PLATE TECTONIC STATUS AND SEDIMENTARY BASIN IN-FILL OF THE NATAL VALLEY (S.W. INDIAN OCEAN)

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ABSTRACT

The Natal Valley lies between the Mozambique Ridge and the east coast of South Africa and Mozambique. A revised reconstruction of South America and south central Africa results from a new fit of the Falkland Plateau and the Natal Valley: euler pole: 46.75°N 32.65°W, rotation angle 56.40°. The reconstruction is corroborated by a) juxtaposition (rather than overlap) of Precambrian crust in the Gulf of Benin and north-eastern Brazil; b) alignment of three pre-break-up inter-continental features - i) the Jurassic Outeniqua Basin (South Africa) with the Falkland Plateau Basin; ii) the Triassic Cape Fold Belt's northern tectonic front and a morphological feature on the Falkland Plateau with which it closely lines up; iii) the Precambrian Pernambuco fault and mylonite belt of Brazil with the equivalent Foumban line of West Africa.

The premise that the southern face of the Tugela Cone marks the continent-ocean boundary is corroborated by the discovery of Mesozoic magnetic anomalies M0-M10 in the Natal Valley. These are offset ∿1300 km from their equivalents in the Cape Basin. These anomalies are used in conjunction with South Atlantic anomaly sets to produce Cretaceous palaeopositions for the Falkland Plateau relative to southern Africa. The offset along the Falkland Agulhas Fracture Zone (FAFZ) remained 1300 km long throughout M10-M0 time and was ∿1270 km by chron 34 time. Therefore no major ridge-jumps had occurred by then. The reconstructions allow events in the Cretaceous Normal Polarity Epoch to be dated more accurately. The continent-ocean boundary in the southernmost Cape Basin implied by the above model has been corroborated by recently published multichannel seismic work.

The northern Agulhas Plateau is not the site of a ridge-jump, but was formed by excessive volcanism together with the Islas Orcadas and Meteor Rises, ∿93 m.y. ago. Meteor Rise and Islas Orcadas Rise, forming a composite feature, separated from the Agulhas Plateau between 93 and 80 m.y.

A revised sequence of Cretaceous and Late Jurassic reconstructions for East Antarctica relative to Africa is generated from published seafloor spreading and fracture zone anomalies. This leads to a revised reconstruction for east and west Gondwanaland: euler pole 1.67°S, 35.99°E, rotation angle 53.43°. The reconstruction is supported by the alignment of a) the Cape Fold Belt with its Antarctic equivalents and b) the western limit of Pan-African reactivation of the Mozambique Mobile Belt with the western limit of the Sverdrupfjella metamorphic suite. This reconstruction places East Antarctica adjacent to the Lebombo Mountains and south of and sub-parallel to the Sabi acid volcanic suite supporting the postulate that these rocks represent Mesozoic plate boundaries. This refit implies that part of the coastal plain of Mozambique is underlain by oceanic crust. Antarctica initially moved along a transform fault lying immediately east of and sub-parallel to the Lebombo Mountains, the eastern face of the Tugela Cone, the south-eastern face of the Falkland Plateau in its reconstructed position, and the eastern flank of the Agulhas Plateau in its reconstructed position. At M10 times a triple-junction formed at the tip of the Falkland Plateau implying that the oldest lineated magnetic anomalies in the eastern Weddell Sea cannot pre-date M10. At M2 times Mozambique Ridge rifted from Astrid Ridge in a ridge-jump episode. Such an event may also explain the presence of magnetic anomalies in the southern Mozambique Basin. Gunnerus Ridge, Conrad Rise and Del Cano Rise are all tectonically associated with the Madagascar Ridge.

The above reconstructions imply that sheared margins occur along the Agulhas Margin, the Lebombo Mountains and the Davie Fracture Zone. Adjacent on-land areas have traditionally been associated with tensional rather than strike-slip-related rifting. A variety of features including en échelon fault configuration, reversal of fault throw along strike, and pull-apart basins suggest that Late Jurassic-Early Cretaceous faulting in coastal Natal is consistent with dextral strike-slip strain associated with Gondwanaland break-up.

Five physiographic sub-provinces are recognised within the northernmost Natal Valley: continental shelf and slope, Limpopo Cone, Inharrime Terrace, Central Terrace and Almirante Leite Bank. The sedimentary succession is divided into five units by five regionally developed acoustic horizons: basement (pre-Barremian), McDuff (~Cenomanian), Angus (Oligocene), Jimmy (Miocene-Pliocene boundary) and 'L' (Within Pliocene-Recent). Structure contour and isopachyte charts are prepared for each horizon and sedimentary unit respectively.

Basement is volcanic and was emplaced sub-aerially. The exposed Almirante Leite Bank and the predominantly sub-surface Naudé Ridge form two basement highs. Sedimentary units 1 (pre McDuff) and 2 (Angus-McDuff) are dominated by up-lap and growth faulting in irregular horsts and graben. They respectively exhibit maximum thicknesses of 1.1 and 1.35 sec. just north of the Naudé Ridge. Unit 3 (Jimmy-Angus) forms a progradation of the Zululand continental slope and the Limpopo cone, with thickness up to 0.7 sec. Unit 4 (Jimmy-'L') is sub-divided: the lower part is dominated by massive slumps; the upper part is a 'slope-front wedge'. Maximum thickness of 0.73 sec. developed in the Limpopo Cone. Unit 5 (post-'L') sediments form the Inharrime Terrace - a slope-front wedge up to 0.44 sec thick with a nested sequence of leveed channels. Total sediment thickness (3.5 km) is less than other marginal basins of southern Africa. Sedimentation rates are relatively high for unit 1, low for units 2 and 3, increasing over five-fold in units 4 and 5. Present-day sediment supply rates are between 25 and 30 times faster than Cretaceous-Recent and Pliocene-Recent accumulation rates, probably as a result of agricultural malpractices.

The Agulhas Current has: a) prevented sedimentation on Almirante Leite Bank seamount flanks; b) exposed Cretaceous and Tertiary out-crops and reworked thin Neogene deposits on the Central Terrace; c) scoured the outer shelf and upper continental slope; d) helped restrict fast Neogene sedimentation to the north-west of the basin; e) controlled glauconite and phosphorite authigenesis under an upwelling cell. Current action may have influenced mid-

Cretaceous (McDuff) and Cretaceous-Tertiary boundary hiatuses. The Agulhas Current is clearly associated with Oligocene (Angus) and Miocene-Pliocene boundary hiatuses. The counter-current and up-welling cell moved east in Post-'L' times in response to depocentre changes. Oligocene and Pliocene changes in palaeo-oceanography leading to cooler sea-temperatures in the Indian Ocean reduced rain-fall over southern Africa and resulted in terrestrial faunal turn-overs at these times.

Sediment instability is demonstrated by slumps, growth faults, tensional graben and diapirs. The largest of the five slumps recognised is at least 20 800 km^2, possibly 34 000 km^2 in extent, rivalling the largest yet described in the world. Three phases of slumping are recognised: Angus - Jimmy (unit 3), Lower Jimmy-'L' (Lower unit 4) and post 'L' (unit 5). Diapirs exist in the earlier two phases. Growth faulting is very common in unit 2 sediments of the Central Terrace. A large growth fault and roll-over anticline complex exists on the southern Limpopo Cone, predominantly in post-Jimmy material. Over 1950 km^2 of units 4 & 5 of the Limpopo Cone and Inharrime Terrace are affected by tensional graben.

ACKNOWLEDGEMENTS

First of all, I thank Professor Richard Dingle, my supervisor for inviting me out to South Africa, arranging a job with the Geological Survey, and initiating this project. Richard acted as chief scientist on our main data-gathering cruises, introducing me to Marine Geoscience à la Thomas B. Davie. He arranged for use of the facilities of the Geology Department of the University of Cape Town where this work was carried out. Furthermore Richard provided reviews of various publications, the essence of which are contained herein, as well as a careful review of an earlier version of this manuscript. Much early work - digging out data, bathymetric compilations, preliminary interpretation of seismics, and gathering of data at sea - was done in conjunction with Steve Goodlad. Although data generated from sediment samples play a small role in this thesis, initial handling of cores, photography, subsampling etc. was again a joint venture with Steve. We also discussed many aspects of the Natal Valley project together. The captains and crews of the R/V's Thomas B. Davie and Meiring Naudé, helped by Bill Siesser, Frances Camden-Smith and particularly Tom Fitton and Jack Engelbrecht made possible the acquisition of samples and geophysical profiles at sea. Our piston corer was built in-house by Eric Bryce, while a great deal of help in handling samples in the laboratory was given by Mathew Smith. Dave Salmon kindly dated key samples micropalaeontologically. The plate tectonic sections of this thesis were carried out using programmes written for the University's Univac computer by Chris Hartnady. Chris' fertile mind provided many ideas and very stimulating discussions. Finance and support was initially provided by the Geological Survey of South Africa through André du Plessis, 'Hannes Theron and Robbie Kleywegt. More recently, support was provided by the National Research Institute of Oceanology (C.S.I.R.) where Burg Flemming kindly incorporated my work into his east coast programme. Burg and I have also enjoyed some very stimulating discussions. Useful discussion were also held with Donne Murray and André du Plessis (magnetics), Gavin Birch and Mike Bremner (phosphorites), John Rogers (palaeoclimatology), and Ivor Van Heerden (seismic interpretation). Various papers which are included herein have been reviewed by C.G.A. Harrison, M.J. De Wit, A.H.F. Robertson, P.E. Mathews, Clive Stowe, Pete Betton, Brett Hendey, Elizabeth Vrba, John McCarthy, Johann Lutjeharms and Dirk Van Foreest. Draughting assistance was given by Shirley Smith, Sue Sayers, Marian Van Meiracker and Marcella Brandstätter, while Grace Krummeck typed and re-typed hefty manuscripts. To all of these people a very appreciative thank-you. Finally, to Judy Elliott, who helped in the lab and drawing office but more especially for tolerating me and dealing with various bouts of Ph.D. blues - thank you.

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PREFACE

"I rather dreaded that, since the obligatory style of dissertations is turgid in the extreme, and I had by now spent nine years trying to write well, I was afraid I simply might not be able to write badly enough to qualify for my degree." - Isaac Asimov.

The usual pattern followed by a research student is to first produce a dissertation and later, possibly, to distil scientific contributions from the dissertation for scientific journals. Emery (1976) laments the fact that many articles which form parts of doctoral dissertations, represent the only publication by an individual in that field. The implication is that only parts of dissertations reach the scientific community at large, while the rest remain relatively inaccessible on the shelves of a single university. Moreover, when a part of Joe Bloggs' thesis is published internationally, it often appears as Bloggs et al.,.

An alternative is to present a thesis in the form of published or publishable papers (Williams 1980 - nine articles, plus six in appendix form; Winter 1981 - two articles).

This thesis is presented in the more traditional format. However many parts of this work have previously been published either in international journals or in the annual reports of the Marine Geoscience Unit, University of Cape Town. Some of these contributions have been multi-authored. Although this situation is similar to the 'Bloggs, Bloggs et al.,' scenario outlined above, I have listed these publications in appendix 2 with notes on the contributions of the various authors.

