distribution of heat flow in the areas where oil wells exist. Outstanding features on the map are as follows:

1. There is an extended region of high geothermal gradient (maybe also high heat flow) in the Gulf of Thailand, eastern Sumatra, and southwestern Java Sea.
2. The region from Luzon to the eastern Java Sea, through the offshore areas of Borneo, and the fore-arc basins of the Indonesian arc, are characterized by low geothermal gradient (or low heat flow).

For the purpose of geophysical investigation, however, it would be desirable, to obtain the direct heat-flow values from these data. Of course, geothermal gradient can be converted to heat flow if the thermal conductivity of the strata concerned is known. Matsubayashi and Uyeda (1979), with the cooperation of EXXON Malaysia and PETRONAS, as a part of the CCOP-IDOE Heat Flow Project mentioned in the Introduction, did this by measuring the thermal conductivity of the core specimens recovered from some oil wells off Malaysia (Fig. 8). It is hoped to increase the heat-flow data in a similar way. As cited in the next section, da Silva Carvalho *et al.* (in press) carried out this type of work more extensively for the onshore oil fields in Sumatra.

This line of data collection has an obvious limitation, that study can be made only where oil wells already exist. In order to investigate the regional distribution of heat flow in the extensive shelf areas where there are no oil wells, it is necessary to develop a new technique by which heat-flow measurement in shallow sea can be made. Three main lines of approach are being followed in Japan at present, under a joint program of the Earthquake Research Institute, University of Tokyo, Department of Earth Sciences of Chiba University, Geological Survey of Japan and the Hydrographic Office of Japan.

The first is to insert into the bottom sediments a several-metres-long probe attached to a long-term temperature-data recorder. It is planned to record, over a long period, the time-variation of bottom-water temperature as well as of sub-bottom temperatures at several depths in the probe. The recorder will be recovered by an acoustic command from the surface at the time of data retrieval. This system is still being tested by the Geological Survey of Japan.

The second line of approach is to make the measurement of the complete temperature profile in a relatively long probe and to eliminate numerically the effect of changes in bottom-water temperature (Matsubara *et al.* in press). In this new technique, the temperature in the sediment is to be measured at every 10 cm to a depth of 7 m (so far only to 5 m) by one thermistor which moves up and down in the probe. From the temperature profile thus obtained, the geothermal gradient is calculated, in the least squares sense, by solving the thermal-conduction equation, i.e. by assuming the annual sinusoidal temperature variation at the sea bottom, of which amplitude and initial phase are parameterized. This system was used in the Gulf of Thailand in 1978 under the CCOP-IODE Heat Flow Project on the Thai Fishery Research Vessel No. 1. Examples of the temperature profile are shown in Fig. 9. Analysis of the data, however, indicated that more reliable results would be obtained if the probe were 7 m long rather than 5 m (Matsubara *et al.* in press). Now, the Japan

Figure 7. Shallow areas in east and southeast Asia. Hatched area shows the area shallower than 200 m in water depth.

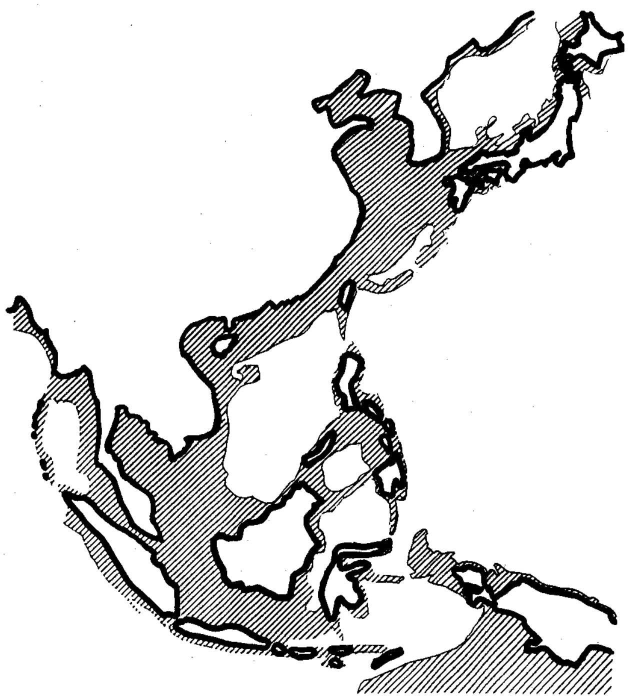
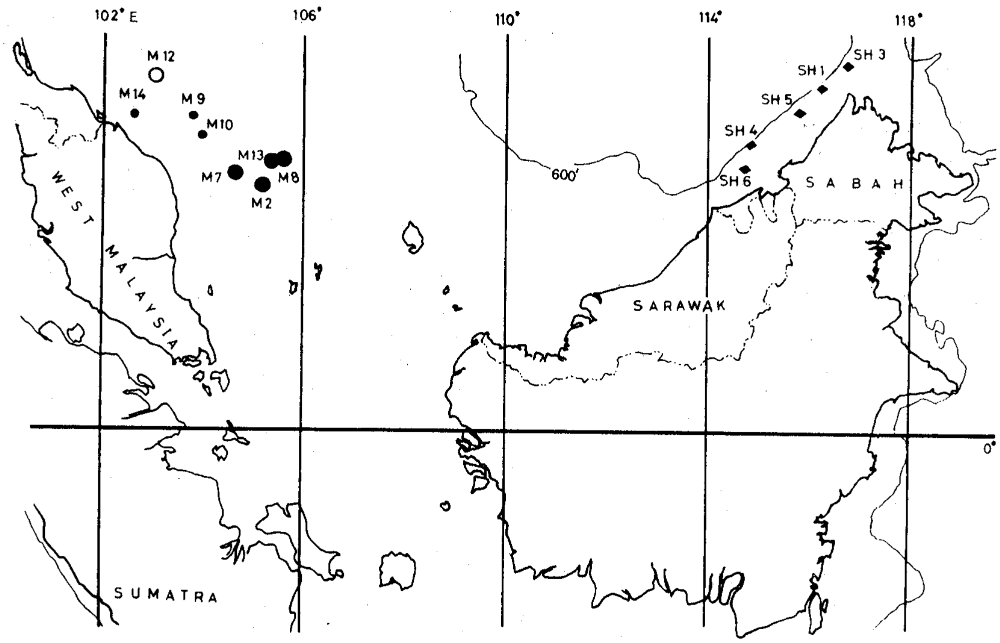


Figure 8. Heat flow in the shallow sea area off Malaysia, deduced from oil-well gradient data: rhombs indicate heat flow less than 60 mW/m^2, small circles 60-80 mW/m^2, large solid circles 80-100 mW/m^2 and large open circles more than 100 mW/m^2 (Matsubayashi and Uyeda 1979).



National Oil Corporation has become interested in this technique, and it is planned to improve the probe and make further measurements in the continental shelves around Japan.

Although this technique is promising, it will not work unless the probe fully penetrates the sediment. Sandy bottom near the land is not always easy to penetrate by simple free fall of the probe. To overcome this problem, the third line of approach is being tried by the Hydrographic Office of Japan, i.e. the vibro-system for penetration (Katsura *et al.* 1980). It is hoped that by combining these new techniques extensive heat-flow measurements in shallow seas will be conducted in the near future.

LAND AREAS

In the land areas of eastern Asia, the Japanese islands have been fairly well surveyed (Uyeda and Horai 1964, Honda *et al.* 1979). Even in this area, however, heat-flow data are now considered insufficient in comparison with other geophysical and geological information, which has been accumulating rapidly in the last decade. The greatest problem in the heat-flow work on land is the non-availability of deep holes. Even when a hole is available, temperature measurements have to be conducted before it is collapsed. Holes often cave in when the casing is removed. But if the measurement is made while the casing is still in the hole, i.e. soon after the drilling, the temperature remains disturbed by the drilling operation. In order to overcome this difficulty, we started sometime ago to bury the cable with the thermistors in the hole before the casing is removed and make the temperature measurements at later times (Uyeda *et al.* 1976). This method permits one to follow the decay of the drilling effect if one performs repeated measurements. One drawback of the method was that the number of thermistors buried is limited by the number of electric conductors of the cable, so that detailed temperature profiles are difficult to obtain. To solve this dilemma, we have recently developed a new technique whereby one can put, theoretically, an infinite number of thermistors in the hole by using only six conductors (Fujisawa *et al.* in press). With this new tool, the land heat-flow program in Japan is underway with the cooperation of the Japan Metal Mining Agency, which drills about 10 deep holes in Japan every year for regional geological surveying (Honda *et al.* in press).

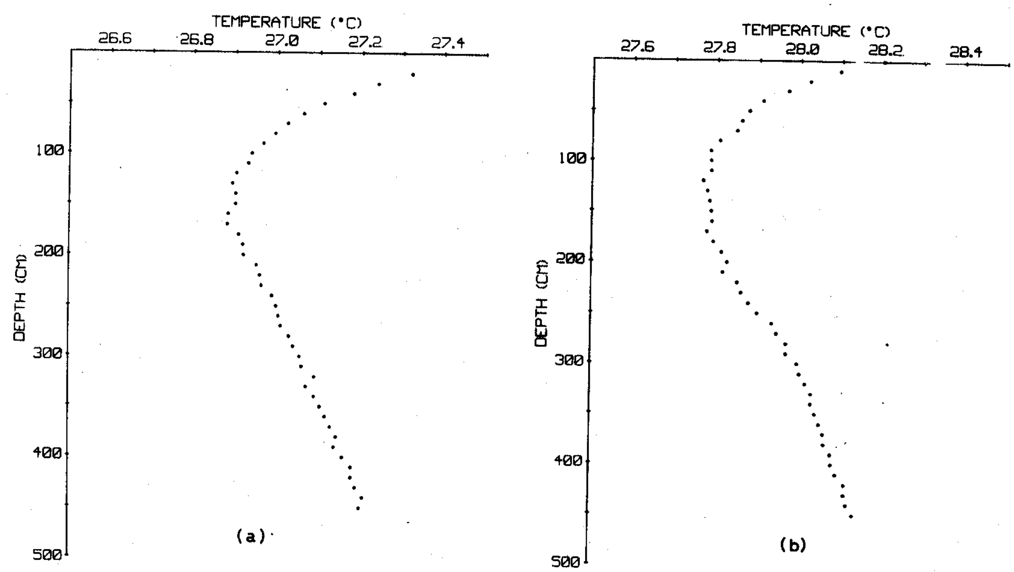


Figure 9. Temperature-depth profiles taken in the Gulf of Thailand; (a) station AM1, (b) station AM2. (Matsubara *et al.* in press).

Heat-flow data on continental east Asia are relatively few. To the author's knowledge there are some for far-east USSR (Lubimova and Nikitina 1978) and for Korea (Mizutani *et al.* 1970). Recently the author has been informed of extensive heat-flow measurements in eastern China by the Geothermal Group, Geological Institute, Science Academy of China. Some of the results will be published shortly (Wang in press).

In southeast Asia, as mentioned in the Introduction, a CCOP-IODE Heat Flow Project is underway. As of the time of writing this report, work has been initiated in Thailand, Malaysia, Indonesia, Philippines and Papua New Guinea. The state of progress is summarized as follows:

Thailand

Heat-flow work has been conducted in cooperation with the staff of Thai Department of Mineral Resources (Thienprasert *et al.* 1978). It was found (Fig. 10) that the Khorat Plateau — framed part in Fig. 10 — has normal heat flow, but there is a belt of high heat flow in western Thailand, running approximately north-south in the extension of the Gulf of Thailand. In the Mae Sot area, west of the Khorat Plateau in Fig. 10, there are hot springs, and some areas are under exploration for geothermal energy (Ramingwong *et al.* 1979).

Malaysia

The Geological Survey of Malaysia has been participating in the Project. Since the great tin resources in this country are in surficial deposits and require no deep drilling for exploration, holes are scarce. In 1976, heat-flow measurements were made in the Sungei Lembing mine, Kuantan, on the east coast of the Malayan Peninsula. Preliminary results indicate a high heat flow. Since more exploration-drilling programs are planned for the granitic rocks of the Peninsula (Choon Seng Ho, personal communication), it is hoped that the holes can be used for heat-flow studies.

As mentioned in the previous section, offshore oil wells have been useful for heat-flow estimation in the shallow seas (Matsubayashi and Uyeda 1979).

Indonesia

In 1976, heat-flow measurement was begun, with the staff of the Geological Survey of Indonesia, in a 12-day field trip visiting 15 sites in Java (Fig. 11). All but two sites were water wells and tended to be insufficiently deep. At one of the metal-exploration sites (numbers 10 and 11 in Fig. 11) a hole 256 m deep was available. Since then, more extensive measurements have been conducted (W. Soebroto, personal communication). The data are still to be worked out, but preliminary analysis indicates very high heat flow in the area near Jakarta on the Java Sea side of the island. It is hoped that the survey will be extended to other islands, such as Borneo, Suluwesi and Timor.

In Sumatra, H da Silva Carvalho and V Vacquier and the staff of PERTAMINA made an extensive study using abundant thermal data from oil fields (da Silva Carvalho *et al.* in press). Their results show clearly that the Tertiary basins in east Sumatra have high heat flow. The situation seems analogous to that on the Japanese arc in that the landward side of the arc has high heat flow. Although Sumatra has no oceanic back-arc basin, the zone of high heat flow is on the southern extension of the Andaman Sea, which is an actively spreading marginal sea (Eguchi *et al.* 1979).

Clearly, the heat flow study in the Java Sea would be of critical importance both from tectonic and economic viewpoints. There, the shallow-sea heat-flow techniques described in the previous section would be needed.

Philippines

Field trips were made in 1975-77 with the staff of the Philippine Bureau of Mines. More than 10 metal mines were visited in Luzon and Cebu islands and the results are under scrutiny by T Watanabe and others preparatory to publication. However, the geological structure of the Philippines is so complex — with multiple arcs and trenches and complicated distribution of volcanoes and earthquakes — that just a few scattered data are far from sufficient to draw any inference on the distribution of heat flow. Fortunately, the Philippines has numerous metal mines with considerable activity in exploration drilling. It

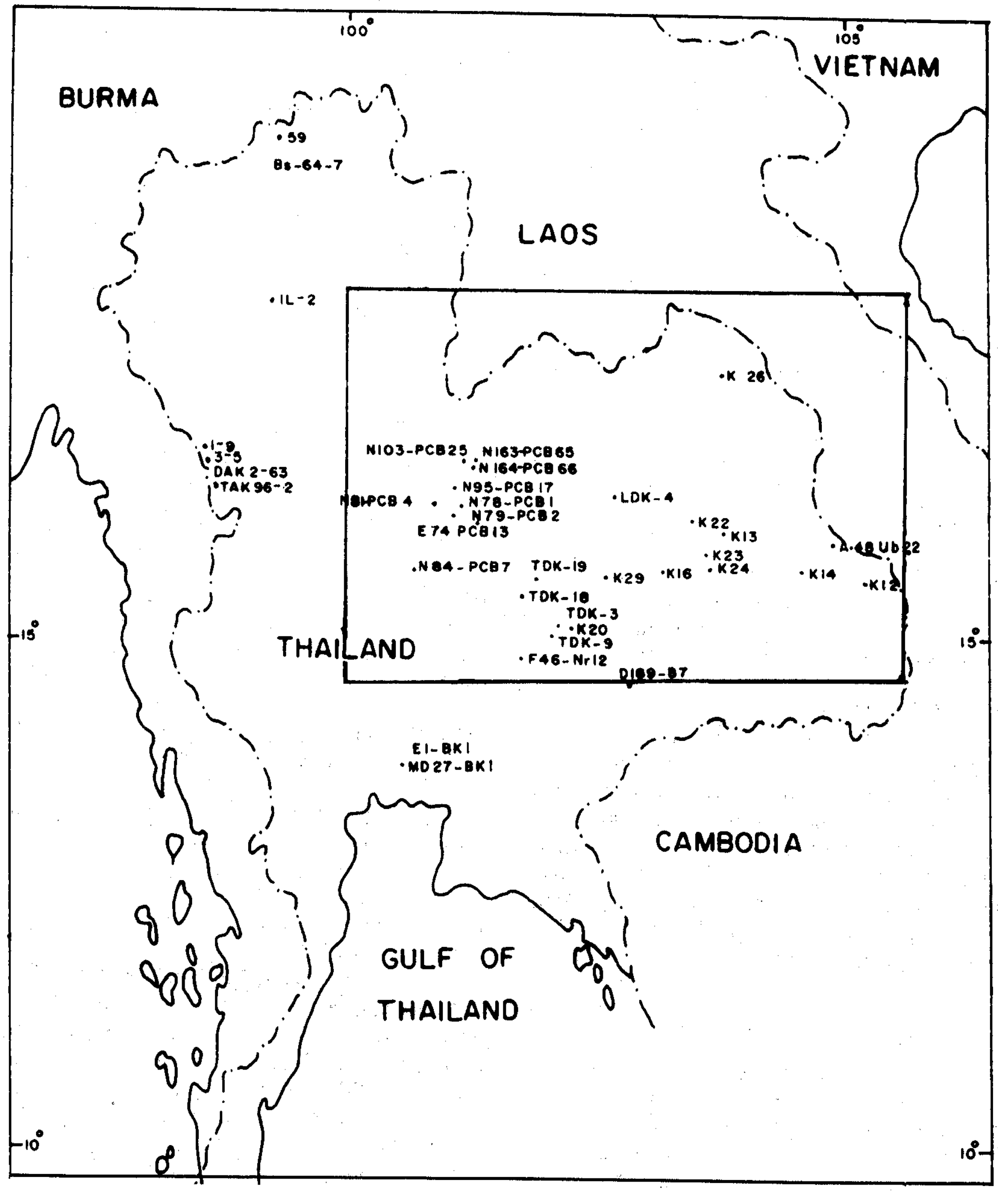


Figure 10. Heat flow stations in Thailand. The area within the rectangle (Khorat Plateau) has been intensively studied. (Thienprasert *et al.* 1978).

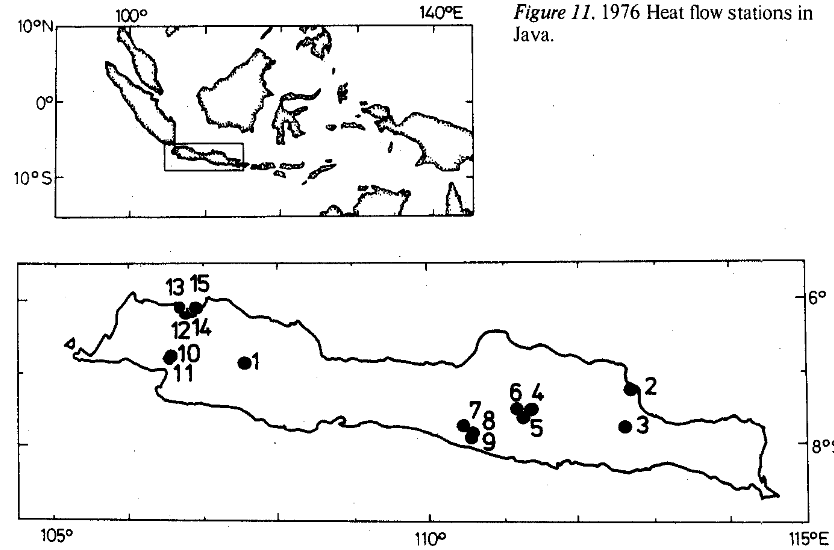


Figure 11. 1976 Heat flow stations in Java.

is hoped that much more work may be done in the near future, to match the rapid rate of data accumulation in other disciplines of earth science in the Philippines, in particular those included in the IDOE-SEATAR program. The use of thermal data from geothermal fields and oil wells should also be encouraged (E V Tamesis, personal communication).

Papua New Guinea

In 1977, a short field trip was conducted with the cooperation of the Geological Survey of Papua New Guinea. With considerable difficulty, three metal mines were visited. The results of the work (by H Mizutani and others; in preparation for publication) show normal heat flow. But again, much more data, including those from oil wells, are needed for this complicated terrain.

CONCLUSION

The current state in heat-flow investigations in the eastern Asia and western Pacific region has been reviewed. The deep-sea regions, where the conventional short-probe technique can be applied, have been studied by several institutions. The results from the western Pacific basin show subnormal values, generally supporting the plate-tectonic models. The trench and trench-arc gap areas show even more subdued heat-flow values. The zone on the arc landward of the front of volcanoes, and the back-arc areas, have generally high heat flow, and, in particular, the actively spreading back-arc basins (e.g., Mariana Trough) show indications of hydrothermal circulation in the crust. The transition from outer low-heat-flow zone to inner high-heat-flow zone appears to lie in the fore-arc zone. But its exact location is still uncertain. The Nankai Trough is a notable exception to the above rule, showing high heat flow.

For the shallow-sea areas, heat-flow information is meagre, and some new techniques are being developed to make measurement possible. Geothermal-gradient information, however, gives some clue to the heat-flow distribution in areas where oil wells exist.

Land heat-flow information for the region is also meagre, except from limited areas such as Japan, Sumatra, Korea, and some parts of the far-east USSR, eastern China, and Thailand, but some endeavors are underway through the CCOP-IDOE program in southeast Asia.

From the presently available data, the distribution of heat flow in the region may be inferred very tentatively, as in Fig. 12. The inference is based on heat-flow and geo-

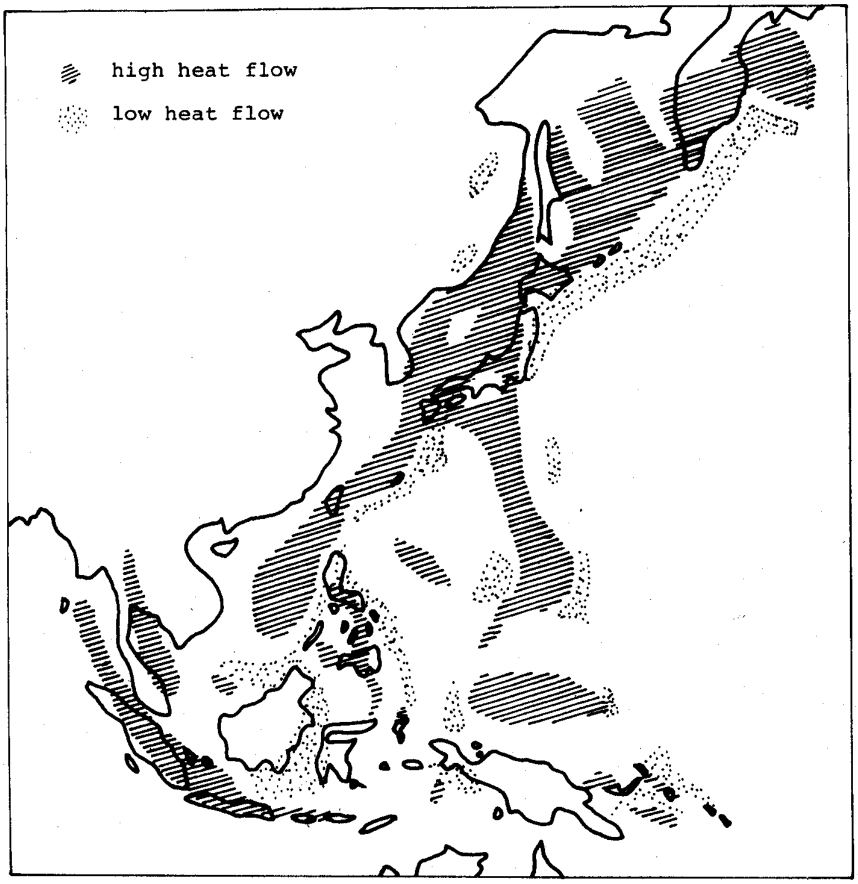


Figure 12. Tentative heat flow distribution in the western Pacific and east and southeast Asia inferred from presently available data. Blank areas are either of normal heat flow or no data. Most of the Philippine Sea is difficult to delineate because of scattered values.

thermal-gradient data and the distribution of active volcanoes. Areas of normal heat flow are not marked in the figure. Therefore, the blank areas are either of normal heat flow or no data: one notable exception to this statement is the major part of the Philippine Sea, for which there are considerable heat flow data but where values are so varied spatially that it is difficult to delineate heat-flow provinces.

ACKNOWLEDGEMENTS

The author is grateful to many colleagues without whose cooperation and kind offices the CCOP-IDOE Heat Flow Project would never have been initiated. I am especially indebted to Drs C Y Li, A Johannas, T Hilde, M Kouno, F Wang, L Hawkins and K Kawada of CCOP for the arrangement of the project. Dr S Sano and Mr H Sakai have also been most helpful in promoting the program. My thanks are also due to the directors and staff of the Thai Department of Mineral Resources, Geological Survey of Malaysia, Geological Survey of Indonesia, Philippine Bureau of Mines, Geological Survey of Papua New Guinea, and Geological Survey of Japan. The names of individuals to whom we owe much are too numerous to be listed here. The work has been partly supported by the Japan International Cooperation Agency.

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UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 171-195

THE MATURATION HISTORY OF THE EPICONTINENTAL BASINS OF WESTERN AUSTRALIA

A C COOK AND A J KANTSLER

ABSTRACT

The continental margin of Western Australia is of 'Atlantic' type, having been produced by Mesozoic rifting and continental breakup. Vitrinite reflectance data from wells drilled in the Perth and Carnarvon Basins show that major variations exist in the pattern of rank distribution indicating significant spatial and temporal variation in heat flow between the basins.

The presence of very high ranks in parts of the Permian section on the Beagle and Harvey Ridges suggests that a Permian to Early Jurassic thermal event, associated perhaps with the initiation of rifting, may be a controlling factor in rank variation in the Perth Basin. Regional uplift and truncation are also significant. Isoreflectance surfaces are deflected over basement highs, where they are also more closely spaced. Depth to magnetic and seismic basement is a major control on rank variation.

In the Carnarvon Basin, depth to the top of the oil mature zone (0.5% R_o) increases from onshore to offshore and this trend is maintained for isoreflectance surfaces at higher ranks. In addition, these surfaces are strongly diachronous indicating that most coalification is related to the Tertiary thermal regime.

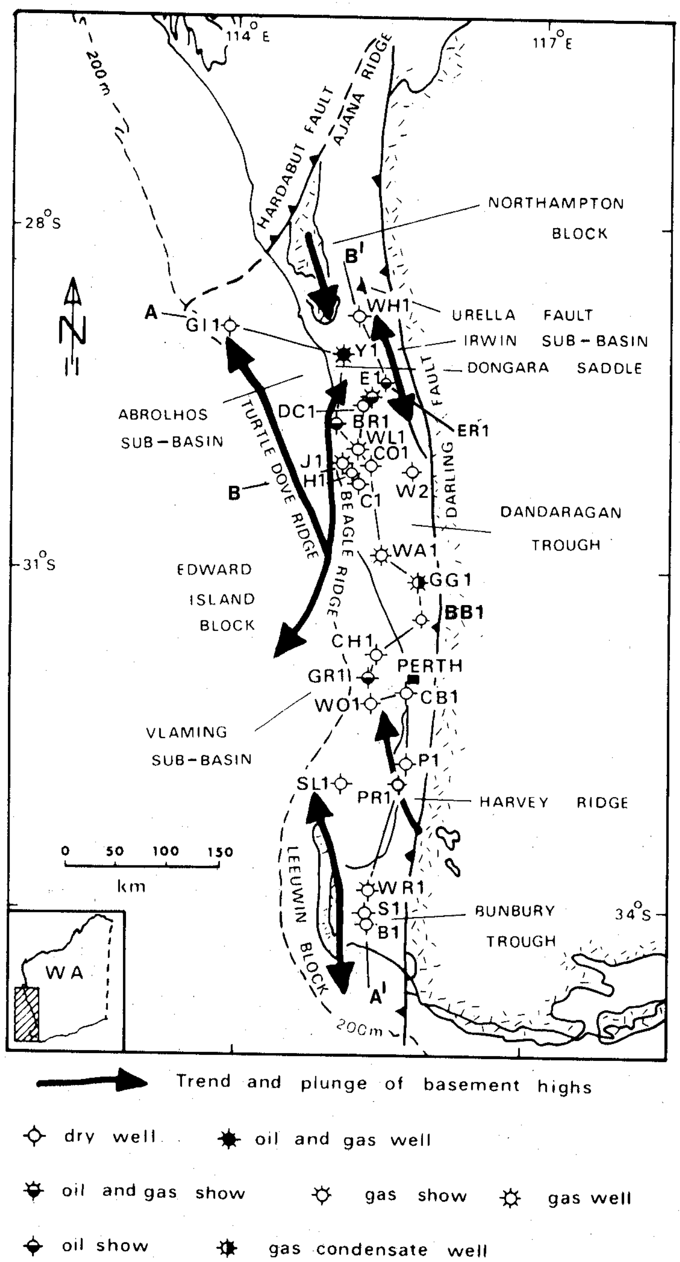
Well temperatures, at any given maturation level, are on average 40°C higher for the Carnarvon Basin as compared with the Perth Basin. In the Carnarvon Basin, high burial temperatures at low maturity are partly associated with Tertiary subsidence which exceeds 3000 m over large parts of the basin. In the Perth Basin, low temperature-gradients are thought to relate to an attenuating post-breakup thermal regime.

Small amounts of gas with minor oil have been found in the Perth Basin whereas major discoveries of gas/condensate and oil have been made in the Carnarvon Basin. In the absence of any profound difference in source rock quality, this difference in hydrocarbon productivity is considered to relate to the overall slow rate of maturation in the Perth Basin and the much stronger and more recent burst of generation in the Carnarvon Basin. The high gas-oil ratio of both basins may be related to the predominance of terrestrial source material but factors such as slow rates of maturation and high temperatures in mature and over-mature sequences are likely to be of significance.

INTRODUCTION

The Perth and Carnarvon Basins form part of the same rifted continental margin and have broadly similar structural and depositional histories. Differences relate to the long history of rifting in both basins, whereby the most intensive tensional phases are not time-equivalent over the entire area. This has resulted in the development of parallel grabens, which are considered to be aborted rifts. Both basins contain appreciable volumes of mature source rocks buried at temperatures suitable for the generation of hydrocarbons. The Perth Basin has recoverable reserves estimated at 452×10^9CFG and 2×10^6bbls oil as compared with the northern Carnarvon Basin where recoverable reserves include 15.33×10^{12}CFG, 264×10^6bbls oil and 384×10^6bbls condensate (West. Australia Dept. Mines 1978). Differences in hydrocarbon productivity between the two basins are related primarily to the timing and rate of source rock maturation rather than to differences in source or reservoir rock potential. This paper explores the nature and extent of that variation.

In our study, the reflectance of vitrinite is used as an indicator of source-rock maturation. The vitrinite is present in coals or as dispersed organic matter (DOM) in other sedimentary rocks. The data presented relate to mean maximum reflectance of vitrinite in oil referred to as $\%R_o$). All samples were prepared on an 'as received' basis since the nature of the organic matter, particularly vitrinite, can be more accurately identified in samples which are not demineralized. Such an approach, supported by observations in fluorescence mode, also yields more definitive information on the quantity and type of organic matter present (source-rock potential).



B1	_ Blackwood _1	▬
BB1	_ Bullsbrook _1	◆
BR1	_ BMR Beagle Ridge _10A	
C1	_ Cadda _1	◨
CB1	_ Cockburn _1	⊗
CH1	_ Charlotte _1	◆
CO1	_ Coomallo _1	◇
DC1	_ Donkey Creek _1	×
E1	_ Eneabba _1	△
ER1	_ Erregulla _1	+
GG1	_ Gingin _1	✖
GI1	_ Gun Island _1	●
GR1	_ Gage Roads _1	✧
H1	_ Hill River _1	
J1	_ Jurien _1	▽
P1	_ Pinjarra _1	◩
PR1	_ Lake Preston _1	■
S1	_ Sue _1	▼
SL1	_ Sugarloaf _1	○
W2	_ Warro _2	◓
WA1	_ Walyering _1	▭
WH1	_ Wicherina _1	◘
WL1	_ Woolmulla _1	◒
WO1	_ Warnbro _1	H
WR1	_ Whicher Range _1	▲
Y1	_ Yardarino _1	

Figure 1. Structural provinces and well localities in the Perth Basin. (Modified from Kantsler and Cook 1979). Symbols refer to the data of Figure 11. Well abbreviations are as given above.

PERTH BASIN

Geological and structural setting

The Perth Basin is a north-south-trending elongate trough which contains clastic sedimentary rocks of Silurian to Recent age. Located in the south-west of Australia, it covers an area of approximately 103 000 km^2, of which slightly less than half is onshore. A large crustal feature, the Darling-Urella Fault System, which has an overall vertical displacement >10 km, forms the eastern margin (Fig. 1) and has been a dominant influence on the structural and depositional history of the Perth Basin. Several shallow basement ridges divide the graben or half-graben structure of the basin into a number of troughs and sub-basins. The western margin of the basin is formed by the continental slope. The Northampton Block, a structural high composed of Precambrian rocks, separates the Perth Basin from the Carnarvon Basin to the north.

Playford *et al.* (1975, 1976) and Jones (1976) have described the stratigraphy, geological history and petroleum geology of the Perth Basin in some detail. A generalized stratigraphic column is illustrated in Fig. 2. The basin functioned as an active graben or half graben from at least the Permian until the Neocomian and three major tectonic episodes, resulting from large movements on the Darling-Urella Fault System are recognized: Late Permian; Late Triassic–Early Jurassic; and Late Jurassic–Early Cretaceous (Jones 1976). These tectonic events relate to a long

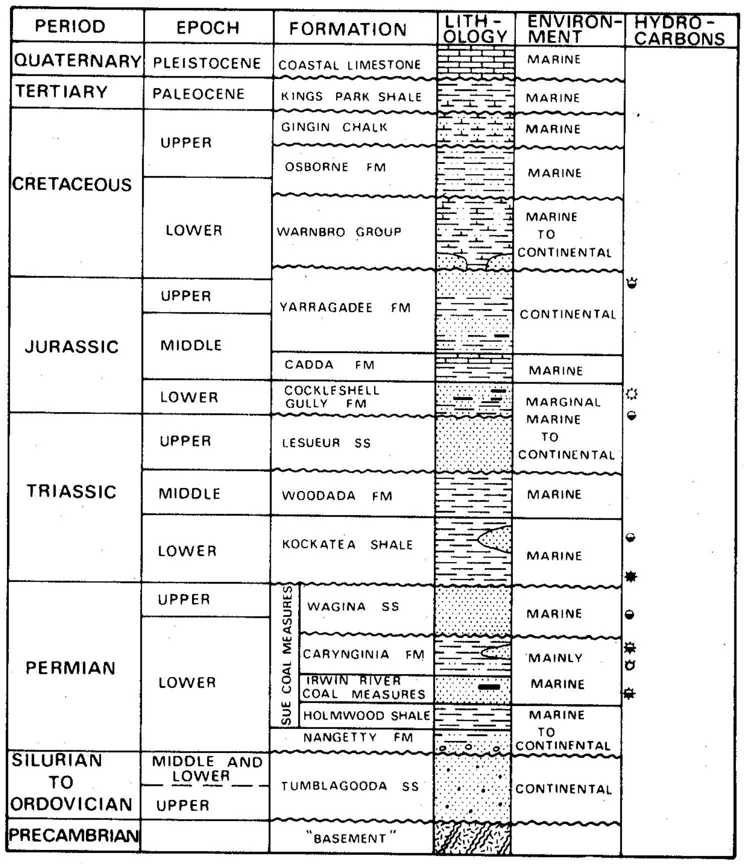


Figure 2. Generalized stratigraphic column for the Perth Basin (modified from Beddoes 1973).

history of rifting, which culminated in the separation of India from Western Australia in the Early Cretaceous (Thomas 1979). The sequence including that in the offshore parts of the basin, is dominated by rocks laid down under paralic to continental conditions. Over 15 000 m of sedimentary section occurs in structural lows such as the Dandaragan Trough and Vlaming Sub-basin. Greatest movement on the major bounding faults occurred during the Late Jurassic and Neocomian, when up to 4500 m of Yarragadee Formation sediments were deposited (Thomas 1979).

Five gas/oil fields have been discovered in the Dandaragan Trough and the Dongara Saddle — the only two areas of commercial production in the basin. The two Dandaragan Trough fields ceased production after a short time owing to a rapid decline in production rates and reservoir pressure. Numerous shows of hydrocarbons occur in sedimentary rocks of Permian, Triassic, Jurassic and Cretaceous age throughout the basin but reservoir quality is poor and porosity declines rapidly with depth.

Geothermal gradients

Lateral and vertical variation of vitrinite reflectance is primarily temperature related although this relationship is complicated by differences in burial history. The establishment of present-day patterns of geothermal gradient variation is, therefore, a necessary first step in assessing the role of temperature variation in coalification and hydrocarbon generation.

Figure 3 shows that present-day geothermal gradients in the Perth Basin range from 20°C/km to >40°C/km. In general, the known high gradients (>30°C/km) occur on, or on the flanks of, the Beagle Ridge and the Northampton Block, where they are associated with a shallow depth to crystalline basement. A less marked regional increase in temperature occurs over the more deeply buried Harvey and Turtle Dove Ridges. Lowest gradients are in the Bunbury and Dandaragan Troughs which have thick sandstone sequences of presumably higher thermal conductivity than the shale-dominated sections on the Beagle Ridge.

Most geothermal gradients were estimated from maximum bottom-hole temperatures (BHT) recorded during well-logging operations and corrected by adding 10% to the recorded temperatures from the first logging run. Thomas (1979) has found that such estimates approach extrapolated (Horner Plot) BHTs in wells in the northern Perth Basin.

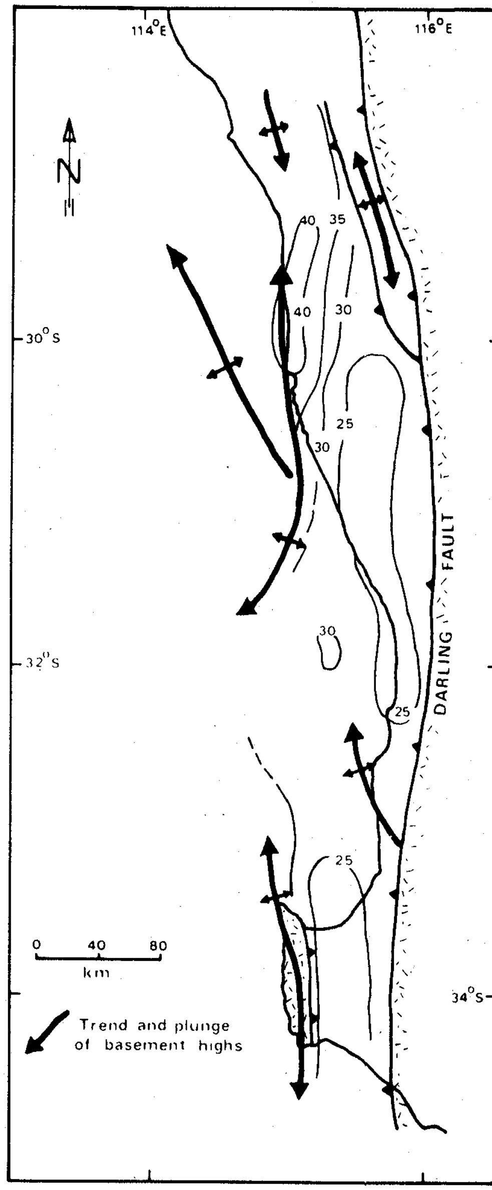


Figure 3. Sketch map of temperature gradients (°C/km) in the Perth Basin. (After Kantsler and Cook, 1979).

Hydrocarbons

Powell and McKirdy (1976) and Thomas (1979) have described the nature of the hydrocarbon occurrences found throughout the Perth Basin. The Dongara and Mondarra gas fields (Dongara Saddle) produce lean gas from Early Triassic and Permian sandstones. The gas is associated with minor oil and condensate which are light (35°–47° API), highly paraffinic and extremely waxy. In the central Dandaragan Trough, multiple pay sands (gas/condensate) have been found at Gingin and Walyering within the Lower Jurassic Cockleshell Gully Formation. These condensates, together with those from the Bunbury Trough, differ from those in the Dongara Saddle by being paraffinic/naphthenic and having intermediate pristane/phytane ratios typical of a mixed source.

Organic-matter type

Descriptions of organic-matter content and type have been published by Kantsler and Cook (1979) and Thomas (1979). They are reviewed below.

Permian

Geochemical data from the *Holmwood Shale* indicate good source potential for gas Thomas 1979). The *Irwin River Coal Measures are predominantly a sandstone unit* with several intercalations of coal, shale and siltstone. The coals and DOM are mostly subhydrous although some are exinite rich. Rank varies from immature to overmature, but the bulk of the Irwin River Coal Measures occurs at depth and at relatively high ranks and is therefore gas prone. However, one low-rank, exinite-rich occurrence (BMR 10A well) is associated with a small oil show. The overlying *Carynginia Formation* is a marine sequence of siltstones and shales which contains abundant plant remains preserved chiefly as inertinite (hydrogen poor). Most intersections of the Carynginia Formation contain DOM of relatively high rank, and it is also considered to have greatest potential as a source for gas.

Triassic

The *Kockatea Shale,* a marine sequence of shale and siltstone containing a rich assemblage of phytoplankton is thought to source much of the oil and gas found in the Dongara Saddle area. The fluorescence intensity of acritarchs and dinoflagellates has proved useful in establishing maturity. In the Abrolhos Sub-basin and Dongara Saddle areas, these microfossils are characterized by strong yellow fluorescence colours at vitrinite reflectances in the range 0.7% to 0.9% R_o, whereas they emit no fluorescence at 1.3% R_o in the Eneabba-1 well. According to Thomas (1979) the Kockatea Shale is rich in organic matter at its base and becomes leaner upwards.

Jurassic

The *Cockleshell Gully Formation* is a sequence of sandstone, siltstone, claystone and shale with an upper coal-bearing unit, the Cattamarra Coal Measures Member, which reservoirs and probably sources the gas found in the Walyering and Gingin fields. DOM is mostly coaly with some layers rich in sporinite and resinite.

The *Cadda Formation* is a marine to paralic sequence of shale, siltstone and sandstone with some lenticular limestones. Marine phytoplankton achieve local predominance but the DOM is mostly humic. The Cadda Formation is considered to have more potential as a seal.

The *Yarragadee Formation* is dominated by sandstones with small amounts of DOM, but shales and siltstones containing appreciable amounts of exinite-rich coaly material are common in parts of the section. Coals are rare, but where present typically contain lenses with $>$ 15% exinite and up to 50% of suberin-rich vitrinitized tissue. The vitrinites commonly have a dull orange-brown fluorescence and are locally veined by exsudatinite (Teichmuller 1974) suggesting mobilization of hydrocarbons within the coal. The organic matter found in the Yarragadee Formation contains a large proportion of hydrogen-rich components such that it must be considered an oil and gas source in those parts of the basin where it is mature.

Cretaceous

The *Warnbro Group* contains a mixed sequence of sedimentary rocks deposited during the Early Cretaceous marine transgression of the Perth Basin. The DOM is commonly humic and immature.

In summary, organic-rich sediments are distributed through many of the units in the Perth Basin. Most of the organic matter is of land-plant origin and this source material is consistent with the high-wax oils discovered to date.

Rank variation

Kantsler and Cook (1979) have plotted vitrinite reflectance data against depth for 24 Perth Basin wells and have shown that most of the data plot in a relatively narrow zone which has a low gradient (0.14% R_o/km). Exceptions to this tendency for low gradients are Woolmulla-1, Cadda-1 and Jurien-1, which lie on or near the Beagle Ridge, and Lake Preston-1 drilled on the Harvey Ridge. Relatively high vitrinite reflectances are found in each of these wells and are most significant at Jurien and Cadda, where they are close to the surface.

Major characteristics of rank variation in the Perth Basin are summarized below.

(1) Wells in the Bunbury Trough and southernmost Dandaragan Trough show linear trends of vitrinite reflectance plotted as a function of depth over the interval penetrated by the drill. Reflectance gradients are low

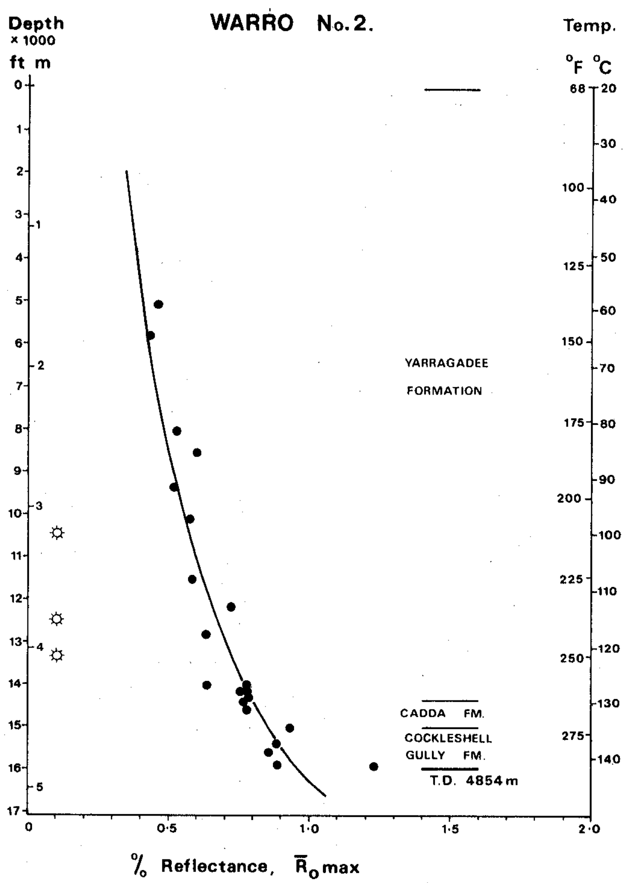


Figure 4. Depth-reflectance profile for the Warro-2 well — central Dandaragan Trough.

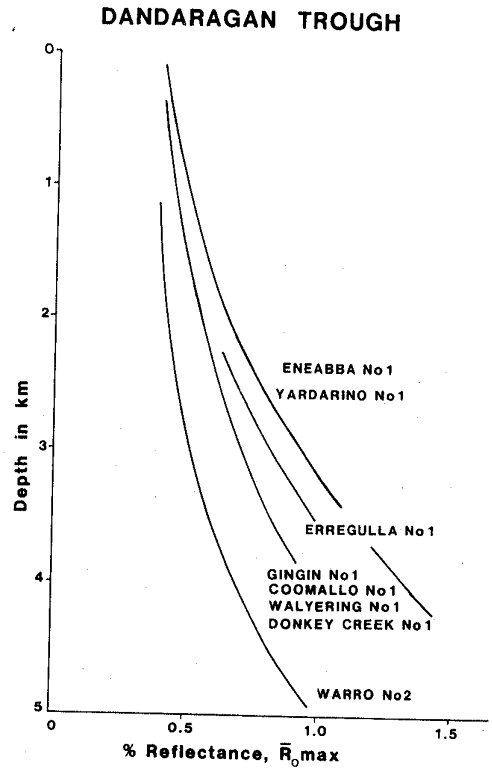


Figure 5. Comparison of selected vitrinite reflectance profiles from the Dandaragan Trough. (Modified from Kantsler and Cook, 1979).

(0.12% R_o/km to 0.17% R_o/km) with the exception of Pinjarra-1 (0.25% R_o/km), implying slow coalification throughout the history of the sequence.

(2) The Dandaragan Trough is characterized by wells with similarly low reflectance gradients and a tendency for this gradient to increase down section. Reflectance profiles show a systematic variation with thickness of Jurassic section and hence proximity to basement. For example, data from Warro-2 (Fig. 4), located close to the axis of the Trough, are significantly offset toward lower ranks as compared with data from wells such as Eneabba-1 and Yardarino-1 drilled in structurally higher areas (Fig. 5). The point of inflexion in Dandaragan Trough wells typically occurs at, or near the level of, the Cadda Formation, suggesting that pre-Cadda sediments underwent most coalification during deposition of the thick Yarragadee Formation.

(3) Strongly curved reflectance profiles occur in the Woolmulla-1 (Kantsler and Cook 1979, Thomas 1979) and Lake Preston-1 (Fig. 6) wells, which abut the Beagle and Harvey Ridges respectively. The Lake Preston profile is complicated by the presence of a large fault at 4139 m, which coincides with a sharp increase in reflectance gradient within the Sue Coal Measures. None of the Sue coals shows any signs of carbonization, and it is presumed that they underwent rapid early coalification before Triassic deposition ceased, since Lower Jurassic coals intersected near the present land surface in the same well have vitrinite reflectances of about 0.20%, implying little uplift of the Harvey Ridge since the Triassic. This phenomenon is equally pronounced at Woolmulla-1, where the point of inflexion in the reflectance profile is in the upper part of the Triassic Kockatea Shale.

(4) Wells in the Vlaming and Abrolhos Sub-basins have curved reflectance profiles. The Gun Island well has a very low reflectance gradient down to 2500 m, but the gradient increases significantly below this depth. The Gage Roads well also has very low reflectances in the upper 2500 m of section, but below this depth the gradient is significantly higher than in the nearby Vlaming Sub-basin wells and in the Gun Island well. This difference appears to be related to the collapse of the regional arch upon which Gage Roads-1 was drilled.

Overall, differences in rank gradients throughout the Perth Basin are thought to relate to:

A period of high heat flow and geothermal gradient associated with the development of rifting in Triassic to Early Jurassic time, followed by a decrease in heat flow and geothermal gradient in the Late Jurassic and Cretaceous.

The effects on post-Cretaceous heat flow of thick Triassic shale sequences being overlain by thick Jurassic sandstone sequences in the deep troughs and sub-basins which flank the structural highs.

Post-Neocomian loss of cover

Many wells in the Perth Basin, with the exception of those in the Bunbury Trough,

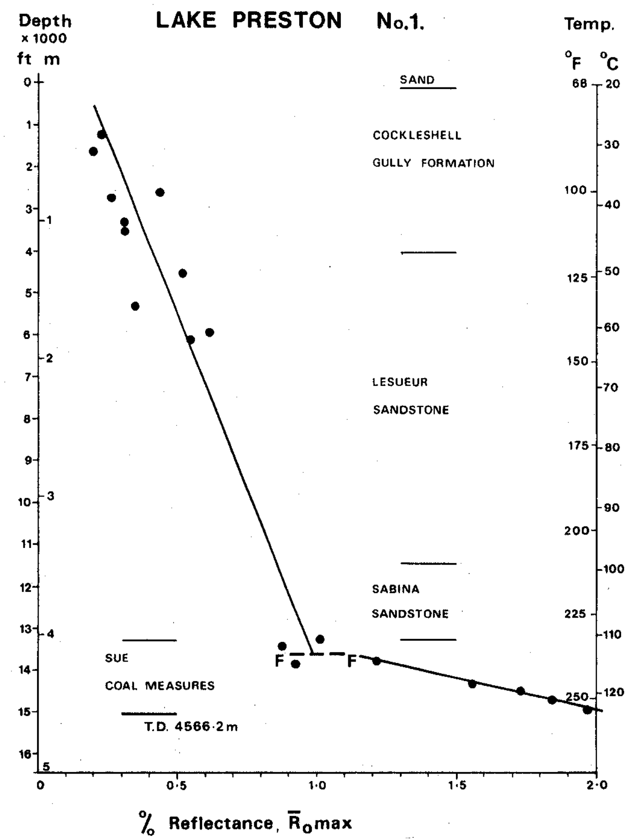


Figure 6. Depth-reflectance profile for the Lake Preston-1 well — Harvey Ridge.

have surface intercepts of their depth-reflectance profiles, which range from 0.4% R_0 to 0.7% R_0 in sediments of Triassic to Upper Jurassic age. When compared to an 'average' surface intercept of 0.2% to 0.25% R_0 from wells in other Australian and overseas basins lacking post-sedimentation uplift, the data imply that regional uplift related to a major tectonic pulse probably took place during the Neocomian. This pulse, centred on the Beagle Ridge, resulted in the uplift of the western flank of the Dandaragan Trough and is thought to represent the end of rifting and the onset of continental drift (Thomas 1979). It is difficult to assess the amount of subsequent erosion because of poor well control and the varying extent of sedimentary onlap. However, it appears certain that at least 1000 m of section have been eroded from the area of the Beagle Ridge near Jurien-1.

Controls over rank distribution

Figures 7 and 8 represent schematic sections of isoreflectance variation through the Perth Basin.

In the Bunbury Trough the 0.5% isoreflectance surface is markedly diachronous. This surface rises over the regional arch drilled by Cockburn-1 (which has a thick but substantially elevated section relative to the troughs nearby) and dips into the Vlaming Sub-basin, where it appears to have been affected by the formation of the Rottnest Trench — an east-west-trending supra-basin of limited extent. It is approximately concordant

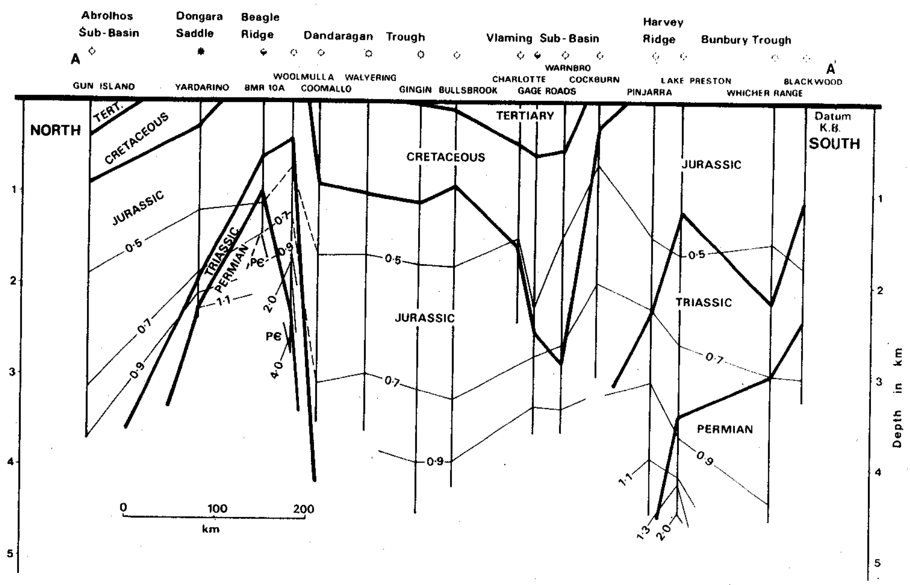


Figure 7. Schematic section A-A' (Figure 1) through Perth Basin with profile of isoreflectance surfaces.

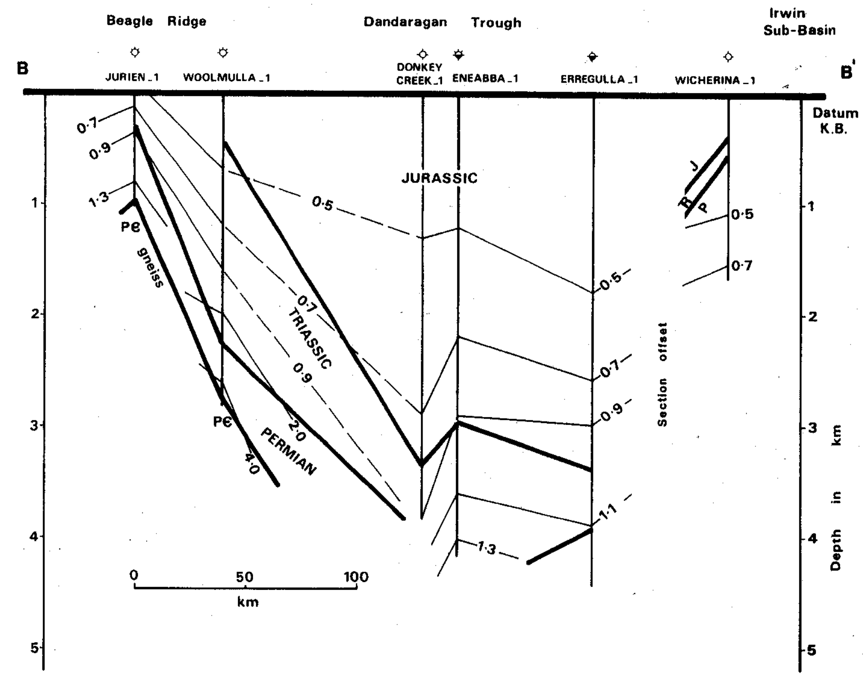


Figure 8. Schematic section B-B' (Figure 1) through Perth Basin with isoreflectance surfaces.

throughout the Dandaragan Trough, rises (discordantly) over the Beagle Ridge and dips into the Abrolhos Sub-basin depocentre. Figure 8 demonstrates how this surface rises onto the Northampton Block. The other isoreflectance surfaces follow this same pattern of variation but become closely spaced over structural highs.

Figure 9 shows that the depth to the 0.5% isoreflectance surface is related to the depth to basement. Comparison of the 0.5% isoreflectance surface with the plot of temperature gradients (Fig. 3) shows that the 'hot' Beagle Ridge and Northampton Block are both associated with least depth to the 0.5% R_0 surface (the zone of initial source rock maturity), whereas the 'cooler' Dandaragan and Bunbury Troughs and Vlaming and Abrolhos Sub-basins are associated with greatest depth to this surface. Present-day temperature gradients show a similar 'direct relationship with depth to magnetic and seismic basement.

Figure 10 represents depth to the 0.75%

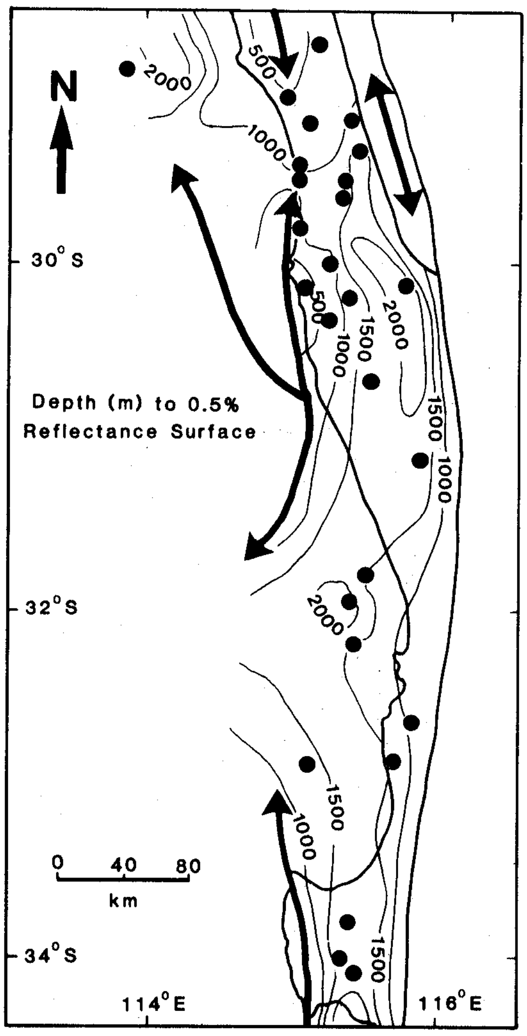


Figure 9. Depth (m) to the 0.5% R_0 max surface, Perth Basin.

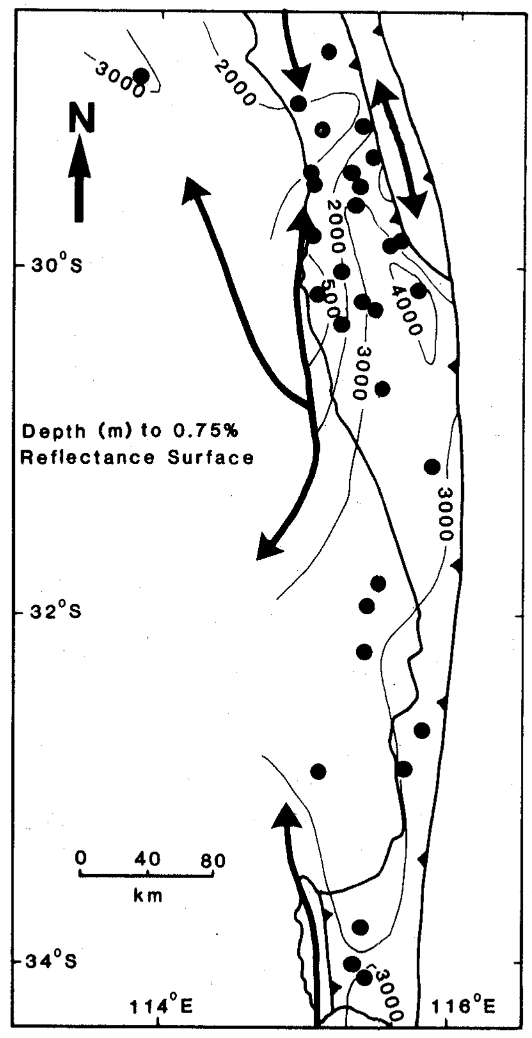


Figure 10. Depth (m) to the 0.75% R_0 max surface, Perth Basin.

isoreflectance surface (which many authors consider represents the zone of intense hydrocarbon generation) and shows a similar systematic variation with present-day geothermal gradient. Shallowest depths to the main oil-generation zone are found on the Beagle Ridge, Dongara Saddle, north Dandaragan Trough and the flanks of the Northampton Block.

Geothermal history

(1) Rank surfaces are diachronous in the Bunbury Trough, implying that most of the coalification in this area of the basin is post-Cretaceous and that the pre-Jurassic sediments underwent only minimal coalification prior to the onset of Jurassic sedimentation. This slow rate of early coalification corresponds to a phase of slow subsidence.

(2) Maximum burial accompanied very rapid trough subsidence and horst-block development in the adjacent shield areas during the Jurassic and Early Cretaceous, such that the Dandaragan Trough and Vlaming Sub-basin were appreciably coalified before Neocomian deposition ceased. The subsequent long exposure time is presumed to have lead to further slow coalification.

(3) The rise in reflectance at the 0.7% R_0 level and higher, from Whicher Range-1 to Pinjarra-1 (Fig. 7), is associated with more rapid early coalification on the Harvey Ridge, relative to the Bunbury Trough, and the growth of the Harvey Ridge in the Late Permian–Early Triassic and post-Early Jurassic.

(4) Isoreflectance surfaces rise markedly over the Cockburn Arch to an extent which cannot simply be considered as 'experimental error'. The 0.5% R_0 surface at Cockburn-1 is elevated by about 1 km compared with either Bullsbrook-1 or Lake Preston-1, the 0.9% surface by 700 m and 400 m respectively. Relative uplift of the Cockburn Arch in late Late Jurassic time may be the cause, since the measured reflectance gradient at Cockburn is low (0.16% R_0/km). This implies that most of the coalification observed at Cockburn occurred before the Late Jurassic and that there was little subsequent coalification. To some extent this conclusion is borne out by the very low porosity and permeability at 3000 m (TD) in Cockburn-1 as compared with porosity of >10% at 3400 m at Bullsbrook-1 (central Dandaragan Trough). According to Thomas (1979) 'The inferior reservoir characteristics of the western margins of the (Dandaragan) Trough are probably the result of deep burial diagenesis'. These interpretations of the data imply regional uplift and erosion of more than 1000 m of Upper Yarragadee Formation at Cockburn-1.

(5) The very high ranks found at relatively shallow depth in the Permian and Triassic rocks of the Beagle Ridge are considered to be a response to deep burial at very high temperatures in pre-Early Jurassic time. The wholly Jurassic section penetrated in the nearby (30 km) Coomallo-1 well is unaffected, lending tacit support to the concept of a temporal and spatial localization of this thermal event. Uplift during the Neocomian has apparently resulted in the loss of up to 500 m of section at Woolmulla-1 and more than 1000 m at Jurien-1.

The Triassic Kockatea Shale in Woolmulla-1 contains dinoflagellate cysts that do not fluoresce at an extrapolated present well temperature of 85°C. Kantsler and Cook (1979) contrast this with the strong fluorescence noted in dinoflagellates in the Kockatea Shale at Yardarino-1 at present well temperatures of 100°C. These data imply that temperatures during the Mesozoic at Woolmulla were substantially greater than present-day temperatures.

(6) Isoreflectance surfaces rise rapidly towards Yardarino from the direction of both Eneabba-1 (Dandaragan Trough) and Gun Island-1 (Abrolhos Sub-basin). The diachronous nature of these surfaces suggests that most of this coalification is post-Jurassic and related to the higher temperature gradients (30°C to 35°C/km) observed in these areas of the basin. There are three other factors that could also affect rank variation in and around the Dongara Saddle:

The degree of curvature of the reflectance profiles at depth suggests that an early phase of high heat flow may have carried over into this region from the Beagle Ridge.

An obvious age difference exists between the Permian to Early Jurassic sections penetrated by wells in the Dongara Saddle and the wholly Jurassic sections penetrated in the Dandaragan Trough proper and the Abrolhos Sub-basin proper.

Surface reflectance intercepts of 0.40 to 0.45% indicate some degree of relative uplift — perhaps only a few hundred metres.

(7) The Gun Island well drilled near the axis of the Abrolhos Sub-basin is characterized by a low reflectance gradient (Kantsler and Cook 1979) offset towards lower ranks as compared with data from the Vlaming Sub-basin but similar to data from the Dandaragan Trough. While a significant amount of coalification occurred during, and immediately after, deposition of the Yarragadee Formation, indications are that post-Yarragadee maturation rates have been low overall.

Estimated palaeotemperatures

Much of the observed variation in rank — particularly in those areas of the basin that underwent rapid and deep subsidence during the Jurassic — can be related to variations in present-day temperature gradient. However, the data of Table 1 show that present differences in geothermal gradient are insufficient to explain all of the observed variation, and it appears that differences in burial temperature must have been even

TABLE 1

Temperature/Age data at various vitrinite reflectance levels in the Perth and Carnarvon Basins together with 'average' figures for each basin.

Well Name	0.5% R_o		0.7% R_o		0.9% R_o		1.1% R_o	
	Temp °C	Age	Temp °C	Age	Temp °C	Age	Temp °C	Age
			PERTH BASIN					
Blackwood-1	61	E Tr	87	L Perm	—	—	—	—
Whicher Range-1	52	E Jur	81	E Tr	109	L Perm	—	—
Sugarloaf-1*	56	E Cret	87	E Cret	112	L Jur	—	—
Lake Preston-1	58	L Tr	81	L Tr	104	L Perm - E Tr	114*	L Perm
Pinjarra-1	56	E Jur	75	E Jur	94	L Tr	113	L Tr
Cockburn-1	36	L Jur	67	E Jur	89	E Jur	—	—
Warnbro-1	62	E Cret	96	E Cret	126†	L Jur	—	—
Gage Roads-1	82	E Cret	96	L Jur	112	L Jur	—	—
Charlotte-1	60	E Cret	—	—	—	—	—	—
Bullsbrook-1	59	L Jur	91	E Jur	112	E Jur	—	—
Gingin-1	64	L Jur	98	L Jur	118	E Jur	—	—
Walyering-1	62	L Jur	94	E Jur	113†	E Jur	—	—
Coomallo-1	44	L Jur	82	E Jur	—	—	—	—
Warro-2	82	L Jur	119	L Jur	135	E Jur	—	—
Donkey Creek-1	51	L Jur	90	E Jur	112	E Tr	—	—
Erregulla-1	70	M Jur	93	E Jur	114	E Jur	130	E Tr
Yardarino-1	47	L Jur	96	E Tr	—	—	—	—
Gun Island-1	77	L Jur	115	E Jur	132†	E Jur	—	—
Eneabba-1	54	L Jur	82	E Jur	104	L Tr	112	M Tr
Basin Average	60°	L Jur	90°	late E Jur	111°	early E Jur	113°	E Tr
			CARNARVON BASIN					
Onslow-1	92	E Tr	135	E Perm	142	E Perm	—	—
Anchor-1	99	L Jur	129	L Tr	—	—	—	—
Pepper-1	108	E Cret	125	L Jur	—	—	—	—
Barrow Deep-1	97	L Jur	138	L Jur	146	L Jur	159	L Jur
Tryal Rocks-1	89	L Cret	126	E Cret	—	—	—	—
North Tryal Rocks-1	85	Paleoc	125	L Tr	—	—	—	—
Rankin-1	106	L Cret	131	L Tr	147	L Tr	—	—
Basin Average	96°	E Cret	130°	E Jur	145°	L Tr	—	—

* Less reliable.

† Extrapolated 100 m to 200 m — well terminated before reaching this reflectance.

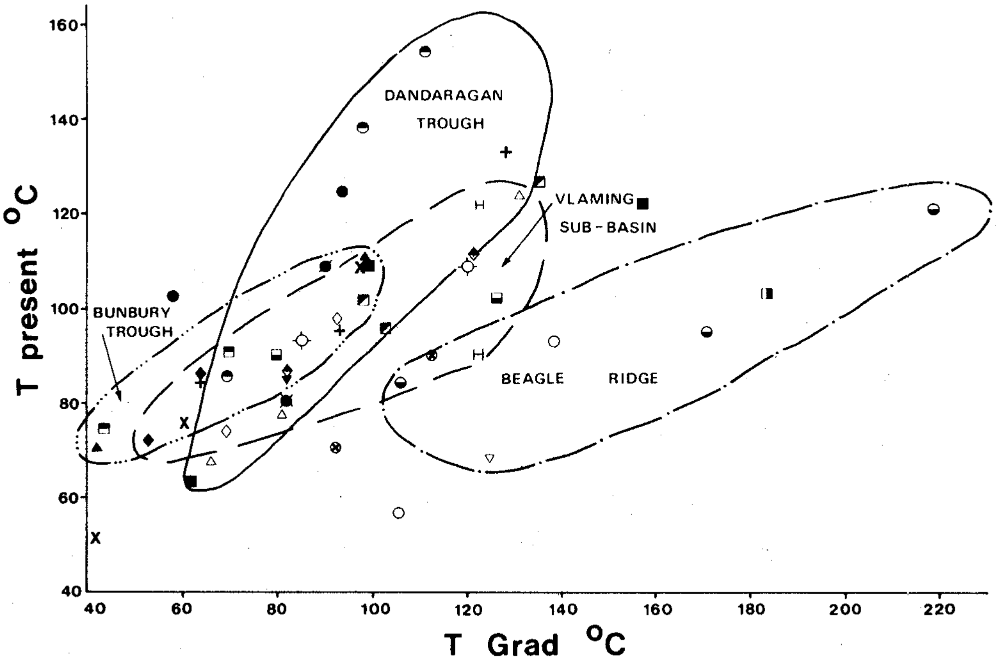


Figure 11. Relationship between temperature estimated from coal rank and coal age data, and present well temperatures for the Perth Basin. Well name symbols as per the caption of Figure 1. (Modified from Kantsler and Cook, 1979).

greater in Late Palaeozoic and Early Mesozoic time. Differences in age are likely to have had only a relatively minor influence on rank variation, for the basin has been virtually quiescent since the Cretaceous and the 'exposure' time can be considered constant.

Model-derived temperatures (Tgrad) were compared with present well temperatures (Tpres) and the results plotted in Fig. 11. Our gradthermal model assumes a history of steadily rising temperature up to a present-day maximum, i.e. Tgrad. If subsidence in the Perth Basin was accompanied by a uniform rise in burial temperature, temperature would have been constant or falling since the Neocomian. Thus Tgrad would be expected to be significantly above Tpres. Figure 11 shows that this is not uniformly so, and also that grouping by structural province occurs. Possible errors in basic assumptions, as well as the model, place limits upon the conclusions which can be drawn, but comparison of data should minimize systematic errors. Such comparisons show, for example, a marked contrast between the temperature history of the Beagle Ridge and that of the Bunbury Trough. The data indicate an early episode of intense heating on or near parts of the Beagle and Harvey Ridges. The timing of this thermal event is thought to be Permo-Triassic. Equally, comparison of data from the Yarragadee Formation of the Vlaming Sub-basin and the Dandaragan Trough indicates that temperatures rose earlier in the Vlaming Sub-basin.

The direction of elongation of the domains in Fig. 11 is related to the direction of the tie lines linking data points near the top and base of the section in each well. It appears that the direction of these lines and their position relative to the line Tpres = Tgrad are significant in relation to the timing of coalification. Lowest slopes are found for the Beagle Ridge wells (where Tpresent < Tgrad) and are considered to reflect the early phase of high heat flow. Moderate slopes occur in wells from the Vlaming Sub-basin and parts of the Dandaragan Trough (where Tpres ≃ Tgrad) and imply a lengthy association of these areas with a temperature regime similar to that which influences them now. Highest slopes occur near the axis of the Dandaragan Trough (Warro-2, Gingin-1) and are associated with the condition Tpres > Tgrad, indicating a possible late rise in temperature. Wells in the Bunbury Trough, Dongara Saddle and Abrolhos Sub-basin (Gun Island) are similarly affected.

Overall, the trends observed by comparison of the model-derived data reinforce many of the concepts of maturation history described previously and, in addition, suggest that in some areas of the Perth Basin (particularly Warro-2) there has been a late increase in temperature which has had little effect, as yet, on vitrinite reflectance.

Relationship of maturation history to oil and gas occurrences

In the north Dandaragan Trough, gas with minor oil occurs in basal Triassic reservoirs in the Dongara, Mondarra and Yardarino wells at 0.7% to 0.75% R_0. The rank data indicate that the overlying Kockatea Shale is at an early stage of oil generation, while the Permian section below is likely to be more mature and in the zone of peak wet-gas generation. Our data suggest that oil generation within the Kockatea Shale is not likely to have begun until the Tertiary, although early migration of hydrocarbons from more mature Permian source rocks in adjacent areas downdip may have begun during the Triassic. Loss of porosity, associated with deep Jurassic burial, is a limit to post-Jurassic migration.

The Permian and Triassic sections on the Beagle Ridge and Cadda Shelf are a mature to overmature source of condensate and dry gas. Major coalification and therefore hydrocarbon generation occurred in Late Triassic or Early Jurassic time. The lack of suitable reservoirs, Neocomian uplift and the long subsequent exposure time of early generated hydrocarbons are possible reasons for the lack of success in this area.

In the Dandaragan Trough, the Early Jurassic Cattamarra Coal Measures Member is associated with vitrinite reflectances of 0.6% to 0.9% R_0 and provides the source and reservoir for the gas/condensate found at Walyering and Gingin. The timing of the main phase of hydrocarbon generation is thought to be Tertiary although generation from the Lower Jurassic section in some areas of the Trough may have begun as early as the Cretaceous. The high GOR relates to the slow rate of maturation in this area of the basin. The Triassic section in these wells is in the wet-gas zone of generation.

The high-wax, paraffinic oil recovered from Gage Roads-2 beneath the intra-Neocomian unconformity (Jones 1976) occurs at less than 0.5% R_0, while gas in the same well beneath the Quinns Shale Member (Yarragadee Formation) occurs at 0.55% R_0. The source is undoubtedly more mature Yarragadee Formation further downdip. Present-day temperatures appear to be maxima, suggesting that the main phase of oil generation took place during the Tertiary.

Shows of gas and condensate, associated with vitrinite reflectances of 0.8% to 1.0% R_0, have been recorded from the Sue Coal Measures in wells drilled in the Bunbury Trough. The low rank and temperature gradients in these wells imply slow rates of coalification and hydrocarbon generation such that early generated hydrocarbons have been subject to a long history of *in situ* maturation. The naphthenic nature of the condensates is consistent with such a history.

CARNARVON BASIN

Geological and structural setting

The Carnarvon Basin is an elongate series of depressions containing sediments of Silurian to Late Tertiary age. It parallels the present western coastline of Australia (Fig. 12) and covers an area of about 240 000 km² (to the edge of the continental shelf) of which slightly less than half is onshore. The southern portion of the basin (mostly onshore) consists predominantly of Palaeozoic strata; the northern part (to which this paper is addressed) comprises mostly Mesozoic rocks underlain by a thick sequence of complexly faulted Palaeozoic sediments. The Precambrian Shield represents the eastern margin of the basin whereas the western limit lies in deep water of the continental slope.

Playford *et al.* (1975), Thomas and Smith (1974, 1976), Powell (1976) and Crostella and Chaney (1978) have described the regional stratigraphy and geological history of the Carnarvon Basin. No single structural feature dominates the basin, but a series of predominantly north-south faults has created three major sub-basins in the area under discussion. The three sub-basins are regarded by Thomas and Smith (1976) as probable depositional entities throughout Jurassic time, the Barrow and Dampier Sub-basins remaining linked until the Neocomian. Maximum thickness of sedimentary section is thought to exceed 15 000 m in parts of the Barrow Sub-basin. The Rankin Platform is a fault-controlled gravity and structural high consisting of Triassic and Early Jurassic strata, over which is draped a Cretaceous, Tertiary and Quaternary sequence which reaches a thickness of 3000 m on its southernmost extremity. The regional stratigraphy is illustrated in Fig. 13.

Powell (1976) has summarized the history of the region as follows, 'The Northwest Shelf is . . . a passive Atlantic-type margin which evolved as a result of major rifting on a continental scale, commencing in the Late Palaeozoic and continuing into the Early Cretaceous. These pull-apart movements resulted in the complete disassociation of eastern Gondwanaland. Epicontinental sedimentation took place in a series of marginal tension basins developed parallel to the principal rift system. Deposition has been controlled by large-scale block faulting, with major adjustments taking place during the

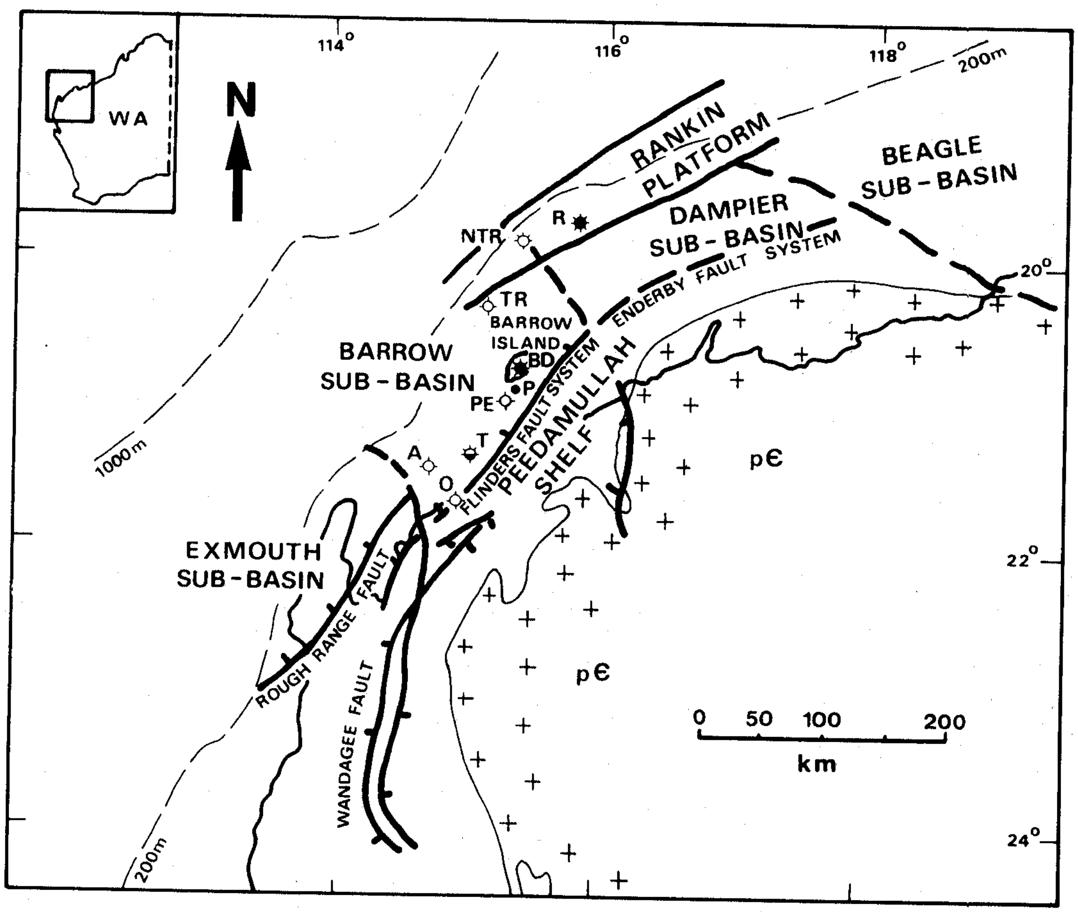


Figure 12. Structural provinces and well localities in the northern Carnarvon Basin. Symbols as for Figure 1. Well abbreviations as follows:

A	Anchor-1	O	Onslow-1	R	Rankin-1
B	Barrow-1, Barrow Deep-1	P	Pasco-1	T	Thevenard-1
NTR	North Tryal Rocks-1	PE	Pepper-1	TR	Tryal Rocks-1

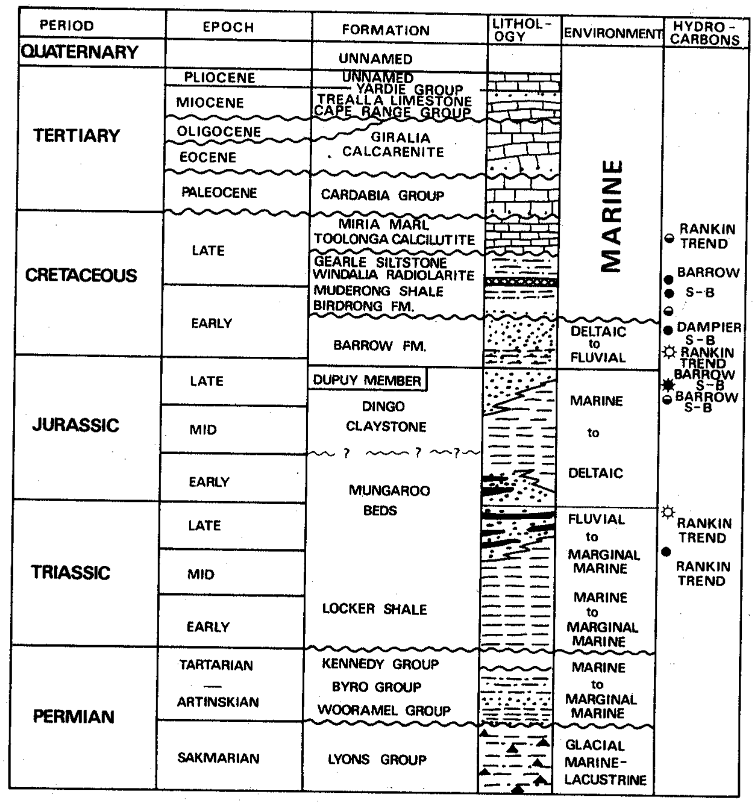


Figure 13. Generalized stratigraphic column for the northern Carnarvon Basin.

protracted breakup period.' Late Tertiary tectonism is related to isostatic instability during collision of the Australian Plate with the Indonesian Arc to the north (Thomas and Smith 1976).

Five commercial gas/condensate fields, one commercial oil field and six sub-marginal gas/condensate fields have been discovered to date in the northern Carnarvon Basin.

Geothermal gradients

Figure 14 is a schematic section of rank variation across the Barrow Sub-basin. Available temperature gradients (°C/km) from individual wells have been included above the well name. Gradients range from 29°C/km to 40°C/km and decrease away from the present coastline. The highest known gradient (45°C/km) occurs in the Muiron-1 well in the Exmouth Sub-basin (Wright and Wheatley 1979) about 50 km SE of Anchor-1.

The geothermal gradients were calculated from BHTs corrected by a factor of 10%. These results generally show good agreement with gradients calculated from reservoir temperatures in discovery wells.

Hydrocarbons

At Barrow Island, the Windalia crude is heavy, aromatic-naphthenic and almost devoid of n-alkanes whereas the crude from the Late Jurassic/Neocomian Dupuy Member 1000 m below is light, paraffinic-naphthenic and highly waxy. The origin of the deeper Jurassic oils is postulated as mixed marine and terrigenous in keeping with the host rocks (Powell and McKirdy 1976). While the marine Muderong Shale and Gearle Siltstone are thought to source the Cretaceous oils of the Barrow Sub-basin (Parry 1967, Crostella and Chaney

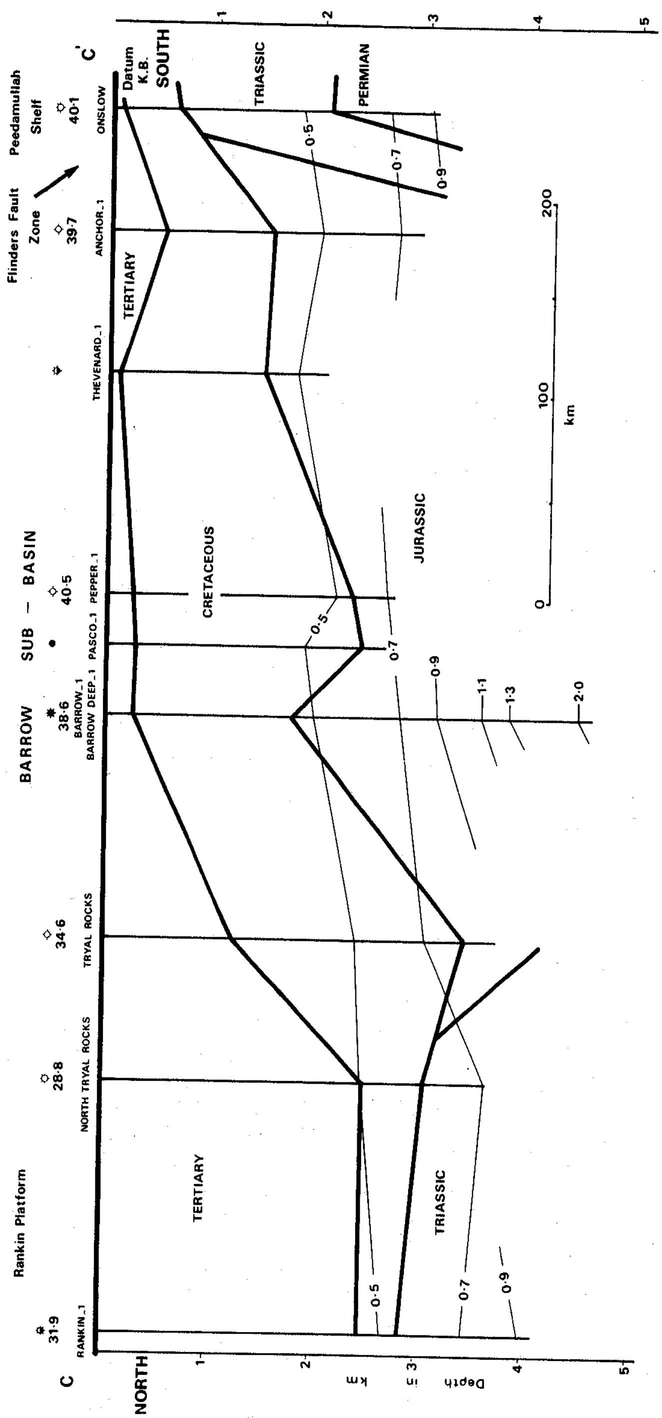


Figure 14. Schematic section C-C' (Figure 12) through northern Carnarvon Basin with isoreflectance surfaces.

1978), Gould and Smith (1978) have pointed out that the Windalia crude may represent the microbiologically altered product of a Jurassic precursor. However, Thomas (1979) has shown that an onshore occurrence of low-gravity, highly aromatic crude of marine origin is probably a migrated (over long distances) and biochemically altered equivalent of an initially light Windalia-like crude generated in the Barrow Sub-basin proper.

Upper Triassic sands (Mungaroo Beds) constitute the main reservoirs for the gas/condensate and oil flows of the Rankin Platform. In the adjoining Dampier Sub-basin, small accumulations of oil and gas/condensate occur in the Upper Jurassic transgressive sequence, and a small oil pool has been discovered in Lower Cretaceous sands. Despite the variations in reservoir age, the light paraffinic-naphthenic oils and condensates of many of the various Dampier Sub-basin discoveries show a remarkable similarity in composition, and a common (Jurassic) source has been invoked by Powell and McKirdy (1976). However, both Thomas and Smith (1974) and Crostella and Chaney (1978) have attested to the source potential of the underlying Triassic sequence. They also consider the heavier naphthenic-aromatic Cretaceous oils, which resemble Windalia crude, to be derived from a marine Cretaceous source.

Organic-matter type (source potential)

Permian (Onslow-1)

The *Lyons Group* is a fluvioglacial deposit comprising greywackes and siltstones. DOM content is generally small and inertinitic in nature. The *Byro Group* consists of interbedded siltstones, fine sandstone and massive shales. Coaly DOM is mostly oxidized, but finely comminuted exinite (phytoplankton/liptodetrinite and rare alginite) is locally abundant, retains autofluorescence and is regarded as having some source potential. The overlying *Kennedy Group* is made up of fine sandstones and siltstones which contain varying amounts of mostly subhydrous coaly DOM and is therefore considered to be gas prone despite its relatively low rank.

Triassic

The marine *Locker Shale* contains at its base abundant exinite (showing strong fluorescence) as microspores, cuticle and microplankton but becomes leaner upwards. Vitrinite is not uncommon. The unit is initially mature at Onslow and is regarded as having significant source potential. It is conformably overlain by the fluvio-deltaic *Mungaroo Beds,* which consist of alternating sandstone and claystones with minor intercalations of coal. The sandstones are mostly barren, whereas the claystones and shales typically contain small amounts of exinite including rare microplankton. Some coaly samples are very rich in organic matter (10%–15%). The common presence of small intergranular bitumens in the associated carbonate-rich rocks testifies to the primary migration of hydrocarbons.

Jurassic

A complete Middle to Late Jurassic sequence has not yet been penetrated, but the Late Jurassic is predominantly an argillaceous marine section represented by the *Dingo Claystone* with an upper unit of shales and siltstones — the *Dupuy Member.* As with most of the Carnarvon Basin sediments, finely comminuted inertodetrinite and oxidized vitrinite are present throughout and are generally regarded as having little or no source potential for oil. Exinite content (sporinite, cutinite, liptodetrinite, microplankton) varies from rare to abundant. Vitrinite occurs sporadically as thin and thick stringers, but may be abundant locally. Fluorescing and non-fluorescing bitumens are sometimes present, often polymerized about uraniferous minerals.

Cretaceous

The *Barrow Formation* is a deltaic sequence of sandstone with thinly interbedded siltstone. DOM content ranges from rare to abundant. Exinite is dominated by sporinite and microplankton, but cutinite occurs with the more coaly (vitrinite rich) samples. Bitumens are not uncommon. The formation contains several oil and gas bearing sands at Barrow Island. The Barrow Formation is unconformably overlain by the *Winning Group*, which is a sequence of predominantly marine claystones and siltstones comprising the *Muderong Shale, Windalia Radiolarite* and *Gearle Siltstone*. The Muderong Shale commonly contains abundant highly fluorescent microplankton and rare alginite in addition to minor amounts of terriginous organic matter. It is immature nearshore but initially mature on the Rankin Platform. Microplankton are rare, occasionally common, in the Windalia Radiolarite and Gearle Siltstone. Immature microplankton are present in small amounts throughout the Toolonga Calcilutite, but the Miria Marl is virtually barren of organic source matter.

Tertiary

The thick prograding wedge of Tertiary carbonate is wholly immature and its source rock potential generally appears to be low.

In summary, initially mature to mature, organic-rich sediments occur in most of the major sedimentary units of the northern Carnarvon Basin. Much of the organic matter is of land-plant origin, contains a significant amount of exinite and is associated with a pervasive marine component which suggests that the high GOR of the Rankin Platform is not primarily source related.

Rank variation

Figure 15 is a broad summary of downhole rank variation in the 10 Carnarvon Basin wells sampled for this study. Most data points lie on curved trends with reflectance gradients ranging from typical values of 0.25% R_0/km at the 0.5% R_0 level up to 0.75% R_0/km at the 1.0% R_0 level in Barrow Deep-1. Lowest gradients (less than 0.10% R_0/km) are indicated from the thick Tertiary section further offshore. With the exception of Onslow-1 (Peedamullah Shelf) where vitrinite reflectances in the Upper Triassic section are offset slightly towards higher ranks, reflectance profiles show little evidence of being displaced at any of the unconformity surfaces illustrated in Fig. 13. This leads us to conclude that any early phase of coalification has been largely overprinted by coalification associated with the present thermal regime. Surface intercepts (and projected intercepts) of depth-reflectance profiles are generally less than 0.2% and indicate no recent uplift of any significance.

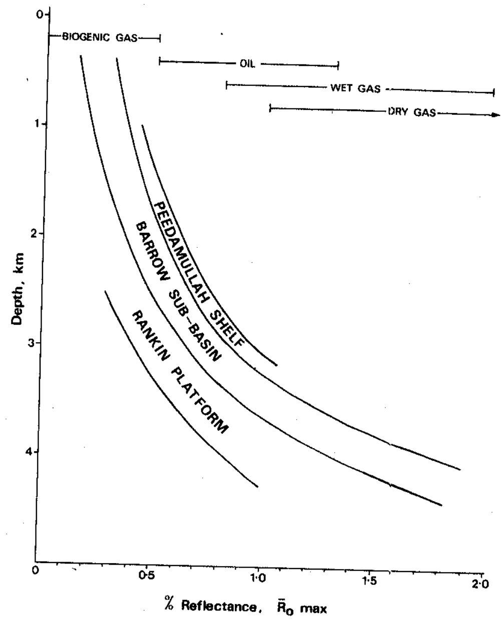


Figure 15. Comparison of vitrinite reflectance data from the various structural provinces in the northern Carnarvon Basin.

As with the Perth Basin, rank data from the Carnarvon Basin show a degree of provincialism which can be related to tectono-sedimentary setting.

(1) The Onslow-1 well (Peedamullah Shelf) is located on an upthrown block of Triassic and Permian strata (showing little evidence of deep burial) overlain by 600 m of Cretaceous and Tertiary sediment. The Peedamullah Shelf overlies Proterozoic basement and has a high present-day temperature-gradient regime. It seems likely that a combination of greater sediment age and relatively high burial temperatures has led to vitrinite reflectances being offset towards higher ranks as compared with data from the Barrow Sub-basin.

(2) The Barrow Sub-basin proper comprises a very thick section of Jurassic sediments overlain by up to 2000 m of Cretaceous section. Sedimentation was essentially continuous throughout the Mesozoic, and progressive subsidence of the continental shelf

from the Late Cretaceous to the present time has allowed the accumulation of up to 3000 m of Tertiary carbonates. Reflectance profiles, as illustrated by Barrow Deep-1 (Fig. 16), are characterized by low gradients down to 0.5% R_o which increase thereafter and are invariably higher than those from equivalent parts of the Perth Basin such as the Dandaragan Trough.

(3) The Rankin Platform consists of several en echelon fault blocks containing Triassic sediments overlain by Cretaceous and Tertiary shales and carbonates. Structural development began in the late Middle Jurassic and continued during the Early Cretaceous. Late Cretaceous and Tertiary subsidence and marine sedimentation led to differential compaction and drape over the pre-existing features. Most reflectance data plot in the broad province represented in Fig. 15, while Fig. 17 shows the spread of reflectance values observed from the Rankin-1 well. Depressed vitrinite reflectance relative to the Barrow Sub-basin relates to the Late Cretaceous foundering and Tertiary burial of the formerly emergent Rankin horst block as well as to an offshore decrease in present-day temperature gradient.

The schematic section of isoreflectance variation shown in Fig. 14 demonstrates that: depth to the various isoreflectance surfaces increases in an offshore direction; rank increases systematically with depth over the line of the section; and rank surfaces are strongly diachronous.

The implication is that observed coalification in the Carnarvon Basin is primarily a function of the Tertiary thermal regime.

Estimated palaeotemperatures

Table 1 includes a compilation of vitrinite reflectance, present temperature, and age data for deep wells in the Carnarvon Basin. The range of temperatures at which 0.5% R_o occurs is about 20°C, despite differences in age of some 160 m.y. At 0.7% R_o the range decreases

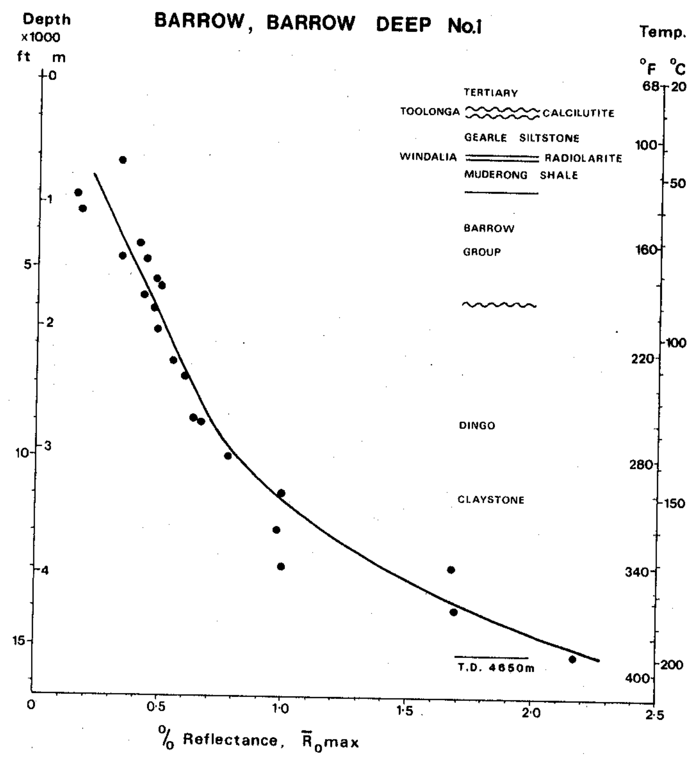


Figure 16. Depth-reflectance profile for the Barrow Deep-1 well, Barrow Sub-basin.