PART 1

CHAPTER 1

INTRODUCTION

This study was initiated by Professor R.V. Dingle in order to extend the Marine Geoscience Unit's work to the east coast of southern Africa. Previous theses had been focussed on the west and south coast shelves (Birch, 1971; 1975; Rogers, 1971; 1977; Bremner, 1978), while Dingle (1973 a,b,c) mapped extensively in these areas. Although the contrast in structure between the sheared south coast margin and the rifted west coast had been recognised (Scrutton et al., 1975; Rabinowitz, 1976) little work had been carried out on the east coast. The bathymetry of the continental shelf and the adjacent margin and basins was partly known (Moir, 1974; Simpson, 1974), while Bang (1968) described some canyons on the continental slope. Seismic refraction and gravity work (Ludwig et al., 1968; Darracott, 1974; Scrutton, 1976b); Chetty & Green, 1977) showed that the southern Natal Valley is underlain by normal oceanic crust, while crustal thickness in the northern Natal Valley and the Mozambique Ridge is intermediate. Following a suggested tectonic scenario for east Africa (Kent, 1972; 1974) vertical tectonics were considered the controlling factor in the northern Natal Valley (Dingle, 1982; Dingle & Scrutton, 1974; Scrutton & Dingle, 1976). Although the general outlines of Indian Ocean sediment distribution were known (Ewing et al., 1969; Kolla et al., 1976) little detail existed for the Natal Valley. Commercial and academic drilling on the coastal plain, continental shelf and offshore plateaux and basins, enhanced by limited seafloor sampling, elucidated the stratigraphy of the area (Saito & Frey, 1964; Saito et al., 1974; Hazzard et al., 1971; Frankel, 1972; Flores, 1973; Du Toit & Leith, 1974; Simpson, Schlich et al., 1974; Vincent, 1976). The lack of knowledge on the Natal Valley is reflected in quite recent syntheses (Nairn & Stehli, 1982). Since the initiation of this study, the situation has improved with the development of an east coast shelf sediment dispersal model (Flemming, 1978; 1980; 1981), some sedimentological and palaeo-oceanographic work in adjacent abyssal areas (Kolla et al., 1980 a+b; Hutson, 1980; Prell et al., 1980), and the publication of parts of this work (appendix 2).

Concerning the east coast continental margin and the Natal Valley, it is obvious that the area is of great interest and many questions remained unanswered. Within this broad geographical region, Professor Dingle allowed a free hand. From a host of possibilities, this study addresses two main facets of the geology of the Natal Valley: 1) its plate tectonic situation; 2) in-fill of the sedimentary basin, with emphasis on the influence of geostrophic currents and mass-gravity processes.

The thesis is arranged in four parts. Part 1, encompassing chapters 1 and 2, introduces the subject and describes the regional setting, giving attention to hinterland geology, bathymetry, climate, and physical oceanography.

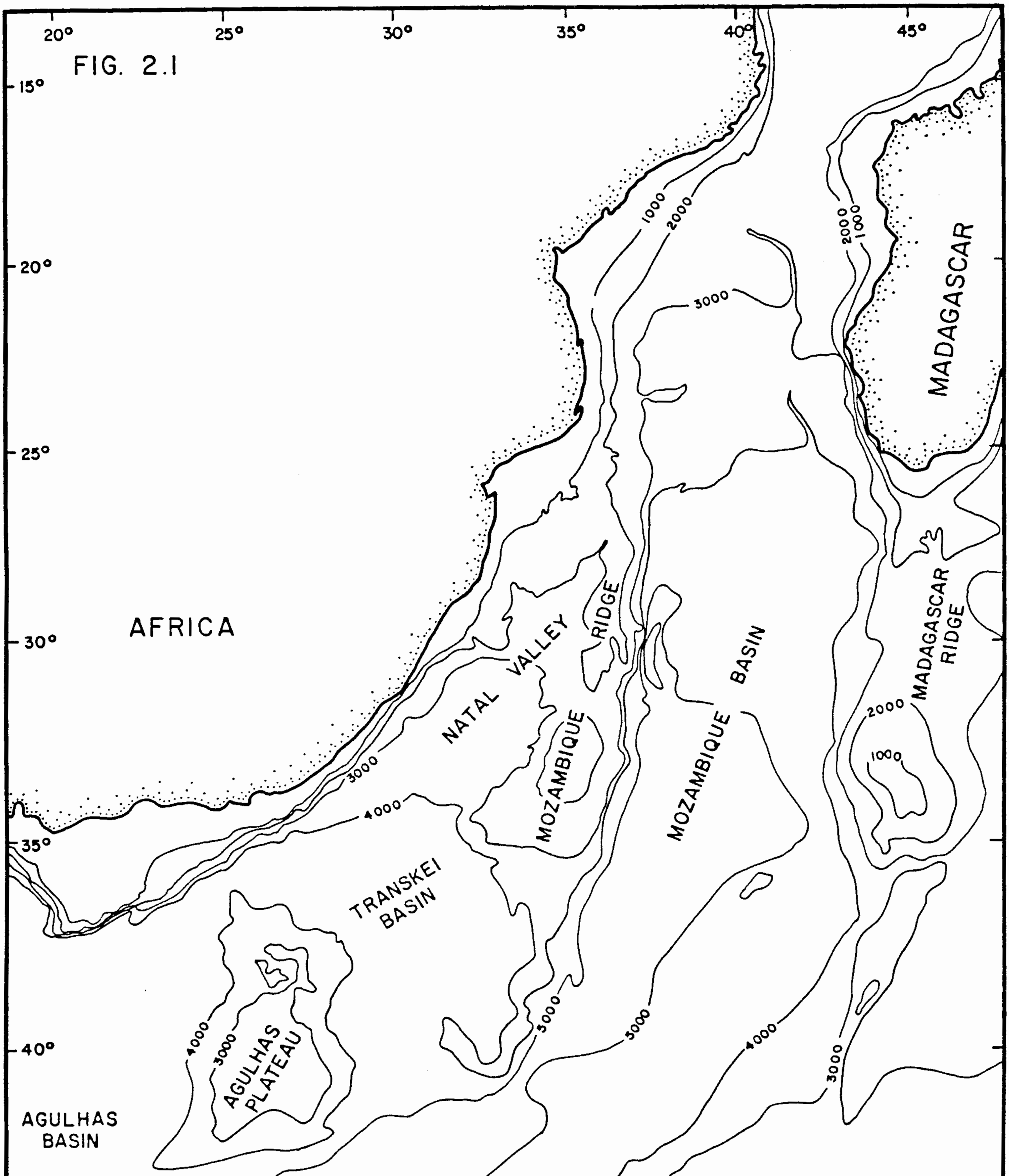
Part 2, comprising chapters 3 to 7, deals with plate tectonic reconstructions and their implications. In chapter 3, a revised reconstruction of West Gondwanaland is presented. This stems from a new refit in the Natal Valley. In chapter 4, I present a re-interpretation of magnetic data, in which for the first time Mesozoic seafloor spreading magnetic anomalies are recognised in the Natal Valley. These lead to a sequence of Cretaceous palaeopositions for the Falkland Plateau relative to South Africa which have implications for a) the evolution of the south-west Indian Ocean and the South Atlantic; b) the formation of aseismic plateaux; c) timing of events in the Cretaceous normal polarity epoch. In chapter 5, a new model is developed for the evolution of the Agulhas Plateau and related ridge-jumps south of the Agulhas Falkland Fracture Zone. Chapter 6 provides a revised reconstruction for East Antarctica and Africa, and a new model for the plate tectonic status of coastal Mozambique, the northernmost Natal Valley and the Mozambique Ridge. Together, chapters 3 to 6 constitute a new refit of Gondwanaland, which implies a dominant role of transform faulting in the development of the continental margins of south-eastern Africa. In chapter

7, I assess the structural style of the hinterland in terms of the new model.

Part 2 provides a framework within which the geologic history of the continental margin may be examined. In part 3 (chapters 8 - 10), I describe the in-fill of the northernmost Natal Valley sedimentary basin. In chapter 8, I present seismic profiles, isopachyte and structure contour charts in order to describe basin configuration and sediment distribution. In chapter 9, I demonstrate the influence of geostrophic currents on sediment dispersal. By assessing hiatus development mechanisms, the palaeo-oceanography of the Agulhas Current is outlined. A new theory is developed relating palaeo-climatic and ecological effects to changing current regimes. In chapter 10, a variety of mass-gravity-related features are mapped and their importance in the development of the continental margin is emphasised.

Part 4 of the thesis, chapter 11, summarises the main conclusions.

FIG. 2.1



20°
25°
30°
35°
40°
45°
15°
20°
25°
30°
35°
40°
45
AFRICA
MADAGASCAR
1000
2000
3000
2000
1000
NATAL VALLEY
RIDGE
MOZAMBIQUE
MOZAMBIQUE BASIN
MADAGASCAR RIDGE
2000
1000
3000
4000
TRANSKEI BASIN
AGULHAS PLATEAU
4000
3000
AGULHAS BASIN
5000
5000
4000
3000

CHAPTER 2

REGIONAL SETTING

2.1 INTRODUCTION

The Natal Valley is a deep submarine valley, lying in the south-west Indian Ocean. South of 30°S it is oriented NE-SW, while farther north it swings to N-S (Fig.2.1). To the west it is bounded by the continental margin of eastern South Africa and southern Mozambique, while to the east it is flanked by a large aseismic plateau, the Mozambique Ridge. To the north it merges with the coastal plain of south-east Mozambique. Deepening to the south, it reaches abyssal depths of 4 - 5000 m as it merges with the Transkei Basin. A description follows on the hinterland geology, bathymetry, climate and physical oceanography. South African geology is sketched very briefly because various age provinces and the tectonic boundaries between them are used in chapters 3 and 6 to test Gondwanaland reconstructions. In addition the onland extensions of the Natal Valley sedimentary basin are outlined. Bathymetry is unusual and is covered in detail. Climate is described so that a new palaeo-climatic theory may be developed in chapter 9. Physical oceanography is given attention because geostrophic currents are controlling factors in sediment dispersal and continental margin development.

2.2 HINTERLAND GEOLOGY

Figure 2.2 is a regional geological map of southern Africa (Brink, 1979). The oldest rocks of the Kaapvaal and Rhodesian cratons (the basement complex of Transvaal and Swaziland - Brink, 1979) date from 3600 m.y. (Truswell, 1970). Most of the rocks are gneissic or granitic while smaller areas of metavolcanics and metasediments occur (Swaziland Supergroup). Several successively younger basins developed on this cratonic area between 3000 m.y. and 1750 m.y. - the Witwatersrand and Pongola Basins, the Ventersdorp Supergroup, Transvaal Sequence and Waterberg Group. The Bushveld Complex was emplaced around 1950 m.y.