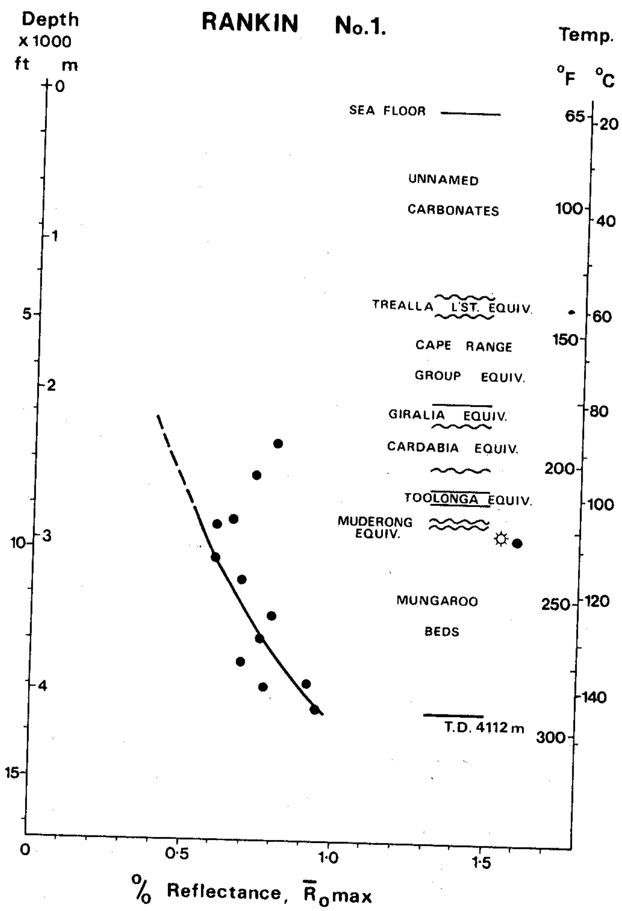


Figure 17. Depth-reflectance profile for the Rankin-1 well, Rankin Platform.

to less than 15°C and 140 m.y., while at 0.9% R_o it is only 5°C and 120 m.y.

Table 2 presents estimates of vitrinite reflectance using the T_{eff} approach of Hood *et al.* (1975) as modified by Bostick *et al.* (1979) and applied further to the nomogram of Karweil (1956) as modified by Bostick (1973). These estimates show quite good agreement with observed reflectances on the Rankin Platform but become increasingly disparate shorewards. The sense of the difference indicates a late rise in temperature, particularly over those parts of the Carnarvon Basin not subjected to deep Tertiary burial.

Table 2 also includes estimates of present temperature from the vitrinite reflectance data assuming:

Isothermal conditions (Tiso) — where a constant temperature is assumed over the entire age of the sample.

Gradthermal conditions (Tgrad) — where temperature is assumed to have increased at a uniform rate over the entire period of burial up to a present-day maximum.

A gradthermal approach using the modified Hood *et al.*/Bostick *et al.* nomogram.

Comparison of these data shows that Tpres > Tgrad throughout the basin, reinforcing the concept of a late rise in temperature. The lack of significant variation in Tpres minus Tgrad between wells contrasts with the data for the Perth Basin.

TABLE 2

Estimated vitrinite reflectances using the Teff approach of Hood *et al.* (1975) together with estimates of present temperature based on a gradual increase in temperature (Tgrad) over the entire age of the rock.

					Est. Coalification (% R_o) using Teff			Temp (°C) to produce observed rank if acting over entire age		
Well Name	Depth (m)	Age	Tpres (°C)	R_o max %	Karweil	Hood	Teff m.y.	Tiso °C	Tgrad °C	Hood °C
Onslow-1	1738	E Tr	90	0.48	1.00	0.78		< 20	< 30	45
	2209	M Perm	109	0.64	1.45	0.88	100	38	55	71
	2899	E Perm	136	0.86	2.05	1.30		59	90	95
Anchor-1	1898	E Jur	93	0.50	0.80	0.68		< 20	< 30	50
	2671	E Jur	134	0.72	1.70	1.15	70	56	85	87
Pepper-1	2218	E Cret	106	0.50	1.10	0.80		< 20	< 30	55
	2453	L Jur	117	0.63	1.30	0.90	70	50	75	78
Barrow Deep-1	2610	L Jur	121	0.60	1.40	0.92		46	68	72
	3354	L Jur	149	1.00	2.30	1.45	70	84	128	118
	4299	L Jur	186	1.70	3.70	2.35		113	172	156
North Tryal Rocks-1	3199	L Tr	110	0.61	0.85	0.78		38	55	70
	3656	L Tr	123	0.72	1.03	0.88	40	50	75	82
Rankin-1	3107	L Tr	117	0.60	0.59	0.66		37	53	68
	3650	L Tr	134	0.75	0.65	0.79	10	53	80	84
	4109	L Tr	149	0.94	0.75	0.95		69	104	105

Relationship of maturation history to hydrocarbon occurrences

On the Rankin Platform the major discoveries of gas-condensate and minor oil occur in Triassic sands (with very good reservoir characteristics) at vitrinite reflectances of 0.50% to 0.65% R_o. The rank data, together with information on organic-matter type, indicate that the overlying Lower Cretaceous claystones are marginally mature and a likely source of oil whereas the Tertiary sequence is wholly immature and has poor source potential. Upper Jurassic claystones in the flanking Barrow and Dampier Sub-basins are mature to post-mature and are generally regarded as having sourced hydrocarbons which migrated into structurally higher traps such as the Triassic horsts of the Rankin Platform. However, we consider the Triassic sequence on the Rankin platform itself to be an equally effective source for gas in terms of both rank and organic-matter type — a view to which Crostella and Chaney (1978) also subscribe. A feature of the mature hydrocarbon occurrences on the Rankin Trend is the low rank of the reservoir rocks, implying substantial migration of oil and gas from more mature source rocks at depth. Both the Jurassic and Triassic claystones have adequate source potential and a dual source is likely.

The main production of oil in the Barrow Sub-basin is obtained from Early Cretaceous sands (Windalia and Birdrong Members of the Muderong Shale) of the post-breakup sequence at Barrow Island. While temperatures at Barrow Island may be adequate for the generation of oil (65°C), indications are that the Muderong Shale there is immature. Available evidence suggests that the Muderong Shale and overlying Gearle Siltstone are good to adequate source rocks and that they reach initial maturity (0.50% R_o) at generative temperatures (80° to 90°C) over large parts of the Barrow Sub-basin west of Barrow Island. Migration of oil is facilitated by the permeability and continuity of the Birdrong Sand Member.

The oil in the Upper Jurassic/Lower Cretaceous Dupuy Member of the Dingo Claystone is found at 0.5% R_o and 100°C at Barrow Island, at 0.6% to 0.7% R_o and 100°C at Pasco. The Dingo Claystone is mature over a large area of the Carnarvon Basin and the present-day high temperatures have accelerated the generation of both oil and gas.

Conclusions and comparisons

(1) Maturity-temperature relationships for the two basins are quite different (see Fig. 18 and Table 1). On average, temperatures in the Carnarvon Basin wells are some 40°C above those from the Perth Basin at equivalent reflectance (maturation) levels. Alexander *et al.* (1979) have observed a similar difference between the two basins based on the percentage of aromatic protons in the aromatic fraction of source rock extracts. Figure 19 indicates that this major difference cannot in general be attributed to differences in the age of the sequences, although uplift may be a factor of significance in the Perth Basin.

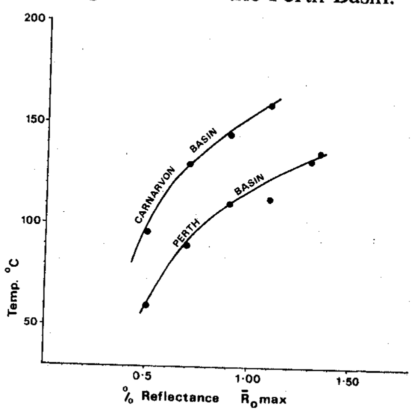


Figure 18. Relationship between average temperature and reflectance in the Carnarvon and Perth Basins.

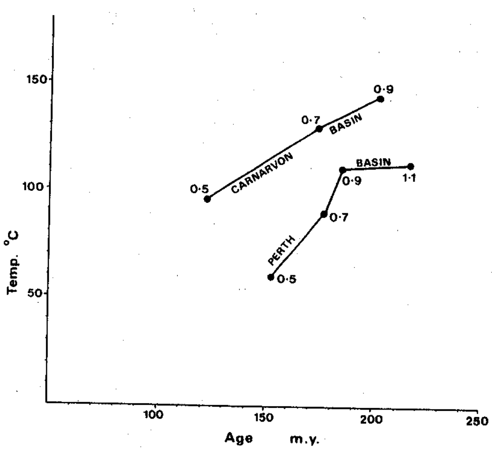


Figure 19. Relationship between average temperature and average age at various vitrinite reflectance levels in the Perth and Carnarvon Basins.

(2) The magnitude of the difference in the maturity-temperature relationships, in the absence of a compensating difference in sediment age, implies a major difference in the overall timing of temperature rise and the main phase of coalification in the two basins. As there is no profound difference in source rock quality and availability, this timing of the rise in temperature is considered responsible for the difference in hydrocarbon productivity between the two basins.

(3) Source rocks in the Perth Basin underwent an earlier rise in temperature (associated with a Triassic thermal event and deposition of the thick Late Jurassic Yarragadee Formation) with concomitant early coalification. Until recently, the basin appears to have been influenced by an attenuating thermal regime. The recent major rise in burial temperatures, however, is of far greater significance in the Carnarvon Basin. On the Rankin Platform this rise in temperature can be attributed to Tertiary subsidence, which has brought the older sediments to maturation levels and temperatures suitable for the prolific generation of hydrocarbons. Onshore, where Tertiary subsidence is minimal, this temperature rise is equally pronounced but is less easily explained. It may relate to lateral dewatering of the basin.

(4) The relatively early phase of coalification and hydrocarbon generation in parts of the Perth Basin leads us to believe that hydrocarbons may have been reservoired for long periods of time with leakage at rates exceeding those of generation and supply. In contrast, the Carnarvon Basin currently appears to be in its most active phase of hydrocarbon generation because of a strong thermal drive, and therefore has a more favourable accumulation balance.

(5) Slow rates of source-rock maturation at relatively low generative temperatures, together with local sources from mature and overmature sequences, may contribute to the high GOR of the Perth Basin. In the Carnarvon Basin the high GOR may relate to mature source rocks being exposed to very high (130°C to 150°C plus) generative temperatures in areas adjacent to the principal trap (the Rankin Platform). Source-matter

type is not thought to exert a major control over GOR, since prolific oil generation from the coal-bearing Latrobe Group in the Gippsland Basin occurs at low ranks (0.5% to 0.7% R_0) and at temperatures between 70°C and 120°C.

(6) The Triassic sequences which contain the major reservoirs in both basins contain good to adequate souce rocks, which are generally late mature to overmature in the Perth Basin but early mature to mature in the Carnarvon Basin. Differences in reservoir quality may be partly rank related and may have contributed to the enhanced accumulation of hydrocarbons in the Carnarvon Basin. However, ultimate constraints on the size of hydrocarbon accumulations are structural and stratigraphic.

(7) Both the Perth and Carnarvon Basins represent trailing-edge continental margins — a basin type supposedly characterized by normal to low temperature gradients (following spreading of the sea floor to oceanic distances) attenuated from a possible high of 60°C to 70°C/km at the inception of sea-floor spreading (Klemme 1975; Siever 1979). While this view can be extended to the Perth Basin, it contrasts with data from the Carnarvon Basin. We would suggest that exploration strategies acknowledge significant variation from such broad trends, since rapid changes in thermal regime and maturation patterns can and do exist over short distances — especially in the Palaeozoic and Mesozoic Basins.

ACKNOWLEDGEMENTS

The authors wish to record their appreciation to the Bureau of Mineral Resources, Australia; the Department of Mines, Western Australia; Offshore Oil N.L., and West Australian Petroleum Pty. Ltd. for supplying the sample material used in the compilation of this paper. The Australian Petroleum Exploration Association Limited kindly allowed us to reproduce and modify figures previously published in the APEA Journal.

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UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 197-219

GEOTHERMAL OCCURRENCES IN THE SOUTHWEST PACIFIC

MALCOLM E COX

ABSTRACT

Geothermal activity occurs on many islands in the Southwest Pacific. The surface expression of this activity ranges from low-temperature seepages to boiling springs and vigorous fumarolic and solfataric activity. Some are related to currently active volcanic systems, and others are in areas in which volcanism has ceased. Many systems are water-dominated, but in some a vapor phase is significant and vapor-only systems are found associated with active volcanism. The estimated temperatures of thermal fluid reservoirs range from around 90 to over 300°C and can generally be related to the geological and tectonic location of each island group. Although most of the thermal fluids are of meteoric origin, their chemistry is varied, largely owing to mixing of seawater and shallow groundwater and to steam heating. Water chemistry is also controlled by the local geology and, in some areas, a magmatic vapor content. The highest estimated reservoir temperatures occur in active volcanic provinces associated with plate margins. A notable exception are the intra-plate Hawaiian systems. The lower-temperature systems are caused by apparently elevated thermal gradients, localized remnant heat in intrusive bodies, or heat associated with tectonic uplift. Based on their chemistry, the thermal waters can be divided into three groups characteristic of vapor systems and high-temperature and low-temperature water systems.

INTRODUCTION

This paper summarizes the recent volcanic history and geology of the countries along the Pacific margin from New Zealand to eastern Papua New Guinea, and describes the numerous, but generally not well known, geothermal occurrences within each area. It outlines some features of the form and distribution of terrestrial heat within the region and the characteristics of the geothermal systems that have developed. A description of an Hawaiian geothermal system is included because of the uniqueness of that occurrence and its Pacific location. It should be noted that 'historic' volcanism in the areas discussed usually covers a period of less than 150 years.

REGIONAL SETTING

The geothermal-volcanic activity of the Southwest Pacific discussed occurs along the marginal zone of the Pacific and Australian crustal plates. This is a region of diverse tectonic features that include crustal subduction and associated island-arc formation, backarc basins, and elevated areas of sea floor in which new crust is being formed, as well as continental-type land masses. The zone is characterized by belts of intense seismic activity: shallow earthquakes (<70 km) occur below many of the land masses and adjacent shallow basins (Fig. 1) and are related to zones of crustal consumption and formation and large-scale crustal faulting.

At the Tonga-Kermadec Trench the Pacific Plate is descending westward below the Australian Plate, and from the New Hebrides to at least New Britain, and possibly into the Papua New Guinea mainland, the Australian Plate is descending north or northeastward beneath the Pacific Plate. Within the Tonga-Kermadec Trench system, Pacific Plate consumption diminishes southward, and the pole of rotation for the convergence of the Pacific and Australian Plates lies to the southeast of New Zealand (172°W, 58°S, Chase 1971; 179°W, 56.2°S, Johnson and Molnar 1972). To the south of New Zealand, along the ENE-trending northern Macquarie Ridge, the Australian Plate underthrusts the Pacific Plate (Johnson and Molnar 1972).

A consequence of these and associated tectonic processes is that most of the landmasses along this zone are of island-arc formation with a volcanic history during the Tertiary and Quaternary. As this is a dynamic tectonic zone, there is an appreciable amount of current volcanic activity. Many of the island groups involved were largely formed by andesitic volcanism, but contemporaneous or later episodes of basaltic volcanism were common and were often associated with crustal migration. The composition of the lavas and the shallow subsurface temperatures can in many cases be related to the depth to the Benioff zone. Basaltic material is now considered to be intruding some of the shallow basins, where it is forming new oceanic crust. This sea-floor formation is occurring in both backarc and marginal basins with the development of numerous, young lithospheric plates (Chase 1971; Johnson and Molnar 1972; Luyendyk *et al.* 1973). These areas are complex in terms of plate tectonics (notably north and south of New Britain and east and west of Fiji) and are areas for which many models, both of components and mechanisms, continue to be presented as more evidence is obtained. These shallow basins have variable heat flow, which is often high (to >10 HFU (μcal/cm^2/sec)) in localized zones which suggests point or linear sources of mantle intrusion, and thermal elevation (Sclater and Menard 1967, Sclater *et al.* 1972, Halunen and von Herzen 1973, Hawkins 1974, 1976).

Many of the terrestrial and submarine volcanoes are associated with crustal fractures and commonly occur at fracture intersections. In some localities these terrestrial structures continue into the adjacent shallow basins and are there associated with sea-floor igneous intrusion.

VOLCANIC HISTORY OF THE REGION

In New Zealand (Fig. 2) the earliest evidence for volcanic activity is in the South Island, where andesitic and basaltic eruptions probably began during the Early Cambrian

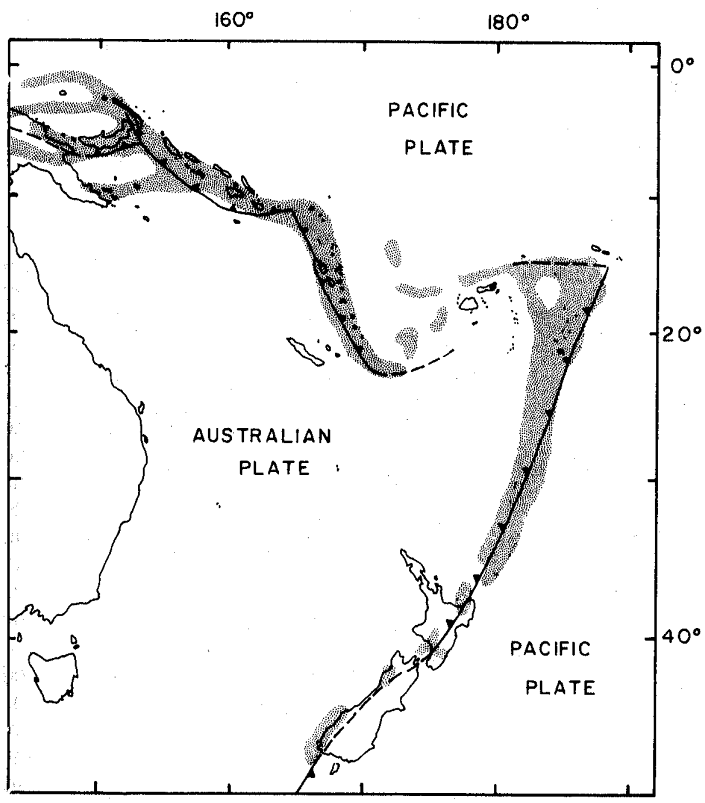

Figure 1. Location of areas of crustal consumption and direction of subduction (triangles); major faults (?transcurrent) (broken lines). Generalized zones of shallow seismicity (< 70 km) are shown by shading (after Johnson and Molnar 1972, Luyendyk *et al.* 1973, Tarr 1974).

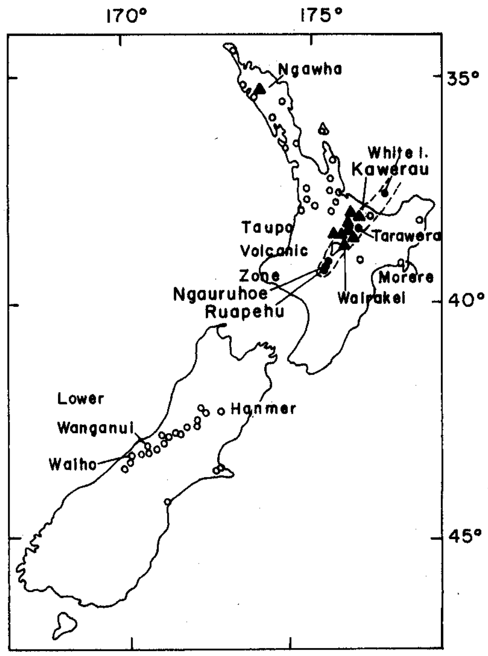

Figure 2. Locations of hydrothermal occurrences in New Zealand. The Taupo Volcanic Zone is outlined (broken line). Sites of active or historically active eruptions, closed circles; high temperature hydrothermal occurrences, triangles; low temperature thermal springs, open circles. (After Mahon 1964, Grindley and Williams 1965.)

(from Brown *et al.* 1968). Activity continued intermittently in the South Island throughout the Paleozoic and probably began in the North Island during the Permian. Mesozoic volcanism occurred on both islands, with considerable activity during the Early Cretaceous in the north of the North Island.

From the Early Tertiary until the Late Pliocene or Early Pleistocene, basaltic volcanism continued in the South Island, with trachytic then phonolitic flows in the southeast from Late Miocene to about Early Pleistocene. Activity was considerable in the Northland Peninsula area of the North Island in the Late Tertiary, and was largely andesitic and basaltic during the Miocene. In Pliocene times, dacitic, followed by rhyolitic, eruptions took place in eastern Northland; eruptions in western Northland were basaltic. By the Late Pliocene or Early Pleistocene, basaltic and andesitic volcanoes were very active in the central North Island.

Quaternary tectonic activity was accompanied by considerable magmatic mobility and volcanism. Alkaline basaltic eruptions continued on the Northland Peninsula, but volcanism in the North Island began to migrate southward, and most Pleistocene activity took place in the Taupo Volcanic Zone, where the earliest eruptions were andesitic. Late Pleistocene and Holocene basaltic eruptions occurred in the central north (Auckland city region). In the west of the North Island, andesitic and dacitic volcanism moved southwards to Mt Egmont, ending during the Holocene.

Current activity is within the Taupo Volcanic Zone (from Healy 1962), a graben-like linear depression filled with volcanic material on basement greywackes of Mesozoic and Paleozoic age. Faults, many of which are active, have a common northeast trend and the zone appears to be broken into tilted fault blocks. Rhyolitic and ignimbritic lavas and pyroclastics were erupted from the central part of the zone and are considered the result of shallow melting within the crust; andesites, basaltic andesites and basalts are found at the ends. The andesitic volcanoes at the southwest of the zone and the volcano of White Island to the northeast are currently active. White Island has had numerous eruptions this century, with intervening strong fumarolic activity, during which steam from one fumarole was measured at 750 to 800°C (Clark and Cole 1976). The volcano has been active since the end of 1976; it erupted explosively in March 1978 forming a collapse crater (McClelland and Simkin 1979) and has continued this explosive activity into 1979 with a maximum fumarole temperature of 400°C (L McClelland, personal communication). Ngauruhoe, an andesitic stratovolcano in the southwest of the Volcanic Zone, erupted explosively in 1972, 1973 and 1974 (Nairn *et al.* 1976). Ruapehu erupted in 1975 and experienced a phreatomagmatic eruption in mid 1978. In the central rhyolitic part of the Volcanic Zone the shield volcano of Tarawera erupted in 1886.

The *Kermadec Islands* (described by Richards 1962) comprise four volcanoes on the Kermadec Ridge about 950 km SSW of the Tongan Islands. The southernmost, Curtis Island, is part of an ancient crater and was described as being in a vigorous solfataric stage with areas of boiling water and mud and deposits of sulfur and siliceous sinter. Brimstone Island is a largely submerged volcano, whose first recorded activity dates back to 1825; since then it has been in a fumarolic state. Raoul Island, a compound island volcano of dominant andesitic and some basaltic flows, erupted in 1872 (Richards 1962) and 1964 (source not known); fumarolic activity has been continuous between the eruptions and since 1964. The northernmost volcano in the Kermadecs is submarine, and its last recorded eruption was of pumice, in 1886.

The islands of *Tonga* (Fig. 3) on the northern part of the Tonga-Kermadec ridge are two parallel chains formed by volcanism presumably related to fractional melting of the descending Pacific Plate (Ewart and Bryan, 1972). The western chain is composed of largely andesitic volcanic islands with limited

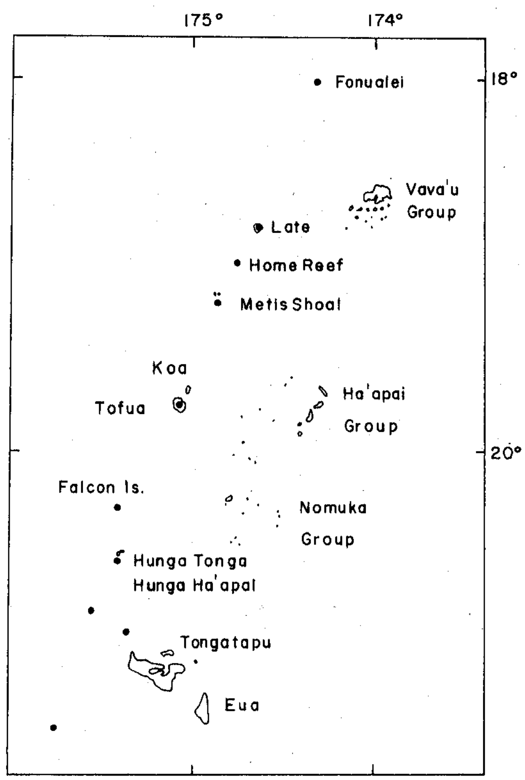

Figure 3. Locations of active or historically active volcanism in Tonga (closed circles). No low-temperature thermal water discharges are reported in the literature cited (after Richards 1962, Bryan *et al.* 1972).

reef formation, and the older eastern chain of elevated islands of both volcanic rocks and well developed reef limestone. Most of the Tongan volcanic rocks are basaltic andesite, and andesite and basalt with some dacite, and are commonly overlain by pyroclastics (Bryan *et al.* 1972). Most of the pre-caldera activity of the western chain was during the Pleistocene, but many lavas are of Holocene age. The oldest known rocks are found in the eastern chain, where basalts occur within pre-Eocene volcanics.

The eastern chain is now volcanically inactive. Possibly, volcanic activity took place along it during the Early Miocene and into the Late Miocene or Early Pliocene and has migrated westward (Katz 1976, 1978). Most of the islands and submarine volcanoes of the western chain have been historically active. Bryan *et al.* (1972) estimate a possible periodicity of volcanic events, other than fumarolic activity, of 20 to 50 years. Summarizing the volcanic activity of the western chain (based on Richards 1962): Fonualei, a strato-volcano, last erupted in 1939; this activity was followed by fumaroles at 88°C discharging SO_2, and several surrounding areas are still hot and steaming and have small sulfur deposits; Late has no recorded eruptions from the main crater but an 1854 eruption from a side crater was followed by weak fumarolic activity; Home Reef, a submarine volcano, erupted in 1852 and probably in 1857; Metis Shoal built up as a submarine volcano from 1852 to 1894, after which time it submerged, but erupted again in 1967 and mid 1979; Kao, the largest volcano, has no recorded eruptions, but the composite volcano of Tofua is recorded as having erupted in 1774 and 1959, and as having subsequent fumarolic activity. The submarine volcano of Falcon Island has a history of emergence and submergence; it is last recorded to have erupted in 1936 and to have had minor fumarolic activity in 1941 and to have been entirely submerged in 1959. Hunga Ha'apai and Hunga Tonga are subaerial remnants of an active cone that erupted in 1912 and again in 1937 and possibly now has minor submarine fumarolic activity. Of two submarine volcanoes to the northwest of Tongatapu, that farthest northwest erupted in 1923 and 1970, the other in 1943. A submarine volcano southwest of Tongatapu erupted last in 1932. Niua F'ou, in the northern Lau Basin last erupted in 1946; its basalts are similar to those in the Lau Basin (Gill 1976). In 1973 a submarine eruption of dacitic pumice took place near Curacoa Reef at the northern end of the Tonga arc (Latter 1976); there have been two subsequent eruptions.

In the *Samoan islands* (Fig. 4) volcanism along the regional rift of the Samoan Ridge has created large shield volcanoes with rift zones and smaller shields and cones along them (Stice and McCoy 1968). Hawkins (1976) considered that the alkalic and strongly silica-undersaturated lavas produced may be from a magma source close to the base of the lithosphere (80 to 90 km) and are associated with a rupture leakage of the deformed Pacific Plate. Along the ridge are three major volcanic piles: (1) Savai'i and Upolu, (2) Tutuila, and (3) the Manu'a Islands. Volcanism in Samoa can in general be considered to have moved east to west (Kear 1967, Stice and McCoy 1968), with activity being greatest from Pliocene to Middle Pleistocene and decreasing by Late Pleistocene to Holocene.

The volcanics of Savai'i and Upolu in Western Samoa are predominantly alkalic basalt, olivine and picrite basalts and olivine dolerite, of Middle to Late Pleistocene and Holocene age. Most of the smaller offshore islands are formed of palagonite tuff of the same age (Kear 1967). These rocks overlie the older, Pliocene or Early Pleistocene, hornblende andesites and trachytes, which are best developed on Upolu. The volcanic cones on both islands follow distinct structural alignments, but on Savai'i numerous flank eruptions also took place. No historic eruptions are recorded on Upolu, but on Savai'i three volcanoes have erupted in historic times (Mauga Afi, Mauga Mu (Aopo) and Matavanu), the last eruptions in 1760, 1902 and 1905 to 1911, respectively. The largest island of American Samoa is Tutuila, which is formed of alkalic volcanic rocks, largely olivine basalts with an associated trachyte cap. Five principal volcanic structures are on a nearly linear trend, the youngest being of Holocene age and the others Pliocene or Early Pleistocene (Stearns 1944). In the Manu'a Group (Ofu, Olosega and Ta'u), which is mostly olivine basalt and picrite basalt, volcanism has continued to the Holocene, and a submarine eruption occurred between Olosega and Ta'u Islands in 1866 (Richards 1962, Stice and McCoy 1968).

In the *Fiji island group* (Fig. 5) most volcanic activity was during the Tertiary, but in a north-central area of the group it extended well into the Quaternary. There were three periods of volcanism: the first tholeiitic (southern Viti Levu), the next predominantly calc-alkaline (central and southeast Viti Levu, central and eastern Vanua Levu, Kadavu,

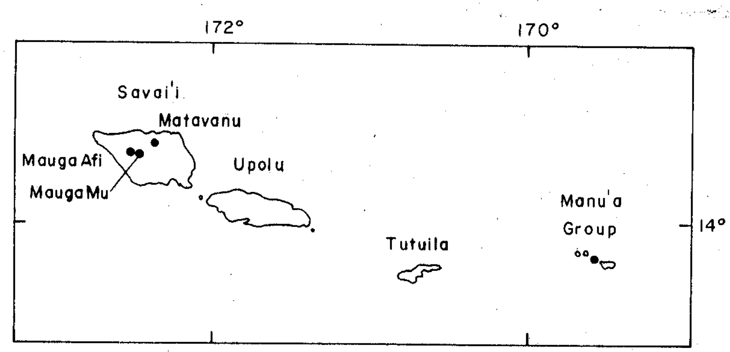

Figure 4. Locations of active or historic volcanism in the Samoan Islands (closed circles). No low-temperature discharges are noted in the literature cited, but the existence of well developed rift zones on the larger islands is considered to indicate a suitable environment for the formation of thermal fluid reservoirs similar to that on the lower east rift of Kilauea, Hawaii.

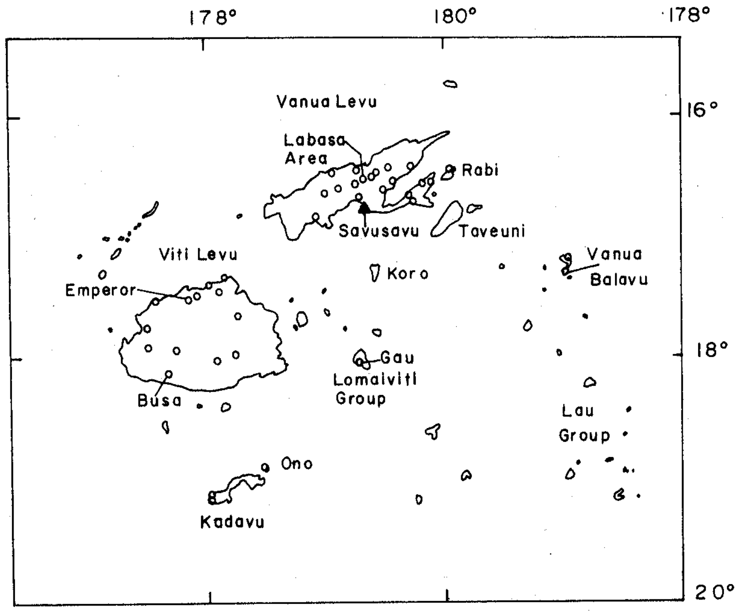

Figure 5. Locations of thermal springs in the Fiji Island Group. All are typically low temperature (open circles) except those at Savusavu (triangle). The northeasterly structural trend can be seen in Vanua Levu and in the location of thermal springs on the small islands from Kadavu to Vanua Balavu. The age and rift structure of the island of Taveuni also suggest a suitable environment for the formation of a thermal reservoir.

Lau), the youngest, basaltic (northern Viti Levu, western Vanua Levu, Taveuni, Lomaiviti, Lau). Much of the volcanic activity was in an island-arc environment, and many of the volcanic centers are associated with crustal structures.

The oldest known rocks of Viti Levu are Eocene to Middle Miocene volcanics, of which the lower section is basaltic in composition, and the upper section largely andesitic, but ranging from basalt to dacite (Rodda and Band 1967). This Group forms much of southern Viti Levu and is intruded by gabbro, biotite tonalite and trondhjemite of Early Eocene and Late Miocene age. In places it is unconformably overlain by Upper Tertiary andesitic to rhyolitic volcanics and volcaniclastics and several marine sedimentary groups. By the Pliocene, there was major basaltic activity at five or more large volcanic centers in northern Viti Levu. Minor differentiated andesitic volcanoes were also formed, and there was some later-stage trachybasalt and trachyandesite formation in some of the larger centers (Rodda 1967). Many of the smaller, largely basaltic, island volcanoes around Viti Levu were also formed during this volcanic period, as were numerous intermediate intrusions (Rodda 1967). The islands of Kadavu to the south are largely andesitic with minor basalt and are considered to be of Pliocene age (Woodrow in press).

The second largest island, Vanua Levu, is formed of three coalesced Tertiary volcanic fields, where there was activity at at least ten main volcanic centers. A basal volcanic group (Middle Miocene to Lower Pliocene) forms much of the center and east of the island and consists largely of basic andesitic volcanics and volcaniclastics. Eruptions in the northwest were of basic and acid andesites, and activity in the northeast formed dacitic to rhyolitic volcanics and pyroclastics. Upper Pliocene activity in western Vanua Levu, contemporaneous with that in northern Viti Levu, formed a large alkali-basalt shield volcano. The youngest volcanism in Fiji was during the Early-Middle Pleistocene and into the

Holocene, with the formation of the alkalic olivine basalt islands of Taveuni and Koro. Eruptions on Tavenui took place from at least 120 centers along a NE–SW rift zone that parallels major crustal structures below Vanua Levu (Ibbotson and Coulson 1967). In the Lau group of islands, the outcropping basement rocks are Late Eocene to Early-Middle Miocene, largely acid andesites and dacitic flows with some basaltic andesites. These are in part overlain by Late Miocene to Pliocene flows and breccia, largely olivine basalts with minor andesitic differentiates (Ladd and Hoffmeister 1945, Gill 1976).

There is no fumarolic activity in Fiji, and any expression of elevated subsurface temperatures is restricted to thermal-water discharges. Most of these are low-temperature, 40 to 60°C, but boiling springs are found in two areas of Vanua Levu (Fig. 5).

New Caledonia (Fig. 6), one-third of which consists of an ultramafic complex, is of island-arc formation and largely of Late Jurassic to Early Cretaceous age. Avias and Tonard (in Dubois *et al.* 1974) described the oldest identified formations as pre-Permian, and proposed a geological history of marine sedimentation during the Permian, followed by a Jurassic orogenic stage and the formation of Upper Jurassic conglomerates and schists. After sedimentation in the Cretaceous, Eocene rocks were formed (mainly in the west) consisting of sediments, and basalt and dolerite flows. Another orogenic period ensued during the Oligocene, with emplacement of peridotites and compressional folding, followed in Miocene times by deposition of limestone and other sediments.

Several groups of low-temperature thermal springs are found in the southeastern third of the island.

The *Loyalty Islands* (Fig. 6) are basaltic volcanic islands capped by reefs. The last major outflows are of Late Miocene age, but there may also have been Late Oligocene volcanism (ORSTOM *in* Dubois *et al.* 1974). The islands are considered to be a volcanic arc associated with the formation of New Caledonia (Dubois *et al.* 1974).

The *New Hebrides-Santa Cruz* island groups (Fig. 6) are formed by two converging chains that merge at about 17°S and are separated by a

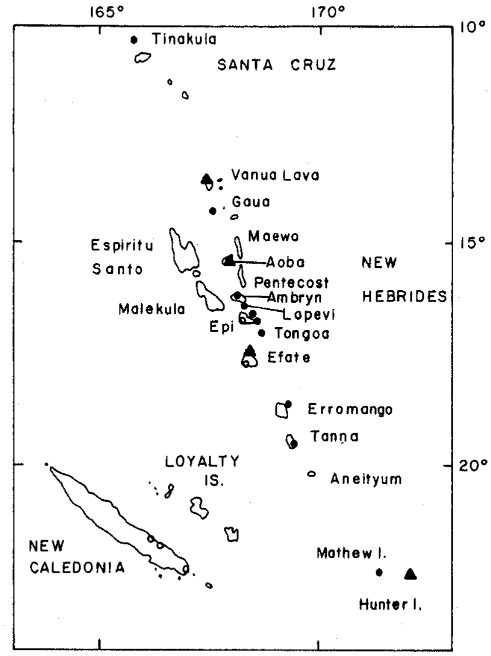

Figure 6. Location of hydrothermal occurrences in New Caledonia and the New Hebrides-Santa Cruz. Active or historic eruptions (closed circles); high-temperature hydrothermal occurrence (triangles); low-temperature thermal springs (open circles). (After Fisher 1957, Coleman 1970.)

narrow extensional zone (Karig and Mammerickx 1972). Active volcanism in this central zone continues to the south. The oldest rocks are in the west of the group and are probably Eocene, but there are common pre-Miocene andesitic and basaltic lavas, flow breccias, pillows and some pyroclastics. Tuffs are abundant. Diorite and gabbro intrusions and ultramafic rocks are present on the large islands such as Pentecost and Santo. Intermittent volcanism and intrusion probably took place through the Miocene, finding their expression mainly in basalts and andesites. The eastern belt is intensely fractured; block-faulting has a pronounced east-west orientation and volcanic vents are sited along fractures and especially at fracture intersections (Coleman 1970). Rocks of the northern islands are mainly basalts, but those of the center and south are olivine basalts, basaltic andesites, andesites and pyroclastics (Warden 1967).

Volcanism generally becomes less mafic with time and some recent eruptions have included dacitic pumice and clasts of rhyolite (Coleman 1970). The active, predominantly basaltic volcanoes in the northern interarc basin are younger than those in the south of the chain but have also produced mildly alkaline basic rocks (Colley and Warden 1974).

Tinakula in the Santa Cruz Group last erupted in 1951; Vanua Lava and Gaua (Banks Group) and Aoba all have solfataric activity and areas of thermal springs, and Gaua recently erupted ash; Ambrym, a very active strato-volcano, has been intermittently active from 1958 and experienced new explosive activity in early 1979; Lopevi was intermittently active from 1939 to 1960; submarine volcanoes occur east of Epi (erupted 1953) and north of Tongoa (eruptions 1949 and 1952, followed by fumarolic activity) and Tongoa has solfataric areas with gas temperatures to 100°C. Both Epi and Efate have thermal springs. A submarine volcano near Erromango was possibly active in 1881 and a strato-volcano on Tanna has been almost continuously active since 1774. At the south of the arc is basaltic Mathew Island, which was active in 1953 and is still degassing, and Hunter Island, which has minor solfataric activity and may have erupted in 1903 (Fisher 1957, Williams and Warden 1964).

The *Solomon Islands* (Fig. 7) contain a broad range of rock types, but there are two major groups, the Cretaceous-Miocene ultramafics and mafics, and the Pliocene and younger island arc tholeiitic and calc-alkaline series which form a series of strato-volcanoes. Coleman (1970) divided the Solomons into four provinces: (1) the Northern and Southern Atoll provinces; (2) the Pacific province (Malaita and Ulawa), which contains the oldest rocks (highly mafic submarine basalts), which are overlain by Upper Cretaceous deep marine sediments; and (3) the Central province, with the islands from Choiseul to San Cristobal, which have an igneous and

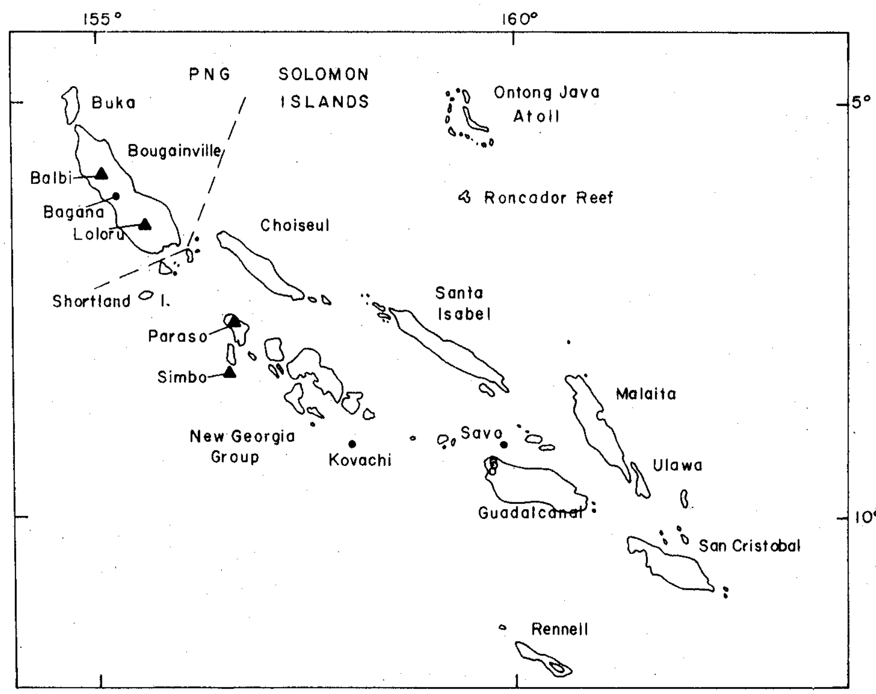

Figure 7. Location of hydrothermal occurrences in the Solomon Islands and Bougainville Island (PNG). Active or historic volcanoes (closed circles); high-temperature hydrothermal systems (triangles); low-temperature thermal springs (open circles). (After Fisher 1957.)

metamorphic Mesozoic basal complex of chloritic 'greenstones', amphibolitic schists and basic lava. Coleman (1970) also suggests the probable existence of small bodies of gabbroic and granitoid rocks and notes that the basal complex is intruded and overlain by Upper Eocene-Oligocene volcanics, largely of basaltic composition, but with some andesites. Ultramafic rocks possibly of Oligocene age occur on the main islands of this province. The fourth, the Volcanic Province, contains the youngest volcanic rocks on the Shortland Islands, New Georgia Group, Russel Group, Savo and northwestern Guadalcanal. Geologically, this province extends to the northwest and includes Bougainville and Buka islands of Papua New Guinea. The volcanic islands are characterized by calc-alkaline volcanism and intensive faulting. Of note is the NE-SW alignment of volcanic centers in the New Georgia Group (Taylor 1976a). Pliocene-Pleistocene and Holocene volcanic centers extend throughout this group and most of the volcanic rocks are olivine basalts and pyroclastics and basaltic andesites. A diorite stock intrudes the lavas of East New Georgia Island.

Current activity is at the submarine volcano Kovachi southeast of the New Georgia Group, and on Savo Island. The former erupted in 1952 to 1953, built a temporary island in late 1976 and also erupted in mid 1977 and again in mid 1978. Savo Island last erupted in 1847 but now has extensive fumaroles and thermal springs within and around the crater. Hydrothermal activity as fumaroles and springs occurs in four main locations in the Solomon Islands: Paraso on Vella Lavella; on Simbo Island; associated with the volcano of Savo Island; and in northwest Guadalcanal.

Bougainville Island, politically within Papua New Guinea, is geologically an extension of the Volcanic Province of the Solomon Islands and is within the same island-arc environment (Fig. 7). The island has a dominant axial chain and is mainly composed of Upper Tertiary and Quaternary volcanics and pyroclastics and their derived sediments, plus Lower Miocene and Pleistocene limestone. Several dioritic to granodioritic intrusions of Miocene to Late Pliocene age occur. Pliocene to Holocene volcanics, calc-alkaline lavas, and pyroclastics, mostly of andesitic composition but with some dacites, cover more than half the island (Bultitude 1976). A central volcano, Bagana, was active from 1971 to 1975 and late 1976 to 1978; since 1947 it has been almost continuously active. Two other volcanoes, Balbi and Loloru, are considered to be dormant, and within their crater complexes vigorous fumarolic activity discharges H_2S and SO_2 at maximum temperatures usually of 95°C (Fisher 1957, Bultitude 1976).

A series of Upper Cenozoic island volcanoes lies off the northeast coast of ***New Ireland*** (Fig. 8), on two of which (Lihir and Ambitle) are areas of solfataric activity and thermal springs. Gas and spring temperatures are commonly 70 to 101°C and some H_2S is discharged (Fisher 1957).

From eastern ***New Britain*** to off the northeast coast of the Papua New Guinea mainland, a chain of volcanoes forms the New Britain-New Guinea arc (Fig. 8), which is considered to be formed of separate western and eastern arcs. The western arc is largely andesitic (Heming 1974) and the most common rocks are hypersthene-normative basalts and low-silica andesites (Johnson 1976). The eastern arc (essentially New Britain island) is composed mainly of Upper Tertiary and Pleistocene volcanic and intrusive rocks and limestone. Basalts are much less common than in the western arc and there are more common high-silica andesites and dacites and rare rhyolites. The eastern arc volcanism, however, is not all typically calc-alkaline, and Heming (1974) noted penecontemporaneous andesitic, dacitic, and basaltic rocks in the Rabaul caldera. Johnson (1976) considers the rocks of both arcs to be chemically similar. The trend of the volcanic centers along the arc is not entirely linear and several offsets occur, notably near Talasea.

Six volcanoes (Ulawun, Langila, Ritter, Motmot (Long Island), Karkar and Manam) erupted between 1972 and 1975. Lavas from Langila and Karkar are low-silica andesite and those from Ulawun, Manam and Long Island are quartz tholeiite (Cooke *et al.* 1976). Ulawun, Langila and Manam also erupted in 1978 and Manam and Karkar again in mid 1979. Bam, in the far western arc, last erupted from 1954 to 1956; Pago, a small cone in central New Britain, erupted early this century and currently has vigorous solfataric activity; an eruption in Lolobau Island took place about 1905, and fumarolic activity that deposited a large amount of sulfur has since diminished. Of the volcanoes that have not recently been active, Talo is probably extinct and has thermal areas with minor solfataric activity and boiling mud pools; Narage, a strato-volcano, is probably extinct and has thermal springs with geyser activity; Garove (or Witu) volcano is largely extinct, with several areas of steaming and hot ground (Fisher 1957). Benda may have erupted early this century, Bola is possibly extinct, and at Talasea there are several geysers and solfataras; at Garbuna are solfataras, thermal springs, and mud pools; Walo has mudpools and solfataras, which are not directly related to any volcanic activity. At Galloseulo there is minor steam only, and the strato-volcano of Bamus has no record of eruption but has minor fumarolic activity (85°C); neither volcano can be considered extinct, however, (Fisher 1957). Most of this dormant or late-stage activity is typified by steam and gas discharges at 80 to 90°C (but up to 104°C) and thermal springs with temperatures of 45 to 98°C. Most of the gases contain H_2S, with some SO_2 and CO_2, and in some localities there are minor deposits of sulfur.

The Rabaul Caldera, the most studied in this

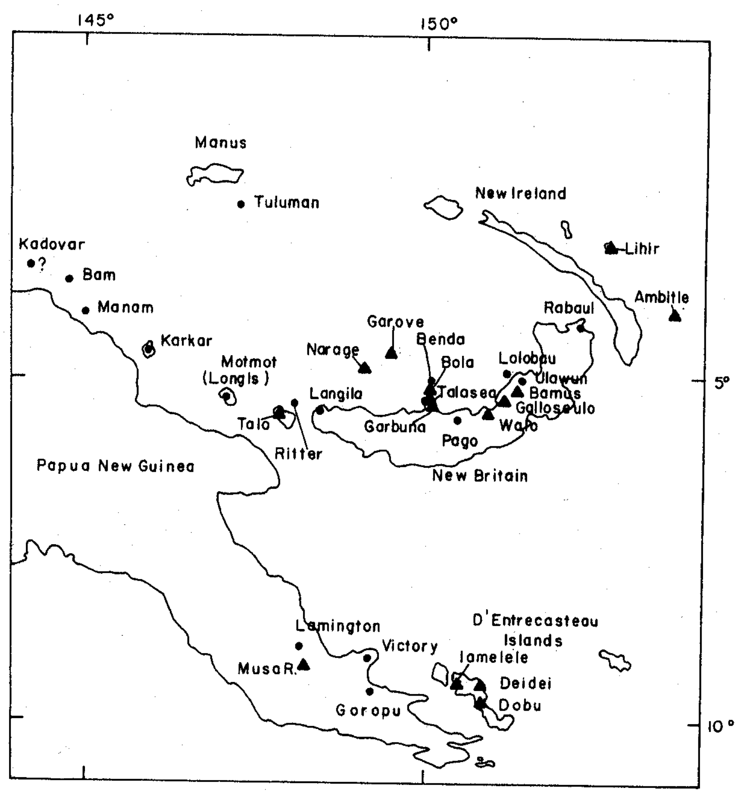

Figure 8. Locations of hydrothermal occurrences in the eastern Papua New Guinea region. Active or historic eruptions (closed circles); high-temperature hydrothermal occurrences (triangles). The location of two still thermally active volcanic centers in the Central Highlands (Johnson *et al.* 1978) is not known. With the intense current and recent volcanism in this region it is highly likely that other, both high- and low-temperature, thermal-fluid discharges occur that are not reported or are reported in literature not cited. (After Fisher 1957, Cooke *et al.* 1976.)

area, has several active sites. The Vulcan crater in the west and Tavurvur in the east erupted simultaneously in 1878 and 1937, and Tavurvur (or Matupi) was active from 1941 to 1942 and now has fumarolic activity with temperatures of less than 100°C. Fumarolic activity also occurs at Rabalanakaia crater and Sulphur Creek in the east (Fisher 1957). The Rabaul caldera lies close to a postulated junction between three crustal plates and a transform fault separating the Pacific and Bismarck Sea plates (Heming 1974, Johnson 1976).

In *eastern Papua* at least 11 Miocene to Holocene volcanoes are situated at the boundary between the Australian and Solomon Sea plates. Three of these have been historically active (Lamington, 1951; Goropu, 1943-1944; and Victory, possibly in 1880). The volcanic rocks here are divided into a southern belt, largely trachybasalts, and a northern belt, dominated by K-rich andesites but containing rocks ranging from basalt to rhyolite (Johnson *et al.* 1978). At Musa River, near Lamington, is a series of hot springs and gas discharges, but they have no obvious connection with any recent volcanic activity (Fisher 1957).

Volcanism in the Central Highlands area was largely of Middle to Late Tertiary age, but two centers of Quaternary volcanism are still thermally active (Johnson *et al.* 1978).

In the *D'Entrecasteaux Islands* solfataric and thermal-spring activity occurs at Iamelele, Deidei, and Dobu. The spring temperatures are up to 98°C and those of the gases to 100°C. There are also deposits (mostly small) of sulfur and siliceous sinter (Fisher 1957).

An initially submarine volcano, Tuluman, is located southeast of Manus Island in the Bismarck Sea. First recorded in 1883, it has now built a small island of silica-rich rhyolite lava, pumice, and pyroclastics; it last erupted from 1953 to 1957 (Fisher 1957, Reynolds and Best 1976).

EXAMPLES OF GEOTHERMAL OCCURRENCES

Geothermal activity can include all those phenomena associated with subsurface thermal conditions whether they are anomalous or 'normal' — normal usually being taken as an average thermal gradient for the crust of 30°C per km, or a heat flow of 1.1 HFU. In the region discussed, geothermal fluid systems range from hydrothermal systems of active volcanoes to those of areas of low-magnitude subsurface thermal anomalies in extinct volcanic terrains. The geography of these areas also varies from active island volcanoes to mature islands with continental characteristics.

A selection of areas is presented that indicates the variation in type, the magnitude and source of heat, and the geological-tectonic associations of different geothermal fluid systems. This also includes estimates of subsurface temperatures of thermal fluids by chemical geothermometers (e.g. SiO_2, Mahon 1966, Fournier and Rowe 1966; Na-K-Ca, Fournier and Truesdell 1973). These geothermometers are best applied to hot water systems and provide an estimate of the temperature of the reservoir fluids at last equilibrium with surrounding rocks. The estimates, however, can be affected by steam heating and mixing of other waters.

South Island, New Zealand

Most of the thermal springs occur in a linear belt east of the northeast trending Alpine Fault on the western margin of the Southern Alps (Fig. 2). The springs in the south of the belt are within the central schist zone and those in the north are largely within greywackes and argillites. On the central east coast, several groups of springs occur within pre-Carboniferous high-grade crystalline rocks and volcanics, and to their south another group discharges from unconsolidated Quaternary sediments. In all, there are about 20 localities. The springs typically are small (1 to 10 l/sec) and of low temperature (40 to 85°C). They are associated with faulting or minor crush zones, and their chemistry (pH near neutral, high HCO_3 and low dissolved constituents) is similar to that of local stream water (Grindley and Williams 1965) and indicates low-temperature reactions (Figs 9 and 10). Collins (in Barnes *et al.* 1978) suggests that the water is heated by deep circulation in rock that is heated by both the natural thermal gradient and fault movements. The SiO_2 geothermometer provides subsurface water temperatures of 100 to 120°C, and a very rough approximation of total heat loss by the springs is 4000 Kcal/s. Grindley and Williams (1965) suggest that heating in this largely nonvolcanic terrain is due to tectonic events that uplifted warmer deep rocks creating a steep thermal gradient.

North Island, New Zealand

There are at least 50 areas with surface thermal occurrences in the North Island, but virtually all the fields with development potential (and boiling springs) occur within the Taupo Volcanic Zone (Fig. 2). The exceptions are the Rotorua Caldera at the northwest edge of the zone and the Ngawha field (natural heat flow of 8000 Kcal/s) in the northwest of the island (Healy 1962). Most of the other occurrences are of low temperature and low to moderate discharge, heated by deep circulation in areas of normal or near normal geothermal gradient. Those to the east of the Taupo Volcanic Zone are similar to the thermal springs of the South Island and are related to active transcurrent faulting. They discharge from Mesozoic greywackes or sandy sediments associated with diapiric anticlines of Tertiary and Cretaceous sediments. Although their temperatures are low (50 to 60°C) they have high concentrations of dissolved constituents, typical of migration through marine sediments (Grindley and Williams 1965) (Figs 9 and 10).

The geothermal localities in the Taupo Volcanic Zone are found more within the permeable volcanic rocks flanking the volcanic centers than within the massive lavas and so are not related to individual volcanoes as are many of the other circum-Pacific occurrences (Healy 1976). The whole volcanic zone has a heat flow of 20 times normal, and temperatures of the deep reservoirs are usually in the order of 260°C; the source of heat is considered to be a body of cooling magma heating deeply circulating groundwater by conduction (Healy and James 1976).

At least 12 geothermal fields in the Taupo Volcanic Zone have been confirmed by drilling. One typical field is Kawerau in the north of the zone. There, production was initially from rhyolitic lava and breccia within interbedded volcanic strata: later production (and highest temperature) was from a deeper fractured andesite, which is capped by the low-

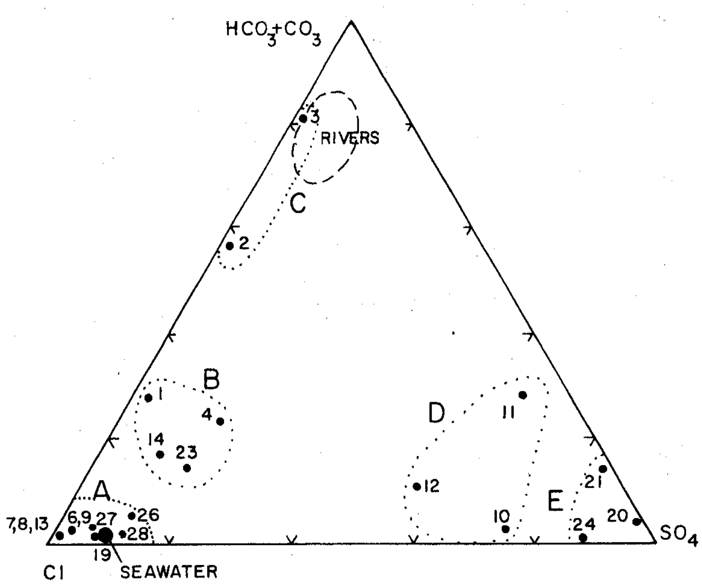

Figure 9. Relative proportions of major anions (approx. total carbonate, SO_4 and Cl) for thermal waters in the region, listed in Table 1. Seawater is an average of five samples from Fiji; river water, eight samples from throughout the region. Five broad groups occur: A, high Cl, both high temperature reactions and/or seawater mixing; B, mixed high-Cl and local surface water, generally low-temperature reactions, some seawater or connate water; C, low-temperature reactions, local meteoric water; D, high sulfate, low-temperature reactions, local meteoric water (migration through marine sediments); E, high temperature, with vapor-derived (H_2S) high SO_4.

permeability rhyolites (Healy and James 1976). The field covers 6–10 km², has temperatures generally of 260 to 270°C (maximum 285°C) and a natural heat discharge of approximately 25 000 Kcal/s (Macdonald and Muffler 1972).

Probably one of the best known fields in the world is Wairakei, which has been exploited for more than 20 years. This field covers an area of about 20 km² and is generating 148 MW (about 35 500 Kcal/s) but has an estimated optimum maximum of 192 MW; its natural heat discharge is well over 100 000 Kcal/s (Healy and James 1976). It is a typical high-temperature, water-dominated (slightly alkaline, high Cl, low SO_4 and HCO_3) field, with most production from a rhyolite pumice breccia (400 to 760 m thick) overlain by a mudstone-shale caprock and underlain by a thick ignimbrite layer. The fluid reservoir is apparently extensive and of relatively uniform chemistry. The temperature of the aquifer is commonly 260°C (maximum 274°C) and the water rises vertically to it through fissures within the underlying ignimbrite, below which is another pumice-breccia aquifer underlain by andesite lavas (Grindley and Williams 1965, Healy 1976). Faults are integral to the system and many producing wells intersect them. Within the southern part of the field are small accumulations of dry steam that have apparently been produced as a result of exploitation, possibly by reduction of the groundwater level. This has caused steam-heating of local groundwater and alteration to an acid high-SO_4 type.

Fiji Islands

In the Fiji islands more than 60 localities with thermal springs are widely distributed over the two main islands, Viti Levu and Vanua Levu, and also on the smaller islands, Kadavu, Ono, Gau, Vanua Balavu and Rabi (Fig. 5). Most of the occurrences are on the island of Vanua Levu. The thermal discharges vary from minor seepages through river alluvium to vigorously boiling pools, but most are springs with termperatures of from 40 to 60°C and low flow rates (> 3 l/sec). The waters discharge from a wide variety of rock types. Most of the occurrences are at relatively low elevations or in coastal areas and are associated with faulting and within the drainage systems.

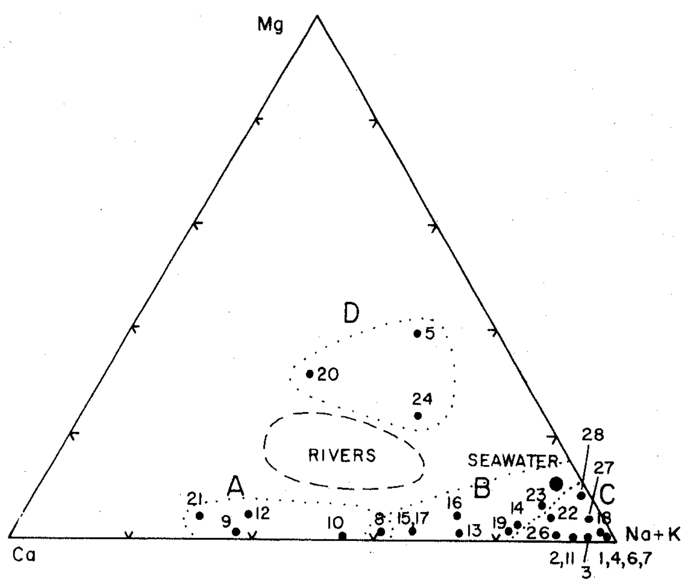

Figure 10. Relative proportions of major cations for thermal waters from the region (Na + K, Ca and Mg) from Table 1. The plotted data can be divided into four general groups but these are transitional: A, migration through marine deposits; B, component of seawater; C, essentially terrestrial meteoric water; D, steam heated.

TABLE 1

Chemical Analyses of Thermal Waters

No.	Location	Ref.	Water Type	pH	°C	Na	K	Ca	Mg	Cl	SiO_2	SO_4	Approx. HCO_3 + CO_3	Fe	Mn
	SOUTH ISLAND N.Z.														
1	Hanmer	1	Spring	8.4	53	388	5.4	6	0.09	476	46	25	191	0	0
2	Lower Wanganui R.	1	Spring	6.5	40	215	14.7	15	1.9	197	79	6	276	0	0.14
3	Waiho R.	1	Spring	6.2	43	494	26.5	26	1.9	224	63	1	1013	0.75	0.19
	NORTH ISLAND N.Z.														
4	Ngawha	2	Pool	6.4	83	830	63	7.8	2.5	1250	178	347	490	—	—
5	White Island	2	Pool	acid	—	7670	1000	2560	7310	61 840	180	10 500	—	11 480	260
6	Wairakei, 190	2	Spring	7.5	98	950	62	20	0.05	1596	245	56	38	—	—
7	Wairakei, hole 44	2	Deep water	8.6	260	1320	225	17	0.03	2260	640	36	19	—	—
8	Morere	2	Pool	6.7	62	6100	100	3900	137	16 000	28	21	25	—	—
	FIJI ISLANDS														
9	Savusavu, Vanua Levu		Spring	7.8	99.5	1000	44	1737	3	4762	158	236	21	0.08	0.26
10	Labasa, Vanua Levu		Spring	6.0	98	200	4	172	<0.02	161	61	520	21	<0.02	<0.02
11	Upper Busa, Viti Levu		Spring	8.8	59	87	1	7	<0.02	17	68	125	52	<0.02	<0.02
12	Emperor Gold Mine, Viti Levu		Deep water	7.3	45	160	5	261	23	404	45	667	132	0.08	0.14
13	Mocemocenitoa, Vanua Balavu	3	Spring	7.1	41	2700	120	987	20	7702	60	56	142	0.06	0.54
	NEW CALEDONIA														
14	Crouen River	4	Spring	9.0	43	23	0	4	1	29	75	4	7	Tr	—
	NEW HEBRIDES														
15	Sulphur R., Vanua Lava	5	Spring	1.5	83	170	44	112	2	1080	540	2017	—	—	—
16	Upper Teuma R., Efate	6	Spring	8.2	31.5	315	35	120	22	640	82	61	—	—	—
17	E. Takara, Efate	6	Spring	7.0	66	4740	200	2560	38	10 700	215	170	—	—	—
18	Pt. Resolution, Tanna	7	Spring	8.6	91	375	38	8	2	390	170	400	—	—	—

	SOLOMON ISLANDS														
19	Paraso, Vella Lavella	8	Spring	5.6	56	1207	178	289	26.6	2340	150	205	6	—	—
20	Paraso, Vella Lavella	9	Mud pool	2.5	48	1.3	0.6	2	1.8	11	168	1288	59	6.7	0
21	Voghala, Savo	8	Spring	7.3	74	77	18	221	14	6	232	649	122	—	—
22	Nggurara R., Guadalcanal	9	Spring	7.2	51	2080	89	226	125	3932	166	74	—	0.5	0.07
	NEW BRITAIN														
23	Sulphur Crk, Rabaul	10	Spring	8.1	39	660	38	68	60	1028	104	237	207	0.05	0.02
24	Tavurvur flank, Rabaul	10	Spring	3.5	36	1000	78	400	490	575	138	4458	5	1.48	11.3
	PAPUA														
25	Deidei, D'Entrecasteau Is.	11	Spring	7.7	95	3250	—	39	13	4900	—	0	—	—	—
	HAWAII														
26	HGP-A, Puna, Hawaii	12, 13	Deep water	4.4	300	800	149	96	1.0	1150	501	176	45	—	<0.1
27	Test # 3, Puna, Hawaii	13	Shallow water	6.4	89	2025	193	79	56	3684	156	325	30	—	—
28	Pohoiki, Puna, Hawaii	13	Coastal spring	7.3	35.5	2080	87	32	219	4062	89	530	59	—	—
	Av. seawater, Fiji			8.2	27	10 063	397	444	1297	21 500	5	2181	149	0.25	0.05

1, Barnes *et al.* 1978; 2, Ellis and Mahon 1964; 3, Rodda 1979; 4, Bontems 1949; 5, Barsdell and Radford 1976; 6, Hochstein 1977; 7, Heming 1977; 8, Giggenbach 1978; 9, Taylor 1976b; 10, Green *et al.* 1978; 11, Beevers 1965; 12, D M Thomas pers. comm. 1979 (preliminary data); 13, Kroopnick *et al.* 1978.

Chemical (Fig. 9 and 10) and stable isotope data show that the thermal waters are essentially similar and of low-temperature water systems in which local meteoric water has circulated through deep zones of normal thermal gradient and returned to the surface, largely via fractures (Healy 1960, Cox in press, c). Most of the waters are near-neutral, with low Cl and high SO_4 (probably largely derived from reactions with marine sediments), but in springs in coastal areas various degrees of seawater mixing occurs. Estimates of subsurface temperature by both SiO_2 and Na-K-Ca methods are in reasonable agreement with temperatures of 90 to 115°C for most spring groups. In two areas on Vanua Levu, Labasa and Savusavu, however, there are groups of boiling springs from which sinter (largely $CaSO_4$ and SiO_2, respectively) is being deposited. Estimates of reservoir temperature from the Labasa water are around 120°C and those for Savusavu 150 to 160°C. The calculated heat loss from the discharges at Labasa is 3400 Kcal/s (Cox in press, b), for Savusavu 4000 Kcal/s (Cox in press, a), and a total for the Fiji islands of about 9500 Kcal/s. The greater geothermal activity on Vanua Levu suggests that the geothermal gradient below that island may be higher than below Viti levu.

The springs at Savusavu occur in an area of about 10 km^2 around the shoreline of a narrow, fault-controlled block-faulted peninsula of andesite lavas and breccia. Some of these spring waters contain up to 25% seawater, which probably enters the system at shallow depth. The source of heat for the Savusavu occurrences is postulated as a Pleistocene basalt intrusion.

Springs are absent in the basaltic terrain of western Vanua Levu and on the (Quaternary) island of Taveuni, not necessarily owing to inadequate subsurface heat, but possibly because the permeability of the basalts is sufficient to prevent any thermal water that may be present from migrating to the surface. The locations of the springs on Vanua Levu and on the smaller islands seem to be controlled by regional northeast-trending structures that also appear to have controlled the location of volcanic centers. Magnetic data indicate the existence of intrusive bodies near many spring localities; such bodies could be the heat source for at least some of the springs.

New Caledonia

Low-temperature thermal springs are recorded in four areas of New Caledonia (Fig. 6): Prony Bay in the southeast (Carenage Bay 42°C and Kaoris Bay 34°C); in the central northeast two areas of seepages near Nakety; several localities in the Thio area discharging through Mesozoic schist (31°C); and the main area of springs near Canala, at the junction of the Crouen and Negropo Rivers (Koch 1958, Guillon and Trescases 1972, Lozes and Yerle 1976).

At Crouen is a group of some 14 springs with temperatures of 40 to 43°C within an area of folded Permian to Jurassic greywacke and schist. The waters discharge from metamorphosed schist and associated faults. The total discharge is about 2.5 l/sec. Adjacent to these are numerous other springs (21 to 26°C) discharging from faults along the courses of the rivers (Avias 1949, Koch 1958).

The total natural heat discharge for Crouen (estimated from the literature) is about 55 Kcal/s and that for all the thermal springs in New Caledonia is less than 100 Kcal/s. On the basis of the SiO_2 content a subsurface water temperature for Crouen of 121°C can be estimated. The waters are alkaline with high Cl, and indicate low-temperature water/rock reactions and a minor component of seawater-mixing (probably at depth).

New Hebrides

Most of the hydrothermal occurrences in the New Hebrides are of fumarolic type typically associated with active volcanism. Others are associated with quiescent or waning volcanism. Both steam-heated systems (e.g. Vanua Lava) and hot water systems (e.g. Efate) are present. Greenbaum (1974a) noted thermal springs on the islands of Santo, Maewo and Pentecost.

On Vanua Lava, extensive solfataric activity is to be seen on the eastern and southern flanks of the Mt Suretamatai volcano. Numerous thermal springs and boiling pools have temperatures commonly around 75°C and very low pH (1 to 1.5). An estimated 5000 tons of sulfur have been deposited, and temperatures of 100 to 110°C have been measured in some of the sulfur mounds (Williams and Warden 1964). Hochstein (in Barsdell and Radford 1976) considers the waters typical of acid-sulfate types related to a vapor-dominated system, but notes the anomalously high Cl and suggests a degassing magma at intermediate depths and the possibility that most of the mineral constituents come from a small, shallow reservoir of condensates with temperatures of the order of 240 to 280°C. A heat flow of 40 000 Kcal/s can be calculated (from Greenbaum 1974b) for the main solfataras area; the value would be higher if the peripheral and crater areas were included.

In the east of Tanna Island, boiling springs occur on the outside of the cone of Yasour volcano; on the adjacent coast in Port Resolution harbor are several geysers (to 99°C), and along the shoreline are areas of thermal seepages. A fumarole (to 99°C), also occurs inland from the geysers. The thermal waters appear to be goundwater with a small seawater component heated by vapor, and the minimum estimated temperature (SiO_2) for the reservoir is 170°C (Heming 1977). A small steaming area is found in the northwest of the island (Williams and Warden 1964).

Most of the detailed studies have been on the island of Efate on which the town of Vila is situated. The geothermal occurrences on this island have the characteristics of water-dominated types. The thermal springs are noticeably associated with the well developed north-northeast to northeast faulting that appears to have produced fault blocks. Of importance is the narrow Teuma graben, which transects the island and probably plays a role in circulation of the thermal waters. There are four general areas of thermal fluid discharges, the main ones on the northeast coast at Takara, the others in the south of the island. The latter have temperatures of 28 to 58°C and generally low flow rates (1 to 3 l/sec) (Greenbaum 1974a).

In the Takara area thermal springs, mild fumaroles, and areas of hot ground are distributed in some 2 km^2 of the coastal plain and shoreline. The main springs have a maximum temperature of 79°C, and in some areas the ground temperatures are between 80 to 100°C (Greenbaum 1973, 1974a). Based on these data, estimates of heat loss from thermal fluids for the whole island of 3600 Kcal/s (Hochstein 1977) to 6000 Kcal/s can be calculated. Estimates of temperatures of deep thermal fluids from the spring waters give 180 to 200°C (Na-K-Ca) and 160 to 185°C (SiO_2) (Hochstein 1977). In the light of dilution effects, temperatures of just about 200°C appear likely.

Overall, the springs are near neutral, with relatively high chloride, which indicates seawater mixing, especially in the coastal areas. Greenbaum (1973, 1974a) considers that the thermal waters are of meteoric origin and derive their heat from deep circulation into areas of hot rock (possibly intrusive) related to a volcanic center in the north of the island and indicated by regional gravity and magnetic data. The youngest volcanics on the island are the Pleistocene basalts in the north. It is likely that a steeper than normal thermal gradient exists in the basement rocks and that temperatures may be locally elevated.

Solomon Islands

Hydrothermal activity in the Solomon Islands is essentially associated with andesitic volcanism and there is a predominance of systems of the acid-sulfate type. There are, however, near neutral, high-chloride springs on Vella Lavella and warm brines emanating from dioritic plutons on Guadalcanal (Taylor 1976a).

On Simbo Island, acid thermal springs rich in Fe and Mn occur at the edge of a saltwater lagoon. Adjacent to these on the southwest coast is an area of strong solfataric activity (to 100°C) which is depositing sulfur; other springs (50 to 60°C) are found at sea level on the east of the island. These occurrences all appear to be associated with faults (Taylor 1976a).

On Savo Island, a volcanic cone, hot springs and fumaroles are found on the east and southeast flanks; and fumarolic activity also occurs within the volcano (90 to 109°C). The thermal fluid discharges are in a northeast-trending fault system associated with the crater. The springs are low-chloride, high-sulfate types of local meteoric water, possibly with minor seawater contamination, and are likely to be volcanic-steam heated. They do, however, have a near neutral pH, which suggests that there is no oxidation of H_2S. Silica geothermometry estimates give subsurface temperatures of 170 to 180°C, whereas Na-K-Ca yields substantially higher estimates to 295°C; however, both are suspect in thermal waters of this chemical type (Taylor 1976a).

In the northwest of Guadalcanal 24 saline thermal springs have been reported (Hackman, in Taylor 1976b). The springs are alkaline and temperatures 20 to 58°C; the hottest is within a zone of linear faulting of Pliocene to Holocene andesitic lavas. The springs are commonly associated with diorite plutons and presumably heated by them; they are highly mineralized and deposit sinter and travertine (Taylor 1976a, 1976b). Taylor reports low heat flows for the different groups, typically 10 to 17 Kcal/s, with subsurface temperatures likely to be in the range of 140 to 180°C.

The most extensive geothermal area in the Solomons is at Paraso (Vella Lavella); its features are summarized from Taylor (1975, 1976a). Three volcanic centers (one possibly quiescent, not dormant) occur within a northeast-trending fault-controlled graben in the north of the island in which the Paraso area forms the most recent feature. It is within a fault-controlled river valley and a tectonic depression within the larger graben structure. Indications of anomalous heat flow occur over an area of 1.7 km^2.

Within this are several hundred hot pools and appreciable emission of gas (90 to 100°C) containing H_2S, mudpools of acid-sulfate type with varying concentrations of chloride, and near neutral, lower-sulfate hot springs near the margin of the area. The mudpools are from 70 to 95°C and their chemistry is characteristic of vapor-dominated systems. The high-chloride springs have water-dominated characteristics. Giggenbach (1978) suggests that separate gas and thermal water phases coexist and can move independently, and that the system is intermediate between vapor- and water-dominated. He considers that the high sulfate content of the mudpools is from oxidation of fumarolic H_2S and that most other constituents are from the interaction of secondary acid fluids with shallow rocks. Giggenbach also suggests that most of the thermal springs can be considered to have steam-heated waters with steam containing CO_2 and H_2S being absorbed in shallow groundwater. A further conclusion is that the system overall has a low liquid-water content, but some of the high-Cl waters suggest a high-Cl water phase at depth, with temperatures over 200°C (Glover 1975, Giggenbach 1978).

Taylor (1976, 1976c) estimated a total natural heat flow of 36 152 Kcal/s, and geothermometry gives estimates of 110 to 155°C (SiO_2) and 200 to 220°C (Na-K-Ca), the latter probably being more reliable.

Rabaul Caldera

The caldera, on a narrow peninsula, is breached on the southeastern side, forming Blanche Bay. A dominant caldera ring fracture approximately follows the edge of the bay indicating the caldera rim, and a graben structure (with which all of the most recent volcanoes are associated) is postulated to trend northwest through the caldera (Heming 1974). Most of the thermal springs are along the northeastern shoreline of the bay (in Matupi Harbour) and in Sulphur Creek (in a fissure about 1 km west of Rabalanakaia). Small steam vents and patches of steaming ground are seen between the shoreline and Tavurvur crater, and minor springs and steam vents also occur near Vulcan crater on the west of the bay. The included analyses are from springs in a gully half way up the west slope of Tavurvur and in Sulphur Creek. Some of the springs have temperatures to 65°C. Fumarolic activity in the craters is mild; temperatures are usually 99°C (commonly 87 to 99°C).

Studt (1961) suggests that the hot water rises to the water table close to the eastern caldera wall, then percolates to the bay. The hydrology of the thermal waters is not known, but the spring discharges are not necessarily directly associated with the fumarolic activity in Tavurvur and Rabalanakaia craters. Both fumarole and spring temperatures are reported to be sharply affected by heavy rain, and airborne infrared imagery has not located hot areas other than those of surface discharges of thermal fluids (Perry and Crick 1976). From their chemical and isotopic work Green *et al.*

(1978) consider it likely that the waters are conductively heated and that they are composed of mixed local meteoric water and seawater, the latter entering near the shoreline. These acid (pH around 3.5) hot waters probably experience relatively shallow, but extensive interaction with the Quaternary basalt-andesitic pyroclastics. This interaction and a relatively slow migration could account for the heavy metal enrichment, especially of Mn, Fe and Zn, as well as the discoloration of Matupi Bay observed by Studt (1961) and visible in aerial photographs. Green *et al.* (1978), however, note a lack of positive oxygen isotope shift in the thermal waters, and such a shift usually takes place with interaction at high temperature. They also consider that volcanic sulfate may be added to some of the thermal waters and that the enhanced SO_4 and low pH of the gas condensates are probably due to dissolved oxy-sulfur gases from oxidation of H_2S. The weak fumarolic activity appears to be dominated by local meteoric waters modified by evaporation, but it has not been satisfactorily determined whether the fumaroles and springs have a common origin.

Green *et al.* (1978) considered that the chemical geothermometry may be misleading, but it provides estimates of subsurface temperatures of about 150 to 160°C. The heat flow cannot be accurately calculated on the data available but is possibly in the order of 600 to 900 Kcal/s for the thermal springs and at least 4000 for the total steam discharges. This suggests a total natural heat flow of at least 6000 Kcal/s.

Hawaii

The Puna geothermal area is on the lower east rift of Kilauea volcano (island of Hawaii) 35 km from the summit. Initial indications are that the thermal fluid reservoir is rift-contained (3 to 4 km in width) by both high fracture-permeability and dike-impoundment. The water chemistry and isotope data show that the system is dominated by water of meteoric origin with approximately 5% seawater mixing (McMurtry *et al.* 1977, Kroopnick *et al.* 1978). The water is of acid (pH around 4.4), high-chloride, low-SO_4 and low-HCO_3 type.

The source of heat to the groundwater is postulated to be intrusive material which has been emplaced along the rift zone, originating from below the summit of Kilauea. The most recent eruptions in this lower part of the rift were in 1955 and 1960. The maximum temperature in geothermal well HGP-A is 358°C. The only surface expressions of geothermal activity (apart from eruptive features) are several mild fumaroles and areas of steaming ground (to 85°C) and some thermal coastal springs (to 36°C). Current work in Hawaii suggests that structures of volcanic origin, such as rift systems, have the greatest potential for the formation of thermal fluid reservoirs. This work further shows that the required permeability is essentially structural and not lithological.

DISCUSSION

From chemical characteristics of the various thermal waters shown in Figs. 9 and 10, three characteristic water-system types emerge: (1) acid, high SO_4, low Cl and HCO_3, indicating vapor systems or steam heating; (2) near-neutral, high Cl, low SO_4 and HCO_3 of high-temperature water systems, but including systems with appreciable seawater mixing; (3) neutral to slightly alkaline, generally high HCO_3 and Cl, and low SO_4 of low-temperature water systems with variable minor seawater mixing (also included in this group would be the slightly alkaline Fiji thermal waters, which have high SO_4). Other transitional chemical groupings are caused by degrees of mixing of different waters.

Apart from the magmatic vapor addition, the thermal fluids appear to gain the dissolved constituents from water/rock reactions at various temperatures (e.g. Ellis and Mahon 1964, Mahon 1967). However, in many of these island environments, there is a component of seawater mixing often associated with thermal perturbation of the Ghyben-Herzberg lens (basal saline/freshwater interface) and also chemical changes due to superficial phenomena such as evaporation, and oxidation. There is, however, evidence that some systems are of intermediate character described as 'gas dominated' (Giggenbach and Lyon 1977). In such systems, high gas contents allow a separate gas phase to be maintained within the zone of upflow and there is independent movement of the liquid and gas phases.

All the high-temperature geothermal systems are within provinces of active or historic volcanism. The overall linear tectonic belts in which these provinces occur are areas of regional thermal anomalies, with localized temperature elevations. Some of the larger landmasses in the region have low-magnitude regional thermal anomalies, with localized highs in some cases. However, from the aspect of geothermal power potential, the presence of anomalous heat or geothermal gradient itself is not as important as the heat flow. This requires that both geoloical and hydrological conditions must exist to create a mechanism by which heat energy can be rapidly transported to the surface. Healy (1976) notes that although most of the stored heat is in the rock, 'a geothermal system suitable for energy production should contain rocks with adequate permeability, saturated with hot water or steam, within economic drilling depth'. Other factors can also influence suitability, such as very acid thermal fluids or excessive solution deposition. Temperatures of at least 200°C are usually required.

As shown by the above examples, a variety of geological and hydrothermal environments exists in the region. Those geothermal systems immediately associated with active volcanism tend to be vapor-dominated systems, typically with fumaroles, solfataras and steaming ground and are, at least in part, heated directly by magmatic steam. These systems develop where the meteoric recharge entering the system is vaporized owing to either restricted supply or a great amount of heat. They often have large heat flows and steep geothermal gradients from extremely elevated deep temperatures, but thermal-fluid reservoir temperatures are usually a maximum of 230 to 240°C.

Geothermal systems not directly associated with active volcanoes but within active provinces can have relatively low geothermal gradients but high heat flow if there is a high thermal capacity and suitable hydrogeological conditions. In these largely hot-water systems the heat source is most likely to be a cooling magma body, and reservoir temperatures are commonly of the order of 230 to 270°C. It is interesting to note that in Hawaii (an intraplate basaltic volcanic province), this type of system occurs within an active volcanic rift with a reservoir temperature of 300 to 350°C.

In areas of recently extinct volcanism, low-temperature water systems can develop by deep circulation and conductive heating from steep geothermal gradients or more localized thermal anomalies associated with intrusions. In older, nonvolcanic terrains heating of deep groundwaters can be related to plutonism and tectonic events, but the heat flow is typically low owing to low permeability of the rocks and restricted circulation of water within faults (Grindley and Williams 1965).

CONCLUSIONS

Within the Southwest Pacific region, extensive active volcanism is found above zones of crustal consumption, and a variety of geothermal-hydrothermal systems exists. Those systems with the highest reservoir temperatures are within active volcanic provinces, and low-temperature systems occupy recently extinct volcanic terrains and non-volcanic environments that have undergone tectonism. The type of surface activity and the chemistry of the thermal fluids characterize the type of system and can provide an indication of deep terrestrial thermal conditions. These thermal conditions and the type of system can be related to the tectonic environment and structure.

Many of the geothermal systems in this region are at present unsuitable for power production because of isolation, unfavorable fluid chemistry or the possibility of eruption; however, several undeveloped areas show promise as geothermal fields.

ACKNOWLEDGMENTS

Review of, and comments on the text by Dr John Sinton and Dr Donald Thomas are greatly appreciated. I especially thank Mr James Eade for presenting the paper in my absence. Hawaii Institute of Geophysics Contribution no. 1031.

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SUBMERGED MARGIN EAST OF THE NORTH ISLAND, NEW ZEALAND AND ITS PETROLEUM POTENTIAL

H R KATZ AND R A WOOD

ABSTRACT

The East Coast Fold Belt (ECFB) of New Zealand is the southward continuation of the Tonga-Kermadec island-arc system. As such it borders an active continental and plate margin. The Kermadec Trench, however, and its associated features terminate off the northernmost part of New Zealand's East Coast. The profound changes in geological structure and history between Kermadec Ridge and ECFB, as well as an abrupt change in geophysical parameters, suggest that the simple subduction model does not adequately explain the complex situation of the ECFB.

The submerged eastern part of the ECFB is 500 km long and 50–150 km wide. Its boundary or tectonic front towards the undeformed abyssal plain lies at 2500–3500 m below sea level. The sedimentary sequence on land is all marine, approx. 7000 m thick, and ranges from Mid Cretaceous to Plio-Pleistocene. Owing to great tectonic mobility through most of its history, there are rapid facies and thickness changes and local gaps in the column. Overpressured shales are common in the Late Cretaceous to Early Tertiary sequence causing extensive diapirism with numerous mud volcanoes associated with hydrocarbon gas seepages. Diapirism is widespread also in the offshore part, suggesting that similar overpressured shale formations extend far to the east.

Offshore structure is dominated by asymmetric, elongate 'tectonic protrusions' of Miocene and older rocks. They often have acted as source areas for younger sequences deposited in a landward direction. Continuous diapiric uplift has locally counteracted regional down-faulting to the east and general subsidence of the submerged part of the ECFB.

Pronounced segmentation of the tectonic front with regard to strike, position and structure is most noticeable. In places there are wide open folds which gradually die out towards the abyssal plain as the sediment thickness decreases dramatically. Elsewhere thick underformed sedimentary aprons and foredeeps extend beyond the tectonic front. Sedimentary continuity from deformed into undeformed zone is well established. The sediments are mostly land-derived, continental slope deposits which have suffered in-place deformation. There is no indication of an accretionary prism formed by offscraping and imbrication of ocean-derived sediments. Thus there is no subduction complex. Multiple evidence, however, exists of former landmasses to the east of the present coast which have since disappeared. Similar situations have been found along other active margins, and foundering and a kind of tectonic erosion of continental blocks seem to be a more viable model for the plate boundary along the ECFB.

If so, the quality and amount of reservoir sands in the submerged margin eastwards may be improving, particularly in the Cretaceous but also in the Miocene section. Since organic-rich, undercompacted shales which on land are associated with numerous oil and gas shows appear to occur extensively also in the submerged part of the ECFB, a good source rock potential is inferred offshore as well. But the structural mobility of the ECFB makes the entrapment of hydrocarbons a problem. However, gas hydrates may have formed along the lower slope and effectively trapped free gas. Higher up on the slope, Late Cenozoic sandstones and bioclastic limestones draped over or flanking the 'tectonic protrusions' and local diapirs could provide trapping reservoirs for hydrocarbons migrating from older sources. As the shale diapirs are themselves likely sources their associated traps are particularly attractive.

INTRODUCTION

The Tonga-Kermadec island arc system continues southward into the East Coast Fold Belt (ECFB) of the North Island, New Zealand (Katz 1974 a; 1976). Tectonically highly active for most of its history, which covers the period from Mid-Late Cretaceous to Recent (Katz 1968), the ECFB lies east of the Main Ranges of the North Island and represents the deformed continental margin of the New Zealand North Island. Along the Main Ranges, the sedimentary series of the ECFB unconformably overlies the strongly indurated and intensely folded, Lower Mesozoic greywacke series, but no basement is exposed nor has been drilled into anywhere further east.

Over 500 km long, the ECFB is exposed on land across a width of some 30–40 km and extends offshore for another 50–150 km until it borders against the undeformed sediments of the abyssal plain at a water depth of 2500–3500 m. From seismic sections the overall structural pattern of the submerged part of the ECFB is seen to be closely similar to the pattern observed on land.

The present study is mainly based on an analysis of data collected by the 1972 marine seismic survey of Mobil Oil Corporation, in which the senior author participated, and the 1973 marine seismic survey of Gulf Oil Company (Fig. 1). Most of the multichannel airgun recordings have been reprocessed as multi-trace stacked sections and are available at New Zealand Geological Survey, Lower Hutt, (Open-File Petroleum Reports nos. 587, 614, 738). Sea-bottom samples used for this study were collected by the senior author (Katz 1975) and workers from New Zealand Oceanographic Institute (Pantin 1966, Lewis 1974).

STRATIGRAPHIC OUTLINE OF THE ECFB

Sedimentation began in the Aptian-Albian (Speden 1975) and is marine all through to the Pleistocene. However, the great tectonic mobility of the ECFB with frequent diastrophic events caused rapid facies and thickness changes and many local gaps and unconformities in the sequence. Besides massive shelf sandstones at the base of the Cretaceous and in the Miocene, there are many flysch-type turbidites deposited in

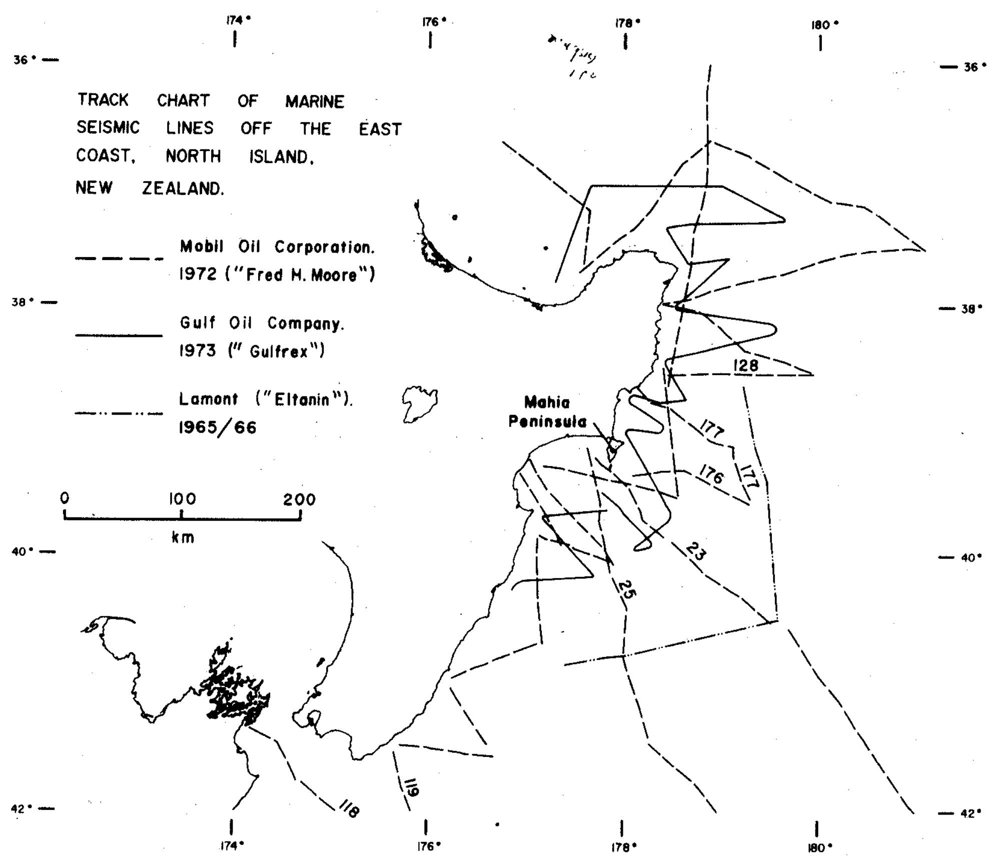

Figure 1. Track chart of main seismic lines used for this study.

rapidly deepening basins particularly in the Late Cretaceous and Mid Tertiary sections. The Plio-Pleistocene section is characterized by shallow-water bioclastic limestone and bathyal fine-silty mudstone (Katz 1968; 1976). A relatively quiescent period affecting a considerable area lasted for about 50 m.y. during the Late Cretaceous and Early Tertiary, when the adjacent land surface was extensively worn down and deeply weathered. Terrigenous clay-mud, partly siliceous, was deposited over very great areas and has given rise to the widespread occurrence of undercompacted shale and mud diapirism.

Average sediment thicknesses onshore are 3000–3500 m for the Cretaceous, 1000 m for the Early Cenozoic, and 2500–3000 m or more for the Late Cenozoic. There is no evidence for a fundamentally different pattern or mode of sedimentation nor for a radical change in average sediment thickness in the submerged eastern part of the ECFB, although the entire section appears somewhat thinner to the east. Evidence from analysis of the lithology and composition of Cretaceous and Miocene sections onland near the coast, and from the Plio-Pleistocene portion of offshore seismic sections (Figs. 2, 3), however, indicates in several instances a sediment source to the east. In these seismic sections, older formations generally constitute acoustic basement and do not allow any correlation. Partly because of a pronounced loss of seismic energy in the pre-Miocene overpressured shales (Katz 1975), and particularly because of the intense tectonic reorganization of the strata, older rock masses quite widely appear homogeneous seismically and do not allow the resolution of distinct formations and structures.

STRUCTURAL ANALYSIS

Shelf and slope

Across the shelf and continental slope the submerged part of the ECFB is characterized by a series of elongate and asymmetric tectonic protrusions which are step-wise downfaulted to the east. Seismically these structural highs invariably have acoustic basement exposed and sampling has shown that their core consists of lowermost Pliocene (Lewis 1974), Miocene (Fig. 3; Katz 1975), and older Tertiary sediments (Pantin 1966).

The tectonic protrusions are formed by narrow and tight fault-anticlines and it has become obvious that often they are associated with diapiric uplift. Diapirism indeed is an extremely important factor in the structural deformation both onland (Ridd 1970) and offshore. Piercement domes and pillows have developed at depth, bending the overlying strata upwards (Figs. 4, 5), while largely along pre-existing fault planes 'diapiric walls' have become prominent features which appear to control many elongate ridge systems (i.e. tectonic protrusions, Dean *et al.* 1976). Diapirs break through to the surface in places and occur even on the outer slope where they form sharply shaped mountains standing high above the surrounding sea floor (Figs. 6, 11A). A similar situation was reported by Shepard (1973) from an area adjacent to the Magdalena River delta off the north coast of Columbia.

The broad basins between the tectonic protrusions are filled with an unconformable sequence of strata which generally form wide, open synclines. In cross section many of the basins are asymmetric and inclined landwards, obviously as a result of subsequent rotation with differential uplift of the high that lies to their seaward side. However, there is little doubt that most of the highs were formed prior to the deposition of the sedimentary fill in the adjacent basins. Many appear to have been deeply eroded and levelled off (Figs. 3, 7) and to have acted as a source for a set of thick foresets developing from them in a landward direction (Fig. 2). This configuration may repeat itself several times across the ECFB, pairs of similar basins and horsts being successively faulted down eastwards to a lower level (Fig. 3). Obviously, most of this downfaulting and basin rotation must have taken place after erosion of the highs and deposition in the adjacent basins.

The age of the landward-directed foresets, and of the unconformable basin fill, is Plio-Pleistocene (Katz 1975). The tectonic protrusions therefore were uplifted above sea level in the Late Miocene to Pliocene. Clearly, their erosion is not a matter only of wave effect during glacially lowered sea level, as was assumed by Lewis (1974); it occurred much earlier and is the effect of tectonic events of greater implications. Indeed, a Mio-Pliocene unconformity is a regional feature widely recorded in the central East Coast, where bathyal Miocene rocks have also been uplifted and deeply eroded far inland, before being covered again by marine Pliocene strata (Katz 1973; 1976). A similar situation is well shown on the outer coast at Mahia Peninsula and can be followed offshore along the broad and

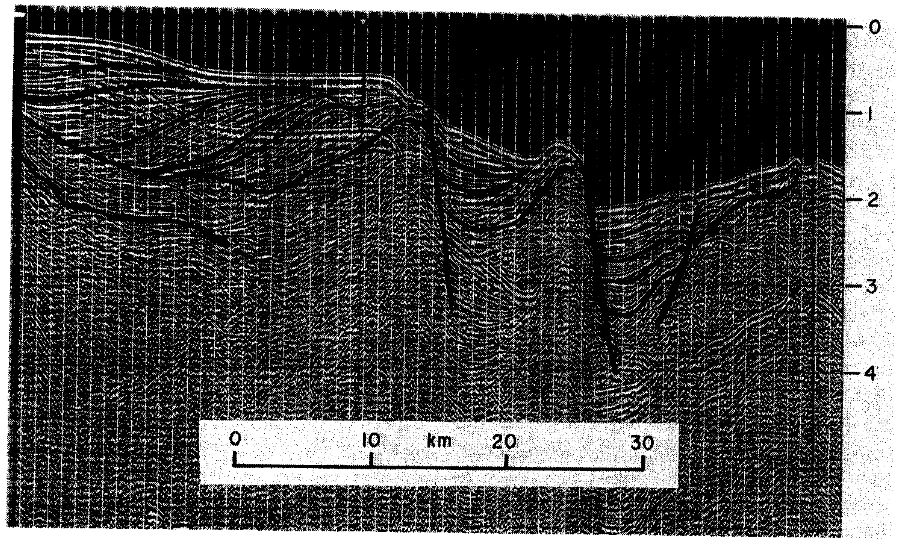

Figure 2. Seismic section showing zone between shelf and upper slope, central East Coast. Note thick foresets developing in a westward (left) direction from structurally high and eastward downfaulted tectonic protrusions and, in upper left corner, easterly prograding wedge of unconformable younger shelf sediments.

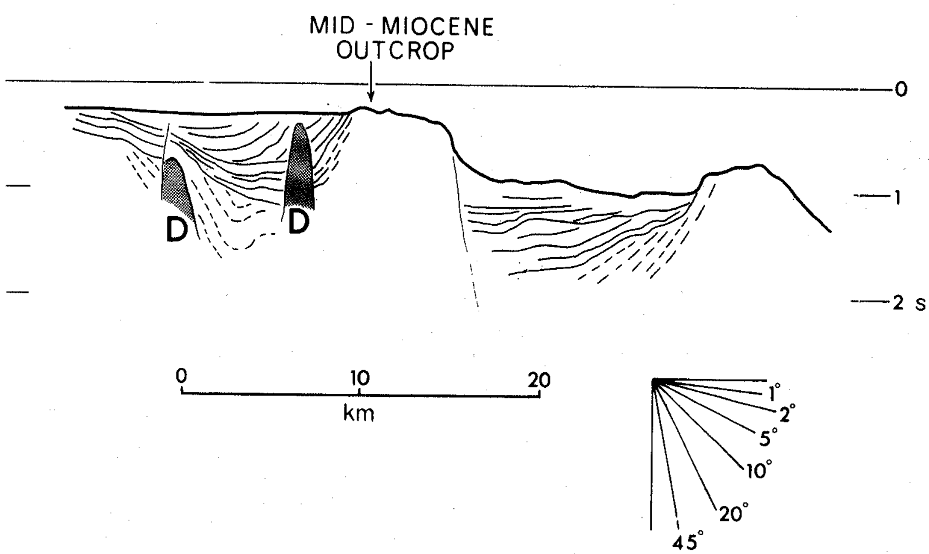

Figure 3. Miocene outcrop on tectonic protrusion at Ariel Bank, Gisborne (from Katz 1975). Plio-Pleistocene basin fill on their landward-side (left). D = diapiric intrusions.

elongate feature, Lachlan Ridge, the crest of which has been eroded down to the Eocene (Fig. 7; Pantin 1966).