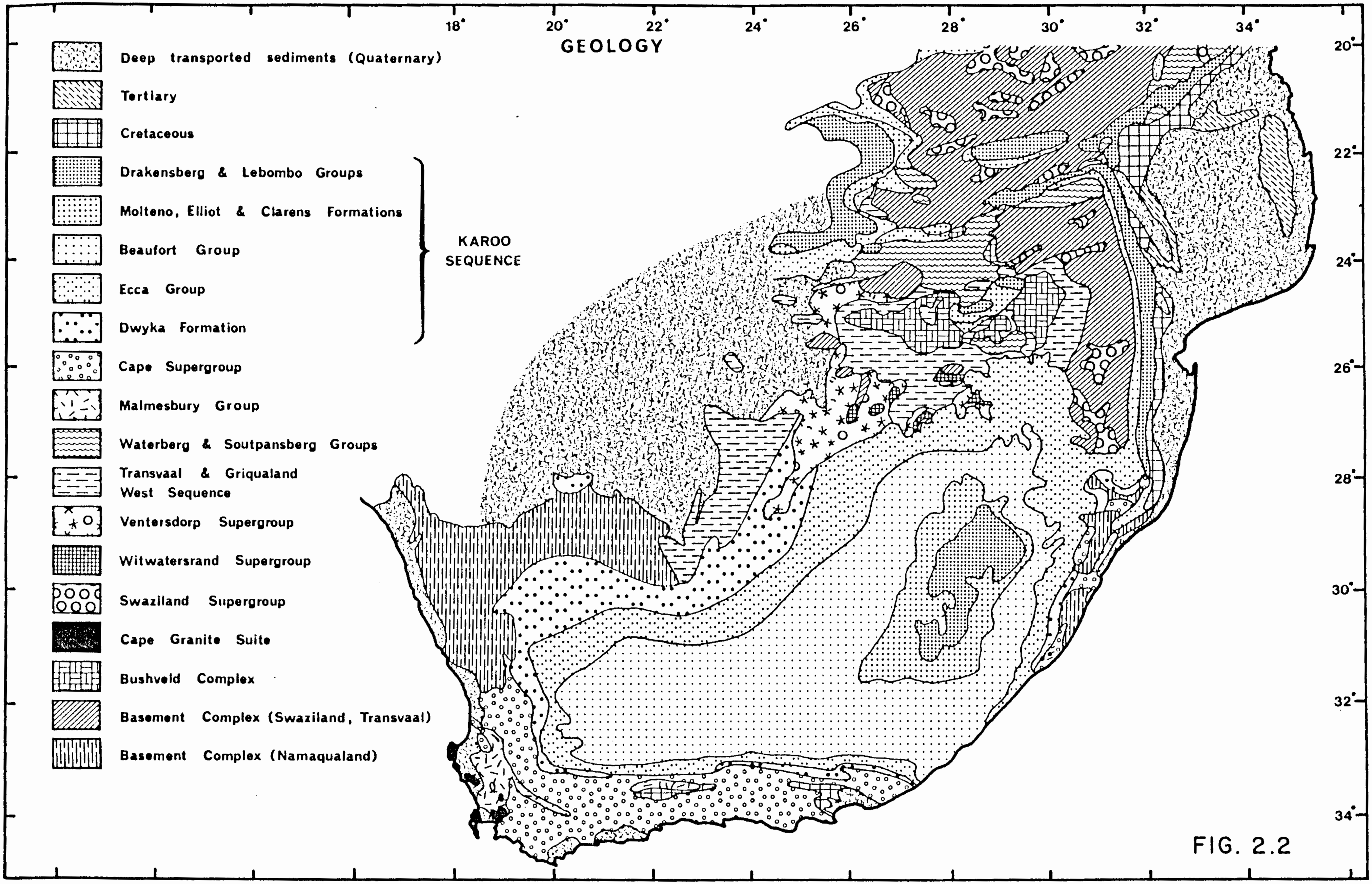

GEOLOGY
18°
20°
22°
24°
26°
28°
30°
32°
34°
Deep transported sediments (Quaternary)
Tertiary
Cretaceous
Drakensberg & Lebombo Groups
Molteno, Elliot & Clarens Formations
Beaufort Group
Ecca Group
Dwyka Formation
KAROO SEQUENCE
Cape Supergroup
Malmesbury Group
Waterberg & Soutpansberg Groups
Transvaal & Griqualand West Sequence
Ventersdorp Supergroup
Witwatersrand Supergroup
Swaziland Supergroup
Cape Granite Suite
Bushveld Complex
Basement Complex (Swaziland, Transvaal)
Basement Complex (Namaqualand)

FIG. 2.2

The $\sim$2700 m.y. old Limpopo mobile belt, runs east-north-east separating the Rhodesian and Kaapvaal cratons in the region of 22°S. At the southern end of the Kaapvaal craton the Namaqua-Natal mobile belt which is dated around 900 - 1100 m.y. outcrops in Natal and is thought to extend westwards under the Karoo basin to Namaqualand.

Sediments affected by the 550 m.y. Pan-African event are represented by the Malmesbury Group near Cape Town and its equivalents farther east. Similar age rocks exist in Namibia. The Malmesbury rocks are intruded by Cape Granites which yield ages between 610 and 505 m.y. (Allsopp & Kolbe, 1965; Schoch & Burger, 1976).

Large areas of southern Africa are covered by the Cape Supergroup and the Karoo Sequence. The Cape Supergroup ranges in age from Ordovician/Silurian to Carboniferous and outcrops in the Cape Fold Belt of the southern Cape and in an elongated area of coastal Natal. In the southern Cape, these rocks have been deformed into E-S-E trending basins and swells by the Cape orogeny of Triassic age (Newton, 1973; 1980). These curve more south-easterly offshore, and the basins were later in-filled by Jurassic-Recent sediments. On the east coast, the Cape Supergroup rocks (Natal Group) outcropping between Port St Johns and Zululand have been interpreted as in-fill of a trough or 'failed arm' (Hobday & Van Bruun, 1979).

The northern front of the Cape Fold Belt is best marked by the southern outcrop limit of the Dwyka Formation, the basal formation of the Karoo Sequence. Karoo rocks of Late Carboniferous to Jurassic age were laid down in two main basins - the Karoo Basin and the Zambezi Basin to the north. The basalts and rhyolites of the Drakensberg and Lebombo mountains form the top of the Karoo Sequence. The Lebombo mountains extend 700 km north from Zululand into southern Zimbabwe, where around 22°S, volcanic rocks extend north-eastwards and south-westwards.

The main watershed in South Africa lies along the escarpment of the Drakensberg mountains in Lesotho, continuing northwards through the Transvaal around longitude 30'E. Natal rivers are thus quite short and erode into a variety of rock-types. The Limpopo River crosses the Lebombo mountains at their northern end before flowing south across the Mozambique coastal plain.

The coastal plain of Mozambique and the Natal Valley are here considered to be post-break-up features. The Lebombo mountains and associated volcanic outcrops which range in age from 190 m.y. - 133 m.y. (Flores, 1973; Cleverly & Bristow, 1979) form the geographical boundary and the older age limits to the formation of this sedimentary basin. Initial sediments are continental, while marine and paralic rocks date from the Barremian to the Recent, interrupted by several important regional hiatuses (chapters 8 & 9).

2.3 BATHYMETRY

The south-west Indian Ocean is a complex ocean basin with various aseismic plateaux separating abyssal plains (Fig.2.1; Simpson, 1974; Fisher, 1982; Shackleton, 1982). The Natal Valley is a shoaler northerly continuation of the Transkei Basin (Figs.2.1 & 2.3), lying between south-east Africa and the Mozambique Ridge. In a preliminary report on the northern Natal Valley, Dingle et al. (1978) divided the area into eight physiographic provinces: continental shelf, Mozambique Ridge, deep ocean basin, Tugela Cone, Limpopo Cone, Central Terrace, Inharrime Terrace and Almirante Leite Bank. Within the northernmost Natal Valley (N.N.V.), the more detailed bathymetric chart presented here (Figs.2.4 & 2.5) encompasses five of these physiographic subprovinces - continental slope, Limpopo Cone, Inharrime Terrace, Central Terrace and Almirante Leite Bank. The maps and profiles (Figs.2.4 - 2.14) have been compiled from echograms collected on R.V. THOMAS B. DAVIE cruises 157, 246, 258, 267, 277, 350 and 371; on R.V. MEIRING NAUDÉ cruise 76/28; on R.V. AFRICANA II Walter's Shoal cruise, and on R.V. CHAIN cruise 99. Additional data sources were the South African Hydrographic Office sheet 373, the Portuguese Hydrographic Institute sheet 401 and Lady Glorita cruises N805-N806 sponsored by the Southern Oil Exploration Corporation. Satellite-navigation was used on THOMAS B. DAVIE cruises 350 and 371 and on R.V. MEIRING NAUDÉ cruise 76/28, and these provided an accurate frame-work of fair chart lines (Fig.2.4). Where bathymetric crossover discrepancies occurred, less accurately navigated cruise-tracks were adjusted to this framework. Isobaths have been drawn at 20 m intervals except in areas where either no accurately navigated lines were available, (in the east and north-east), or where the relief is too complicated to be resolved fully with the track coverage at hand (the eastern region of the Almirante Leite Bank).

Although the area lies on a passive continental margin, its physiography is unusual. Classic Atlantic-type rifted margins are linear, with a gently-sloping continental shelf separated from the abyssal ocean basins by a continental slope of varying gradient (1:5 - 1:25) (Heezen, 1974). The N.N.V. descends from the Mozambique coast step-wise to the ocean basin 450 km to the south. The region is an aseismic marginal plateau (c.f. west Australia - Veevers & Cotterill, 1978). Within the marginal plateau, physiographic features are distributed in a simple asymmetric pattern which comprises two main terraces: one lying in the north-western part of the basin is shallow (generally less than 800 m) and comprises the Limpopo Cone and the Inharrime Terrace, while the second, lying between 1300 m and 2000 m is formed by the Central Terrace and occupies the southern and eastern parts of the region. The two terraces are separated by a steep slope running diagonally across the area from NE to SW.

2.3.1 The Limpopo Cone

This is a smooth-surfaced feature stretching 300 km south from the Limpopo River mouth. Its upper reaches form a gently-sloping terrace between 400 and 800 m in depth, with a smooth-floored shallow valley on its western flank. The only irregularity on the surface of the cone is a shoal (55 m) at the base of the continental slope off the Limpopo River. Below 800 m the steep

FIGURE CAPTIONS

Fig.2.3. Bathymetry of the Natal Valley and Mozambique Ridge north of 31°S. 100 m intervals. Compiled by S.W. Goodlad and A.K. Martin: taken from Dingle et al. (1978).

Fig.2.4. Track-chart for detailed bathymetry north of 29°S.

Fig.2.5. First bathymetric map of part of the southern African margin at 20 m intervals. Satellite-navigated tracks were used as a framework for correcting cross-over discrepancies involving less accurately navigated tracks. Contour interval of 20 m was discontinued either where topography was too complex (Almirante Leite Bank) or where track coverage did not justify the more detailed treatment.

Figs.2.6-2.14. Tracks and profiles reduced from echo-sounder traces for the region north of 29°S. These are intended to help elucidate the physiography and microtopography of the area. Profiles for the Almirante Leite Bank are shown in more detail in Figures 2.12-2.14.