In Early and Mid Pliocene time, a new transgression advanced from east to west across an area of considerable relief. This is shown by great variations in facies and in thickness of sediments (Fig. 14), and has become particularly evident in the course of oil exploration and well drilling (Katz 1971 a; 1974 b). While subsequent tectonic movements undoubtedly were locally complex and differentiated, the general tendency since the Pleistocene has been one of regional rotation about an axis not far from the present coast: i.e. regional uplift to the west and subsidence to the east. There is good evidence for this from the geology on land, while offshore a clear indication is seen in the step-wise downfaulting eastwards. Even where some of the originally high protrusions had become partly buried again by sediments of the Pliocene transgression, continued differential rotation of individual blocks finally resulted in their disruption and downdrop to the east. Regional subsidence to the east is well substantiated also from the reversal of sediment transport across the shelf in the youngest sequence, which in seismic sections is seen to form a wedge prograding eastwards, above and over the older foresets which face west (Fig. 2). The large downdrop of the Hikurangi Trough in Plio-Pleistocene time (Katz 1974 a; 1979; Fig. 12) is a further demonstration of regional subsidence of the submerged margins of the ECFB.

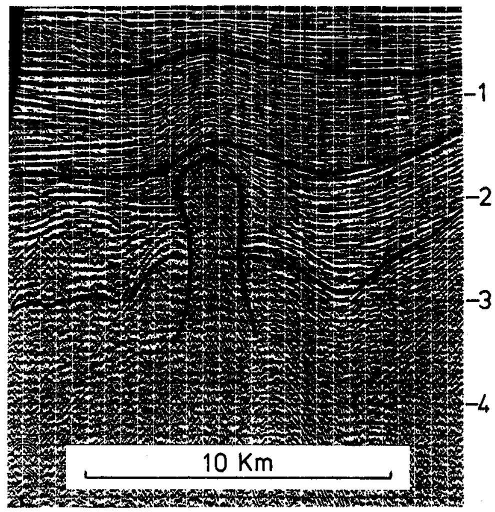

Figure 4. Shale diapir, originating in the acoustically opaque Lower Tertiary section, intruding and bending upwards Miocene and Plio-Pleistocene beds. Offshore central East Coast.

Superimposed on this regional tilt to the east, further diapiric uplift may have occurred locally, as is probable for some of the structural highs on the outer slope (see Fig. 10 B, C. Recent anticlinal folding as shown by Lewis (1971) may be a drape effect above similar diapirs at depth (cf. Fig. 4, 5). Indeed, for both

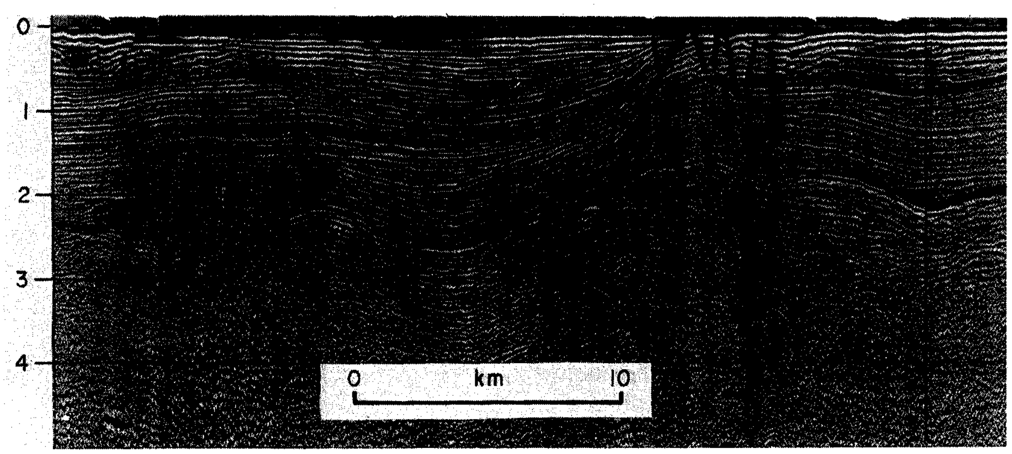

Figure 5. (As for Fig. 4).

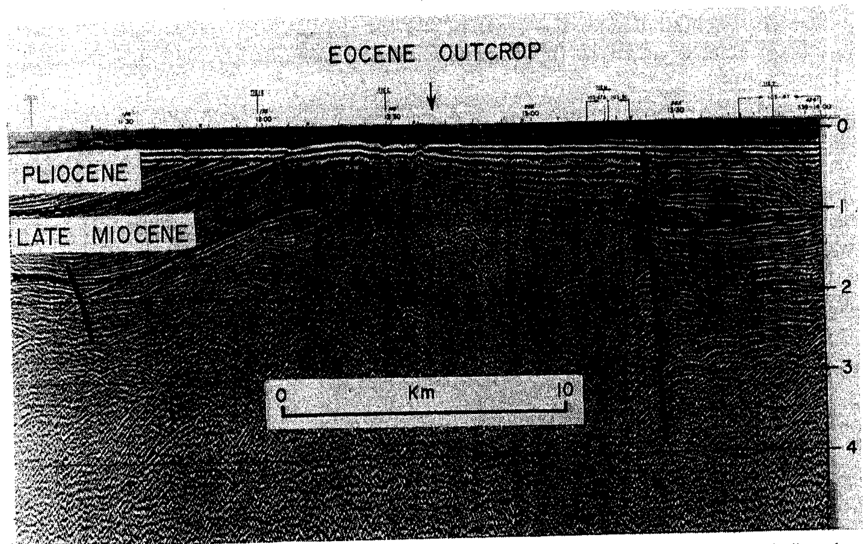

Figure 6. Piercement structure on continental slope east of Mahia Peninsula, Hawke Bay. Believed to represent Lower Tertiary shale diapir which has broken through to the surface.

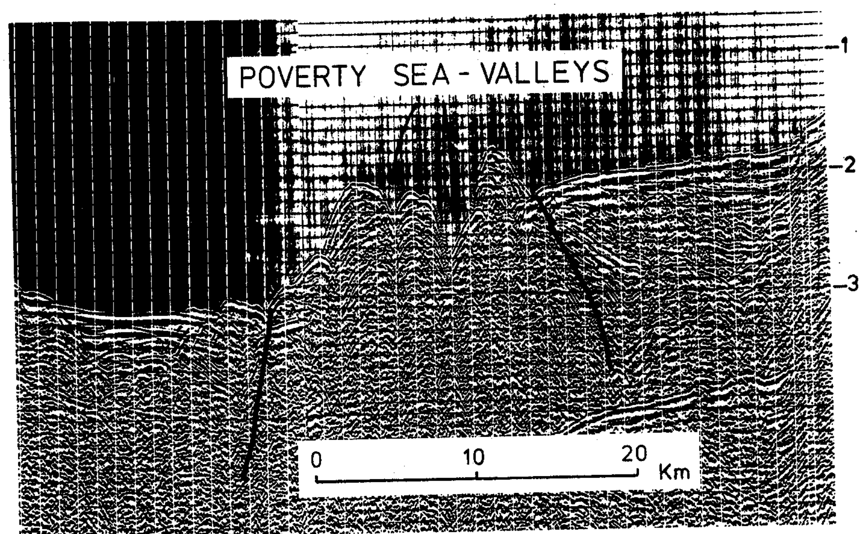

Figure 7. Seismic cross section of Lachlan Ridge south of Mahia Peninsula, central East Coast. Note Mio-Pliocene unconformity and lack of observable structural features in tectonic protrusion of central ridge.

the Kidnappers and Lachlan Ridges described by Lewis (1971), Dean *et al.* (1976) have demonstrated that 'diapiric activity has been a major factor contributing to their formation'.

In summary, we conclude that the younger tectonic evolution is considerably different and more complex — demonstrably so on land and probably so offshore in the submerged part of the ECFB — than that envisaged by Lewis (1974 and in press).

Structure of the tectonic front

Our work has shown that the Kermadec Trench terminates opposite the northernmost part of the East Coast. There is no trench further south, neither structurally nor morphologically. To the contrary, there are a number of seamounts, apparently consisting of igneous oceanic basement, immediately south of the head of the Kermadec Trench (Fig. 8), while the easternmost portion of the tectonic front of the deformed ECFB is located about 50 km further west and has a different strike direction from the axis of the Trench. It is notable that the structural elements of the ECFB immediately west of here are similarly disposed; in particular, the composite East Cape Ridge, which develops out of the large Pouawa diapir north of Gisborne and the Tolaga and Tokomaru Bay anticlines, cuts all other structures to the south of it in a strikingly discordant manner. Nor does the negative gravity anomaly which accompanies the Tonga-Kermadec Trench continue further south (Fig. 9). Its North Island counterpart, which previously was thought to form its continuation has been recently shown to be 'substantially isolated from . . . the Kermadec system and displaced to the west' (Hatherton and Syms 1975). Immediately to the north of the North Island there is a corresponding discontinuity of seismicity (Eiby 1977), while earthquake locations underneath the North Island down to 100 km depth show a very diffuse pattern with no clear definition of a Benioff zone (Adams and Ware 1977). A Benioff zone is well defined only at intermediate depths between 100 km and 250 km, but at a depth of 100 km it lies 200–300 km away from the continental margin or tectonic front of the ECFB and plunges at a relatively steep angle of 50°. If subduction does occur underneath the North Island, therefore, a rather low angle underthrust is required for a distance of some 200–300 km, and then a sudden steepening of the thrust angle. We are at a loss to give a satisfactory explanation for such a feature, and in particular — in terms of a uniform subduction model — for the pronounced segmentation in both structure and orientation of the tectonic front and continental margin south of the Kermadec Trench termination discussed below. It is obvious, however, that subduction — for as much as it may continue south of the Kermadec Trench — has markedly changed its character in this area, its effects becoming less and less clearly defined to the south. The gradual dying-out of calc-alkaline volcanism to the south, in the very centre of the North Island, seems to underline this fact (Fig. 9).

The tectonic front of the ECFB south of East Cape, between 38°30'S and 41°S, trends S 10°W and is marked, particularly in its northern part, by a steep uninterrupted slope from the shelf to a depth between 2500 and 3000 m, with no further slope break (Fig. 10 A). It sharply abuts an undeformed foredeep filled with up to 1.4 s of flat-lying, turbidite-type sediments which onlap eastwards against a series of basement seamounts of the abyssal plain. The sediments in this basin (Poverty Basin, Katz 1974 a; Figs. 10 B, C) apparently are land-derived and have been trapped behind the basement highs. From both seismic and magnetic data the oceanic basement can be traced underneath the foredeep westwards to immediately below the tectonic front (Fig. 10), where it comes up again in some other, shallow structures, which are covered by the youngest sediments of lower slope and foredeep basin. No basement is recognized further west.

Whereas in the northern part the steep slope is virtually homogeneous seismically, basin and horst structures become progressively better developed and recognizable southwards and clearly strike obliquely to the tectonic front and plunge northeastwards. The result is an ever-widening belt of apparently lesser compression to the south, to about 41°S.

At this point (41°S, 178°30'E) a sharp kink occurs in the boundary line between the ECFB and the undeformed abyssal plain, and the strike of the tectonic front changes to west-southwest. The fold-fault structures of the lower slope again strike obliquely to it, this time in the opposite sense, and plunge to the southwest. Thus the area of this kink is the widest portion of the ECFB. The narrow, tight fold structures observed both further north and southwest become in this region a series of well developed, simple concentric folds which gradually die out towards the abyssal plain (Fig. 11 B). This decrease in deformation intensity is indicative of a corresponding decrease in compression.

The folds along the tectonic front to the

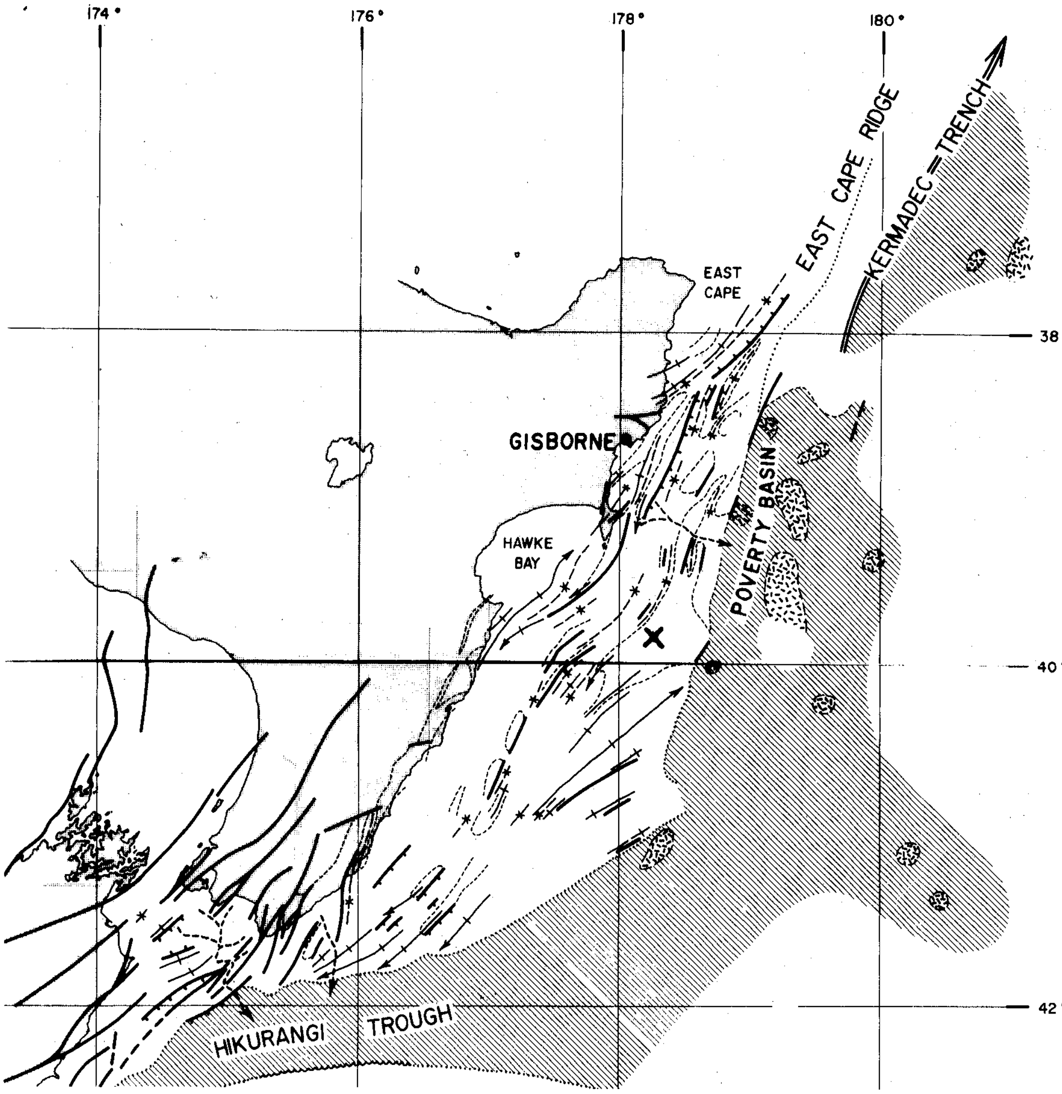

Figure 8. Structural map of submerged part of East Coast Fold Belt, showing main anticlines, synclines and faults. Thin dashed line is stratigraphic boundary between Plio-Pleistocene and older rocks. X denotes one large area of such older rocks exposed on the outer slope (see text). Diagonal hatching represents region outside tectonic front, which south of Kermadec Trench is tectonically mainly undisturbed. Stippled areas indicate seamounts of oceanic basement outcrop. Broken lines with arrow head represent submarine canyons.

west-southwest become progressively tighter again and more pronounced, forming a pattern of en echelon structures which one by one disappear by gradually passing, along their WSW-plunging axes, into the undisturbed fill of the Hikurangi Trough. Off Cook Strait, to the southeast of Wellington, the ECFB is narrowest and exhibits strong tectonic compression. Its boundary with the flat-lying sediments of the Hikurangi Trough (Katz 1974a) is a simple fault of about 7 km throw (Fig. 12). The Hikurangi Trough itself, which

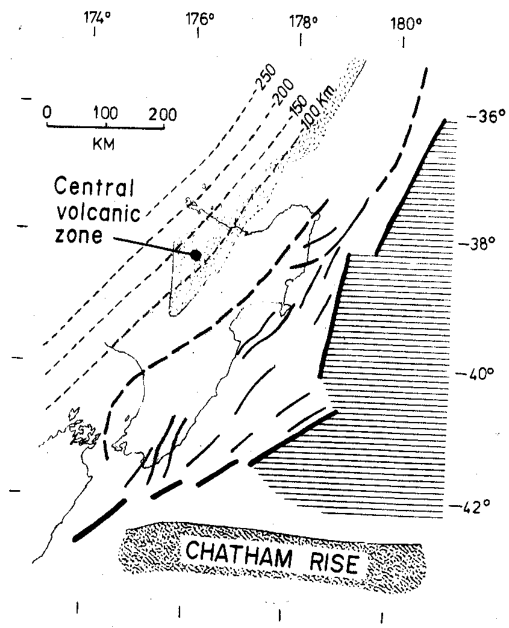

Figure 9. Tectonic framework east of North Island, New Zealand. Heavy line: tectonic front of East Coast Fold Belt (ECFB). Thin lines: main structural alignments within ECFB. Horizontally hatched: sediments and underlying oceanic crust outside ECFB. Note that Chatham Rise is underlain by continental crust. Dashed heavy line: axis of negative gravity anomaly. Dashed thin lines: depth contours of Benioff zone (after Adams and Ware 1977).

is a young foredeep of the ECFB, is filled with 4–4.5 km of turbiditic, Plio-Pleistocene sediments which onlap the continental foreland of the Chatham Rise.

Conclusions

In the extreme north and south of the ECFB the tectonic front is tightly compressed. In the central part where the ECFB is widest and where the boundary of the tectonic front against the undeformed abyssal plain changes direction and forms a sharp kink, broad open folds occur across the lower slope. Although the intense reorganization of the rocks in the north and south has given rise to a seismically quite homogeneous medium, there is nevertheless no evidence for thrusting and imbrication of oceanic formations. To the contrary, there is good evidence in several sections of a continuity of rock formations across the tectonic front with only a change in the degree of deformation. The change from an area of intense deformation into one of little or no deformation, from west to east, may be sharp and sudden or gradual. Décollement in deeper parts of the crust (Walcott 1979) is still possible, but at this stage remains a matter of speculation.

In the central part of the ECFB tectonic front, at least, décollement of any significant amount does not occur in the section, even at the level of oceanic basement (Fig. 11 B). What does happen, however, is that the intensity of folding gradually diminishes across an area where sediments are thinning dramatically. This configuration suggests that within this zone the sediments are continental slope deposits (i.e. terrigenous) which have suffered in-place deformation (Scholl *et al.* 1977), and are not oceanic-derived sediments which have been scraped off and accreted. Thus the entire belt right to the tectonic front is interpreted as a fully continental domain, there being no basic tectonic difference between the lower and upper parts of the slope as is required by the imbricate thrust model of ocean-continent collision (Karig and Sharman 1975); in other words, there is no subduction complex here. Where local foredeeps have developed outside the tectonic front, as in the north and southwest (Poverty Basin and Hikurangi Trough, Katz 1974 a), they are filled with thick turbidite-type sediments obviously derived also from a nearby land source, but have so far remained undeformed. Thus there seems to be a correspondence between the outer limits of deformation and continental slope deposition (i.e. accumulation of mainly land-derived sediments), with the two either coinciding or the latter extending beyond the zone of deformation. This may not be fortuitous: there is a striking resemblance to the common picture of an advancing tectonic front gradually and successively involving its own foredeeps, or foreland basins. A tectonic study of the ECFB shows that similar situations have existed before, if only locally (Katz 1974a).

In addition, there is multiple evidence for the existence, during several intervals of the geological history, of easterly landmasses which have since disappeared. Late Miocene to

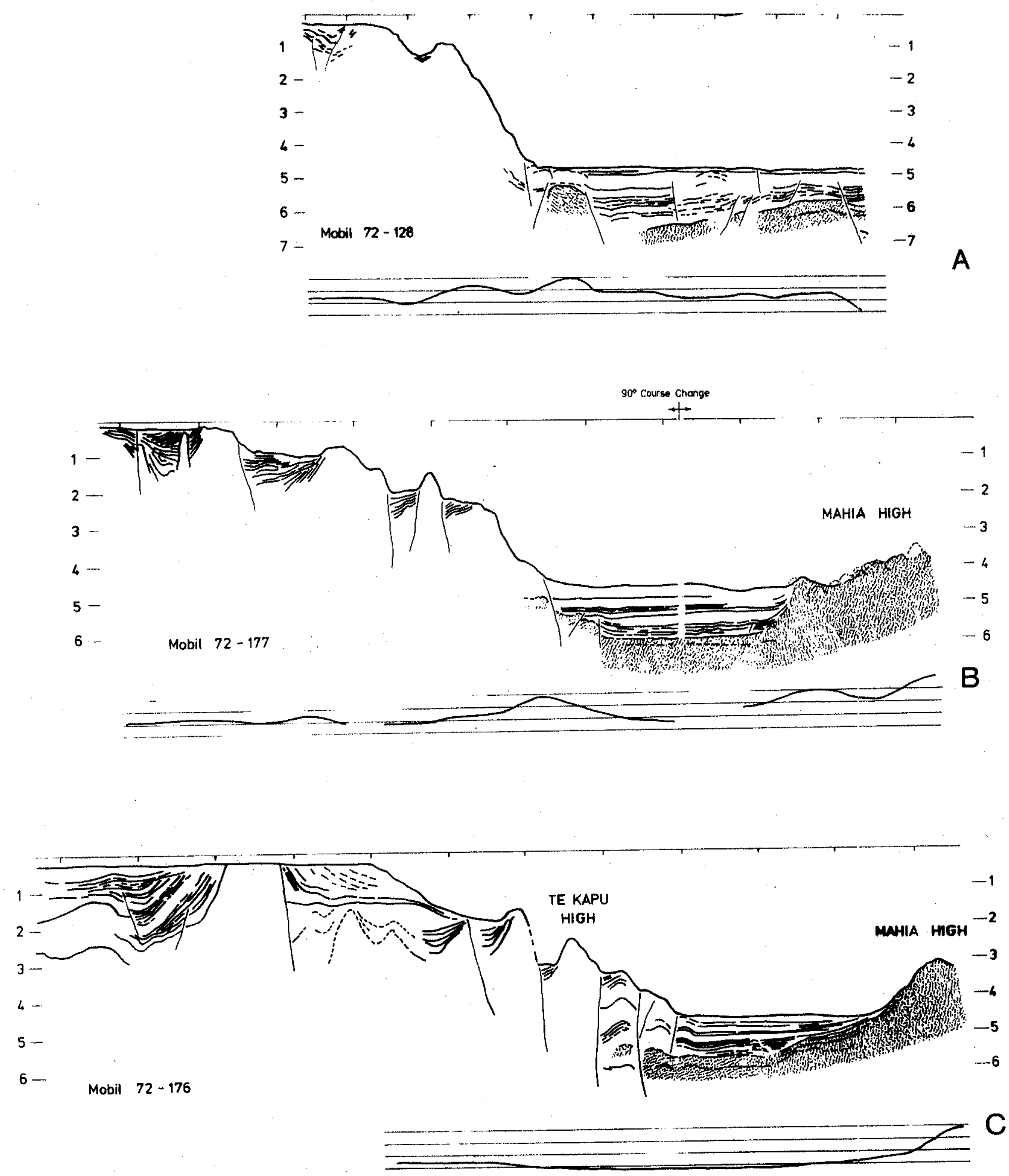

Figure 10. Line drawings of seismic sections in northern part of ECFB (for location see Fig. 1). Poverty Basin foredeep with thick sediments ponded behind Mahia High. Marbled pattern denotes inferred oceanic basement. Line spacing of magnetic profiles beneath seismic sections is 200 gammas.

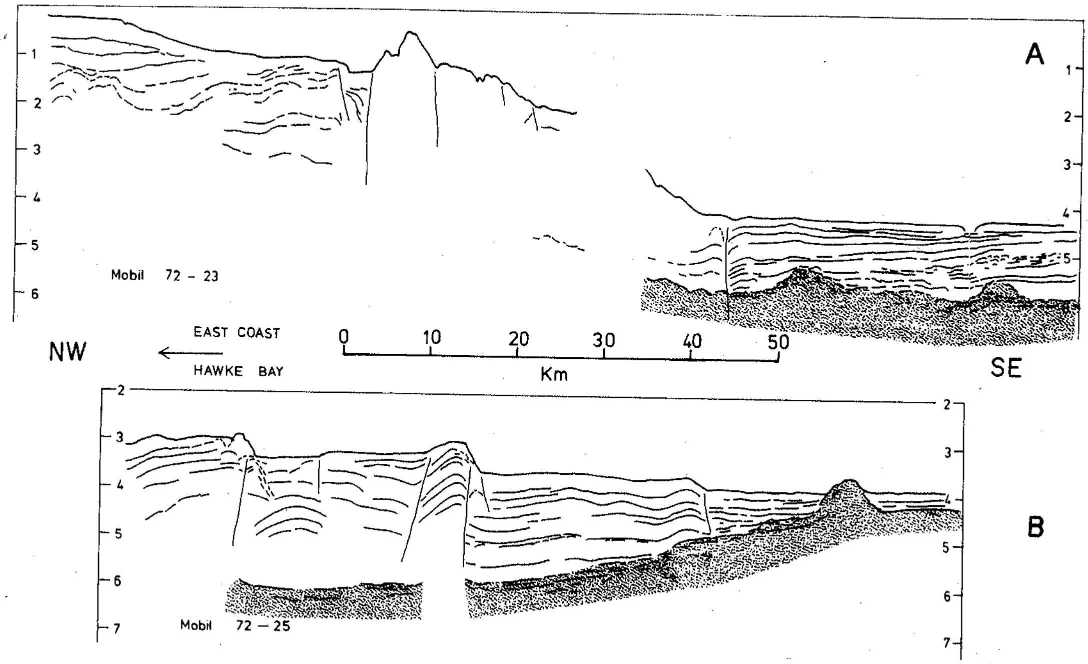

Figure 11. Line drawing of seismic sections across tectonic front southeast of Hawke Bay (for location see Fig. 1). Marbled pattern denotes inferred oceanic basement.

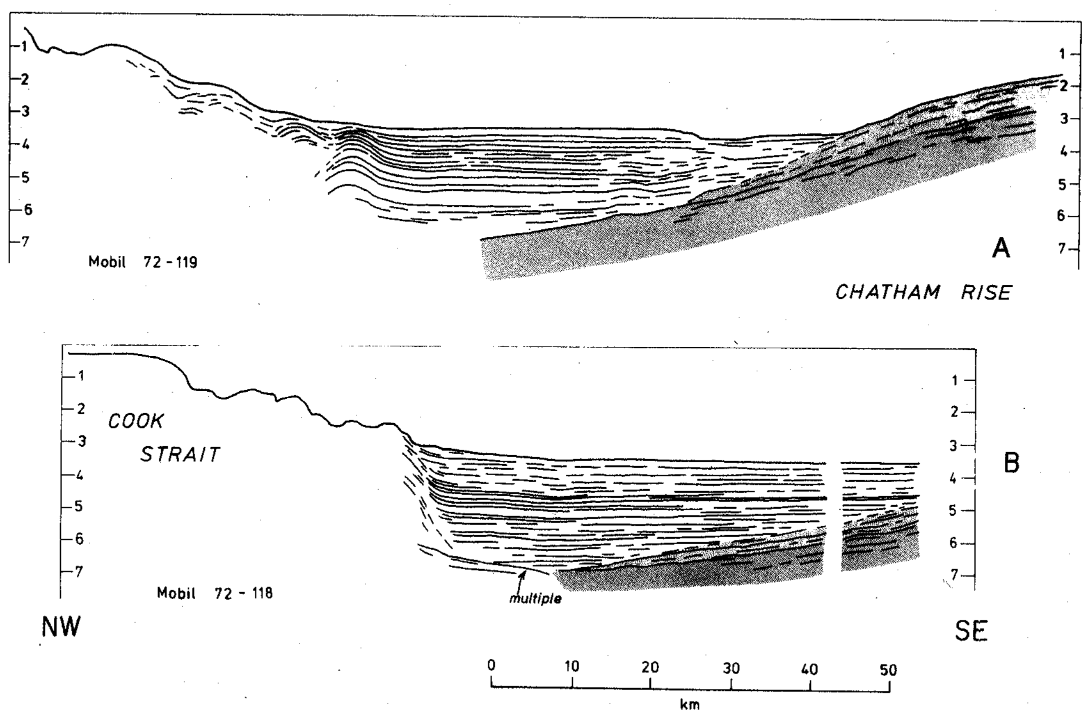

Figure 12. Line drawings of seismic sections across Hikurangi Trough southeast of Cook Strait (for location see Fig. 1). Shaded areas: Miocene and older rocks of Chatham Rise north slope, forming basement underneath undisturbed sediment fill of Hikurangi Trough. Note tectonic front is gradual and transitional in A, but is distinct and faulted in B.

Early Pliocene erosion of structural highs east of the present coast line has been mentioned above. Late Miocene turbidites in the Makara basin south of Hawke Bay are derived from the presently sea-covered area to the east (Kingma 1958), possibly from a similar setting as the Plio-Pleistocene beds in Figs. 2 and 3, i.e. from structurally high areas under erosion which later have become downfaulted and submerged. From other lines of evidence an easterly source has been advocated for the rhyolitic components in the Late Miocene Makara beds as well as for many other, both older and younger, rhyolitic tuff occurrences along the East Coast (van der Lingen 1968). For the high-metamorphic (amphibolite-facies) pebbles and boulders in the Ihungia conglomerates (Ongley and Macpherson 1928) a derivation from the east is most likely, too. The same probably holds true for the Late Cretaceous coarse conglomerates with exotic pebble composition (largely granitic-rhyolitic) which are widespread along the coast south of Hawke Bay but do not occur further inland. North of Hawke Bay the fine-grained flysch sequences of Late Cretaceous age penetrated in the Opoutama-1 well (Fig. 13) have been shown to represent an eastern facies (Zimmermann and Faber 1967).

On the whole, we thus feel that there is rather compelling evidence for the former existence of land areas to the east of the present coast, as has been suggested by a number of authors (cf. Kuenen 1960; Kingma 1974). Perhaps the large structural high about 80 km south-east of Mahia Peninsula and immediately west of the southern end of Poverty Basin — well represented on bathymetric maps where a particularly extensive block of older, acoustically opaque rock formation lies close to the outer margin of the ECFB, (X on Fig. 8) — has some bearing on this and may contain remnants of an older, perhaps Cretaceous, landmass.

In conclusion, we are lead to believe that the subduction model of continental accretion by imbricate thrusting of oceanic sediments is not applicable to the ECFB. There is not the place here for a more complete synthesis, but the currently available data lead us to believe that the situation may rather be similar to that of many active margins (Japan, Mariana, Central America, Peru-Chile), where recent investigations have found evidence not for continental accretion, but for tectonic erosion of continental blocks (Katz 1971b; Scholl *et al.* 1977; Sisskind 1977; Karig *et al.* 1978; von Huene *et al.* 1978; Hussong 1979; Mrozowski and Hayes 1979; von Huene 1979).

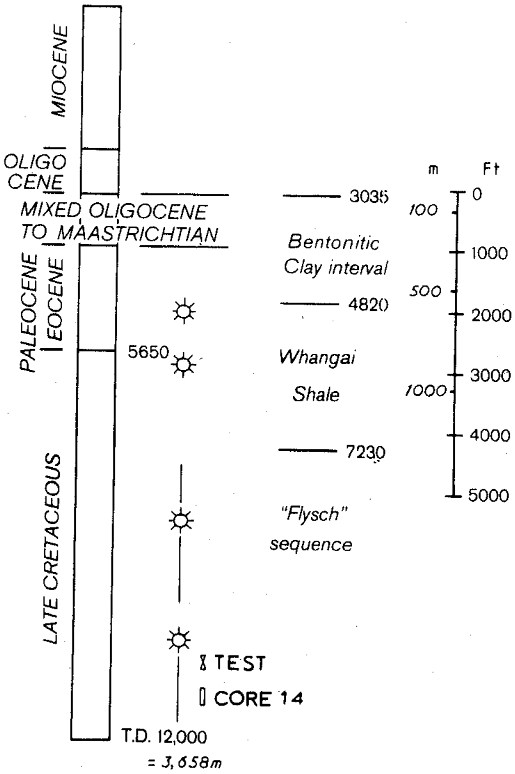

Figure 13. Well section of Opoutama-1 (after Zimmermann and Faber 1967; for location see Fig. 14). Note zones of good gas shows mainly in Cretaceous sequence.

PETROLEUM PROSPECTS

The foregoing analysis suggests that the submerged part of the ECFB is basically similar in sedimentary composition and structure to the onland portion. In particular, the absence of a subduction complex formed of highly imbricated and ocean-derived, pelagic sequences, and the in-place deformation of land-derived sediments right down to the lower slope indicate that the hydrocarbon potential of the ECFB as a whole is similar to that found on the shelf and onshore, where both oil and gas are known to have been generated extensively (Katz 1974b).

Diapiric structures on seismic sections suggest that over-pressured shales in the Lower Tertiary and Upper Cretaceous sequences, well known on land, occur widely in the submerged part of the ECFB as well. These shales generally have a high organic content and are associated with methane generation (Hedberg 1974). Gas-charged diapirs and mud volcanoes discharging hydrocarbon gases are abundant on the East Coast (Ridd 1970), and we believe that, likewise, hydrocarbons have been, or are, widely originating offshore along the continental margin. With existing tectonic conditions there is no reason to assume, for the offshore continental margin area, a past environment less favourable for the deposition of source material than there was onshore; further, the propagation of the tectonic front with locally intense disruptions and reorganization of the sediments may have caused a sufficiently sharp increase of pressure and temperature to accelerate maturation (Montecchi 1976). In the undeformed foredeeps, filled with mainly land-derived sediments several km thick, the very rapid deposition (Katz 1979) of a considerable proportion of shale components in these turbiditic sequences may have led to additional generation of appreciable quantities of hydrocarbons beyond the tectonic front through burial metamorphism alone. Following prevailing pressure gradients, much of the diagenetic fluids from these sequences, including hydrocarbons, would tend to move up-slope into the ECFB. However, the general tectonic mobility and ever changing structural patterns may have prevented the formation of lasting traps, with the result that fluids would be continuously further dispersed through the partly homogenized rock mass and eventually escape. On the other hand, the pressure-temperature relations in sediments at great

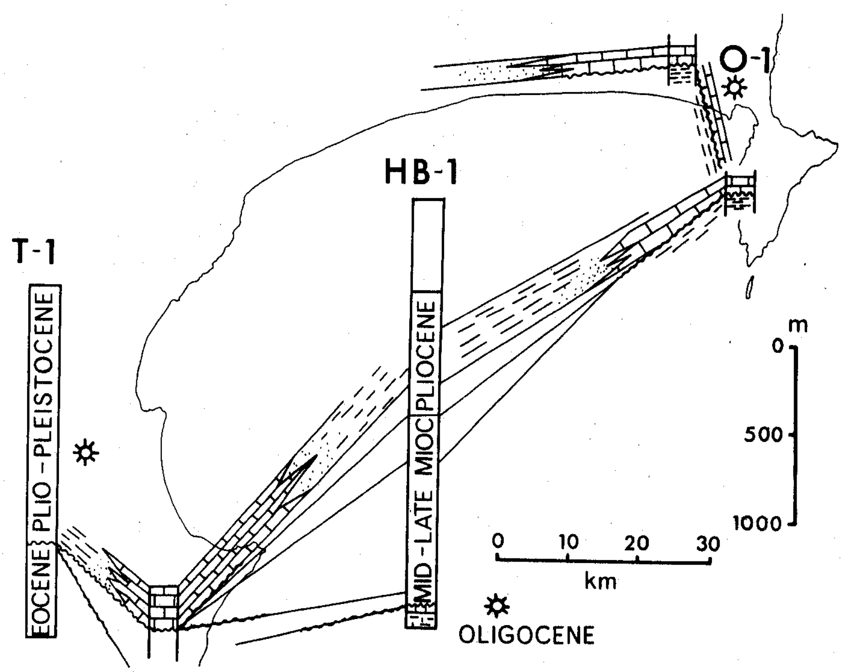

Figure 14. Fence diagram across Hawke Bay area, central East Coast, showing facies change between Plio-Pleistocene shallow-water bioclastic limestone and deep-water mudstone. Well sections with good gas shows, Taradale-1 (T-1), Hawke Bay-1 (HB-1); and location of Opoutama-1 (cf. Fig. 13). After Heffer *et al.* 1976.

water depth, but relatively shallow sub-bottom depth in the lower slope and foredeep areas, may have resulted in the formation of gas hydrates which could effectively have trapped free gas. If so, the possibility exists for vast quantities of gas being locked-in under the continental margin (Shipley *et al.* 1979).

Areas with favourable structural traps may exist in some of the less disturbed zones, such as in the central part of the ECFB southeast of Hawke Bay (Figs. 8, 11). Sands interfingering with the sourcing mudstones would constitute the main reservoir potential. If the indication of an easterly source of some of the Cretaceous formations on land is correct, more extensive porous sandstones can be expected nearer this source area under the present-day submerged margin. In this light the increasingly good gas shows downwards in the thick but mainly shaly-silty-muddy Cretaceous 'flysch' sequence of Opoutama-1 (Fig. 13; Zimmermann and Faber 1967), right on the coast near Mahia Peninsula northeast of Hawke Bay, are particularly important and encouraging: associated sandstone formations of better reservoir quality may exist offshore to the east.

Fluids migrating up-slope may also have been trapped within sandstones and bioclastic limestones of the Late Cenozoic blanketing complex which drapes over and flanks the elongate tectonic protrusions, solitary diapirs, and other fold and fault structures. In particular, the Plio-Pleistocene limestones which in the Hawke Bay region consist mainly of barnacle plates are of high porosity and permeability (de Caen and Darley 1968). These rocks are concentrated along pre-existing structural highs but laterally grade into bathyal mudstone and would be excellent reservoir rocks (Fig. 14). Considering the structure and tectonic evolution of the offshore region, there is good reason to believe that similar limestones and/or calcareous sandstones have locally developed also in the submerged part of the ECFB and could be favourably located for the entrapment of hydrocarbons.

Structural traps associated with shale diapirs have an added attraction in that the diapirs themselves are a potential source for hydrocarbons. Traps formed in strata adjacent to diapirs (Figs. 4, 5) may be found in many places on the continental shelf and slope. Even where good reservoir quality in overlying beds is lacking, modern drilling technology may allow profitable production if a sufficiently thick gas-charged section can be put on stream.

ACKNOWLEDGEMENTS

We would like to thank Gulf Oil Company for providing us with copies of their offshore seismic data, and the Mobil Oil Corporation for their prompt response to our request for copies of their reprocessed seismic reflection data.

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SOME ASPECTS OF RESERVOIR CHARACTERISTICS IN JAPANESE OIL AND GAS FIELDS, WITH SPECIAL REFERENCE TO THEIR DEPOSITIONAL ENVIRONMENTS

H MIYAZAKI, Y IKEBE AND M UKAI

ABSTRACT

Oil and gas fields have been discovered mostly in the Neogene sedimentary basins located in the Uetsu geosyncline onshore and offshore along the coast of the Japan Sea, in the Northeastern Honshu Arc. Reservoirs are composed chiefly of volcanic complexes ranging from lavas to sedimentary tuffs, and of deep-sea sandstones, which are nearly all tuffaceous and derived by turbidity currents.

Volcanic reservoirs providing vug/fracture type pore systems are composed of rhyolite to basaltic andesite of early Miocene to early Pliocene age. Hydrocarbons are preserved in the piles of volcanic units, some of which seem to be enveloped individually by sedimentary tuffs and marine pelagic mudstones and others amalgamated. It seems likely that most of the reservoirs have originated by subaqueous volcanic activity. Following subaqueous eruption, movement of the lavas and associated rocks, accompanied by autointrusion and autobrecciation, must have taken place, bringing about the formation of primary vugs and fractures.

Almost all of the sandstone reservoirs in Japan are turbiditic in origin. The Neogene sequence reveals a single sedimentary cycle in a broad sense since early Miocene, and the hydrocarbon-bearing sandstones were deposited while the area was widely covered by sea. At the end of Pliocene time, strong tectonic movement took place and most of the shallow marine sequences deposited in marginal areas were deeply eroded. Some sedimentological studies have been carried out in the oil and gas fields in the Niigata sub-basin with the aim of making environmental interpretations of reservoir sandstones.

Volcanic reservoirs have been scarcely recognized in the world but recent indications are that the hydrocarbon potential in small basins containing volcanic complexes situated on and near island arc systems is worth investigating.

INTRODUCTION

The Japanese islands are situated on a volcanic arc, and the sedimentary sequence as a whole is characterized by the presence of volcanic rocks. Commercial hydrocarbon accumulation has been found only in the Uetsu Geosyncline — filled with a thick Neogene sequence — in the Northeastern Honshu Arc (Fig. 1). The reservoirs are composed of volcanic complexes (ranging from lavas to sedimentary tuffs) or deep-sea sandstones which are nearly all tuffaceous.

Five of 12 major oil and gas fields discovered during the last 25 years have coarse volcaniclastic reservoirs with brecciated and fractured lavas (Table 1) and they represent some 20% of the initial reserves found in Japan. It may be noted here that the term volcanic reservoir as used in this paper implies 'volcanic complex'.

Recent information on the widely outcropping volcanic complex indicates that most of the volcanic reservoirs are products of submarine volcanic activity. It is also evident that the flysch-type sedimentary rocks in the area under discussion were produced mainly by turbidity currents. This paper reviews and discusses recently acquired knowledge of the depositional environments of reservoir rocks in the Niigata sub-basin of the Uetsu geosynclinal basin.

GEOLOGY

The Uetsu geosynclinal basin is divided into two sub-basins: Akita and Niigata sub-basins (Fig. 1). A Neogene sequence overlies Paleozoic sedimentary rocks and granites, possibly of Cretaceous age. Both sub-basins seem to have followed a fairly similar geologic development. The Neogene sequence, with an estimated maximum thickness of 10 000 m in the Niigata sub-basin, represents a single sedimentary cycle in a broad sense (Table 2). The geosynclinal basin itself had its beginnings in fault-bounded grabens in early Miocene time, when severe volcanism was followed by subsidence. During Pliocene time, strong tectonic movement occurred regionally, caused by block movement and compressive stress (Kitamura 1976). The movement determined the present geologic features as well as most of the structural traps for final hydrocarbon accumulation.

Most of the shallow-marine sedimentary rocks, deposited especially in the eastern marginal portions, were deeply eroded. It is assumed that, for this reason, only deep-sea reservoirs have been preserved in the area. Most of the oil fields are situated in the Akita sub-basin and almost all of the gas fields in the Niigata sub-basin. Hydrocarbon-bearing horizons range from early Miocene to early Pliocene in age.

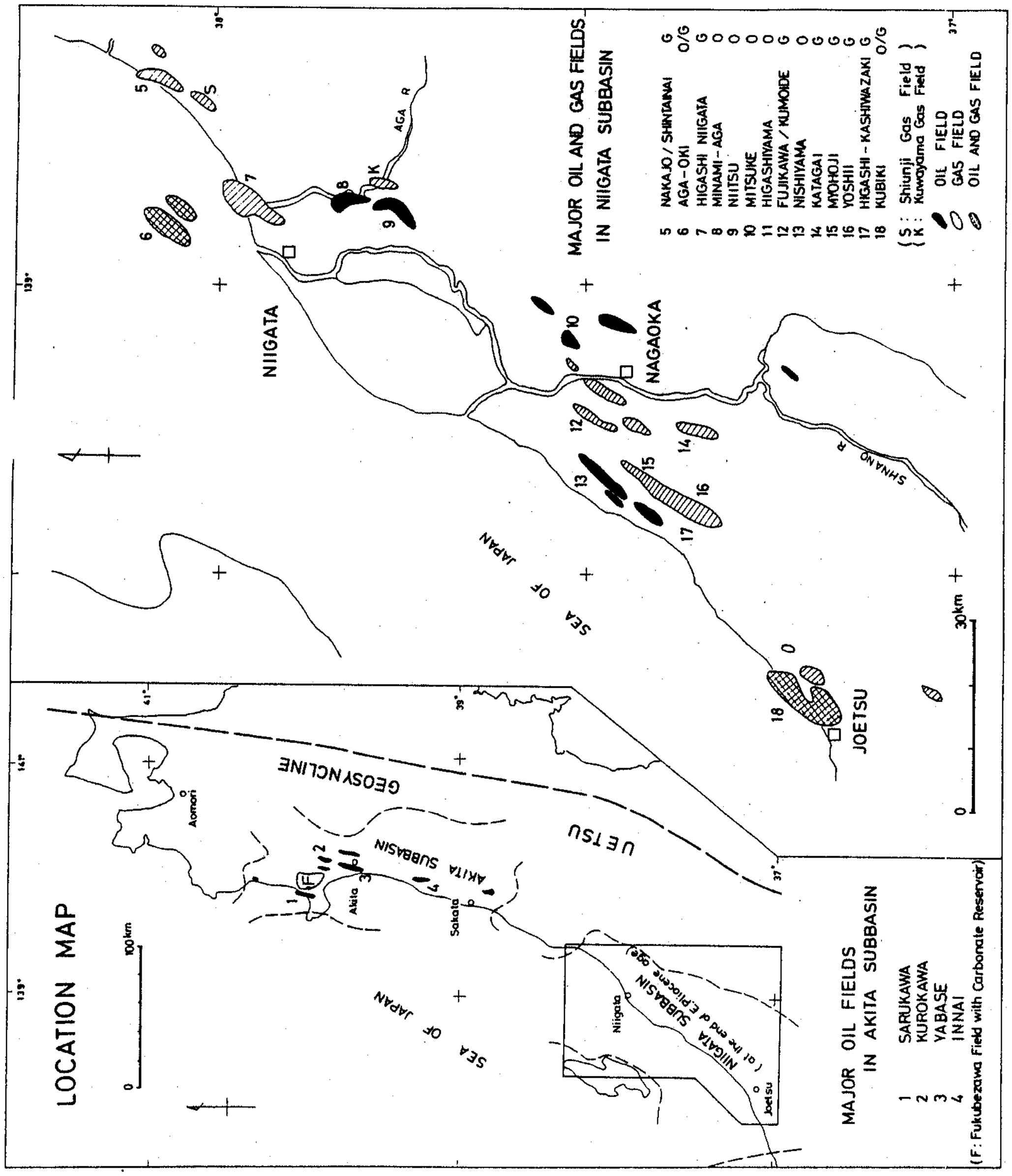


Figure 1. Location map of oil and gas fields in Japan.

TABLE 1

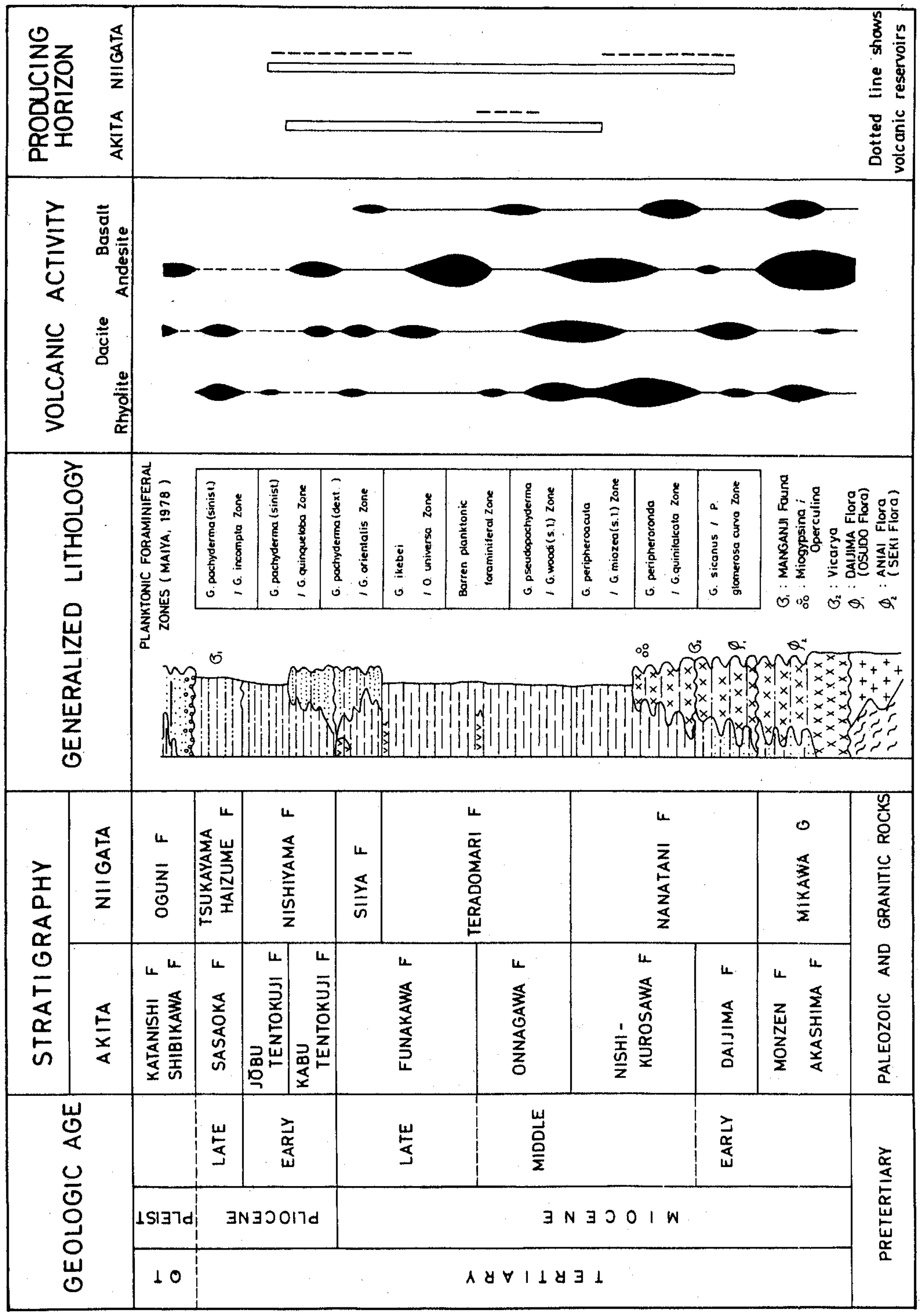
List of oil and gas fields in Japan. Oil and gas fields with reserves of 1 million kl or more are listed; 1000 m^3 of gas is equivalent to one kl of oil for practical purposes.

Field		Basin and	Year of Discovery	Producing Horizon	Lithology of Reservoir	Depth of Reservoir
NISHIYAMA	OF	N	1883	Sy - Td	Tuffaceous Sandstone	-400 / -900
HIGASHIYAMA	OF	N	1888	Sy - Td	Sandstone	+400 / -700
NIITSU	OF	N	1896	Sy - Td	Tuffaceous Sandstone	-50 / -850
KUROKAWA	OF	A	1913	On	Andesite	-250 / -300
INNAI	OF	A	1923	Kb - Fu	Tuffaceous Sandstone	-600 / -1300
YABASE	OF	A	1933	Kb - On	Tuffaceous Sandstone	-350 / -1700
KUBIKI	O/GF	N	1955	Td	Tuffaceous Sandstone, Tuff	-400 / -1900
SARUKAWA	OF	A	1958	Kb - On	Tuffaceous Sandstone, Tuff	-600 / -1500
MITSUKE	OF	N	1958	Nt	Dacite, Dacitic Tuff-breccia	-1500 / -2000
HIGASHI-NIIGATA	GF	N	1959	Ny - Td	Sandstone, Tuffaceous Sandstone	-1200 / -3400
KATAGAI	GF	N	1960	Ny , Nt	Andesitic Agglomerate	-800 , -4100
NAKAJO/SHINTAINAI	GF	N	1961	Ny - Sy	Sandstone	-800 / -2000
FUJIKAWA / KUMOIDE	GF	N	1962	Ny - Td	Dacitic Tuff-breccia, Sandstone	-1400 / -3000
MINAMI-AGA	OF	N	1964	Sy	Tuffaceous Sandstone	-2250
YOSHII	GF	N	1968	Nt	Andesite, Andesitic Tuff-breccia	-2600
HIGASHI-KASHIWAZAKI	GF	N	1969	Nt	Andesite, Andesitic Tuff-breccia	-2600
MYOHOJI	GF	N	1969	Nt	Andesite, Andesitic Tuff-breccia	-2600
AGA-OKI	O/GF	N	1972	Ny - Sy	Sandstone	-1900 / -2400

NOTES A : Akita Sub-basin (Kb : Kabu-Tentokuje Fm, Fu : Funakawa Fm, On : Onnagawa Fm)
N : Niigata Sub-basin (Ny : Nishiyama Fm, Sy : Shiiya Fm, Td : Teradomari Fm, Nt : Nanatani Fm) Depth of Reservoir in meters
OF : Oil Field, GF : Gas Field, O/GF : Oil and Gas Field.

TABLE 2

Neogene stratigraphy of Akita and Niigata sub-basins.



VOLCANIC RESERVOIRS

Lithology

The volcanic rocks acting as reservoirs are rhyolite to basaltic andesite of early Miocene to early Pliocene age. Lithologically, they consist of lavas and volcaniclastic associates ranging from breccia to sedimentary tuff. Basalts and intrusive rocks predominantly developed in the Miocene sequence have not been recognized as reservoirs so far.

Conduit system

The pore system of the reservoirs is in general of fracture/vug type with total porosity of 15% to 20%. In the volcaniclastic rocks, a part of the pore system seems to be based on the original intergranular pores, secondary fracture pores and dissolution casts of individual components. The presence of fracture/vug type porosity was indicated by comparison of porosity values from sonic log and density log on sandstone and volcanic reservoirs (Ukai *et al.* 1972). Fluorescence along fractures in cores immediately after their recovery from wells has often been reported. Productivity is extremely high in the middle of the brecciated lavas at the Yoshii and Mitsuke fields but decreases towards the perimeter, where the reservoir is composed of fine-grained volcaniclastic rocks.

Geometry and depositional environments

Hydrocarbons are preserved in the piles of volcanic units; some of the units seem to be individually enveloped by sedimentary tuff and/or marine pelagic mudstone, others may be amalgamated. Microscope studies on thin sections of volcanic-reservoir rocks from the

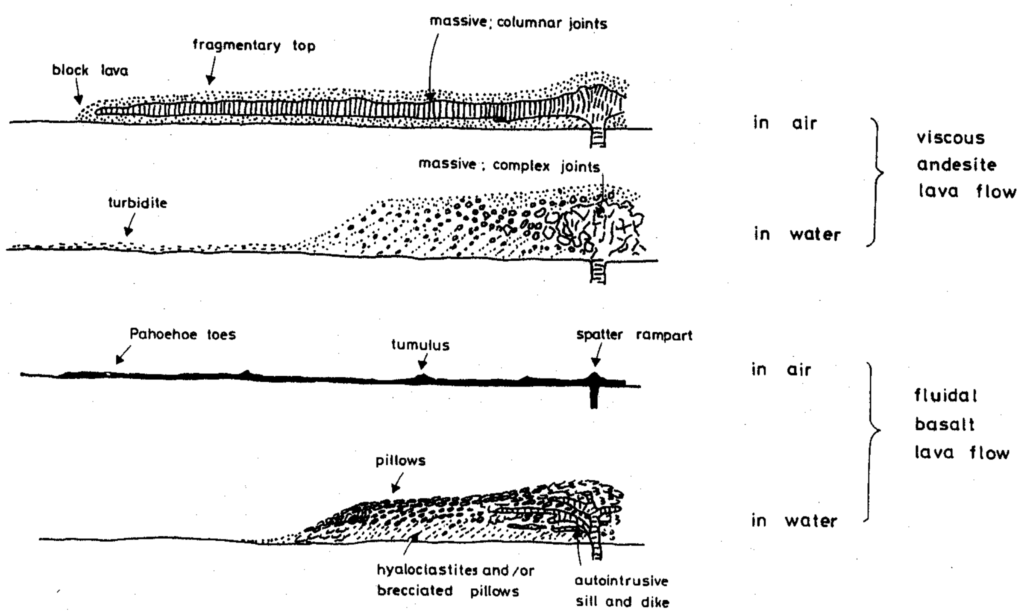


Figure 2. Schematic sections of internal structures of volcanic bodies produced by viscous andesite and fluidal basalt lava flows, for eruptions above surface and under water (After Aramaki and Nakamura 1967). A practical assumption is that eruption takes place in soft sediments under water and that the volcanic substances are usually mixed with the sediments. Some varieties of hyaloclastite may have accumulated with large particles located close to the source, particle size gradually diminishing with distance from the centre.

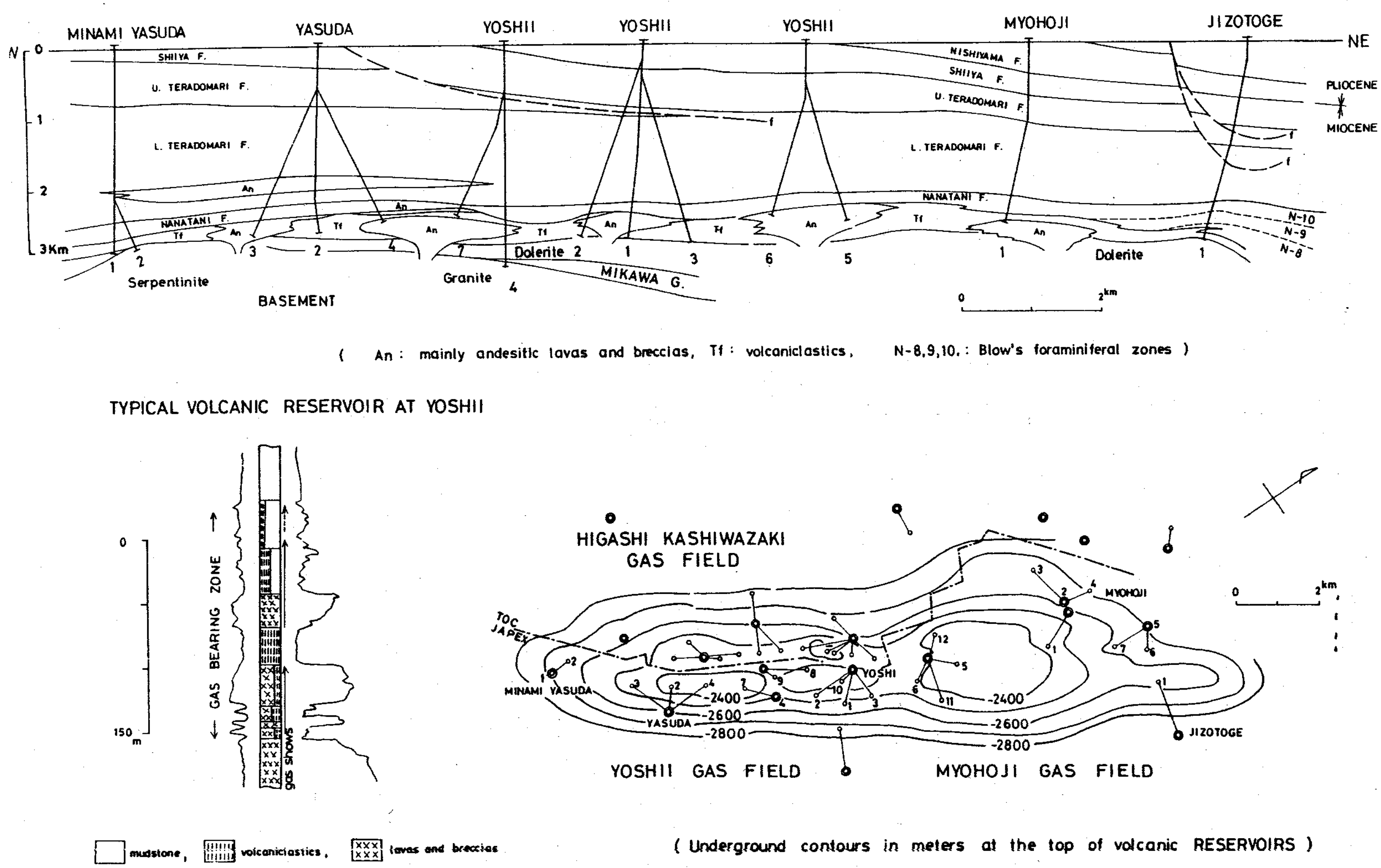


Figure 3. Longitudinal section of the Yoshii and Myohoji gas fields. Some highly productive bodies composed chiefly of lavas and breccias are assumed to exist independently in the volcaniclastics. Blow's zones N.8 to N.10, overlapping in order towards the top of the volcanic complexes, are shown at the righthand end of the section.

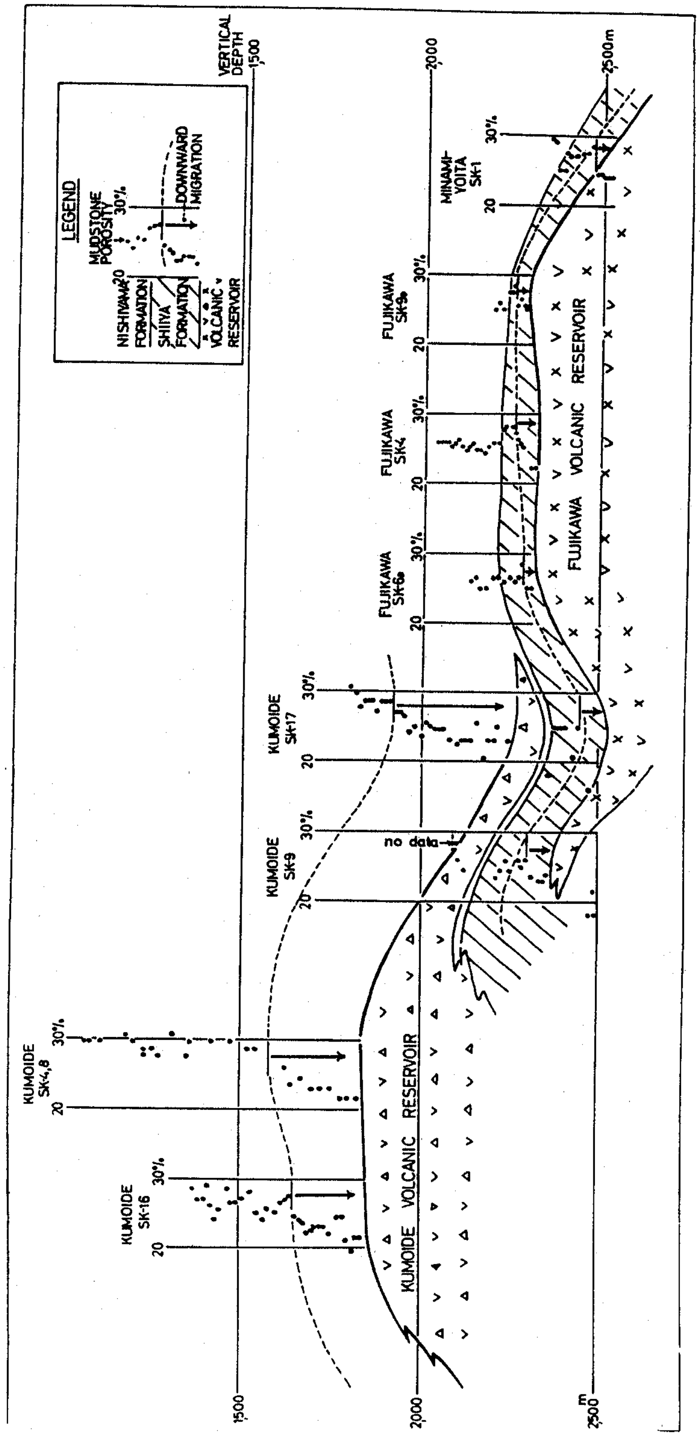


Figure 4. Vertical porosity distributions of mudstone in Fujikawa/Kumoide trend. (After Magara 1968.)

Yoshii field showed them to be fine-grained and glassy andesites with rare phenocrysts and many amygdales. A small amount of celadonite and palagonite are recognizable as altered green minerals formed by reaction with sea water.

From the presence of fossiliferous marine mudstones interbedded in the reservoirs or enclosing them it seems likely that most of the volcanic-reservoir rocks in the basin originated from subaqueous volcanic activity, i.e. they are hyaloclastites as redefined by Kawaguchi *et al.* (1976).

Subaqueous volcanic products have been studied at outcrops in the basin and distinctive features described (Matsuda and Nakamura 1970, Tamiya *et al.* 1973, Yamagishi *et al.* 1979). Following the subaqueous eruptions, some movement and autobrecciation of the lavas and their associates must have occurred, along with immediate cooling, so contributing to the formation of fractures and vugs. Eruptive activity under water would have been small in scale owing to hydrostatic pressure and would have varied according to water depth.

Observation of outcrops suggests that products of subaqueous eruptions sometimes built up prograding fore-set beds, including pillow lavas and slumpings, and formed thick lens-like masses at the edge of which density currents of hyaloclastites might have originated. Lavas cooled in water usually have irregular joint or fracture systems, whereas lavas cooled in air form regular columnar joints (Fig. 2). Autointrusion may have taken place, especially in basic complexes, after a series of eruptions.

The volcanic piles are presumed to have persisted as paleo-highs after the eruptions and were gradually buried by mudstone. This is suggested by the fact that Blow's planktonic foraminiferal zones N.8 to N.10 overlap in order towards the top of the volcanic mass at the Yoshii gas field (Fig. 3).

Hydrocarbon accumulation

The mechanism of hydrocarbon migration into the volcanic reservoirs may resemble that for carbonate rocks. The source rocks must have been the mudstones enclosing or interbedded with the porous volcanic rocks. As mentioned before, some of the volcanic units appear to be individually separate, and continuity in the conduit system of each unit with another must initially have been poor. Sources, therefore, must be close to reservoirs, and long-distance migration is unlikely.

Although continuity of the pore systems in the present fields seems variable, each field has a more or less uniform water table. The present continuity of the pore system was produced by tensional fracture systems developed along the anticlinal axis by the tectonic movements responsible for the present geologic structures.

For the primary migration of hydrocarbons into volcanic reservoirs in the Niigata sub-basin, Magara (1968) proposed the downward migration hypothesis. He noticed abnormal high-porosity zones on the porosity/depth profiles for thick mudstone sequences above hydrocarbon traps (Fig. 4). High-porosity zones correspond to abnormal pressure zones (undercompacted zones), and the compaction current may move downwards below such undercompacted zones and upward above the zones. Downward fluid movement due to increasing overburden and accompanying hydrothermal effect must have forced hydrocarbons generated in nearby mudstones into volcanic reservoirs situated beneath the mudstones, sometimes through the sedimentary tuffs enclosing the reservoirs. It seems that the undercompacted zones also served as cap rocks for hydrocarbon accumulation.

The timing of primary hydrocarbon migration into volcanic reservoirs is very important, because the sedimentary tuffs that usually enclosed each volcanic unit would gradually have become compacted owing to increasing overburden; as compaction increased it would have become difficult for the hydrocarbons to pass through the tuffs into the reservoirs.

Volcanic reservoirs sometimes produce considerable amounts of N_2 and CO_2 gases accompanied by hydrocarbon; they are thought to be of volcanic origin. The deepest volcanic reservoir recently found is at 4100 m below sea level in the Niigata sub-basin; wet gas accompanied by condensate was recovered.

SANDSTONE RESERVOIRS

Lithology

Almost all of the sandstone reservoirs in the Niigata sub-basin are turbiditic in origin. Stratigraphically, they are concentrated mainly in the late Miocene to early Pliocene sequence. The sandstones, as a whole, are markedly tuffaceous and in places pumiceous, so that mineralogically they are quite complex because of subsequent alteration of material of volcanic origin to clay minerals.

Reservoirs

Clay usually has considerable effect on the intergranular pore system as well as on wireline log response. Hydrocarbon-bearing sandstones in the sub-basin are unconsolidated and provide reasonable porosity, but generally low permeability. Zeolite minerals, especially, are low-density hydrous aluminosilicates characterized by large ion-exchange capacity. It is possible, therefore, that the neutron tool would read the porosity units based on matrix-hydrogen index too high, whereas resistivity value is usually low in the zeolite-bearing intervals.

Depositional environments

Reservoir sandstone is interbedded with pelagic mudstone, which usually contains foraminiferal assemblages indicating a deep-sea environment. Turbidites generally contain shallow-marine and/or shelf fauna and are accompanied by volcaniclastic fragments of various grain sizes. Accordingly, it is assumed that the volcanic piles situated topographically as paleo-highs may have supplied source material to gravity flows. Restored sections and isopach maps sometimes indicate that sand bodies were deposited as paleo-topographic lows. A good example is known in the Shiunji gas field.

Studies on turbidites in the basin have been based mainly on outcrop observation (e.g., Honza 1965, Sasaki and Ushijima 1966, Morita *et al.* 1973, Kageyama and Suzuki 1974, Tsuda *et al.* 1976). Morita *et al.* (1973) constructed paleogeographic maps of Shiiya and Nishiyama ages, concluding from plaeocurrent directions that the present trend of folding seems to have developed at the beginning of the Nishiyama time (Fig. 5).

Abe (1978) made a detailed correlation of lithologies in the Minami - Aga oil field and surrounding area, using cores together with micro-log and dipmeter-log resistivity-curve patterns. He successfully delineated the submarine fans formed of the reservoir sandstone (D) of the Shiiya Formation. The direction of source area (south) for these sandstones was also made clear, by petrological examination of associated gravels (Fig. 6). It is possible that the main reservoirs of the Higashi-Niigata/Matsuzaki gas field, which are stratigraphically above the reservoir sandstone of the Minami - Aga field, are parts of the suprafans that have prograded northwards after deposition of the sandstone of the Minami - Aga field.

Gas-bearing sandstones of the Shiunji gas field (of the Nishiyama Formation) are seen only at the saddle of the anticline. Here gross thickness, grain size of the sandstones, and sand/shale ratio were carefully counted or examined. Distribution patterns of these items on each interval of reservoir groups, together with micro-paleontological evidence, led to an interesting environmental interpretation for these sandstones (Fig. 7). The coarse sedimentary rocks contain a shallow-marine assemblage of benthonic forams and an exotic assemblage representing the older formations. These rocks must have been derived from a source area presumed to lie east of the field. Hydrocarbon accumulation seems to be structurally controlled and to be limited to the western half of the inner fan.

In the recently developed Aga-Oki field, detailed layer-to-layer correlation has been carried out by Okabe (personal communication) for the N-6 zone, one of the main reservoir groups of the Nishiyama Formation, using micro-resistivity curves of dipmeter log, 1:20 in scale (Fig. 8). Each unit correlated consists of 2 to 4 single beds. The geometry of individual units has been found to be similar, and the probable direction of the turbidity currents is north to south (Fig. 9). The study is continuing.

As a result of recent deep drilling at the Niitsu oil field, an oil reservoir of mudstone

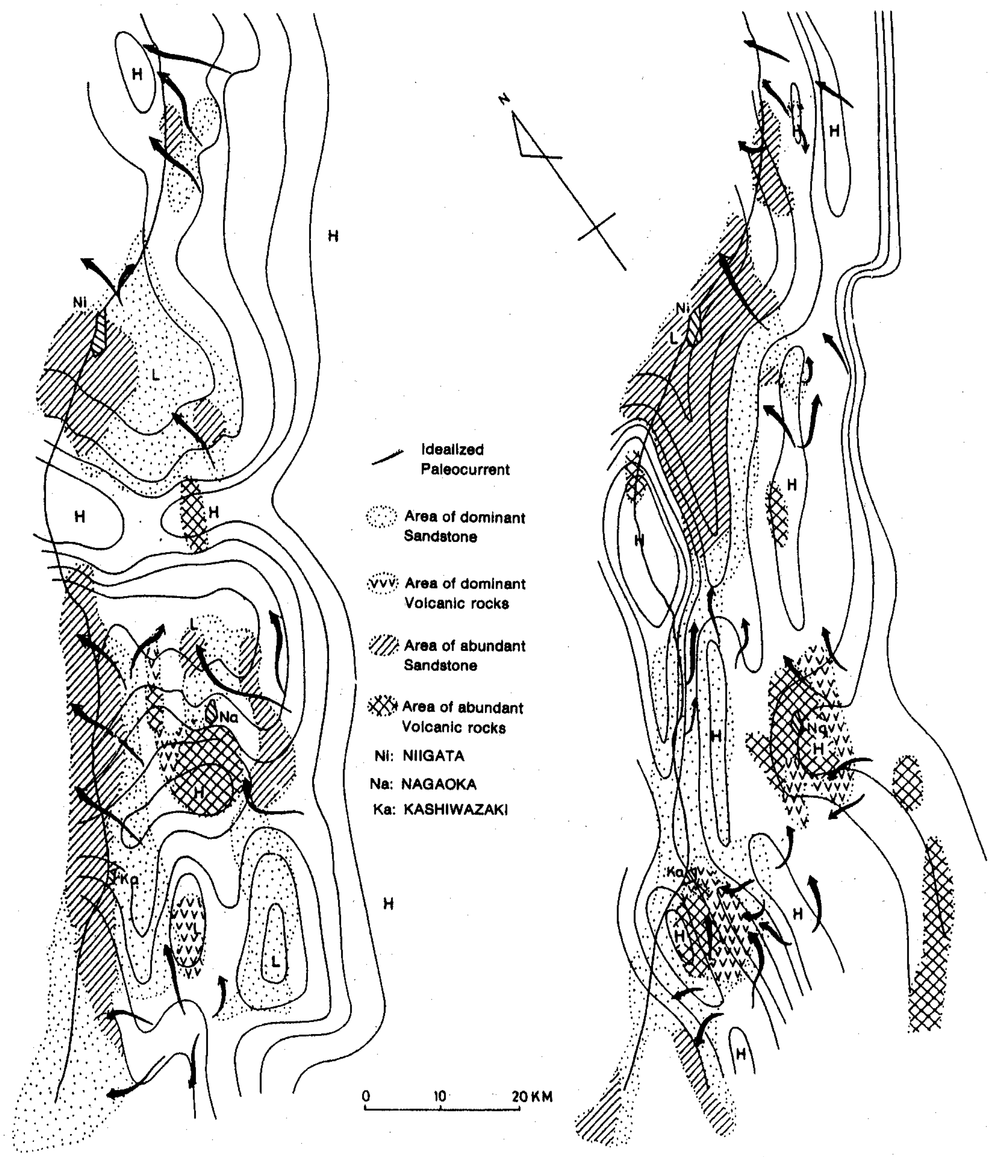


Figure 5. Idealized paleogeographic maps for Shiiya time (left) and Nishiyama time (right). (After Morita *et al* 1973.)

grain size was found in the deep-marine sequence of the Teradomari Formation (Tanaka *et al.* 1979). The reservoir, 10 m thick, is reported to consist of more than 50% of well-sorted quartz grains. A lithologic survey carried out during drilling could not detect the reservoir, but the result of wire-line log analysis justified a test being run. From what is known of the reservoir to date, it may be the product of contour bottom-currents.

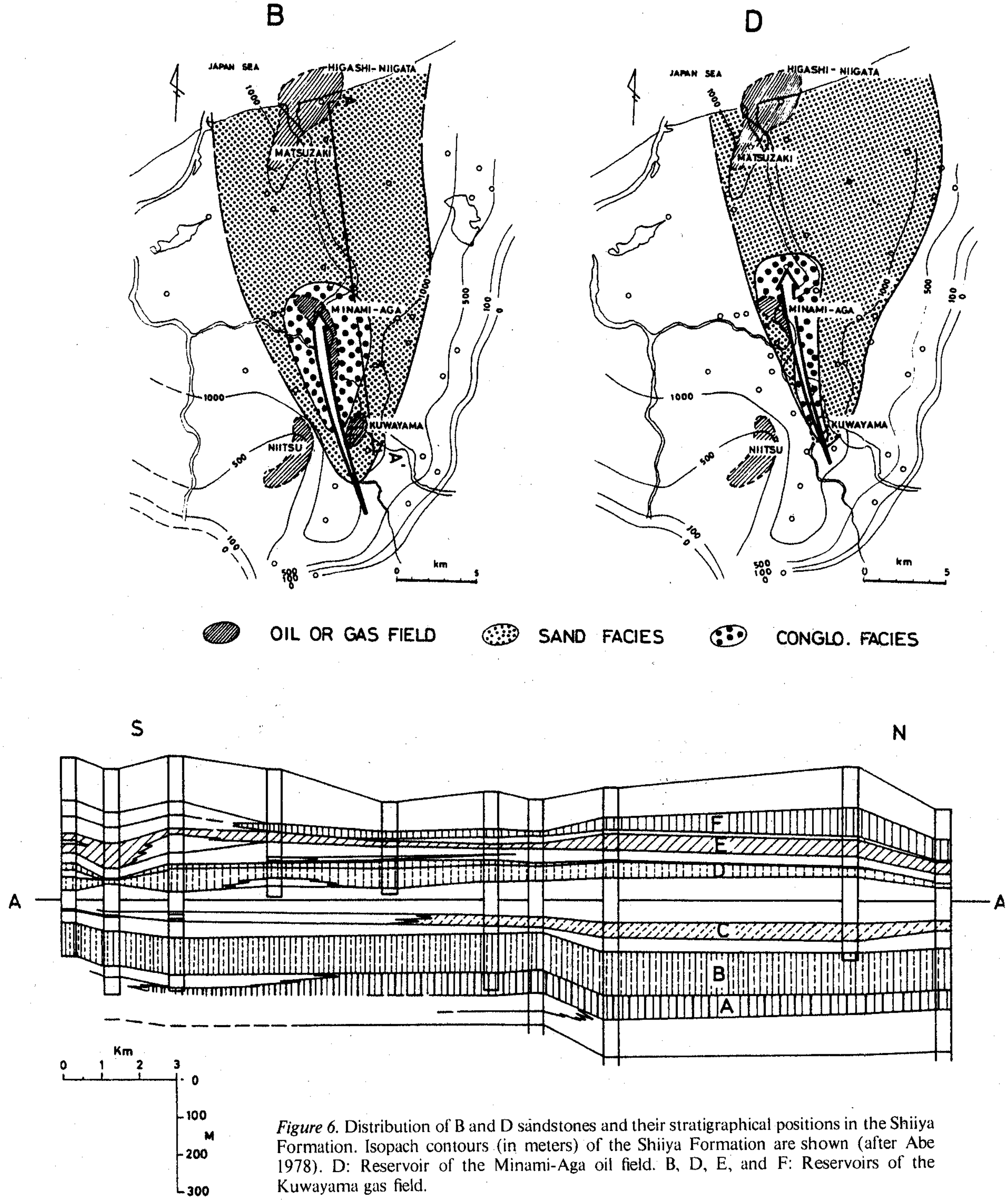


Figure 6. Distribution of B and D sandstones and their stratigraphical positions in the Shiiya Formation. Isopach contours (in meters) of the Shiiya Formation are shown (after Abe 1978). D: Reservoir of the Minami-Aga oil field. B, D, E, and F: Reservoirs of the Kuwayama gas field.

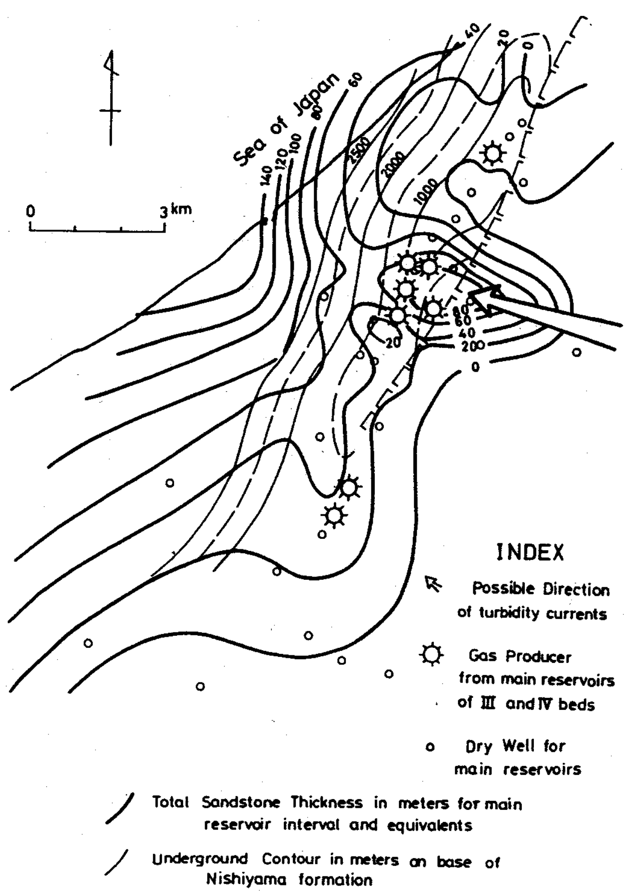


Figure 7a. Thickness of main reservoir sandstones around the Shiunji gas field. Hydrocarbon accumulation seems to be structurally controlled.

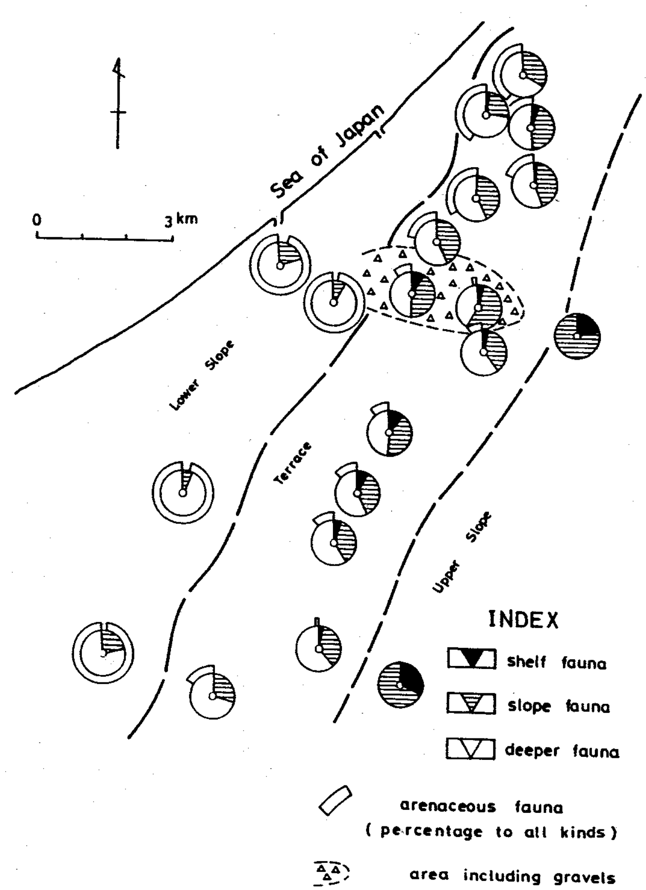


Figure 7b. Paleoenvironments during the deposition of main reservoir sandstones around the Shiunji gas field as deduced from benthonic foraminiferal assemblages.

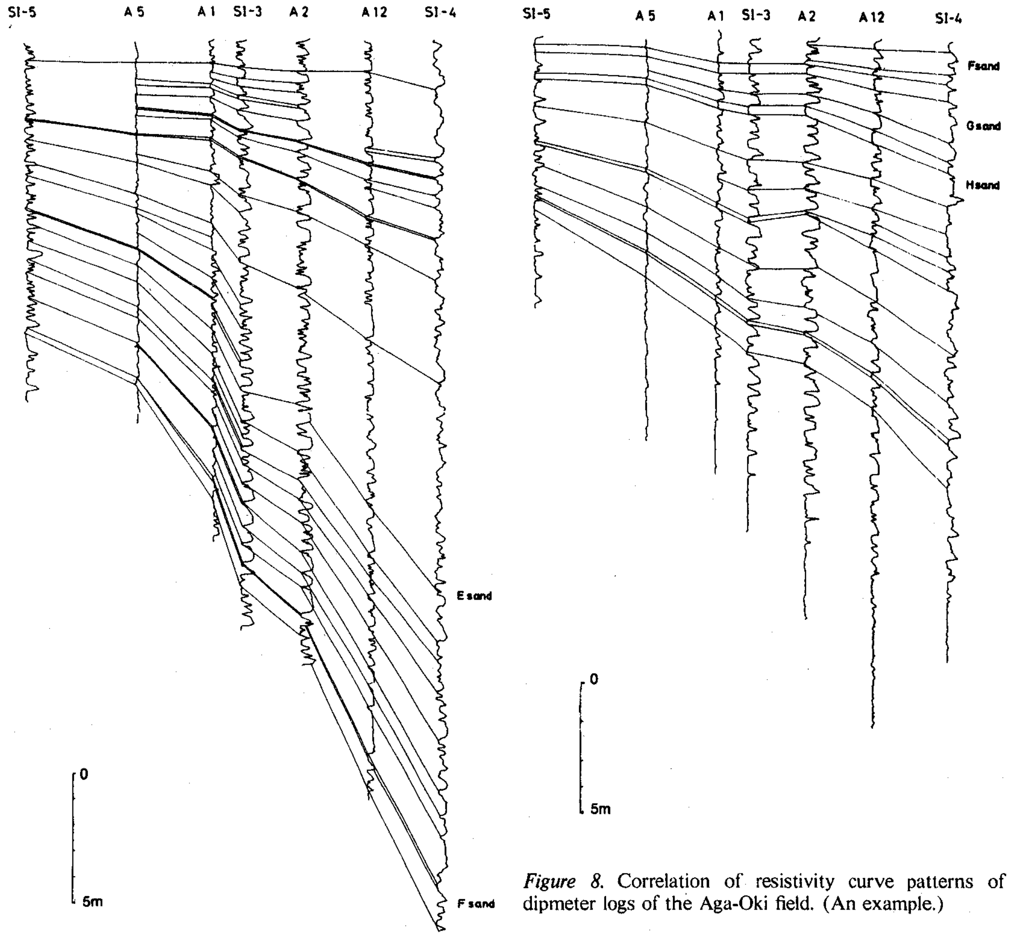


Figure 8. Correlation of resistivity curve patterns of dipmeter logs of the Aga-Oki field. (An example.)

CARBONATE RESERVOIRS

Although carbonate is rarely observed in the Neogene sequence in the basin, a small commercial trap of carbonates (Fukubezawa oil field) was recognized in the Akita sub-basin (Fig. 1). These rocks are made up mainly of skeletal debris of diatoms radiolaria, foraminifers and nannofossils (Aoyagi and Kazama 1971) and are enclosed by pelagic marine mudstones.

SUMMARY

Hydrocarbon accumulation in the volcanic reservoirs seems to have occured under more turbulent conditions than in the sandstone and carbonate reservoirs. The two most important factors for hydrocarbon accumulation may be the reservoir quality and the nature of the association of reservoirs with source rocks. Productivity is high in the fracture/vug type reservoirs of lavas and breccias, but generally lower in the finer-grained volcaniclastic rocks, such as volcanic sand and sedimentary tuff. In volcaniclastic rocks with intergranular pores, alteration of unstable components will reduce or eliminate the original porosity (Surdam and Boles 1979) in the early stage of diagenesis (hydration and carbonization). The major diagenetic control is with the chemistry of pore fluids, and the significant hydration reactions are volcanic glass to zeolite and plagioclase to

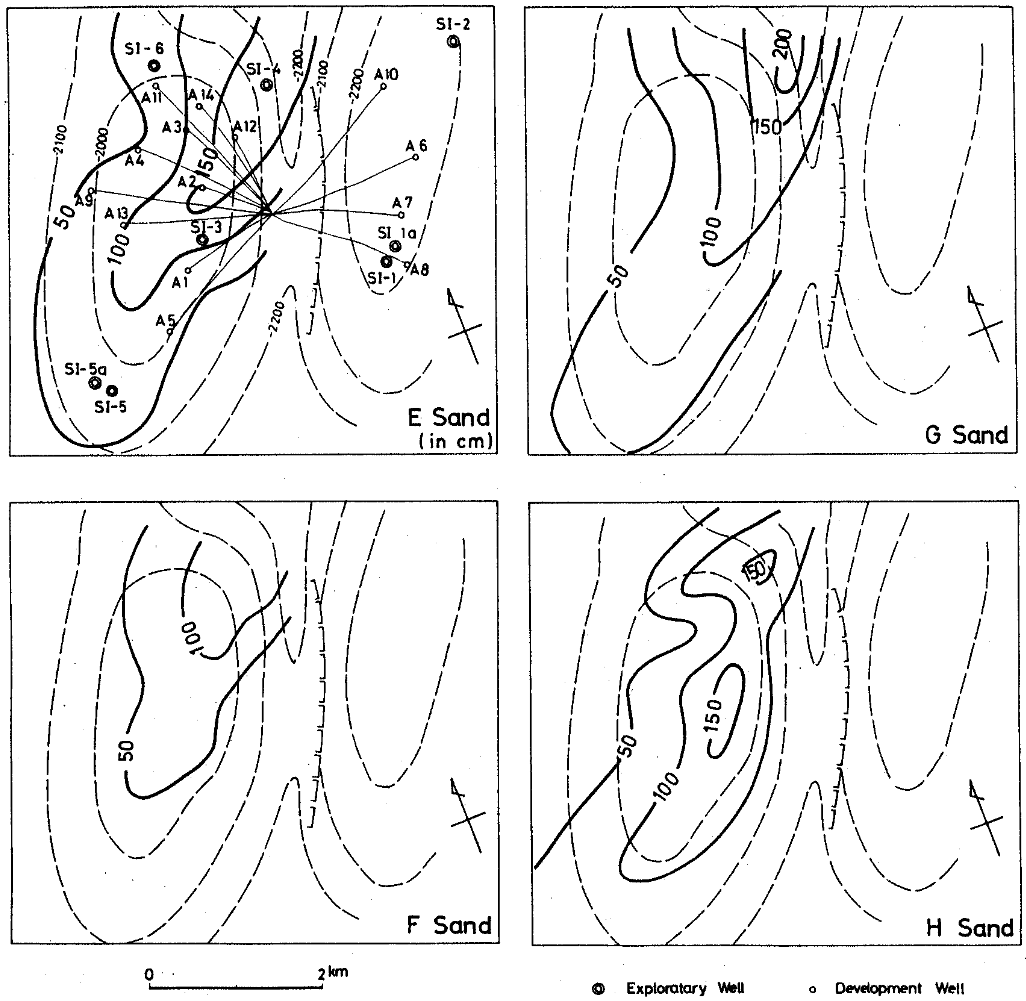


Figure 9. Distribution pattern of each sand unit on the Aga-Oki field (an example). Structure contours on the top of the Shiiya Formation are shown in meters. Source area is located probably to the north.

zeolite (Davies *et al.* 1979). As diagenetic porosity and permeability reduction occur earlier in the volcaniclastics than in the sandstones, the timing of primary migration into volcanic reservoirs must be another important factor for hydrocarbon accumulation.

Volcanic reservoirs have been scarcely recognized in the world, but the experience of recent years has shown the worth of exploring the hydrocarbon potential in small basins containing volcanic complexes situated on and near island-arc systems.

It is evident, too, that environmental and geometrical studies of sandstones are very important and useful in pursuing the prospects of sandstone reservoirs.

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EXPLORATION AND DEVELOPMENT OF THE NIDO REEF COMPLEX OIL DISCOVERY, PHILIPPINES

ALLEN G HATLEY AND ROBERT Y HARRY

INTRODUCTION

At the present time (late 1979) the Philippines produces approximately 40 000 bbls/d from the Nido Reef Complex Oil Field, offshore Northwest Palawan. While this production represents only some 18% of the Philippines daily petroleum requirements, less than a year ago Philippine production was nil and no alternative existed to the importation of 100% of the nation's crude-oil requirements. Additional development projects in the Philippines are in the evaluation stages and exploratory drilling continues in a number of both onshore and offshore basins in that nation.

ACQUISITION OF PROSPECTS

The exploration and development of the first oil production in the Philippines began with disappointment in the results of an exploration well drilled by operating subsidiaries of Citco International Petroleum Company and Husky Oil International Inc. in another basin in the Philippines. Cities Service management considered this exploration well to be sufficiently disappointing as to significantly downgrade additional exploration prospects in that basin. Continuing regional evaluation by the company, however, highlighted a possibly more favorable basin offshore Northwestern Palawan. Previously, this basin had been largely ignored by foreign oil companies, owing to the prior drilling, by a Philippines oil-exploration company, of two dry exploratory wells.

Cities Service regional exploration office in Singapore completed the evaluation of the basin, and in November 1975 recommended the acquisition of a new area offshore Northwest Palawan. This was approved by Cities Service management and that of Husky Oil International Inc. By early December, 1975, agreements had been completed with four Filipino oil companies (Basic Petroleum & Minerals Inc., Landoil Resources Corp., Oriental Petroleum & Minerals Corp., and Philippines Overseas Drilling & Oil Development Corp.), who held petroleum-exploration concessions over parts of this area and with the then authorized Philippine Government agency, The Petroleum Board. On 17 December 1975, the Northwest Palawan Service Contract (Fig. 1) was executed by all parties.

EXPLORATION OF THE NORTHWEST PALAWAN BASIN

Stratigraphy and basin history

The generalized stratigraphic column for the Northwest Palawan Basin is shown on Fig. 2.

The pre-Eocene stratigraphy and geological history in the Northwest Palawan Basin is even today, not well understood; however, Cities Service has for many years considered that the northern portion of much of onshore and offshore Palawan (north of the Ulugan Fault) represents continental crust, initially rifted in the Mesozoic and moved southward from the China continental margin.

In Lower Oligocene times a broad carbonate platform or shelf existed in parts of the basin. Locally, in Upper Oligocene and Lower Miocene time, a number of carbonate 'reef-like' complexes developed, and these are the reservoirs of today's offshore oil discoveries. Fine-grained clastic (typically claystone and mudstone) strata were deposited contemporaneously, flanking the carbonate reefal complex in some parts of the basin. Near the end of the Lower Miocene, conditions conducive to carbonate deposition ended, and

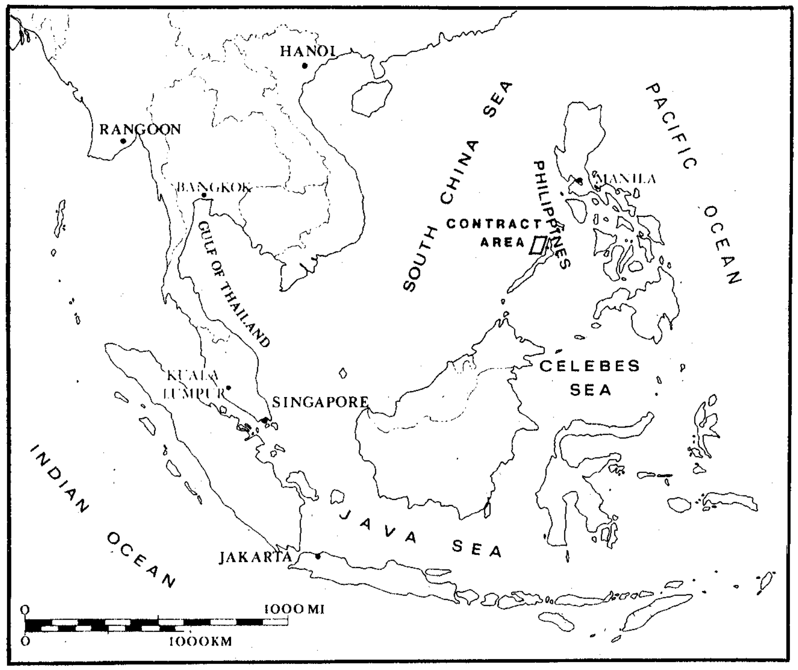

Figure 1. Location of the contract area.

by Middle Miocene times claystone and several very coarse clastic units covered the previously deposited carbonate section.

An important unconformity exists in the Middle Miocene section, at least in the 'Nido Reef Complex' area, and marks a major break in sedimentation. It was caused by uplift, erosion and a shallower water environment of deposition resulting in an overlying, predominantly coarser, clastic sedimentary sequence.

In the Pliocene to Recent, coarse clastics and shallow-water carbonates predominate.

Exploration drilling (for locations of wells see Fig. 3)

The first exploratory well, Nido-1, drilled in the Service Contract by Cities Service and Husky's operating subsidiaries, Philippines-Cities Service Inc. and Husky (Philippines) Oil Co., was spudded-in on 31 January 1976. It was drilled to a total depth of 9026 ft, in 372 ft of water, using the drillship *Tainaron.* At 5960 ft the top of the expected Lower Miocene carbonate build-up or reefal complex was encountered. Shows of oil and gas were encountered near the top of the carbonate section. Upon reaching total depth, Nido-1 was plugged back to 6040 ft and $9^5/_8$-inch casing perforated to test the best potential oil-productive zone in the well. The well was acidized and the zone flowed crude oil at a maximum estimated rate of 1440 bbls/d through a $^5/_8$-inch top choke. At the time, this was by far the most significant amount of oil ever tested in the Philippines.

Nido-1 eventually proved to be non-commercial, with only 17 ft of net pay. It was, however a catalyst for increased exploration interest in the Philippines.

Seismic work during and following the drilling of Nido-1 indicated several more nearby carbonate reefal-complex prospects, and resulted in the delineation of a much larger potential trap, approximately $3^1/_2$ miles updip to the north of Nido-1. North Nido-1, was drilled in late 1976 on this prospect. The top of the Lower Miocene carbonate section was encountered at 5549 ft, approximately 430 ft higher than the same unit in Nido-1. Oil shows were encountered below the top of the carbonate and the next 110 ft were continuously cored. Good shows of oil were encountered in the upper part of the first core. The well was drilled to a total depth of 7614 ft and plugged back for testing. Eight drill-stem tests were run, but producible amounts of oil and gas were not present. The well was plugged and abandoned as a dry hole.