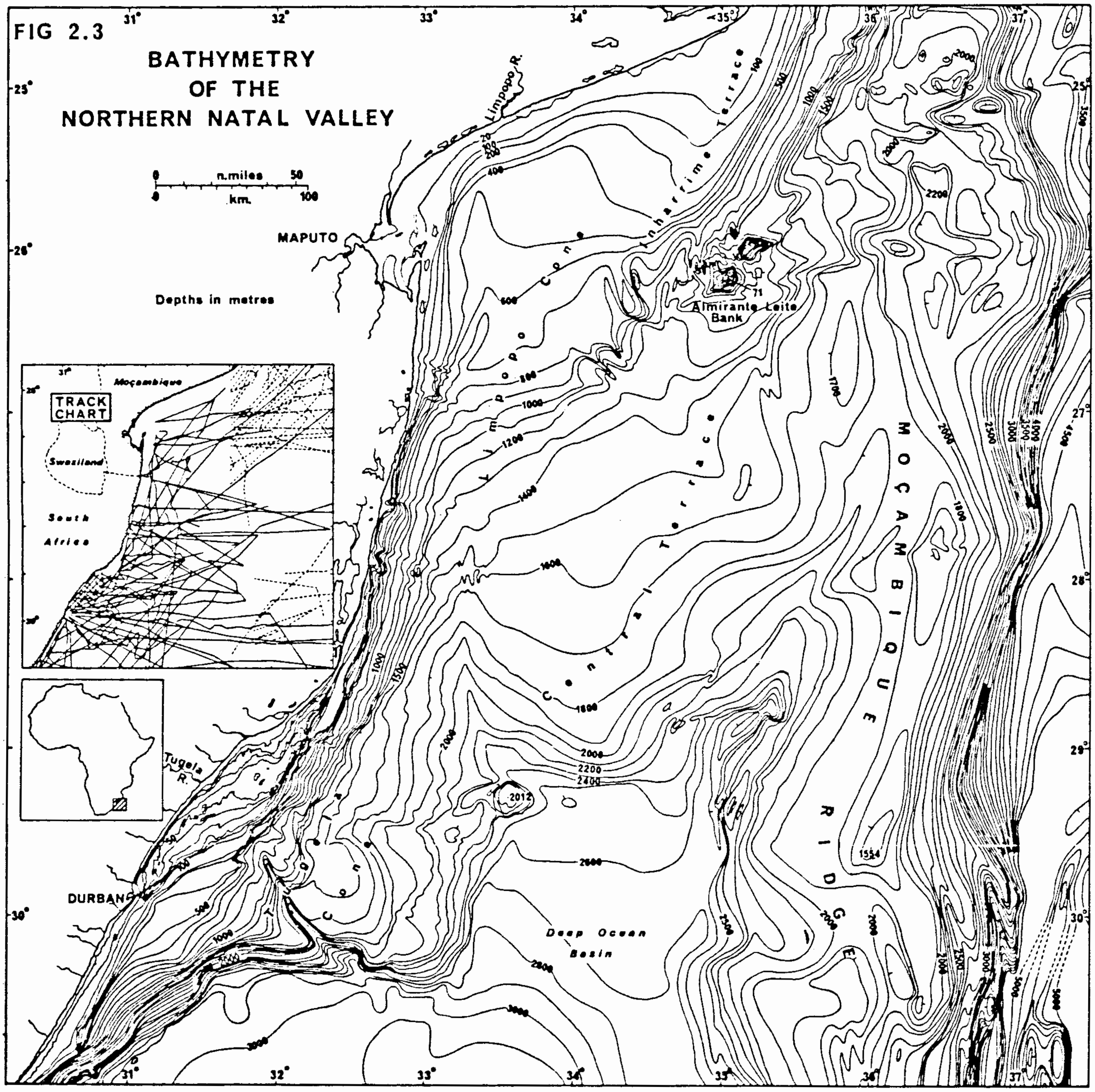


FIG 2.3
BATHYMETRY
OF THE
NORTHERN NATAL VALLEY
n.miles
km.
MAPUTO
Depths in metres
TRACK
CHART
Moçambique
Swaziland
South
Africa
Limpopo R.
Inharrime Terrace
Limpopo Cone
Almirante Leite
Bank
Central Terrace
MOÇAMBIQUE RIDGE
Tugela R.
Tugela Cone
DURBAN
Deep Ocean
Basin

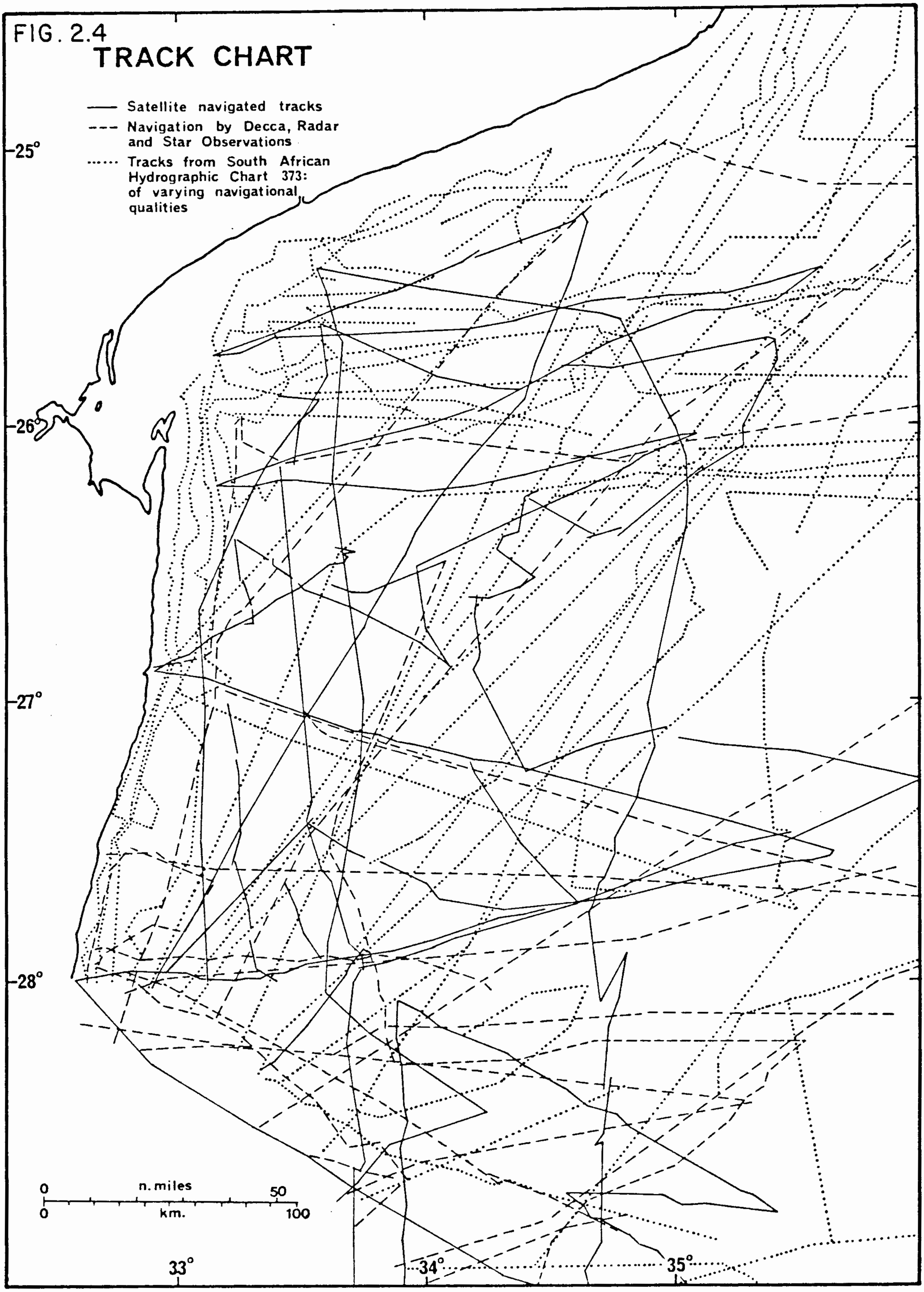


FIG. 2.4 TRACK CHART

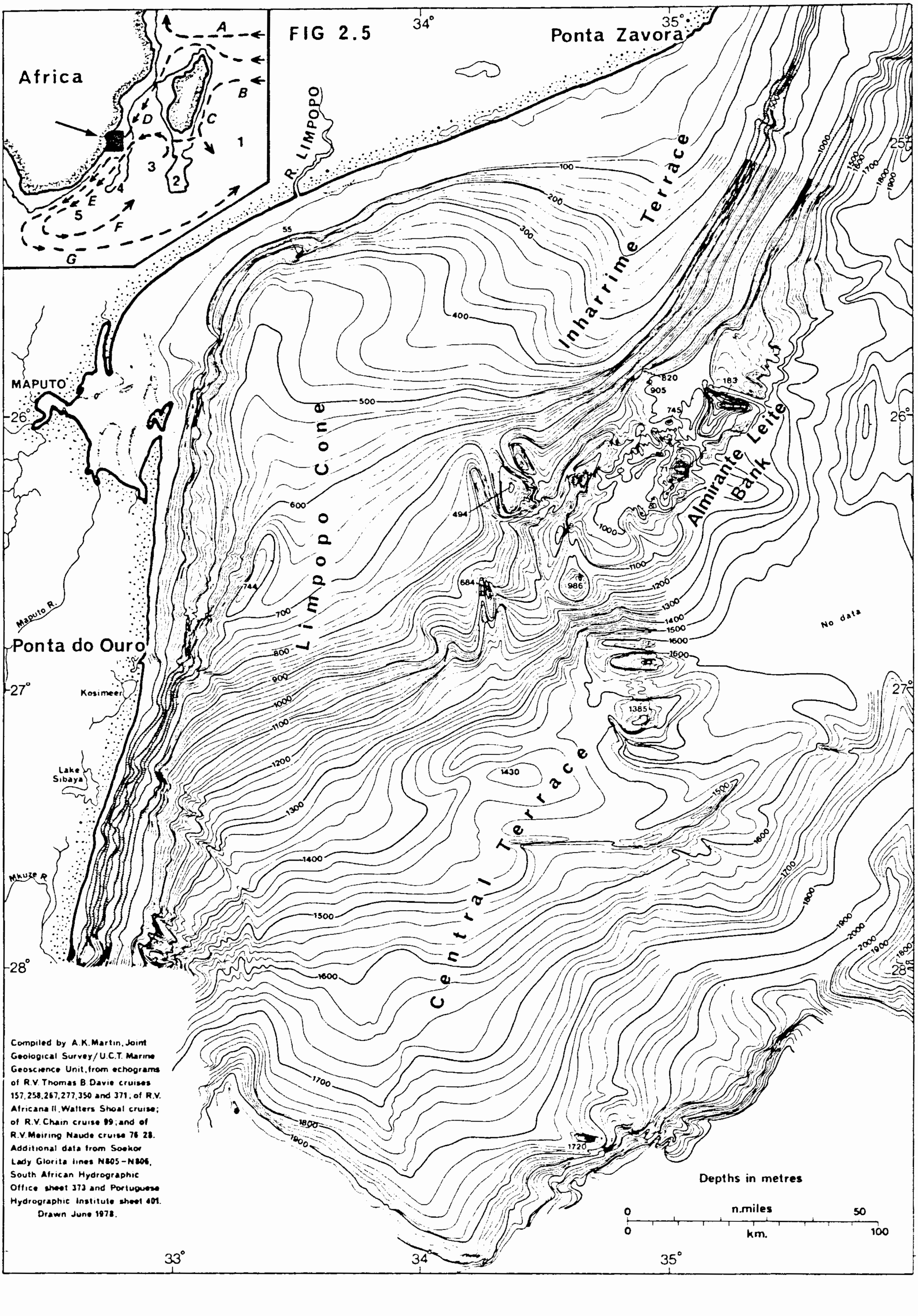


FIG 2.5
Africa
A
B
C
D
E
F
G
1
2
3
4
5
Ponta Zavora
R. LIMPOPO
MAPUTO
Maputo R.
Ponta do Ouro
Kosimeer
Lake Sibaya
Mkuze R.
Limpopo Cone
Inharrime Terrace
Almirante Leite Bank
Central Terrace
No data
33°
34°
35°
26°
27°
28°
Compiled by A.K.Martin, Joint Geological Survey/U.C.T. Marine Geoscience Unit, from echograms of R.V. Thomas B Davie cruises 157, 258, 267, 277, 350 and 371; of R.V. Africana II, Walters Shoal cruise; of R.V. Chain cruise 99; and of R.V. Meiring Naude cruise 76 28. Additional data from Soekor Lady Glorita lines N805–N806, South African Hydrographic Office sheet 373 and Portuguese Hydrographic Institute sheet 401. Drawn June 1978.
Depths in metres
n.miles
0
50
km.
0
100

FIG. 2.6

TRACKS OF PROFILES TRENDING GENERALLY EAST-WEST

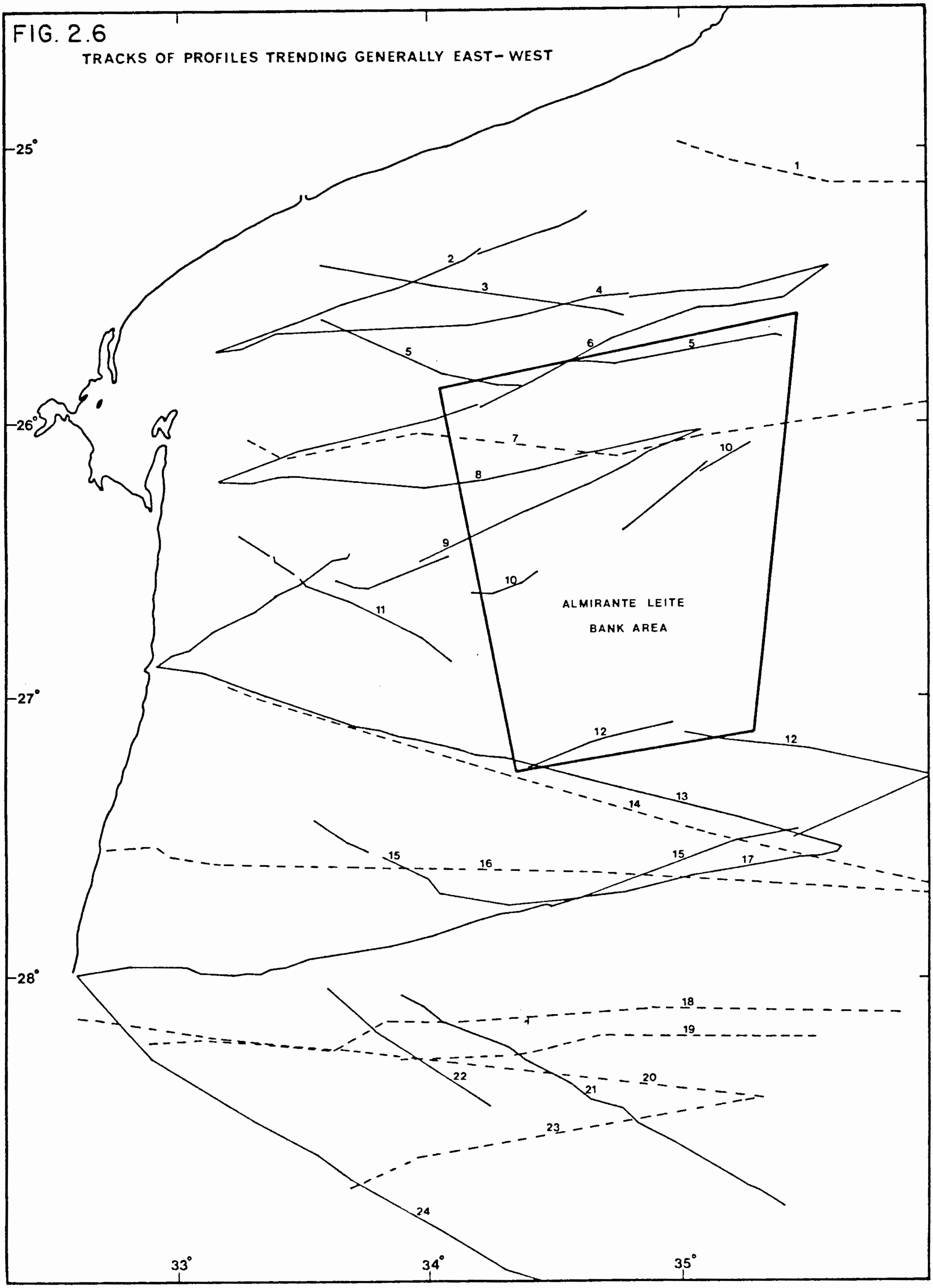
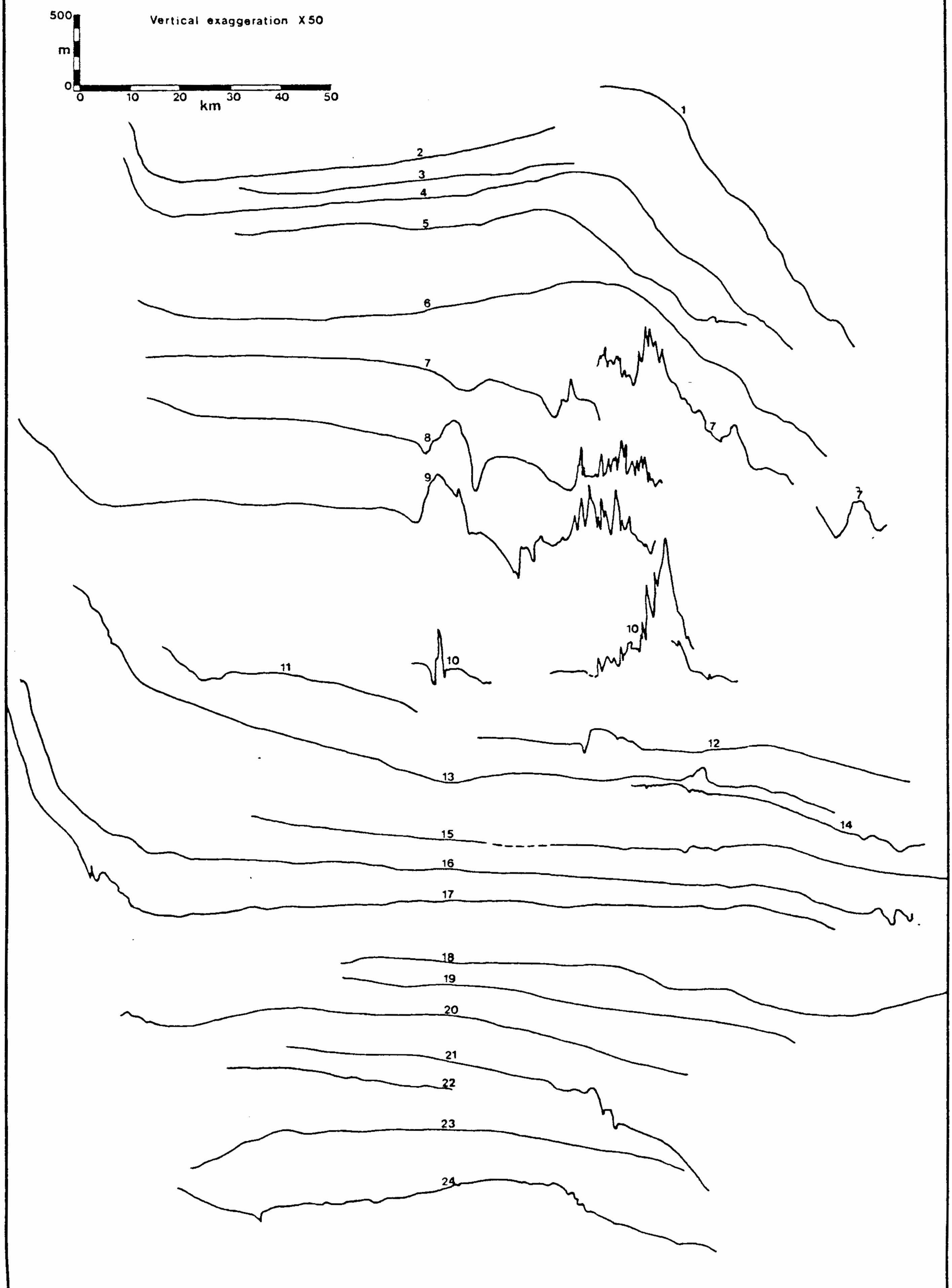


FIG. 2.7

BATHYMETRIC PROFILES TRENDING GENERALLY EAST-WEST



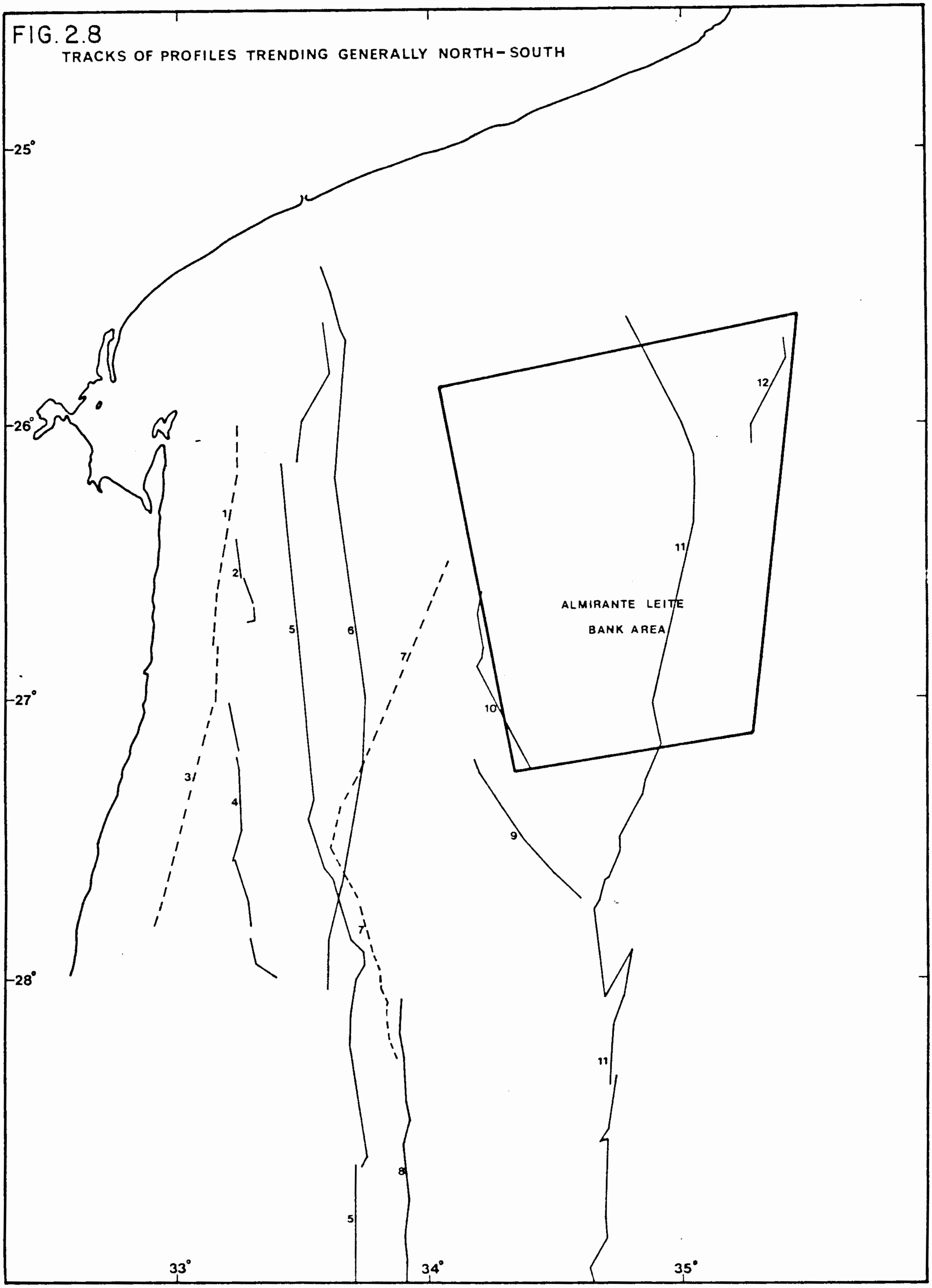


FIG. 2.8
TRACKS OF PROFILES TRENDING GENERALLY NORTH-SOUTH
-25°
-26°
-27°
-28°
33°
34°
35°
1
2
3
4
5
6
7
7
8
5
9
10
11
11
12
ALMIRANTE LEITE
BANK AREA