In mid 1977, the Cities Service Consortium (with the exception of Husky) participated in drilling the South Nido-1 exploration well on a separate reef-like carbonate buildup, approximately 3½ miles southwest of Nido-1. The top of the Lower Miocene carbonate was encountered at 6796 ft, some 837 ft lower than in Nido-1. Oil shows were soon encountered, and the carbonate section was cored and drilled to a total depth of 8010 ft. Five tests were

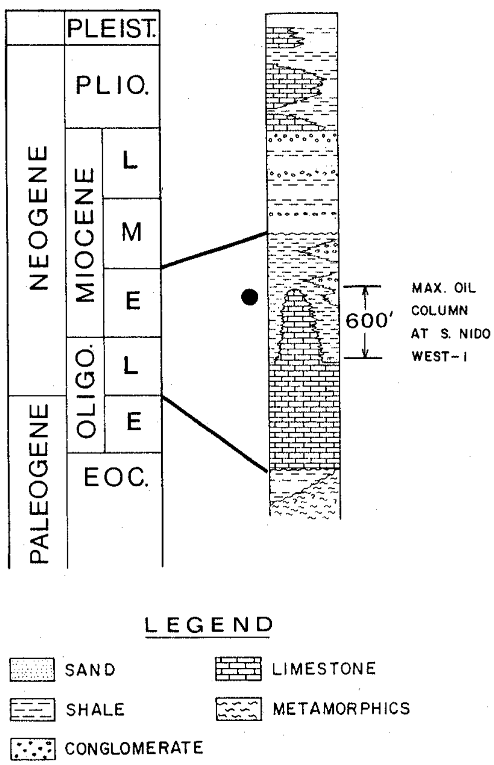

Figure 2. Stratigraphic column of the Northwest Palawan Basin.

conducted through perforations in the 9⅝-inch casing. The well flowed at rates up to 7343 bbls/d.

Although South Nido-1 was a discovery well, a number of problems and some strong reservations regarding the commerciality of the discovery remained. As a result it was decided to undertake a development feasibility study, and further, to conduct a major three-dimensional (3-D) seismic survey in the Nido area. One of our problems was the irregularity of the water bottom, caused by the presence of surface and near-surface reefs of Pliocene to Recent age; interpretation of seismic data was influenced greatly by these shallow features.

Another exploratory well, Guntao-1, was drilled after the South Nido-1 well, in order to test a somewhat different concept and was located 3½ miles north of North Nido-1. It went to a total depth of 7685 ft, and although it had shows of oil in the Lower Miocene carbonates, it failed to flow oil when tested. It was plugged and abandoned as a dry hole.

The development feasibility study, undertaken late in 1977, was a major contribution to exploration and development. In the first place, initial studies indicated not only the necessity for drilling an additional exploratory well on a separate carbonate buildup on the Nido Complex, but also offered the Cities Service Consortium 26 options for development: without an additional significant discovery no commercial development could be undertaken of the South Nido-1 discovery.

South Nido West-1 well, completed in late February 1978 resulted in a second discovery on a separate Lower Miocene–Oligocene carbonate buildup encountered at 6916 ft. The well flowed oil at a maximum rate of 9880 bbls/d in an extended testing program.

Figures 4 and 5 show the relationship between the major producing and non-producing features on the Nido Reef Complex Oil Field.

DEVELOPMENT OF DISCOVERIES

Decision to develop Nido Reef Complex

With the discovery of additional reserves in South Nido West-1, the risk involved in undertaking a possibly unwarranted production feasibility study was more than

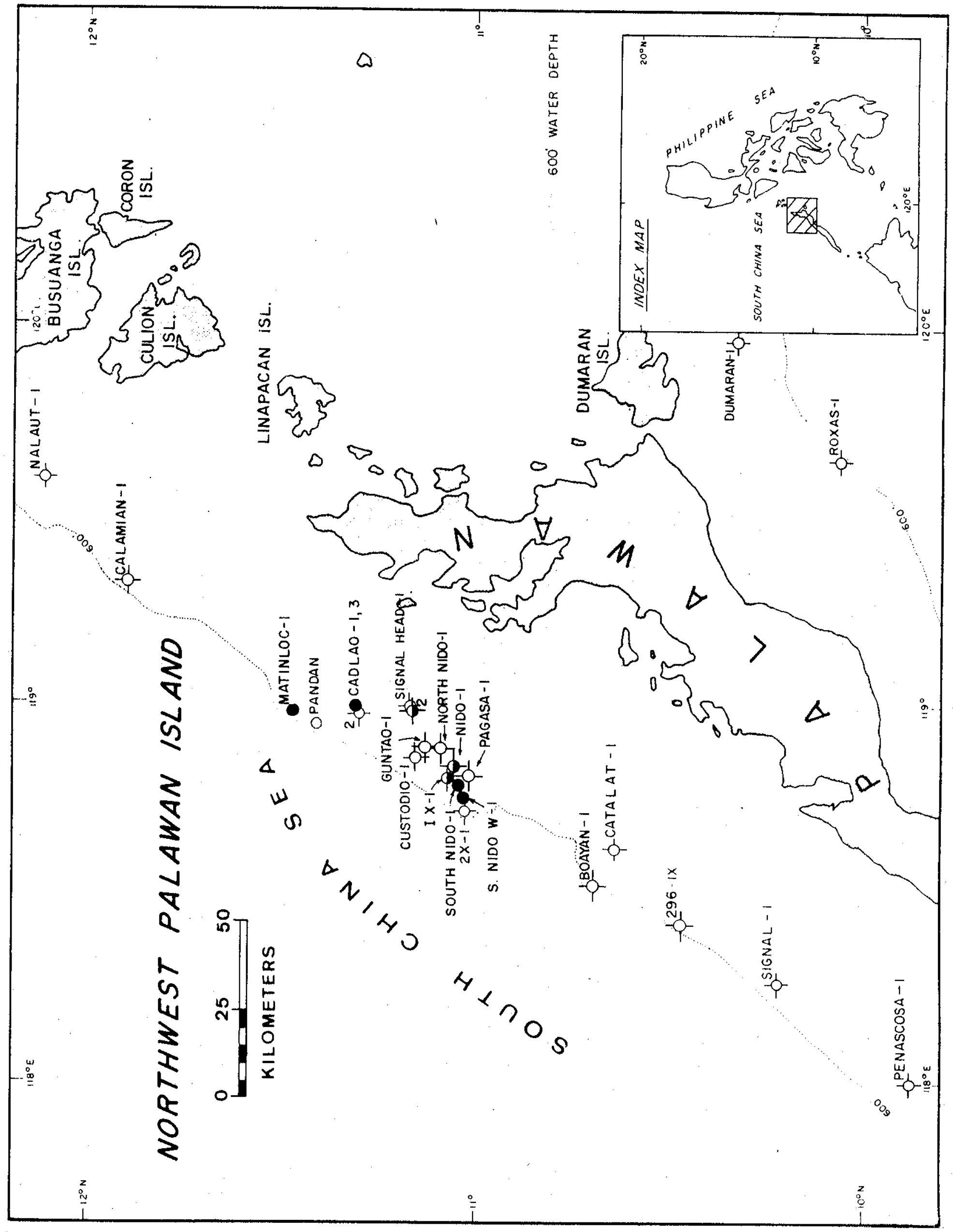

Figure 3. Location of wells.

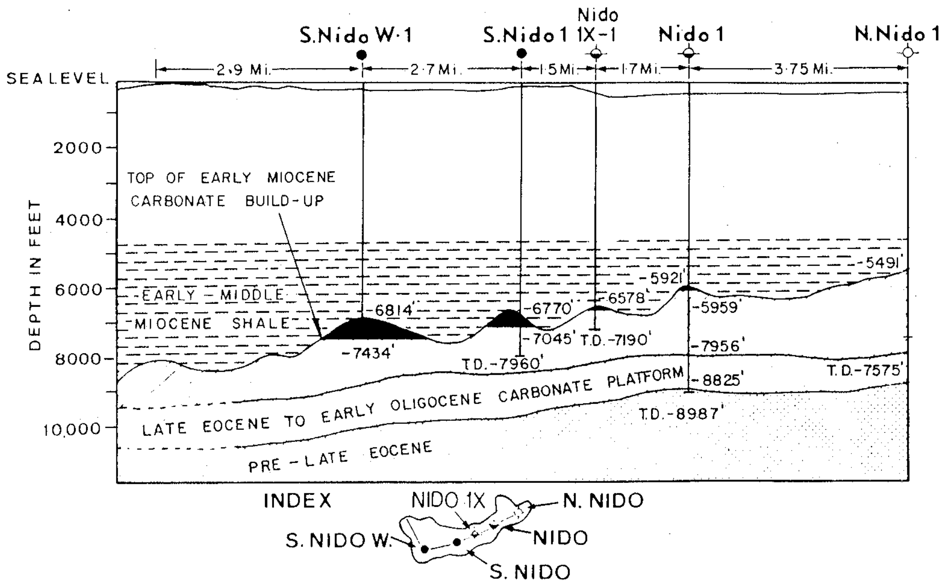

Figure 4. Cross section, Nido Reef Complex Oil Field.

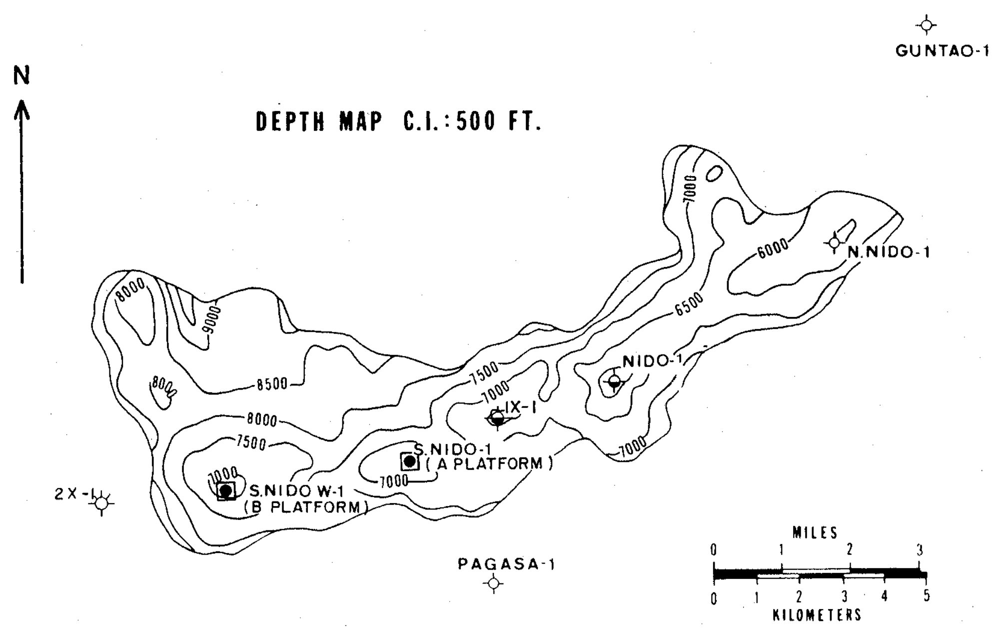

Figure 5. Top of carbonate build-up, Nido Reef Complex Oil Field.

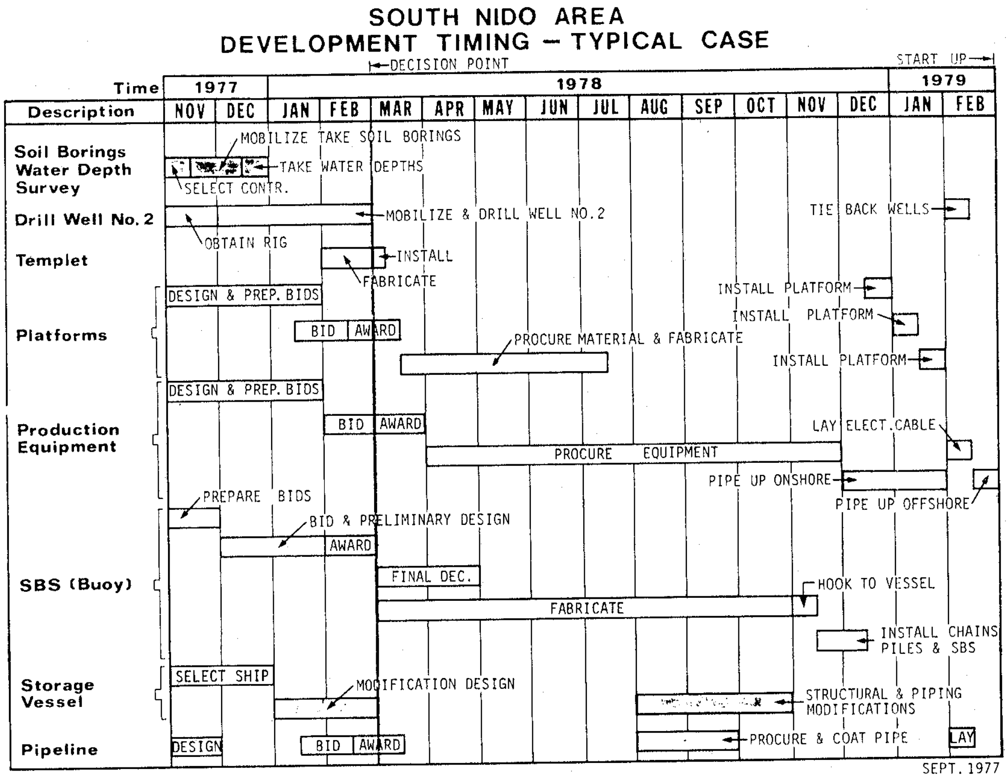

Figure 6. Development timing.

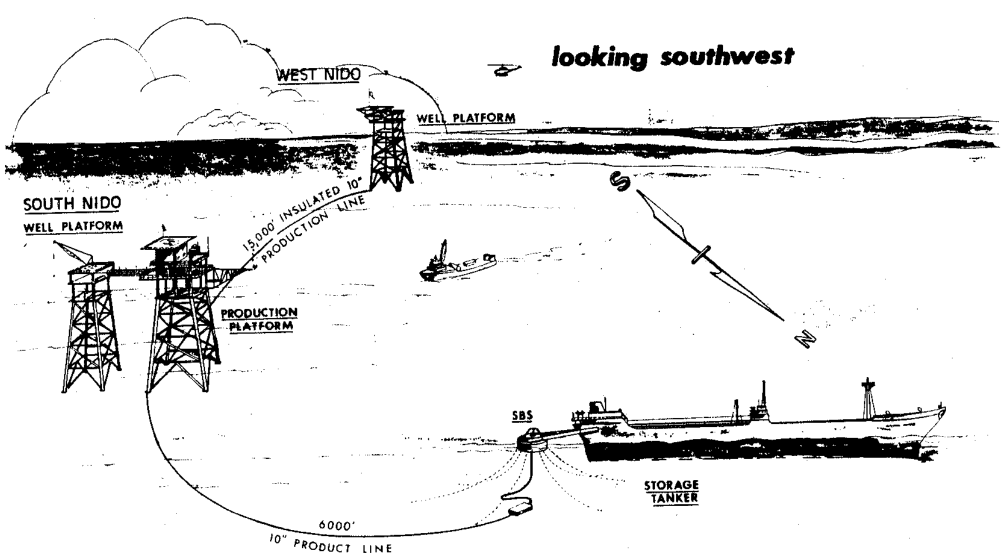

Figure 7. Production facility, South Nido Development.

justified, the most obvious result being the time saved in bringing onto production the Nido Reef Complex Oil Field. Figure 6 shows the various activities, and the timing of each, that were undertaken as part of the production feasibility study. Intensive engineering design, environmental, and reservoir studies made before the discovery in South Nido West-1, allowed the first well to be brought onto production only 11 months after the final decision to develop the discoveries.

One of the significant factors contributing to the success in the design, procurement, construction and supervision of development, was the use of key technical personnel on a 'task-force' basis from both within the company and through the use of selected consultants, regardless of location. It is obvious when one reviews the timing of many of the activities that it was possible for the 'task-force' to begin preparation of, and in some cases to award, bids for construction, even prior to the completion of testing of the South Nido West-1 well.

Development and production

The production system selected for the Nido Complex (Fig. 7) consists of two, four-pile well jackets located in 140 ft of water for the 'A' reefal feature (South Nido) and in 255 ft of water for the 'B' reefal feature (South Nido West).

Two development wells have been completed on the 'AW' Platform. The A-1 well was completed by tying back to the original South Nido-1 well. The A-2 well was deviated into the 'A' reefal feature. On the 'BW' Platform, B-1 well tied back to the original South Nido West-1 well. B-2 and B-3-A are two deviated development wells.

Nido crude oil is typically black-brown in color with an average API gravity of 27°. The reservoirs are undersaturated, with solution ratios of only 7 to 10 SCF/stock-tank barrel. The crude oil contains 1200–2000 p.p.m. H_2S, which renders it both corrosive and extremely dangerous to handle within the close confines of platforms and storage vessels.

Both well jacks and the storage vessel are equipped with helidecks. A two-deck, 70 × 70 ft, four-pile production platform, capable of processing 70 000 bbls of fluid daily, is located 150 ft from the AW well jacket and is connected by a bridge. Process equipment (Fig. 8) consists essentially of two 12 × 30 ft combination free-water knockout Chemelectrics. All gas is separated and burned through a cantilevered flare because of its toxic H_2S content. The production platform contains additional equipment to remove entrained oil

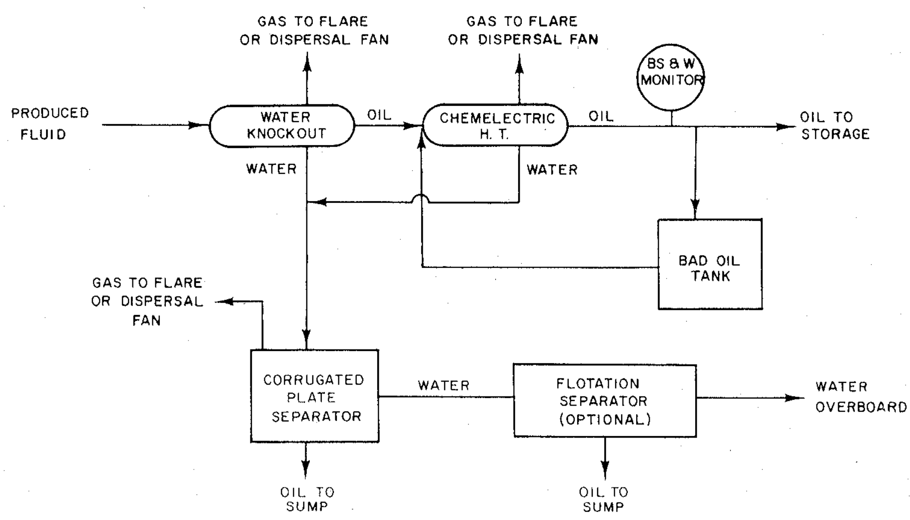

Figure 8. Oil-processing equipment and preliminary flow diagram.

from produced water, prior to safely disposing of the produced water in the ocean.

All processed crude is stored aboard a storage vessel, attached to an SBS which allows weather vaning around the buoy. The storage vessel, *J Ed Warren,* is a converted 103 550 DWT tanker. It has been equipped to offload crude, either with side-by-side berthing, or by an innovative employee-designed bow-to-bow tandem-loading method. The vessel has been equipped with a flue gas generator and remote gauging to minimize the danger of exposure to H_2S. Living accommodation has been provided for up to 90 people, and suitable office space is also provided. All oil is metered prior to offloading the *J Ed Warren.*

Location of, and access to, the storage vessel have been a critical part of the development program, by reason of the extreme irregularity of the sea bottom, mentioned earlier in this paper.

CONCLUSION

Today the Nido Reef Complex Oil Field is producing approximately 40 000 bbls/d. The discovery of this field is the result of innovative exploration thinking and action, performed with full management support and recognition of the inherent risks involved.

The rapid development and earlier production than normally expected result from a no smaller degree of management support. The approval of an early development feasibility study and a 'task-force' approach to development, performed and led by experienced engineers and logistical and management personnel, along with recognition and acceptance of other development and production risks, represent a major technological and management accomplishment.

ACKNOWLEDGEMENTS

Special acknowledgement is made to the managements of Cities Service Company, Philippines-Cities Service, Inc., Husky Oil International, Inc., Oriental Petroleum & Minerals Corp., Philippine Overseas Drilling & Oil Development Corp., Basic Petroleum & Minerals, Inc., and Landoil Resources Corp., as well as the Philippines Government and, especially, the Bureau of Energy Development (BED), whose cooperation and trust before, during and after the drilling of the Nido wells have made it all possible.

Most of all, I would like to acknowledge the contribution of Mr P W J Wood, former International Manager and now Executive Vice President of the Energy Resources Group for Cities Service Company, without whose understanding and support there would have been neither a Northwest Palawan Service Contract nor a Nido discovery.

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GEOCHEMICAL TECHNIQUES FOR PETROLEUM EXPLORATION

LEO HORVITZ

ABSTRACT

Geochemical prospecting methods are all based on the premise that hydrocarbons move from oil and gas reservoirs to the near-surface where recognizable distribution patterns are produced. Some of the techniques employed include: (1) direct measurement in the near-surface of the light, saturated hydrocarbons, methane through pentane; (2) detection of fluorescent substances believed to be indicative of the presence of aromatic hydrocarbons; and (3) the measurement of inorganic compounds that may be concentrated in the near-surface as a result of hydrocarbon migration.

Evidence supporting movement of hydrocarbons from subsurface oil and gas deposits to the near-surface is furnished by the facts that (1) near-surface hydrocarbon patterns overlie petroleum accumulations; (2) the composition of hydrocarbons in near-surface soils and sediments is generally similar to the composition of reservoir gases; and (3) the C^{13}/C^{12} isotope ratios of methane in near-surface anomalies are in the same range as those of methane in reservoir gases.

INTRODUCTION

Geochemical prospecting was first introduced by Laubmeyer (1933) who reported that near-surface soil air over gas fields is richer in methane than is the soil air in barren areas. Russian workers, headed by Sokolov (1935), improved upon Laubmeyer's technique and found heavier hydrocarbons as well as methane in soil air. Research in this field began in the United States in 1936 but, from the very beginning, the soil itself (Rosaire 1938, Horvitz 1939) was investigated as the sampling medium, instead of soil air. It was anticipated that analysis for the adsorbed hydrocarbons would yield higher values than were reported for soil air. Another significant reason for selecting soil as the sampling medium is that it can be employed in practically every area to be explored. Soil-air sampling, on the other hand, is limited to arid areas. Furthermore, soil or sediment sampling makes possible the application of geochemical methods to water-covered areas.

Laubmeyer and the Russian investigators reported near-surface hydrocarbon patterns in which the higher concentrations were found in the soil air directly over productive areas. We found, from the outset, that the principal hydrocarbon distribution pattern is one in which the near-surface soils that overlie productive areas contain relatively low concentrations, and the near-surface soils over the edges of the producing areas contain anomalous amounts of hydrocarbons. This became known as the 'halo' pattern. Although it is the most common type found over important oil and gas fields, other distribution patterns are also observed. Anomalies that consist of areas of relatively high hydrocarbon concentration, surrounded by small amounts of near-surface soil hydrocarbons, are often encountered over productive areas, but they usually occur over deposits that are less important than those that produce the halo pattern. A third pattern is the crescent type which is found over accumulations that are trapped against the upthrown side of a fault. In this type, high hydrocarbon concentrations are found in the near-surface soils over oil-water or gas-water contacts, with relatively small amounts of hydrocarbons in the near-surface soil over the intercept of the subsurface accumulation and the fault. Leakage across the fault plane appears to be restricted. Hydrocarbon leakage over pinchout edges of accumulations also tends to be limited. Apparently, the principal leakage occurs at or near oil-water or gas-water contacts.

To show how geochemical methods are applied to petroleum exploration, three examples of hydrocarbon surveys will be presented. Two of the surveys were conducted on land and one was made offshore. Other techniques that will be discussed, and that can enhance light-hydrocarbon surveys, involve carbon-isotope and fluorescence measurements.

ONSHORE HYDROCARBON SURVEYS

Hastings Survey

Figure 1 shows the data of an early hydrocarbon survey. It is of the Hastings oil field in Brazoria County, Texas and is associated with a structure of high relief. The survey was conducted in 1946, 12 years after the field was discovered. Ethane and heavier hydrocarbon values, expressed in parts per billion (p.p.b.)* by weight (grams of hydrocarbon per billion grams of dry soil), of 125 samples collected from depths of 8 to 12 ft are shown. Briefly, the analytical technique used in obtaining the data involved removing the hydrocarbons by heating about 100 grams of soil in a partial vacuum with phosphoric acid at 100°C (Horvitz 1969). The extracted gases were then separated into two fractions, one containing methane and the other, ethane and heavier hydrocarbons. The separation was effected by using high vacuum, low temperature fractionation methods previously described (Horvitz 1954). Methane was

* For this paper, billion (b. in p.p.b.) is defined as 10^9.

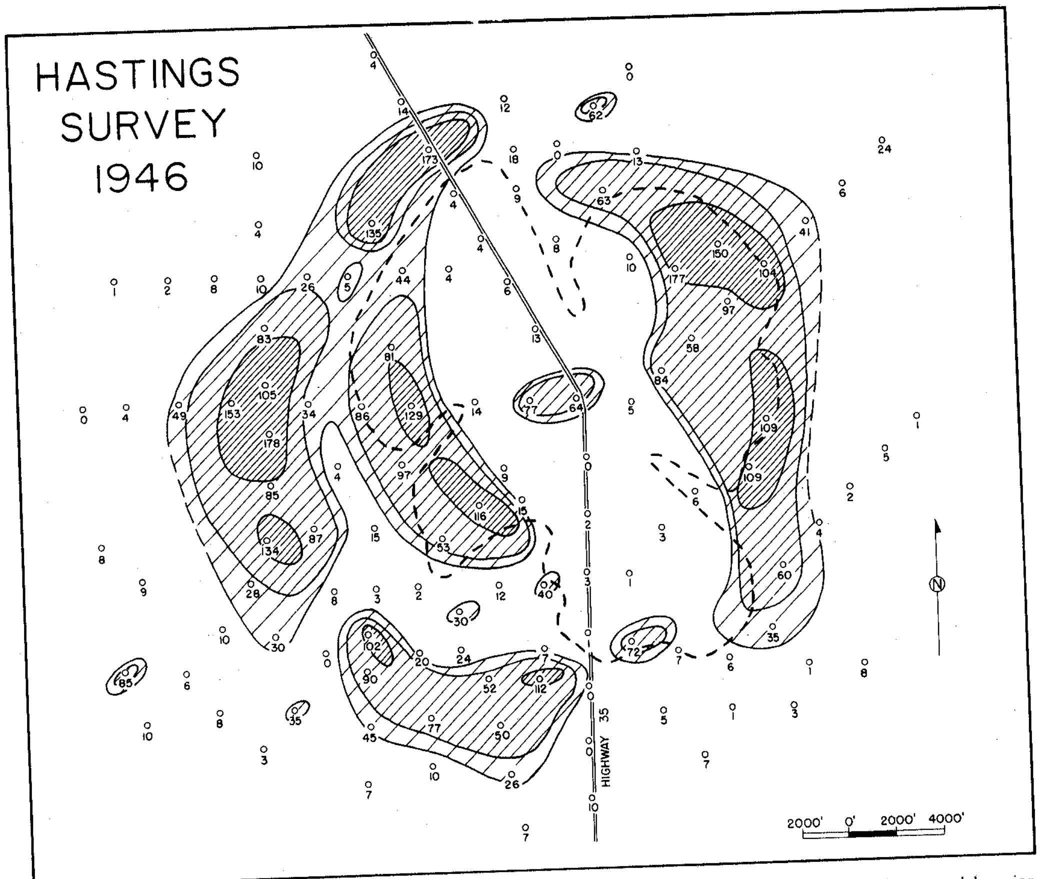

Figure 1. Hydrocarbon survey of Hastings oil field, Brazoria County, Texas, conducted in 1946. Ethane and heavier hydrocarbon values are expressed in parts per billion (10^9) by weight on dry-sample basis. Producing area is outlined by dashed line. Scale in feet. (From Horvitz 1969.) Reprinted by permission of the Institute for the Study of Earth and Man, Southern Methodist University.

separated from the ethane and heavier hydrocarbons because it may be present in near-surface soils as a result of bacterial action on organic matter. On the other hand, ethane and heavier hydrocarbons have no known source other than petroleum. In 1946, the hydrogen flame chromatograph was not yet available and, therefore, the more complicated vacuum analytical technique had to be used.

The ethane and heavier hydrocarbon values range from 0 to 178 p.p.b. Three ranges of values are contained in the contoured areas: 25–49; 50–99; and 100 p.p.b. and above. The lightly shaded areas contain values ranging from 25 to 49 p.p.b.; the areas of intermediate shading include values of 50–99 p.p.b.; and the most heavily shaded areas contain values of 100 p.p.b. and higher. The background contains values below 25 p.p.b. and averages less than 10 p.p.b. The productive area, which covers about 5000 acres, is indicated on the map (Fig. 1) by the dashed outline. The Hastings oil field produces from the upper part of the Frio formation, topped at about 6000 ft, and the estimated reserve of this field is over 500 000 000 barrels of oil. The anomalous soil hydrocarbons aligned themselves in the form of a halo, the typical near-surface pattern associated with important petroleum accumulations.

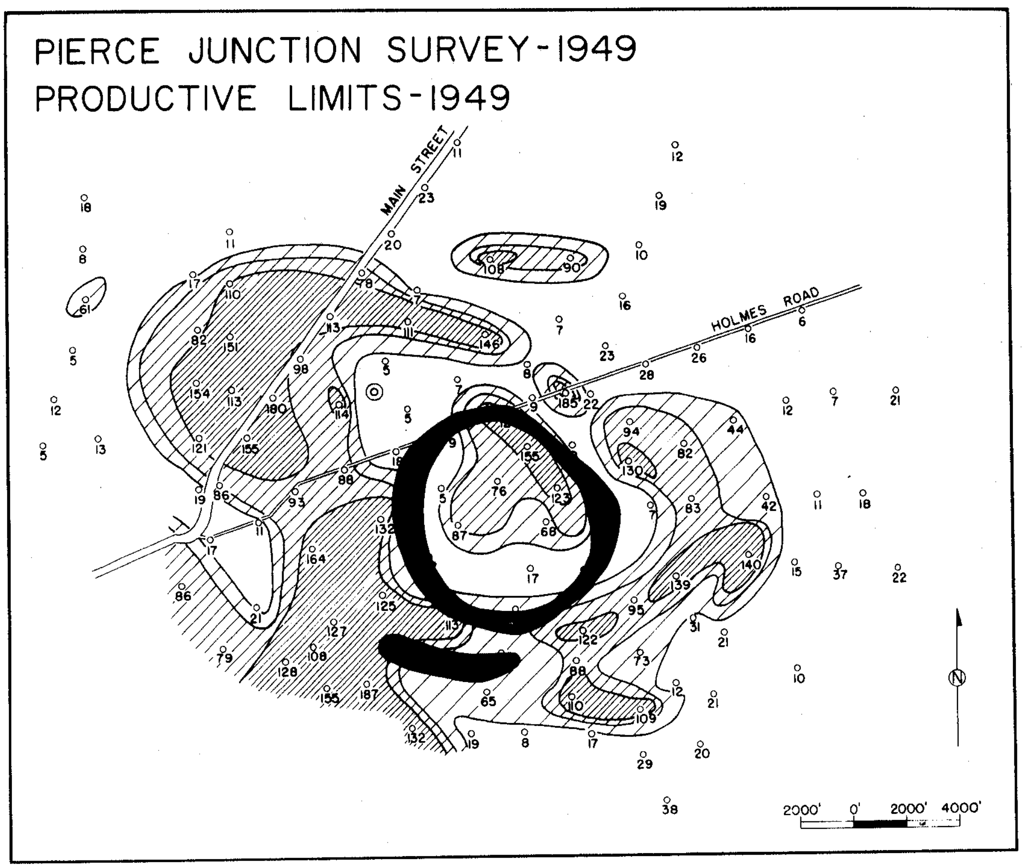

Figure 2. Hydrocarbon survey conducted in 1949 over Pierce Junction dome, Harris County, Texas, together with development (solid black) at time of survey. Ethane and heavier hydrocarbon values are expressed in parts per billion (10^9) by weight on dry-sample basis. Recommended location for flank test is indicated by double circle in northwest part of anomaly. (From Horvitz 1969.) Reprinted by permission of the Institute for the Study of Earth and Man, Southern Methodist University.

Pierce Junction Survey

Figure 2 shows the ethane and heavier hydrocarbon data obtained in 1949 over the Pierce Junction oil field, located in Harris County, Texas. The samples that were used in conducting the analyses were collected at depths of 10 to 12 ft below the surface and were composed of clays and sandy clays. The method of analysis was the same as the one used in obtaining the data of the Hastings survey. The contour lines enclose values ranging from 40 to 74 p.p.b. by weight; values of 75–99; and values of 100 p.p.b. and higher. As in the Hastings survey, a system of shading was used to display the variations in hydrocarbon concentrations.

Figure 2 shows also the producing limits of the Pierce Junction field at the time the survey was made. The field had been producing since 1921 from several zones in the Miocene formation, within the section from 3000 to 4300 ft, and from the Frio beginning at 4800 ft. This production came from the narrow, circular area (black) which flanks the shallow salt dome. To the southwest of the main part of the field, a few wells were producing from the Vicksburg formation (small black area) at a depth of about 6800 ft.

The purpose of the survey was to determine if additional reserves could be expected on the

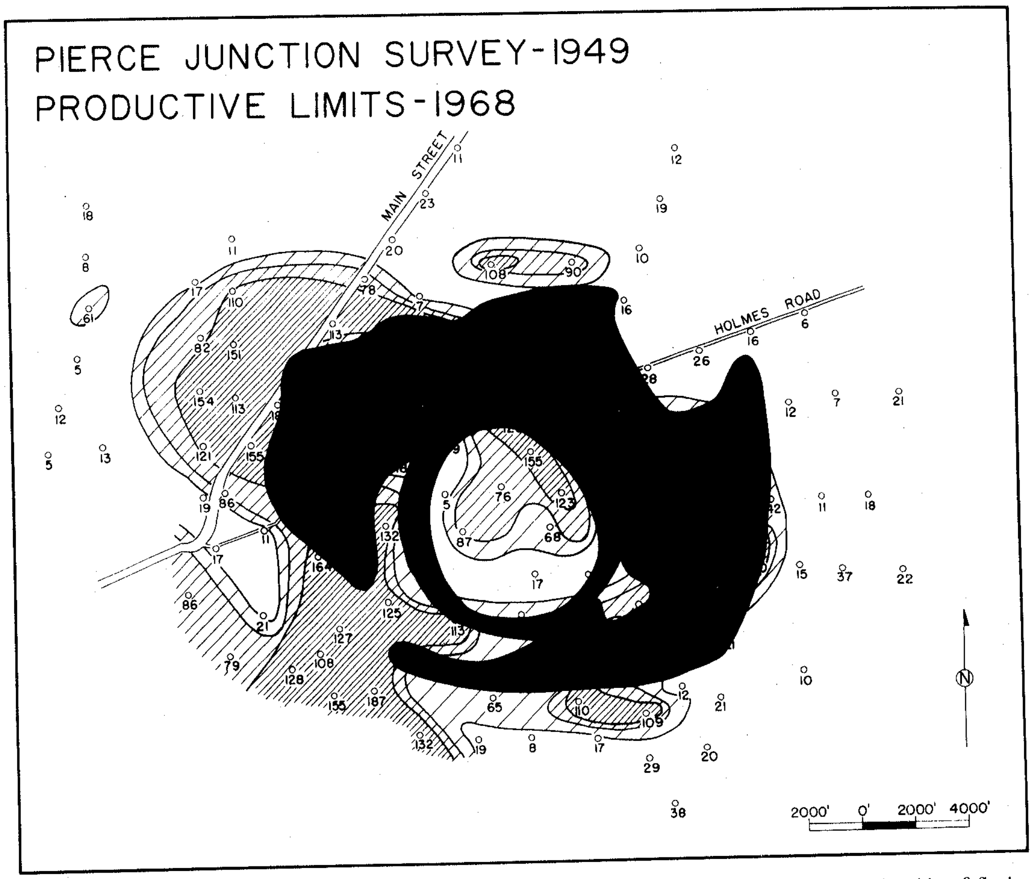

Figure 3. Pierce Junction survey (same area as in Fig. 2) with development as of 1968. Close relationship of flank Vicksburg production to hydrocarbon anomaly confirms prediction made on basis of 1949 data. (From Horvitz 1969.) — Reprinted by permission of the Institute for the Study of Earth and Man, Southern Methodist University.

undrilled flanks of this piercement-type dome. The strong hydrocarbon anomaly of high contrast that developed, and that extends well beyond the oil producing limits, did, indeed, suggest a sizeable extension to the field. This possibility was supported by the fact that only weak anomalies, if any, occur over old fields (Horvitz 1969). Since the Pierce Junction field had already produced about 40 000 000 barrels of oil, the hydrocarbon anomaly was thought to be related to a virgin deposit. The overall anomalous area was recommended for leasing and a location in the northwest part of the anomaly was suggested for a test well.

The oil company that sponsored the survey decided against leasing and drilling after finding that the area was divided into numerous small tracts, some less than one acre, and each tract required a well. However, in 1954, five years after the survey was completed, a well was drilled on the east flank and Vicksburg production was found at about 7100 ft. Because of rapid development, the new Vicksburg limits were established within a few years. Figure 3 includes the development that has taken place after completion of the hydrocarbon survey. The new production represents more than 20 000 000 barrels of oil and substantiates the fact that important accumulations do exist on the sides of structures where drilling is often avoided.

OFFSHORE GEOCHEMICAL EXPLORATION

Since January 1967 geochemical exploration has been applied to many offshore areas around the world including the Gulf of Mexico, the East Coast of the United States, and the North Sea. In fact, a hydrocarbon survey was conducted offshore Fiji during the early part of 1979. Most of our offshore work was done in the Gulf of Mexico where many hydrocarbon anomalies were encountered prior to drilling. The resulting confirmation of more than 50 per cent of these anomalies provided evidence of the usefulness of geochemical methods in distinguishing potentially productive structures from those that are dry. The value of geochemical data in the lease-bidding process also became apparent.

Offshore samples are taken from the upper 6 ft of sediment with a piston-type coring device equipped with a 6 ft core barrel, 3 in. in diameter, into which a plastic liner is inserted. The bottom part of the sample is removed from the liner, on board the ship, and packaged in plastic bags in such a manner as to allow a minimum of air to come in contact with the sample. Stored in this manner, no significant losses of adsorbed hydrocarbons have been noted, even after many months. The analytical methods used are the same as those employed for onshore samples.

Experiments, conducted with samples taken from the same locations at several intervals down to about 12 ft below the mud line, indicated that the depth of sampling is not as critical in water-covered areas as it is onshore. Offshore samples show fairly uniform concentrations of hydrocarbons from the water-sediment interface to 12 ft. Often, the uppermost foot of sediment contains somewhat higher concentrations than does the deeper part of the hole. On land, the first few feet of soil, especially along the Gulf Coast, contain very low concentrations of hydrocarbons and significant values do not usually appear above 6 ft; therefore, samples are taken well below this depth, generally at 10–12 ft. Offshore, the water serves as a blanket which prevents the rapid escape of hydrocarbons from the sediment. Nevertheless, offshore samples are taken at 6 ft, whenever attainable, instead of the immediate surface because surface sediments may be moved around by water currents or storms. At 6 ft the sediment is more likely to have been in place for a geologically long period. Owing to the water blanket, background values in offshore petroliferous areas, especially when the bottom is mainly clay, are usually much higher than those in onshore areas. Background ethane and heavier hydrocarbon values in such areas are, usually, at least 35–50 p.p.b. by weight, whereas onshore they are often below 10. On the other hand, in non-petroliferous offshore areas, values for the ethane through pentane fraction are also very low; frequently they are well below 10 p.p.b. In clay areas, sampling poses no problem, but, in areas of very hard bottom, samples are often missed or only small quantities of rock or reef material are retrieved. In hard, sand bottoms the penetration depth is also less than 6 ft. Furthermore, in these areas the background values are much lower than in areas of clay bottom (Horvitz 1980).

An offshore Louisiana survey

During the fall of 1973 and winter of 1974, a hydrocarbon survey was conducted offshore Louisiana to evaluate certain areas that were to be offered in the March 1974 Offshore Federal Lease Sale. One of the areas that was included is in the Mobile South No. 2 Area, about 100 miles south-east of New Orleans. Figure 4 shows the locations of 256 stations in this area from which sediment samples, composed of clay, were collected at 6 ft. The water depth ranged from 318 ft in the northwest corner of the survey to 1860 ft in the southeast corner. The stations were spaced at half-mile intervals along north-south profiles 1 mile apart. The ethane and heavier hydrocarbon values, shown on the map at their respective station locations, range from 39 to 537 p.p.b. by weight and average 68. Values ranging from 75 to 89 p.p.b. are in the lightly shaded areas and values of 90 and higher are in the areas of heavier shading.

The technique used for extracting the hydrocarbons from the sediment is the same as the one used in conducting the early land surveys. However, the more versatile

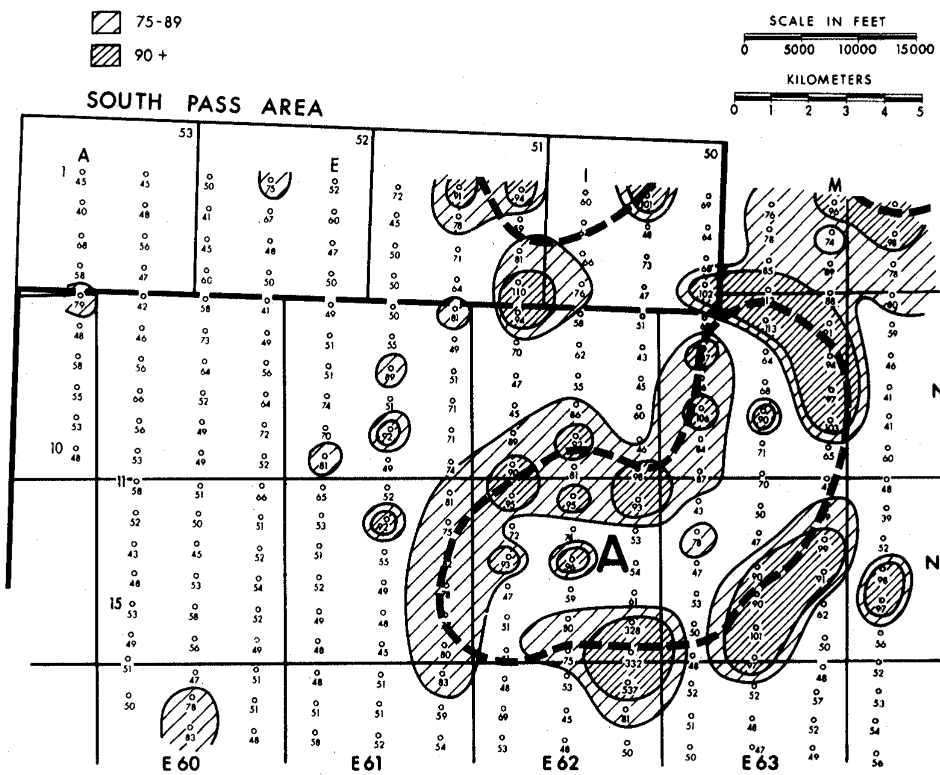

Figure 4. Halo-type concentration pattern produced by ethane and heavier hydrocarbons from sediment samples taken at 6 ft in Mobile South No. 2 area, offshore Louisiana, Gulf of Mexico. Hydrocarbon values in parts per billion (10^9) by weight on dry-sample basis. Water depth ranges from 318 ft at northwest corner to 1860 ft at southeast corner of survey. Interpretation of the data (prior to March 1974 Gulf of Mexico lease sale) resulted in delineation of the area considered most prospective (outlined by heavy dashed line and graded A). (From Horvitz 1980.) Reprinted by permission of A.A.P.G.

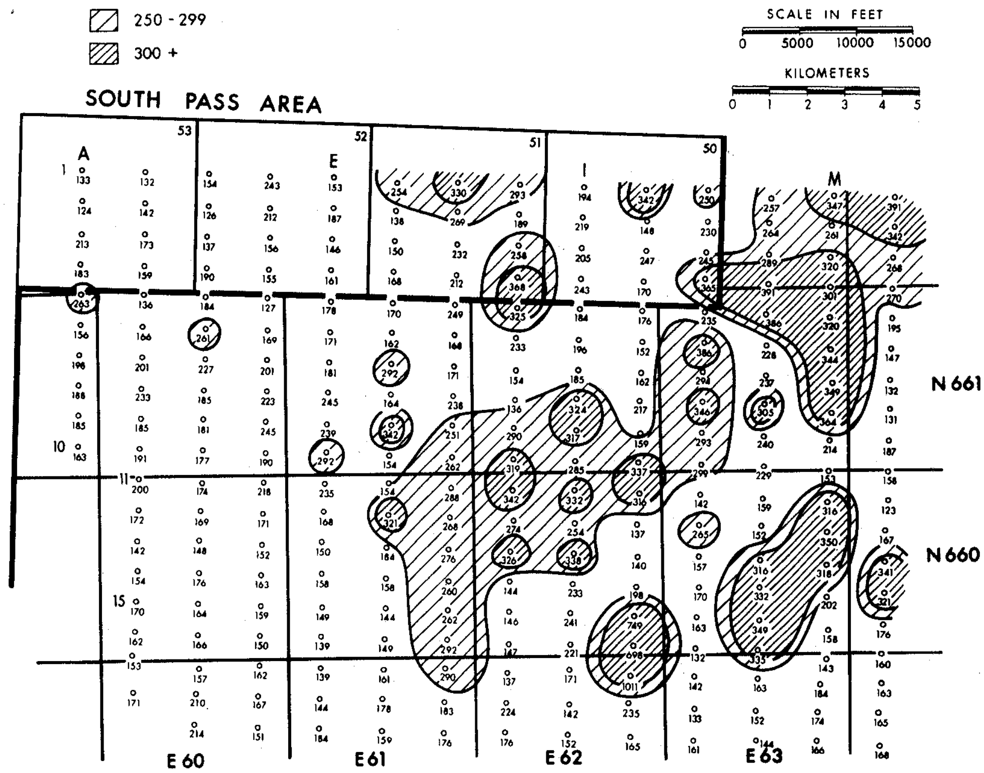

Figure 5. Methane data in Mobile South No. 2 area produced a halo-type anomaly in eastern part of survey which resembles closely the one produced by the ethane and heavier hydrocarbons. (From Horvitz 1980.) Reprinted by permission of A.A.P.G.

hydrogen-flame chromatograph was used to analyze the extracted gases instead of the vacuum analytical-combustion technique.

The interpretation made on the basis of the ethane and heavier hydrocarbon data is also shown in Fig. 4. A well defined halo-type anomaly, graded A, is indicated by the dashed outline. The central part of the anomaly lies under about 1000 ft of water.

The methane data produced a pattern (Fig. 5) very similar to that of the ethane and heavier hydrocarbons. Values of 250–299 p.p.b. by weight are in the lightly shaded areas and values of 300 and above are in the more heavily shaded zones.

Figure 6 is a map of the pentane values, which range from 0 to 101 p.p.b. and average 9.9. Values of 12–14 p.p.b. are in the lightly shaded areas and values of 15 and above are in the more heavily shaded areas. The pattern resembles that produced by the ethane and heavier hydrocarbons. Pentane data are useful in determining if potential subsurface reservoirs contain liquid hydrocarbons. If pentane appears in the near-surface sediment and produces an anomaly, liquid hydrocarbons may be expected in the reservoir.

In July 1975 Shell announced the results of an exploratory well drilled about 1 mile west and a few hundred feet south of the northeast corner of block N660-E62 and indicated in Fig. 7 by a solid black circle. Several gas and oil zones were encountered which indicated that a potentially important discovery had been made. Subsequently, 11 additional exploratory wells were drilled to depths ranging from 9000 to 13 930 ft within the area of the dotted outline (Fig. 7). The field, believed to represent a major reserve, is referred to as Cognac and will be on stream soon.

The outline of the area that contains the exploratory wells bears a very close relationship to the hydrocarbon anomaly. If the latter were shifted about 1 mile to the northwest, an almost perfect fit with the potentially productive limits would result. Such displacements of hydrocarbon anomalies from producing areas are frequently noted. Hydrodynamic effects may be involved (Hitchon 1974) in the process of moving hydrocarbons from depth to the near-surface, and they may cause shifting of near-surface geochemical anomalies. If these effects can be corrected for, the accuracy with which potentially productive areas could be outlined by near-surface hydrocarbon data would be greatly improved. Even without correction, 75 per cent of the potentially productive area is included within the Cognac hydrocarbon anomaly. It is also of interest that the amount of methane, by volume, in the total light hydrocarbon fraction (C_1-C_5) in the near-surface soils of the anomaly, is 89 per cent, and within the range normally present in reservoir gas.

Several miles to the west of the Cognac discovery well, five wells were drilled but no official results have been released. Unconfirmed reports indicate that gas accumulations were encountered there. However, according to the interpretation presented, no anomaly developed in this part of the survey; therefore, any accumulation that exists there is expected to be unimportant, at least when compared to Cognac.