FIG. 2.9

BATHYMETRIC PROFILES TRENDING GENERALLY NORTH-SOUTH

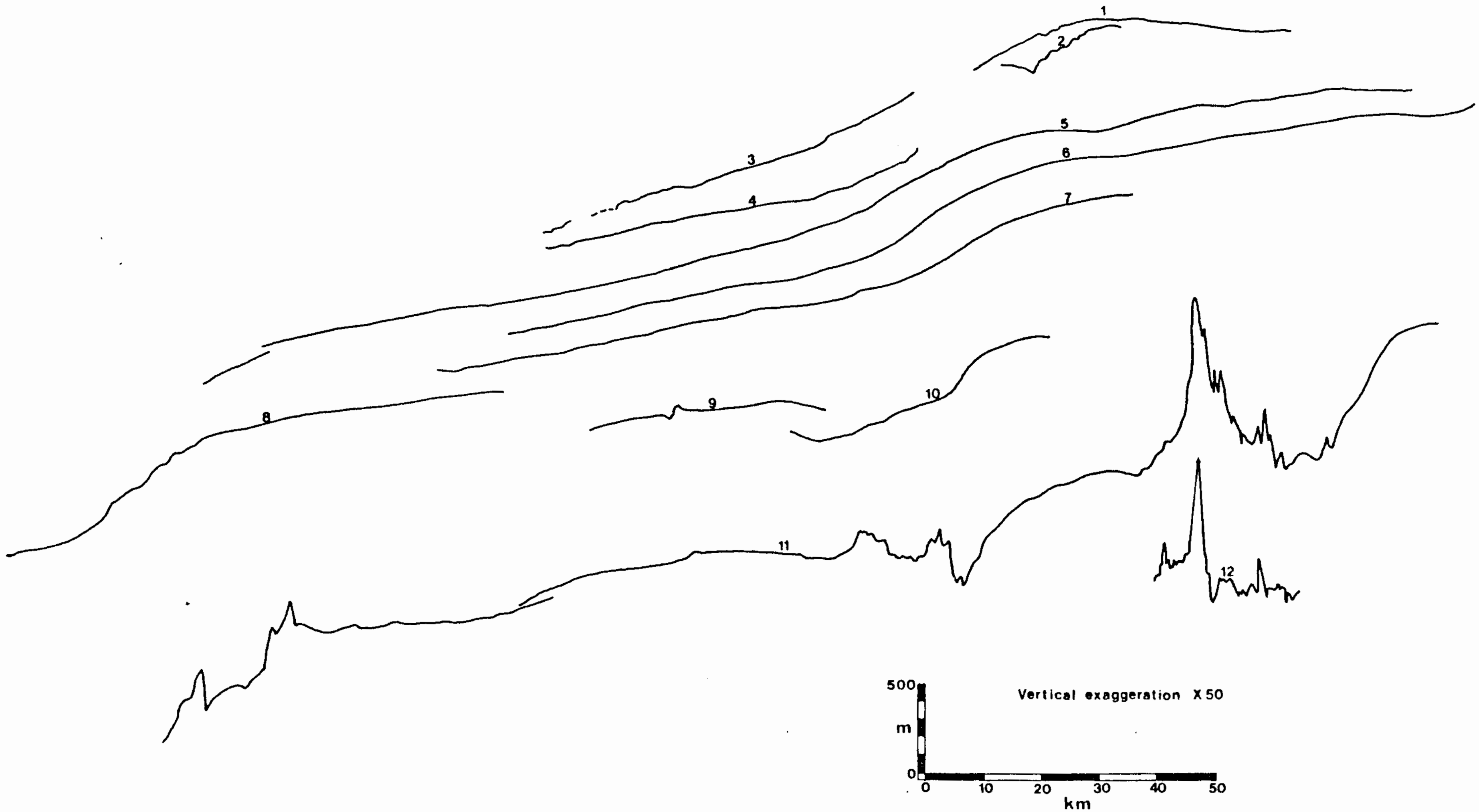


FIG. 2.10

TRACKS OF PROFILES TRENDING GENERALLY NORTH-EAST - SOUTH-WEST

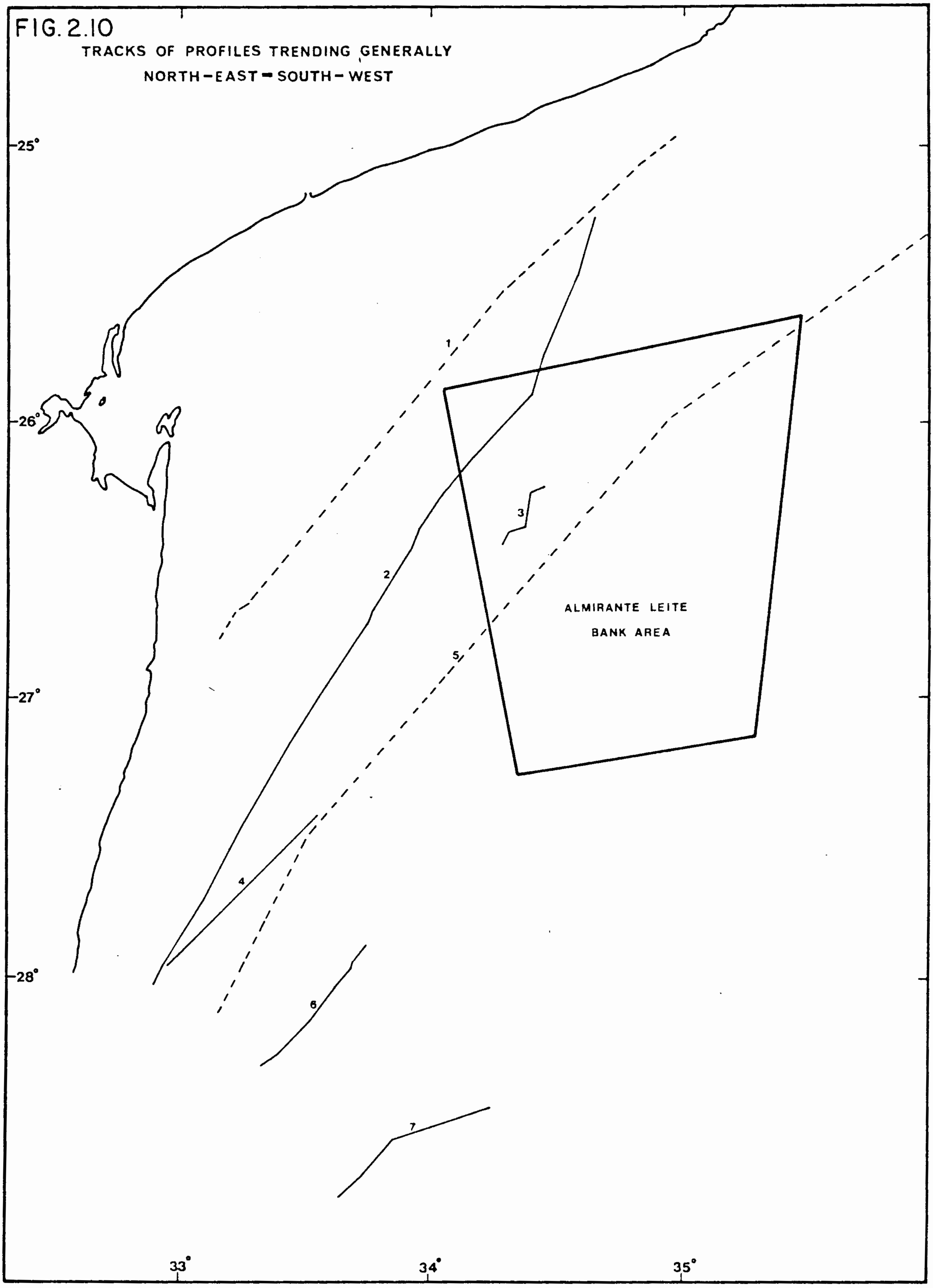


FIG. 2.11

BATHYMETRIC PROFILES TRENDING GENERALLY NORTH-EAST - SOUTH-WEST

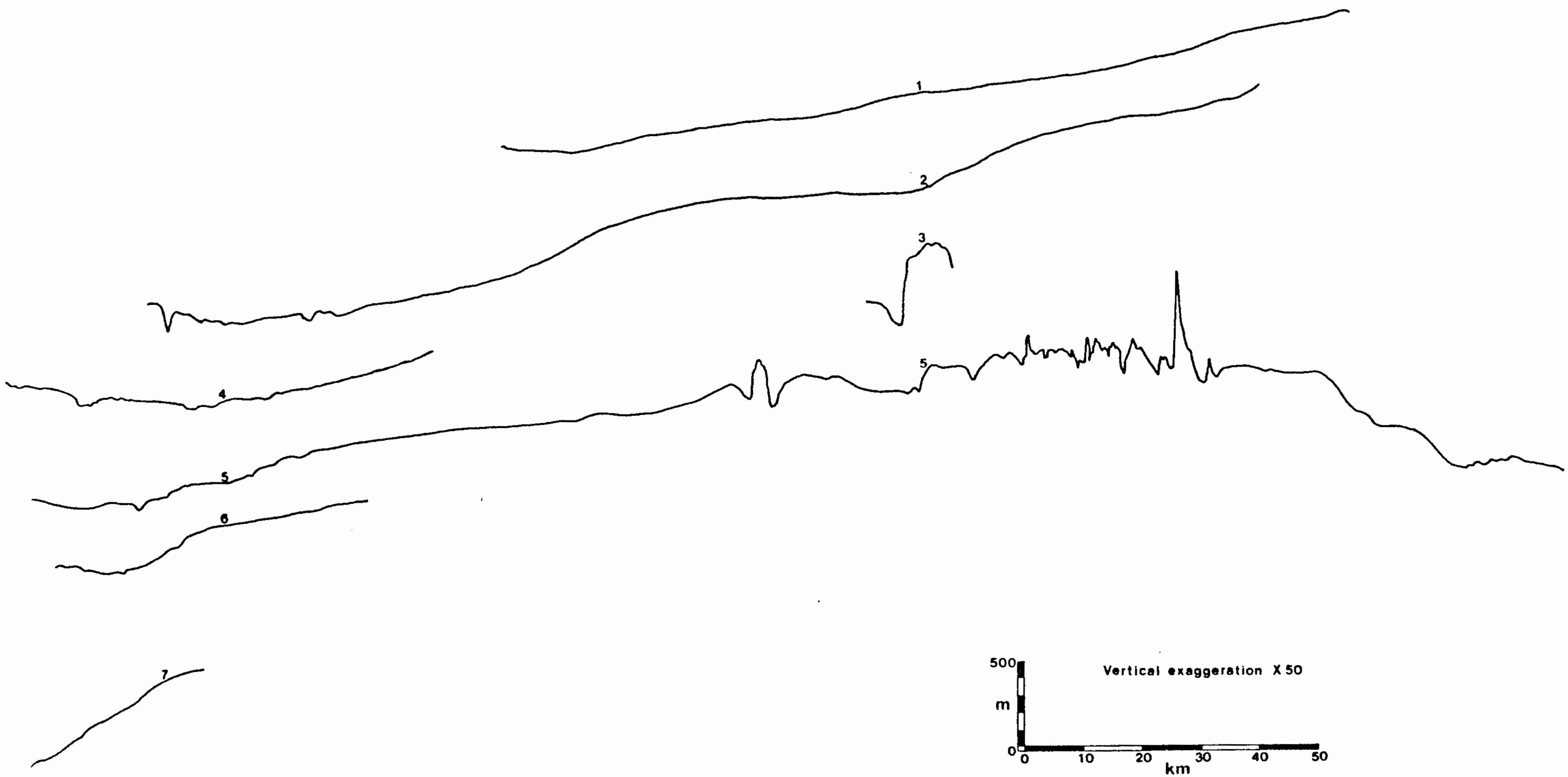
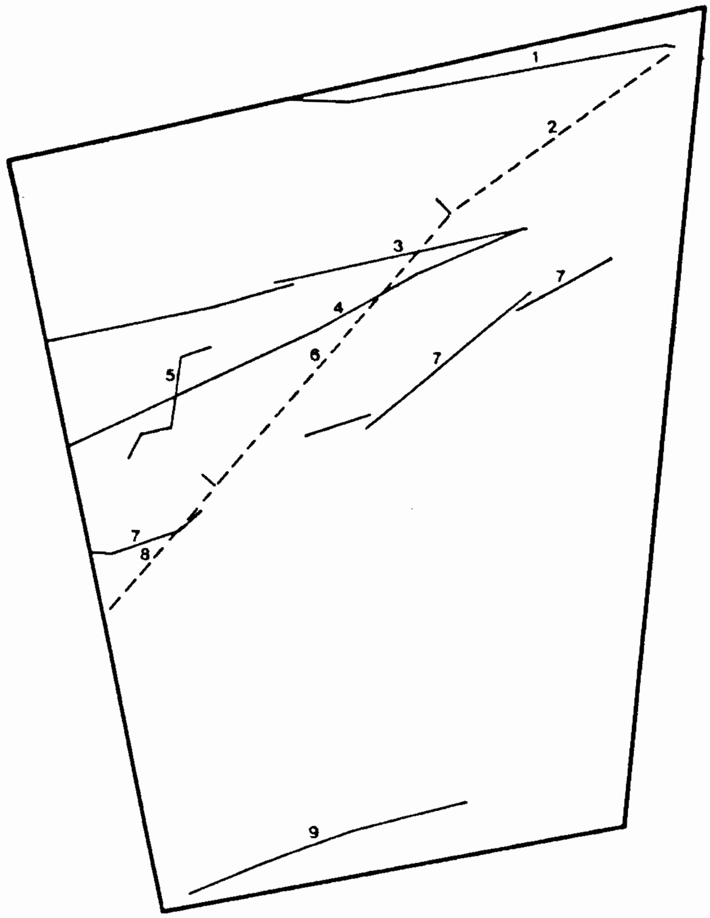


FIG. 2.12

ALMIRANTE LEITE BANK AREA

TRACKS OF PROFILES TRENDING GENERALLY EAST – WEST



TRACKS OF PROFILES TRENDING GENERALLY NORTH – SOUTH

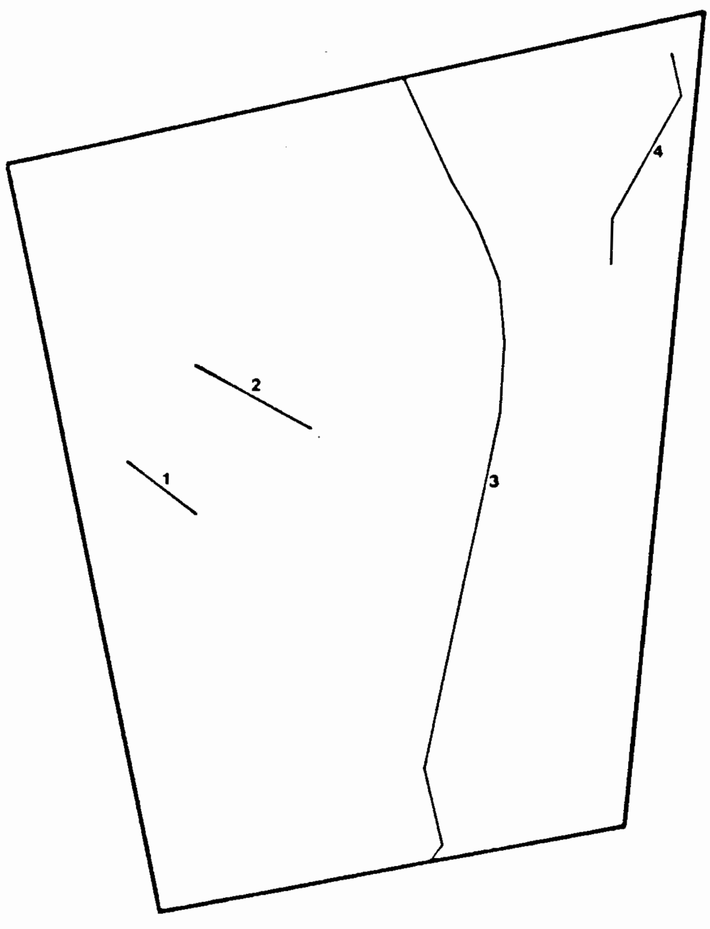


FIG. 2.13

BATHYMETRIC PROFILES TRENDING GENERALLY EAST-WEST — DETAIL ACROSS ALMIRANTE LEITE BANK

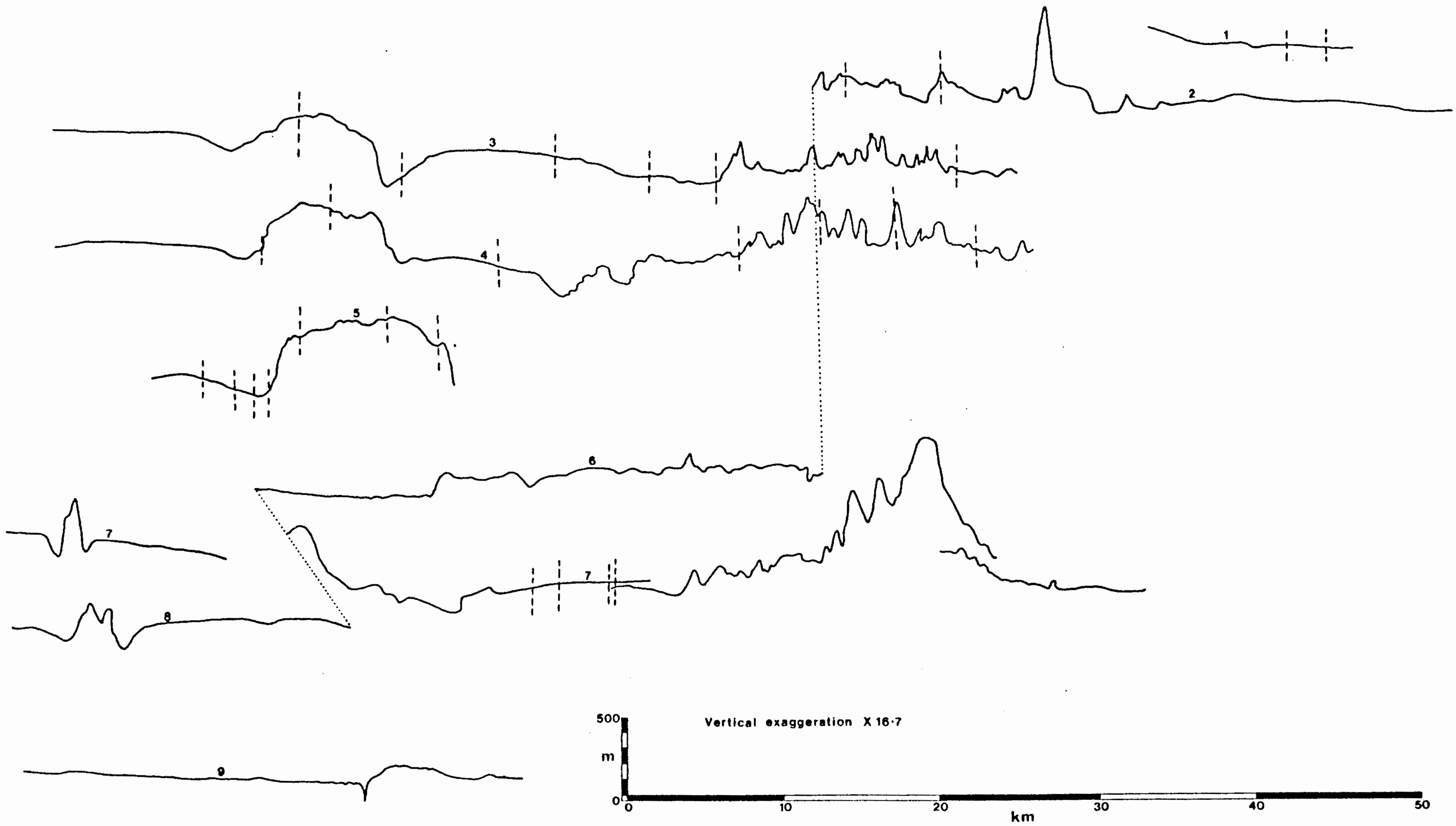
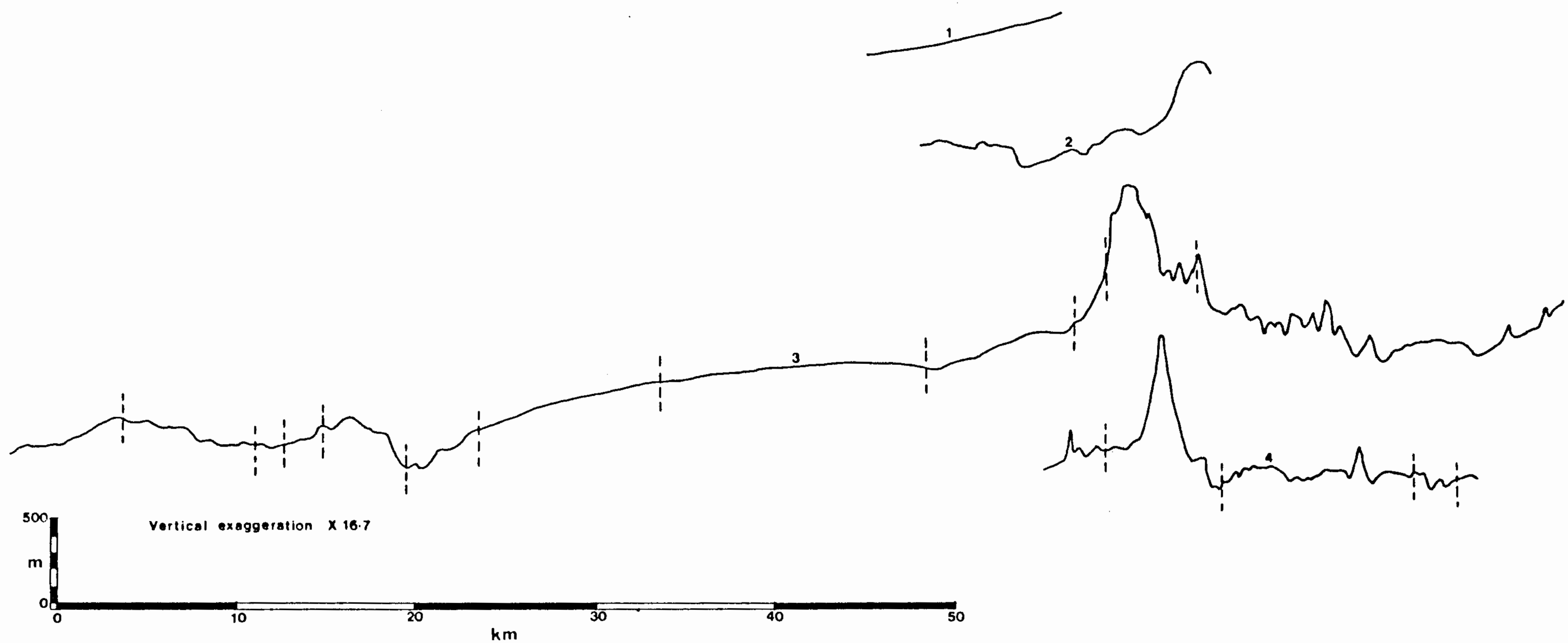


FIG. 2.14

BATHYMETRIC PROFILES TRENDING GENERALLY NORTH-SOUTH
DETAIL ACROSS ALMIRANTE LEITE BANK



southern face of the cone (Table 2.1) dips towards the Central Terrace.

2.3.2 The Inharrime Terrace

To the east, the Limpopo Cone merges with the Inharrime Terrace - a smooth, convex, curvilinear sedimentary feature extending 140 km south-west from Ponta Zavora. Its eastern slope descends towards the northern end of the Mozambique Ridge in two steps separated by a shoulder at 850 - 1000 m. A secondary bathymetric lobe of the Inharrime Terrace curves towards Maputo from its western flank and is related to changes in sedimentation patterns and the stratigraphy of the area (chapter 8). Together the Limpopo Cone and the Inharrime Terrace form a large triangular sediment wedge in-filling the coastal bight between Ponta Zavora and Ponta do Ouro.