CARBON-ISOTOPE DATA

Some of the most important evidence that substantiates vertical migration of hydrocarbons from subsurface reservoirs has

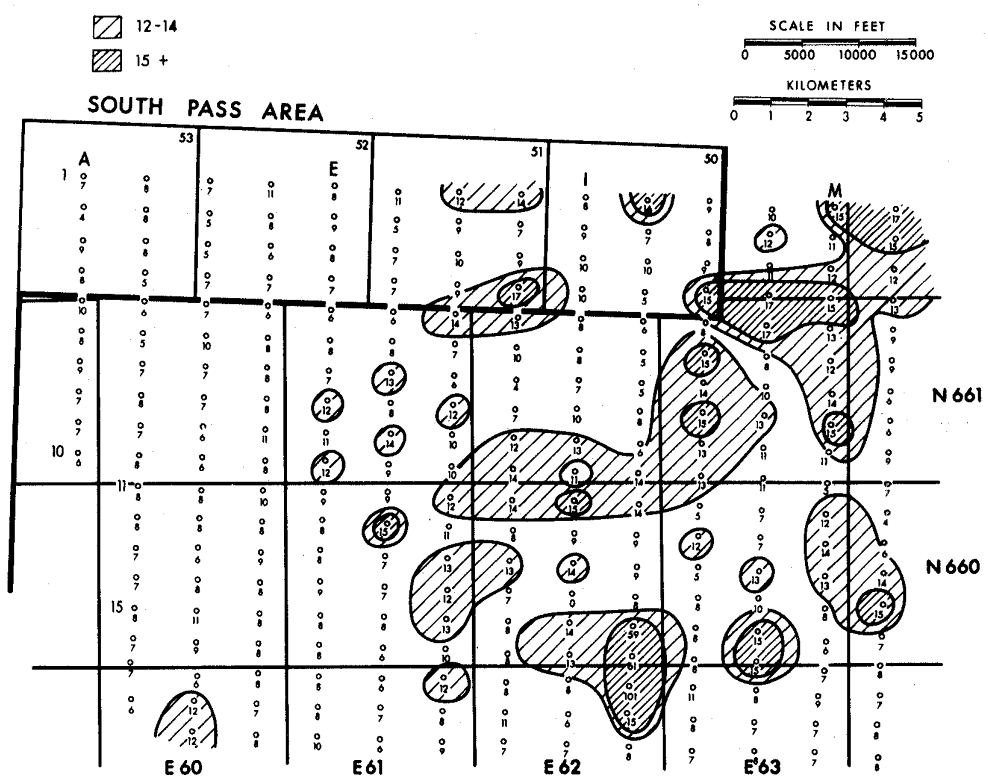

Figure 6. Pentane data produced a halo-type anomaly in Mobile South No. 2 area similar to that of ethane and heavier hydrocarbons (Fig. 4). (From Horvitz 1980.) Reprinted by permission of A.A.P.G.

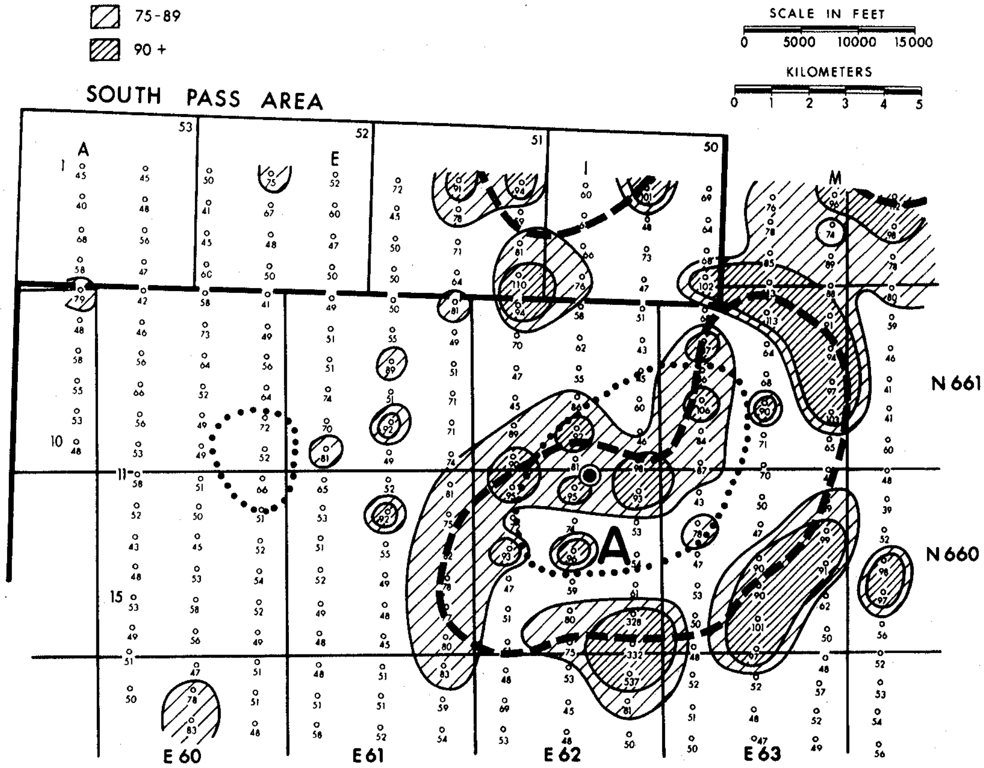

Figure 7. Ethane and heavier hydrocarbon map (Fig. 4) to which the discovery well (large black dot in circle) of the new Cognac field has been added. Within the dotted outline 11 additional exploratory wells have been drilled. The small area (dotted outline) to the west of the A anomaly contains five exploratory wells, some of which may have penetrated a gas deposit. No significant hydrocarbon anomaly is apparent in this part of the survey. (From Horvitz 1980.) Reprinted by permission of A.A.P.G.

come from isotope studies. It is possible to determine if different hydrocarbon samples come from the same source or from different sources if their ratios of C^{13} to C^{12} are known. The technique involves removal from the hydrocarbon of any carbon dioxide initially present, and then converting the resulting purified hydrocarbon to carbon dioxide. Carbon-isotope ratios of this carbon dioxide are determined in a sensitive mass spectrometer, and the results are expressed as $\delta^{13}C$ values, or deviations, in parts per thousand (per mil) of the C^{13}/C^{12} ratio of a sample from that of a standard. A popular standard is belemnite, a carbonate from the Peedee formation of South Carolina, and known as the PDB standard.

The writer's first experience with carbon isotope ratios was in 1973. Measurements were made on methane which was extracted from 12-foot samples, located within a hydrocarbon anomaly over the Francitas oil field in Texas, and $\delta^{13}C$ values of -40.8 and -44.0 per mil, relative to the PDB scale, were obtained. Samples of methane, extracted from the reservoir gas of the field, showed $\delta^{13}C$ values of -41.0 to -43.8 per mil. These values are practically the same as those of the near-surface methane, indicating a thermogenic, and not a biogenic, source for the near-surface methane found over the Francitas field. The $\delta^{13}C$ values of methane from biologic sources range from -50 to -80 per mil. The writer is indebted to Amoco Production Company for the isotope determinations of the Francitas samples.

Carbon-isotope ratios of methane, extracted from three sediment samples of the Cognac survey, were also determined and $\delta^{13}C$ values of -37.3 to -39.2 per mil were obtained. Just as for the case of Francitas, these $\delta^{13}C$ values fall in the same range as those of methane from petroleum gases. The writer is indebted to Tenneco Research for the isotope values of the Cognac samples.

NEAR-SURFACE HYDROCARBON STUDIES BY GULF

Geochemical exploration techniques, especially those involving the detection of light hydrocarbons in near-surface soils, recently received encouraging support from V T Jones III and R J Drozd in an unpublished paper read before the 1979 Annual Meeting of the American Association of Petroleum Geologists, Houston, Texas. The paper discussed near-surface soil-hydrocarbon data obtained by Gulf Research & Development Company which show close relationships between near-surface hydrocarbons and subsurface accumulations. Furthermore, the authors explained how percentage compositions of near-surface soil gases, together with ratios of methane to ethane and propane to methane, can be used to predict whether a reservoir contains oil or gas.

FLUORESCENCE MEASUREMENTS

An interesting geochemical prospecting technique introduced in recent years involves the determination, in soils and sediments, of substances that fluoresce when excited by ultraviolet light. The same types of fluorescence spectra are produced by soils and sediments, in the near-surface over oil fields, as are produced by crude oils. When crude oils are excited with ultraviolet light having a wavelength of 265 n, emissions are produced at 320 n or 365 n, or at both. Using a sensitive fluorescence spectrophotometer, the emission intensities can be measured, and the resulting spectra can be recorded. D B Purvis International Geochemics was apparently the first to employ this technique in oil exploration.

The hydrocarbons of natural gas do not fluoresce, nor do the heavy, straight-chain saturated hydrocarbons. Experiments conducted on a series of 10 fractions that were distilled from a crude oil sample indicated the fluorescent material to be in the two heaviest fractions, those that showed an amber or darker color. The exact compositions of the substances that fluoresce are not yet known, but they probably belong to the high-molecular-weight aromatic-hydrocarbon family.

Unfortunately, the refining process does not eliminate the materials that fluoresce, and therefore it is impossible to distinguish crude oil from refined oils. Because of this fact, fluorescence data should not be used alone in exploration. When used in conjunction with a light-hydrocarbon survey, however, fluorescence data become very meaningful. A strong fluorescence anomaly, which conforms to that of the light hydrocarbons, suggests a subsurface oil reservoir; a strong light-hydrocarbon anomaly, in the absence of a fluorescence anomaly, suggests a gas or gas-distillate accumulation.

CONCLUSIONS

Geochemical exploration has made much progress in recent years, especially in substantiating its basic premise that near-surface hydrocarbons originate in subsurface gas and oil deposits. This is important, not only to hydrocarbon geochemical prospecting, but also to other geochemical techniques including those that measure inorganic substances in the near-surface such as carbonates, sulfides, ferrous iron, etc. They, too, rely on hydrocarbon migration as the cause of near-surface concentration changes. Remote-sensing techniques that claim to measure hydrocarbons at the very surface of the earth, also gain in credibility through proof of hydrocarbon migration from subsurface oil and gas reservoirs.

Geochemical techniques can play important roles in petroleum exploration, but best results are realized when they are used in concert with other valid methods, especially geological and geophysical. However, geochemistry may have to stand alone when a well defined near-surface hydrocarbon anomaly is encountered which is *not* supported by structurally oriented methods. In such a case the petroleum explorationist must not overlook the possibility that the hydrocarbon anomaly is reflecting an important stratigraphic trap.

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UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 273-276

REVIEW OF RECENT ACTIVITIES BY CCOP IN THE FIELD OF HYDROCARBON RESOURCES*

ABSTRACT

The past year has been devoted principally to planning future directions for the CCOP programme in hydrocarbon resources. Traditionally, it has been CCOP's goal to stimulate interest in the offshore basins of the community. This aim has been largely accomplished and the interest of the petroleum industry in the offshore areas of east and southeast Asia is firmly established. Accordingly, it is recommended that CCOP assemble and inventory the results of the past 13 years of exploration by means of an area-wide programme of geological analysis. Emphasis is to be placed on hydrocarbon resources and identifying new areas of exploration interest. Ancillary programmes consist of disseminating information in the disciplines of basin assessment, exploration economics and seismic stratigraphy. Co-ordination has been maintained, and assistance offered where appropriate, with other regional and international agencies.

CCOP has maintained a long continuing interest in the promotion and development of the petroleum industry and has played a significant part in stimulating interest in the prospects of the East Asia offshore regions. It is CCOP's wish to continue to play an active role in the future progress of the industry.

For the present purpose, the petroleum industry may be considered as an interactive association of national oil entities representing the national interest, and privately owned and international oil companies representing commercial interests. Apart from this direct and primary relationship, there are a number of organizations, one of which is CCOP, which represent secondary or peripheral or specialized interests. Other members of this grouping include the ASEAN Council on Petroleum (ASCOPE), the South East Asia Petroleum Exploration Society (SEAPEX), the Indonesian Petroleum Association (IPA), the service companies, advisory organizations and consultants.

In contemplating the role which CCOP might play, it is clear that the nature of the industry has changed over the years. It is also clear that the private oil companies have entered into the offshore realm to an extent and in such a manner that CCOP has to reconsider its traditional role of stimulating geophysical surveys of the offshore shelves and (somewhat less so) the continental slopes and rises.

The unique qualities which CCOP brings to the industry include: (1) the ability to transcend national boundaries; (2) the ability to obtain international support in financing and technical expertise; (3) the ability to assess a problem or to offer advice, acting in a neutral role and without conflict of interest; and (4) the ability to develop a multinational team approach to the solution of problems of national and regional interest.

Turning our attention to specifics, CCOP has been called upon at one time or another to:

1. Stimulate the search for oil and gas;
2. Investigate the frontier regions;
3. Disseminate information on new techniques and developments;
4. Prepare subsurface maps and interpretations of area geology;
5. Maintain inventories of hydrocarbon occurrences;
6. Assess future potential.

Additionally, and from time to time, it has been suggested that CCOP become more resource oriented.

Accordingly, a programme of regional hydrocarbon assessment has been proposed which will have as its fundamental task the solution of limited-scope stratigraphic and sedimentary problems, which can later be

* Document prepared by CCOP Project Office, Regional Offshore Prospecting in East Asia, United Nations Building, Bangkok 2, Thailand.

combined into regional and area summaries. Deliberate emphasis is to be placed on the generation, migration and occurrence of hydrocarbons. It is believed that all objectives of the hydrocarbon programme can be met through this plan. The hydrocarbon assessment project is described in somewhat more detail below, but it is pertinent here to review some of the preliminary steps taken prior to its proposal.

Of fundamental importance in developing and carrying-out a programme was the gathering of the available basic information on the area. For that, a search was made of most area publications, as well as of international journals such as those of the Geological Society of America and the American Association of Petroleum Geologists, South East Petroleum Exploration Society, Petroleum News and others. This work is continuing.

The library of the Project Office has been considerably enlarged. Facilities offered by Petroconsultants have been reviewed and their 'Oil, Gas and Coal Compendium' has been purchased. To maintain some degree of awareness of what is going on in international developments, Petroleum Abstracts, published by the University of Tulsa, has been found useful.

Note was made above that the petroleum industry is changing. In 1966, when CCOP held its first session, daily production of Brunei, Indonesia, Japan and Sarawak was of the order of 579 000 barrels/day; in 1978, production was over 2 000 000 barrels/day (with all of Malaysia standing in place of Sarawak). At least in the offshore, the region has changed from one in which exploration was in a very youthful stage, to one in which exploration is locally in a mature to advanced-mature stage. Which is not to say that there is no further room for exploration.

A new parameter is also involved. Although in 1966 there were regional imbalances in production and distribution of hydrocarbons (which there still are), little thought was given, even by the major oil companies, to the idea that hydrocarbons were of finite quantity and limited distribution. Today, at a time when areas of apparently favorable characteristics have been reworked two, three, four or more times, and the industry is being driven into deep waters and other hostile situations, the global 'energy crisis' must be given serious consideration.

Among the most vital steps that should be taken, whether by the member countries of CCOP, or by the Project Office, or in some sort of joint effort, is that of stock-taking. Where have we been? Where are we now? Where are we going?

Integral and fundamental to the business of stock-taking is the hydrocarbon assessment programme. This programme has as its objectives: (1) to obtain an accurate inventory of oil and gas reserves for the region; (2) to develop information and maps of significant rock units in the region; (3) to develop and demonstrate the mechanisms responsible for the genesis, migration and storage of hydrocarbons; (4) to investigate the marginally productive or non-productive areas which lie *between* the basins; and, finally, (5) to outline new areas of exploration interest. It is believed that this programme satisfies the directives which were set out earlier, and specifically includes the traditional role of CCOP of investigating frontier regions. It is granted that many of these frontier regions are investigative rather than geographic, but at least for the present it is believed that the 'pay-out' will be of equal or greater value than other actions which might be taken.

Other frontiers have become apparent in the 13 years of CCOP's existence. These are in fact tools or concepts which were beginning to emerge in 1966 but which have subsequently developed to the point of being indispensable aids to the explorationist. These are the disciplines of assessment, exploration economics, and seismic stratigraphy.

Assessment, in the present sense, refers to the art of predicting the ultimate hydrocarbon resources which can be expected to be found within a basin, or an area, or, for that matter, the world. Originating with simple and now largely outdated types of volumetric approaches in which ultimate recovery was expressed in terms of millions of barrels per square or cubic mile, assessment has progressed through Delphi methods, barrels of oil to be found for a given amount of exploration hole to be drilled, to methods which use deposit modelling, risk analysis, and

Monte Carlo simulation.

The CCOP Project Office has prepared a conference on assessment, to be co-sponsored by ASCOPE, and to be held in Kuala Lumpur, Malaysia, in the first week of March, 1980.

As part of the assessment seminar, there will be a session devoted to exploration economics, the tool by which one judges which of several exploration decisions is economically most favorable and permits a country or company to direct its actions accordingly. Economic evaluations permeate the entire fabric of the exploration process and are integral to the exploration business.

It is anticipated that seismic stratigraphy will provide material input into the programme of hydrocarbon assessment. Seismic stratigraphy offers a means of recognition and correlation of depositional sequences, interpretation of environment, and identification of lithology directly from seismic cross sections. The Project Office believes it to be a very powerful tool, particularly in the offshore and in areas in which well control is minimal. It is hoped to make arrangements to further disseminate the required skills within the region.

Maintenance of an inventory of area hydrocarbon resources is one of the goals of the Project Office. Although efforts in this direction are still in a rudimentary stage, such an inventory is a logical outcome of the hydrocarbon assessment programme.

The central programme of hydrocarbon assessment is planned to start with a series of limited work projects of individual national or of regional interest, but designed to bring out geological details contributory to an overall appreciation of the hydrocarbon potential of the area. Such limited work projects can be undertaken on a strictly bilateral basis, or within regional organizations, acting with the support of CCOP, or by the Project Office itself. It is suggested that such limited projects be so programmed as to contribute in their totality to a series of long-range objectives which may be briefly summarized as:

1. Basic series of internally consistent maps and cross sections illustrating the geology of the region;
2. Catalog of oil field(s) data;
3. Data base for assessment and economic evaluation;
4. Map products for integration into the programmes of the Commission for the Geological Map of the World and the Circum-Pacific Council for Energy and Mineral Resources, etc.;
5. Geological atlas illustrating stratigraphy, structure, oil-field distribution for southeast Asia.

In short, it is a programme from which everyone stands to gain. The countries acquire a data base for use in future planning; industry obtains the benefits of an area-wide perspective of the geology of southeast Asia; potential newcomers have instant summaries available for studying their areas of interest; and CCOP will have responded to its initial charter of stimulating interest in the offshore hydrocarbon resources.

I might add that this is a programme in being. ASCOPE has initially designed and is now carrying out with CCOP assistance a series of geological-seismic sections which will illustrate the geology of the Gulf of Thailand, the Malay Basin, the Natuna Sea area, the northwest coast of Kalimantan and the area offshore Palawan Island, extending southwest into the Makassar Straits area. This is a short-term and limited project which forms an integral part of the long-range objectives.

The complete programme, as adopted in Bandung, Indonesia, September 1979 is as follows:

1. Petroleum Data Acquisition and Storage
2. Hydrocarbon Assessment Programme
3. Methodology of Assessment of Undiscovered Hydrocarbon Resources
4. Heat Flow Measurements and Geothermal Gradients as Related to Generation and Maturation of Hydrocarbons
5. Pre-Tertiary Petroleum Potential
6. Maintenance of Inventory of Hydrocarbon Occurrences in the CCOP Region.

A final note might be added which is pertinent to this Symposium (Petroleum Potential in Island Arcs, Small Ocean Basins, Submerged Margins and Related Areas). That is, that in considering the petroleum potential of deep-water areas, it would seem to be sound policy to proceed from the known to the unknown. The shelves of southeast Asia are relatively well known; a consideration of their rock geometry, geochemistry, and oil-field systems should contribute directly to an understanding of the environments of deposition and the petroleum potential of the adjacent deep-water realms.

Revised, May 26, 1980
Bangkok, Thailand.

UN ESCAP, CCOP/SOPAC Tech. Bull. 3 : 277-285

TWO YEARS OF DRILLING IN DEEPWATER

W E WHITNEY

The petroleum industry has long been intrigued by the idea of oil and gas production in deepwater; however, there was little economic incentive to pursue deepwater exploration until the dramatic upward surge in crude prices during the last few years. The resulting accelerated search for additional oil and gas reserves has led the industry into increased offshore exploration and additionally into ever increasing water depths.

Offshore drilling experience had slowly progressed from a water depth of 80 m in 1960 to 413 m through 1973. However, since 1973 the record has rapidly increased to 1486 m (4876 ft) at a recently completed exploratory well off eastern Canada drilled by Texaco and its partners using the Offshore Company's drillship, *Discoverer Seven Seas.* Figure 1 presents the progress in drilling-water depth since 1968. The rapid increase since 1973 can be clearly seen.

The definition of deepwater drilling itself has had to undergo refinement through the years as new water-depth records were set. What was once considered deepwater is no longer valid. Some now define deepwater as that water depth where dynamically positioned or unanchored vessels are utilized. This has not proven a definitive term, since some anchored units have been used in water depths beyond the reach of a number of dynamically positioned vessels. Another definition has referred to production technology and assigned deepwater to those water depths beyond conventional platform design. This also has been somewhat indefinite as conventional platforms have been pushed into deeper and deeper water depth, with the recent Shell Cognac platform holding the present record at 312 m (1025 ft) of water.

Some of us prefer to consider any water depth beyond the rating of most floating drilling units as deepwater (which is actually depths greater than about 300 m), and water depths of 900 m or more as ultra-deepwater.

At the same time as the industry has progressed in drilling deeper water depths, there has been an increase in the number of drilling units capable of drilling in deeper water. For instance, there is a rapidly growing list of offshore drilling units capable of drilling in water depths greater than 300 m (usually 300 to 600 m). Some are newly constructed, and a number are older units that have been converted for deeper water than they were originally fitted for. Some 87 drilling units are owner-rated as capable of drilling in 300 m or deeper water.

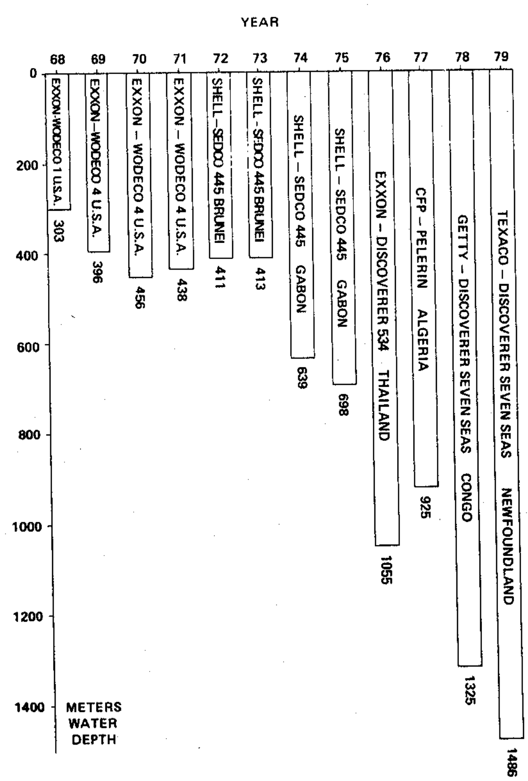

Figure 1. Deepwater drilling experience, 1968–79.

Drillships command a strong majority of the units capable of drilling in water depths greater than 600 m. At 900 m there are only two semisubmersibles in competition with the drillships.

At present some 12 offshore drilling units are rated as capable and equipped for drilling conventional petroleum wells in water depths of 900 m or more. These units and their depth ratings are presented in Table 1. Of these 12 units, 10 are dynamically positioned and 2 utilize anchor mooring. Dynamically positioned units use no anchors. Instead, they rely upon a system of position reference by sonic and mechanical means, which is fed into a computer which in turn directs power to auxiliary thrusters and the main propulsion propellers to hold the ship at a selected position with reference to a spot on the ocean floor.

Although it is possible to anchor-moor to about 1500 m (5000 ft) water depth, the problems of anchor pennant line handling and the complicated design and bulky combination of the wire/chain-anchor line system has limited practical application of anchor-mooring to about 400 m water depth. However, the record for anchor-mooring of an offshore drilling unit is much deeper and was established at a well drilled in 1055 m (3460 ft) of water off Thailand. This well was drilled by Exxon in 1976 using the Offshore Company's *Discoverer* 534.

The ability to drill in deeper water depths has resulted from a number of factors in improved equipment and technology. Briefly the major contributing factors have been:

(1) Extension of combination wire/chain mooring systems such as that used by the *Discoverer* 534.
(2) Development of dependable dynamic positioning systems which eliminate anchoring requirements.
(3) Development of sonar/TV systems to permit timely re-entry on the ocean floor without the use of guidelines.
(4) Development of multiplex electro/hydraulic blowout preventer and wellhead connector control systems to replace dependency on straight hydraulic-operated controls which become slow-reacting and bulky at increased water depth.
(5) Improved riser couplings and riser design plus buoyancy devices to permit proper riser tension in deeper water.

One industry expert, D S Hammett of Sedco, recently stated that our capability to drill in deepwater has advanced such that present equipment and technology are available to drill in 2440 m (8000 ft) water depth. Figure 2 presents a comparison of capability versus the actual experience record during recent years as developed by Mr Hammett.

The ability to drill to the deeper water depths

TABLE 1

DEEPWATER DRILLING UNITS

900 METERS WATER DEPTH OR DEEPER

DRILLING UNIT	TYPE	YEAR MFG.	DEPTH CAPABILITY METERS/FEET	MOORING METHOD
(1) SEDCO 471	DRILLSHIP	1978	900/3000	D. P.
(2) PACNORSE	DRILLSHIP	1979	900/3000	D. P.
(3) BEN OCEAN LANCER	DRILLSHIP	1977	900/3000	D. P.
(4) FERNSTATE	SEMI	1978	900/3000	ANCHOR
(5) PELERIN	DRILLSHIP	1976	1100/3600	D. P.
(6) DISCOVERER 534	DRILLSHIP	1975	1130/3700	ANCHOR
(7) NEDRILL II	DRILLSHIP	1977	1200/4000	D. P.
(8) PENGUIN	DRILLSHIP	1979	1500/5000	D. P.
(9) DISCOVERER SEVEN SEAS	DRILLSHIP	1976	1800/6000	D. P.
(10) SEDCO 445	DRILLSHIP	1971	1800/6000	D. P.
(11) SEDCO 472	DRILLSHIP	1977	1800/6000	D. P.
(12) SEDCO 709	SEMI	1976	1800/6000	D. P.

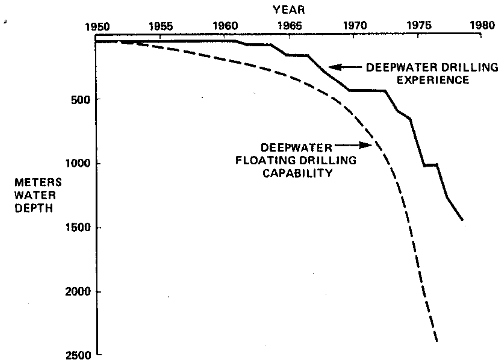

Figure 2. Deepwater drilling capability/experience, 1950–80.

recorded in recent years perhaps received its greatest contributions from two international oil-company/drilling-contractor programs beginning in 1972 with Shell's use of the *Sedco* 445 and CFP/Total's use of the *Pelican.* Both of these units pioneered dynamic positioning or station keeping systems, guidelineless re-entry methods and other currently applied deepwater technology. The units themselves led to a number of present-day deepwater vessel designs with the 445 leading to the *Sedco* 471 and 472 and the *Pelican* becoming the forerunner of other IHC designs: the *Havdril, Ben Ocean Lancer, Petrel, Pelerin, Pacnorse,* and *Penguin.*

Prior to 1974 the number of wells drilled in water depths deeper than 300 m was quite small. However, as with increasing water depth, since 1974 there has been a marked increase in the number of deepwater wells drilled. This ties into deepwater exploration expenditure and the technology developed by the 445 and *Pelican* during 1972 and 1973. Figure 3 presents a graph of the number of deepwater wells drilled each year.

It has been estimated that petroleum prospective areas in water depths from 200 to 3000 m amount to a worldwide total area of about 35 million km^2. This is more than twice the offshore area presently being exploited for petroleum with a water depth of less than 200 m. Existing licences cover only 5 million of the 35 million km^2 of prospective deepwater and only about 280 wells have been drilled in deepwater (200 m or deeper) of which only 30 have been in water depths over 600 m. Therefore, many expect a substantial level of deepwater drilling activity for some time to come.

So far, the results of deepwater drilling have not been very encouraging. The economics of deepwater production, considering water depth, environmental problems, production rates, taxes and royalties, indicate that very large and highly productive fields are required for commercial exploitation of deepwater discoveries. Also, production technology for deepwater production has not kept pace with deepwater drilling capability and is lagging well behind with only limited deepwater production equipment beyond a designing stage. Of an estimated $5.4 billion spent so far on deepwater drilling, only two discoveries have reached a development stage: Exxon's Hondo and Shell's Cognac fields, which were in water depths within the extreme limits of present-day conventional platform installation and completion methods and therefore did not require a radical change in completion equipment.

The costs of deepwater drilling are quite high as compared to those for normal offshore drilling. This is readily apparent when one considers the cost of the deepwater drilling units themselves and the more sophisticated ancillary support services required. Recently constructed dynamically positioned vessels have cost in the range of $60–65 million for the basic drilling unit. Total 'all-in' operating cost for such units run from $100 000 to $130 000 per day. The cost for a single deepwater well therefore is correspondingly high. For instance, the cost estimate for the recent world-record water-depth well off eastern Canada was published at $25 million.

Table 2 presents a typical breakdown of daily drilling costs associated with a dynamically positioned drillship operating in approximately 500 m water depth off the eastern coast of Spain. The total daily drilling cost is approximately $107 000 per day. Table 3 presents a summary of the total well cost for a well at such a location requiring 50 days to drill, resulting in a total well cost of $5.7 million when mobilization and other items are considered.

In spite of the high costs involved, it is

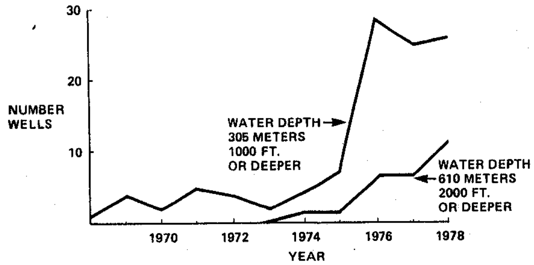

Figure 3. Number of wells drilled in deepwater 1971–78.

TABLE 2

ESTIMATED DAILY DRILLING COST
OFFSHORE SPAIN
D. P. VESSEL 500 METERS WATER

DRILLING UNIT DAILY RATE		$ 53,000
COMPANY SUPERVISION		2,000
FUEL AND UTILITIES		4,500
COMPANY DRILLING EQUIPMENT		1,000
	SUB TOTAL	60,500
TRANSPORTATION	SUPPORT BOATS	6,750
	HELICOPTER	2,500
	AIR AND TRUCK	1,500
	SUB TOTAL	10,750
FIXED RENTALS	WIRE LINE LOGGING	1,000
	TESTING EQUIPMENT	350
	DIVING	1,250
	CEMENTING/TESTING	600
	MUD LOGGING	1,000
	COMMUNICATION	100
	MISCELLANEOUS	150
	SUB TOTAL	4,450
SERVICE COMPANY PERSONNEL		1,500
SHORE BASE		2,000
MISCELLANEOUS SERVICES		400
	FIXED DAILY COST	79,600
EXPENDABLES		27,500
	TOTAL DAILY COST	$107,100

expected that deepwater drilling expenditures during the next few years will continue to represent about 15% of the industry total oil and gas capital and exploration expenditure. Figure 4 presents a curve of actual industry deepwater drilling expenditures by year with a forecast through 1985.

The increasing trend towards acquiring and exploring deepwater areas was evidenced by the recent O.C.S. lease sale No. 48 covering offshore Southern California where out of 55 blocks bid on, only 11 were in waters less than 200 m deep. It is of interest to note that Chevron U.S.A. was high bidder as operator for 21 of the 55 tracts.

TABLE 3

ESTIMATED COST
D. P. VESSEL 500 METERS WATER
2750 METERS WELL DEPTH
50 DAYS – OFFSHORE SPAIN

DRILLING UNIT COST		$2,650,000
SUPERVISION		100,000
FUEL AND UTILITIES		225,000
COMPANY DRILLING EQUIPMENT		31,000
	SUB TOTAL	3,006,000
DRILLING FLUIDS		85,000
TRANSPORTATION		538,000
BITS, STABILIZERS		21,000
RENTALS AND SERVICES		275,000
MISCELLANEOUS SERVICES/EQUIPMENT		120,000
	SUB TOTAL	939,000
WIRE LINE LOGGING		120,000
MUD LOGGING		48,000
CORING		52,000
TESTING		78,000
	SUB TOTAL	298,000
CASING		384,000
SUB SEA PRODUCING EQUIPMENT		137,000
CEMENTING		81,000
	SUB TOTAL	602,000
GENERAL OVERHEAD		400,000
SITE SURVEY AND PREPARATION		100,000
	TOTAL ON-SITE COST	5,345,000
	DAILY COST	106,900
MOBILIZATION AND DEMOBILIZATION		360,000
TOTAL WELL COST		$5,705,000

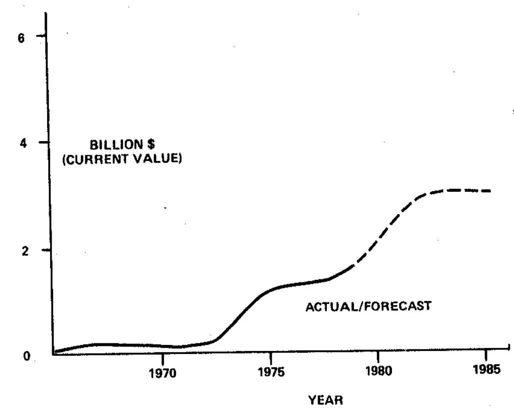

Figure 4. Petroleum-industry deepwater drilling expenditures, 1965–85.

Chevron, like other international oil companies, has been active in deepwater drilling throughout the years either as operator or as a partner in outside ventures. This follows logically since Chevron has drilled more than 2000 offshore wells throughout the years and has contributed heavily to modern offshore drilling technology. In late 1974, a corporate review made of forecasted offshore operations which combined the programs of all Chevron Operating Companies for the next five years indicated a number of deepwater wells to be drilled during the next few years. One operating company, Chevron Overseas Petroleum, indicated a continuous requirement for a unit capable of deepwater drilling for an approximately two-year period beginning in 1976.

As a result of this review, and approaches from ODECO, a commitment was made to contract the *Ben Ocean Lancer*, an IHC-designed, dynamically positioned drillship, which was under construction in Scotland at the time, and scheduled for completion in mid 1976. Like many other offshore construction projects during that period, the delivery date slipped and the vessel was not ready to begin drilling operations until the fall of 1977. As a result, a portion of the original deepwater drilling program fell by the wayside or was accomplished by other units such as the *Pelerin*, a sistership to the *Ben Ocean Lancer*, which was used to drill offshore Greenland during the summer of 1977.

The *Ben Ocean Lancer* is typical of the more modern, dynamically positioned drillships. It has an overall length of 154 m (504 ft), a waterline length between perpendiculars of 137 m (449 ft), a beam of 23 m (77 ft), and a design draught of 7.9 m (26 ft). A comparison of dimensions and power between the *Ben Ocean Lancer* and *Discoverer Seven Seas*, as examples of modern dynamically positioned drillships, is presented in Table 4.

During the time that the *Ben Ocean Lancer* was under contract to Chevron, it was rated and fully capable of drilling in water depths up to 900 m (3000 ft). Major vessel drilling equipment terms included:

(1) 18⅝ in o.d. riser of ⅝ in and ½ in wall thickness and equipped with Emmons Cummins flotation modules. The complete riser was carried on board the vessel.
(2) Gardner Denver 3000 E drawworks.
(3) Two Gardner Denver PZ11 (1600 HP) mud pumps.
(4) Pyramid 160 ft derrick rated at 1 000 000 lb load capacity.
(5) Rucker motion-compensator system.
(6) B J Hughes vertical pipe racking system.
(7) Honeywell ASK (dynamic positioning (d.p.)) system with redundant displays and three Honeywell 316 computers.
(8) Two complete Cameron 16¾ in 10 000 psi

TABLE 4

COMPARISON OF TYPICAL DEEPWATER DRILLSHIPS

	BEN OCEAN LANCER	DISCOVERER SEVEN SEAS
OVERALL LENGTH	154 METERS (504')	163 METERS (534')
LENGTH BETWEEN PERPENDICULARS	138 METERS (449')	148 METERS (486')
OVERALL BEAM	23.5 METERS (77')	24.4 METERS (80')
DRAFT	7.9 METERS (26')	7.5 METERS (25')
PROPULSION	FOUR 1750 HP UNITS ON MAINSHAFTS	FOUR 1500 HP UNITS ON MAINSHAFTS
	FIVE 1750 HP TRANSVERSE THRUSTERS	SIX 2500 HP LATERAL THRUSTERS
POWER GENERATION	FIVE 3400 HP UNITS PLUS TWO WITH 1200 HP TOTAL	SIX 3600 HP UNITS PLUS ONE 1500 HP UNIT
DRILLING VARIABLES CAPACITY	8600 LT	8825 LT

rated blowout-preventer stacks. Both stacks carried on board the vessel.

(9) Cameron multiplex electro-hydraulic control system.

(10) Satellite navigation system.

(11) Marisat satellite communications system with facsimile, teletype and voice facilities.

(12) A diving system utilizing a 3000 ft depth rated, one-atmosphere bell, which was launched in a separate moon-pool.

For storage the mud and cement bulk silos held 19 800 ft^3 and there was also room for 365 tons of sacked material. Liquid mud storage was 2000 bbls, fuel storage was 3150 tons (consumption was 20-25 tons daily), drill-water storage was 370 tons and potable-water storage 260 tons.

The unit was fully capable of carrying complete casing requirements and wellhead supplies for two 10 000 ft wells.

Chevron Overseas became active in working with the drilling contractor while the drillship was still under construction, and Chevron technological and engineering staff participated in several of the new equipment designs. Also, an active Chevron supervisory/management team was appointed and placed at the construction site in Scotland in late 1976 so that they might become intimately familiar with the unit's operating systems, assist ODECO in construction design decisions, and participate in the selection, training and development of crew members well before the unit would begin actual drilling operations.

The onsite Chevron team consisted of a Project Drilling Manager, an Assistant Project Drilling Manager, and two 2-man Drilling Representative pairs. This team continued to function when the unit began actual drilling operations. The team reported directly to the Manager Drilling Operations in San Francisco. Membership on the team provided excellent training exposure and those who served on it have provided a key source of personnel to participate in other deepwater drilling projects during and since completion of the *Lancer* program. The team participated in programming and following the ship's trial and drilling-equipment function tests. This provided further familiarization with the unit and laid the foundation for a later extremely efficient trouble-shooting and problem-solving joint effort between Chevron and ODECO whenever equipment and operational problems developed.

All operations were planned ahead of time and wells were carefully programmed prior to the beginning of individual drilling operations. For example, weather downtime predictions were calculated for each well using historical weather data and vessel motion characteristics in a highly engineered computer program. These analyses proved quite accurate and played a key part in the timing selected for drilling offshore Canada and in the Baltimore Canyon.

It was forecast and later supported that the dynamically positioned drillship's ability to select optimum heading for particular wave and swell conditions might enable the d.p. unit to operate more efficiently than similar-sized anchored units under certain rough weather conditions. Figure 5 presents a vessel's response curves for various headings at a selected wave condition, and it is readily apparent that an anchored unit with a fixed heading could encounter more downtime problems than a d.p. unit. During the period of use by Chevron the *Lancer* proved capable of holding position within 1.5% of water depth with winds exceeding 60 knots and seas of 25 ft. The unit was actually able to continue drilling operations with 40-45 knot winds while holding its position within 0.5% of water depth.

Chevron's deepwater drilling program with the *Ben Ocean Lancer* began in August 1977 off the east coast of Spain, with the first well in 47 m (1544 ft) of water and drilled to a depth of 2800 m (9190 ft) in 66 days. This was believed to be a commendable performance considering that it was the first well drilled with a new unit utilizing a number of new and complicated equipment concepts. Six wells were drilled during the continuous *Ben Ocean Lancer* program in water depths ranging from 471 m (1544 ft) to 866 m (2842 ft), and drilled depths from 2069 m (6790 ft) to 5287 m (17 347 ft). Table 5 presents a listing of the six wells drilled, with pertinent timing and depth information. The total period of time to drill the six wells was approximately 19 months, including moves, which were accomplished at a sailing speed between

TABLE 5

DEEPWATER WELLS
BEN OCEAN LANCER
CHEVRON OVERSEAS PETROLEUM

WELL	SPUD DATE	RELEASE DATE	WATER DEPTH (METERS)	TOTAL DEPTH (METERS)	DRILLING DAYS
(1) MONTANAZO D-1 (SPAIN)	8/13/77	10/18/77	471	2800	66
(2) MONTANAZO C-1 (SPAIN)	10/23/77	1/24/78	673	2896	93
(3) MONTANAZO D-2 (SPAIN)	1/30/78	3/19/78	746	2743	48
(4) ACADIA K-62 (NOVA SCOTIA)	4/01/78	8/02/78	866	5287	124
(5) HOPEDALE E-33 (LABRADOR)	8/10/78	10/01/78	550	2069	52
(6) COST B-3 (EAST COAST USA)	10/09/78	1/24/79	819	4822	107

locations averaging about 11 knots.

No major problems or long shutdowns were encountered in the drilling of the six wells; however, like all operations with new and highly sophisticated equipment, a number of problems occurred which resulted in brief shutdowns. Initially, problems were encountered with the sonic/TV re-entry tool and later re-entries were mostly made with an outside TV camera assembly. Another problem developed when after 21 days the drawworks drum shaft developed a serious crack and had to be replaced. Probably the greatest loss of time resulted from problems encountered with the multiplex system. These mostly concerned shorts in the cabling and the connections. Backup cables were used, and eventually new techniques were applied to cable connections and termination assemblies with resulting great improvement. In spite of the multiplex problems, the stack had to be pulled only once. Problems were also encountered with the original motion compensator, which was of non-standard manufacture and was incapable of proper operation or of being repaired. It was eventually replaced with a proven and well known brand. During the 19 months, two drive-offs occurred as a result of loss of control of the dynamic-positioning station-keeping system. One resulted from a power failure and the other from a technician/operator error. In both instances, disconnect was accomplished without a problem, and no difficulty was encountered in reconnecting to the stack. The power system was corrected and the technician operator learned his lesson.

One of the major accomplishments with the unit during the program was the number of drillstem tests that were effectively performed without mishap. During the six-well program, 16 drillstem tests were made with oil flow rates as high as 9800 bbls/d offshore Spain, and a test offshore Canada yielded 18.5 million ft^3 of gas with 350 bbls/d condensate. The experience gained from these tests developed technology for improved testing in deepwater and established Chevron as one of the most experienced deepwater testing operators. The test of 9800 bbls/d oil at Montanazo D-2 in 746 m (2448 ft) water depth is believed to be the deepest-water high-volume oil-flowing test ever made.

Drilling efficiency with the unit during the six-well program is considered to have been very good overall and with a definite trend toward better performance throughout the program with additional experience. Figure 6 presents a comparison of the third well drilled with the *Lancer* offshore Spain against the 'best time' nearby well which was drilled in shallower water with a modern but well-proven conventional semisubmersible. Figure 7 presents a comparison of the sixth well drilled with the *Lancer* against two nearby wells drilled with other units in the Baltimore Canyon area. As can be clearly seen, the *Lancer* by this time had become an outstanding performer.

In conclusion, it can be said that Chevron Overseas' six-well program with the *Ben Ocean Lancer* clearly supports the premise that the technology is here to effectively drill and test deepwater prospects. The program also

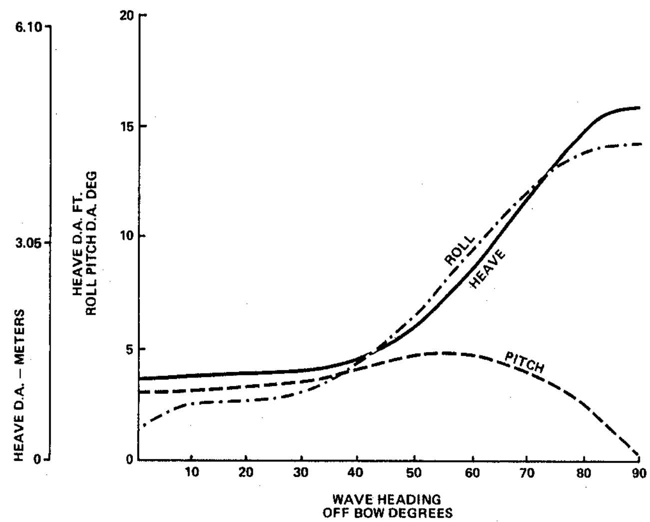

Figure 5. Influence of heading on vessel motion (from G D Smith, 1979); (15.3 ft sig. wave height).

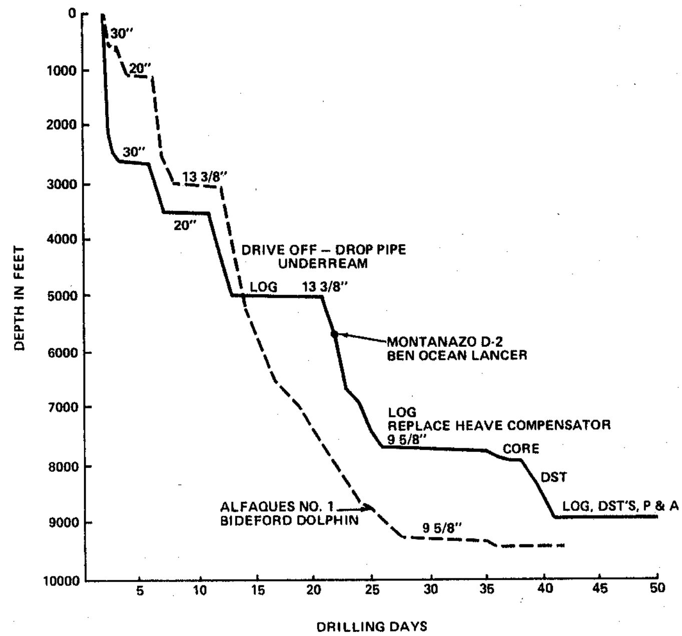

Figure 6. Ben Ocean Lancer drilling progress. (For description see text.)

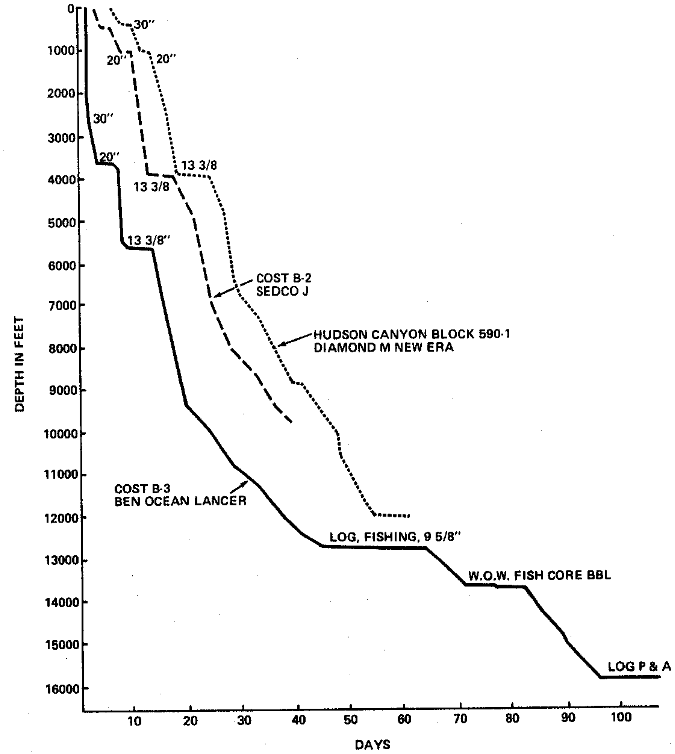

Figure 7. Ben Ocean Lancer drilling progress. (For description see text.)

emphasized that deepwater drilling is expensive, complicated and requires careful planning, detailed engineering and constant high-quality supervision. However, with careful attention to these factors, deepwater drilling efficiencies, comparable with those of conventional shallow-water-depth offshore drilling, can be obtained.

ACKNOWLEDGEMENTS

The author expresses thanks to Mr J W Wright and the Chevron *Ben Ocean Lancer* Supervisory/Management Team for their comments and their contribution to the *Lancer* Project. The author also thanks Chevron Management for permission to present this paper.

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