2.3.3 The Almirante Leite Bank

The rough topography of the Almirante Leite Bank breaks the otherwise smooth outer slopes of the Limpopo Cone and the Inharrime Terrace. This region comprises over 80 rocky, volcanic seamounts spread out over 130 km in a NE-SW trend. The seamounts are relatively steep-sided (Table 2.1) with a maximum relief of 1060 m. The area is characterised by channels, asymmetric sediment mounds and moats on the seamount flanks. The most spectacular of these is a 10 km wide 255 m deep moat on the north-eastern side of the most north-westerly seamount.

2.3.4 The Central Terrace

To the south of the Almirante Leite Bank and the Limpopo Cone, the second main terrace in the area - the Central Terrace - slopes gently down from 1300/1400 to 2000 m. Its southern extension is flanked to west and east by prominent valleys separating it from the continental slope and the Mozambique Ridge, respectively. Most of the terrace is smooth-topped and gently sloping and has 2 convex axes curving towards the south-east. Centrally situated on the northern half of the terrace lies a smooth ovate mound (shoalest point - 1430 m) which is almost completely surrounded by a very shallow wide moat.

Table 2.1

Quantitative Bathymetric Data

	CONTINENTAL SLOPE			SHELF WIDTH (km)
	Max. Relief (m)	Slope	Max. slope	
28°S	1500	1:26 (2.2°)	1:10 (5.7°)	3
PONTA DO OURO	800	1:16 (3.6°)		4
MAPUTO	400	1:23 (2.5°)	1:10 (5.7°)	45
LIMPOPO RIVER	350	1:31 (1.9°)		15
SOUTH OF PONTA ZAVORA	500	1:30 (1.9°)		120
EAST OF PONTA ZAVORA	1900	1:37 (1.5°)	1:12.5 (4.6°)	30
WORLD AVERAGE	4000	1:13 (4.28°)		74

	Max. Relief	Slope	Max. slope (wall)
CANYONS	315		1:3 (17°)
SEAMOUNTS	1060		1:4 (13.7°)
LIMPOPO CONE		1:357 (0.16°)	1:112 (0.5°) (southern face)
CENTRAL TERRACE		1:300 (0.19°)	

To the east of this three rocky outcrops of older sedimentary units (chapter 8) mar the smooth topography. One of these is an arcuate ridge, 130 km long and up to 135 m high. On the upper slopes of the valleys flanking the Central Terrace rocky outcrops of older sedimentary units again lead to broken topography. To the south-east these outcrops are also flanked by scour-moats.

2.3.5 Continental shelf and slope

Between 28°S and Maputo the continental shelf is steep (Table 2.1) and very narrow (2-7 km wide), contrasting markedly with world averages (Shepard, 1963). North and east of Maputo, the shelf broadens, bulging seaward off Ponta Zavora as an expression of the Inharrime Terrace. Because of the sedimentary in-fill of the Limpopo Cone, the continental slope near Maputo descends to depths of only 350 m. To the south, the slope descends to progressively greater depths. Slopes vary from 1:10 to 1:26, with the steepest slopes coinciding with slump scars (chapters 8 & 10). Southwards of 27°S the slope is dissected by a series of canyons which appear to coalesce in a dendritic pattern. The canyons are narrow, V-shaped and steep-sided with relief decreasing seaward from 315 m to ∿40 m where the canyons coalesce.

2.3.6 Microtopography

Bathymetric profiles (Figs.2.6 - 2.14) indicate that over much of the N.N.V. the seafloor is relatively smooth and gently-sloping. Three areas of distinctive microtopography occur: a) the continental slope, b) the Almirante Leite Bank, c) rocky outcrops on the Central Terrace. Rough topography on the slope is related to basinward sediment movement via slumping and canyon-fan systems (chapters 8 & 10). Steep slopes coincide with main and subsidiary glide-plane scars, fissures or canyon systems. On the Almirante Leite Bank, rough relief is due to the many volcanic seamounts and the asymmetric sediment mounds and channels on their flanks (chapter 8). Outcrops of older strata on the Central Terrace also generate rough relief. Some of these outcrops are associated with slumps and upfaulted blocks, while others are related to current-moulding and scouring.

2.4 CLIMATE

Over much of southern Africa, the climate is sub-tropical with a summer wet season, the exception being the south-western Cape which has a Mediterranean climate with a winter wet season. The Natal Valley and its hinterland, lying between 25°S and 35°S are sub-tropical. Heating in the tropics leads to up-welling air, and inflow of air at ground level forming the inter-tropical Convergence Zone. To compensate for upwelling air in the tropics one would expect down-welling air in polar regions. The spin of the earth complicates this picture, breaking the meridional air movement into three cells. Prior to Coriolis' work, these spin effects were recognised by Hadley, and the meridional circulation patterns are termed "Hadley cells". The down-welling limb of the tropical cell results in the belt of sub-tropical high pressure around 25°S (Tyson, 1969, Fig.2; Nicholson & Flohn, 1980, Fig.6). Two main high pressure anticyclones occur - one centred in the South Atlantic seaward of the west coast, and one centred over eastern Transvaal, southern Mozambique, and the adjacent south-west Indian Ocean. Weather results from alternative dominance of this system and east-moving cyclones from the circumpolar westerlies (Tyson, 1969; Jackson & Tyson, 1971; Grindley, 1979; Heydorn & Tinley, 1980). Typical synoptic charts are shown in Figure 2.15. When the south-west Indian Ocean high pressure cell is centred over the land-mass of southern Africa, the down-welling air results in a prolonged dry spell or even drought (Fig.2.15a). If the high pressure cell lies predominantly over the ocean, cool moist air is advected in towards Zimbabwe and Botswana (Fig.2.15b). This results in showery weather in the interior, and warm 'Berg Winds' on the west coast. Much of the rainfall in the interior and even in Namibia is derived from Indian Ocean air advected in this way. Because the anticyclonic high pressure cells are weaker in summer, inflow of maritime air is more prevalent, and rains fall mainly in summer (Fig.2.16). Conversely, the south-western Cape receives its rain when

FIGURE CAPTIONS

Fig.2.15. Typical synoptic weather charts for southern Africa (after Heydorn & Tinley, 1980) 2.15a shows high pressure over the continent. Down-welling air within this anti-cyclone leads to dry weather or even drought. When the high pressure is displaced to the south-east it brings moist air from the Indian Ocean to the interior of southern Africa causing rain (2.15b). The passage of a cyclone which brings rain to the southern Cape is shown in 2.15c and d.

Fig.2.16. When down-welling air of an anti-cyclone dominates conditions over the continent, a temperature inversion occurs over coastal and adjacent areas. In winter the base of the inversion is usually below the level of the escarpment of the Drakensberg Mountains, and westward transport of moisture from the Indian Ocean is inhibited. In summer, the anti-cyclonic high pressure cell is displaced (Fig.2.15b) and moist air is transported over the escarpment, leading to rain in-land (after Tyson, 1969). These features of the climate are outlined so that the palaeo-climatic effect of changes in the Agulhas Current may be discussed in section 9.5.2.

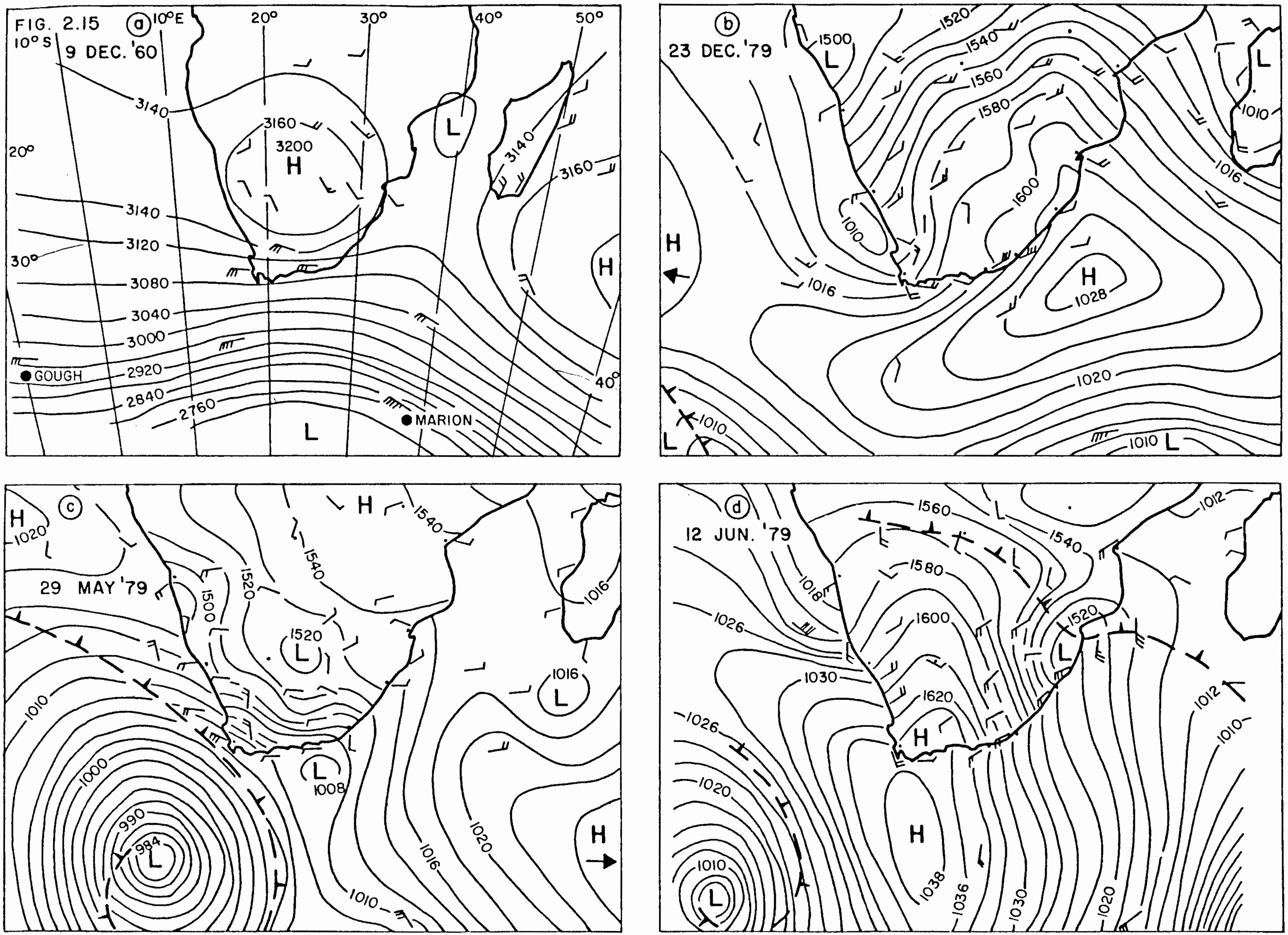


FIG. 2.15
a
9 DEC. '60
10° S
10°E
20°
30°
40°
50°
20°
30°
40°
3140
3160
3200
H
L
3140
3160
H
3140
3120
3080
3040
3000
2920
2840
2760
GOUGH
MARION
L
b
23 DEC. '79
1500
L
1520
1540
1560
1580
1600
1010
L
1010
1016
H
1010
1016
H
1028
1020
1010
L
1010 L
c
29 MAY '79
H
1020
H
1540
1520
1540
1500
1520
L
1016
1016
L
1010
1000
990
984
L
L
1008
1016
1020
1010
H
d
12 JUN. '79
1560
1540
1012
1018
1580
1600
1520
L
1026
1030
1620
H
1012
1010
1026
1020
H
1010
L
1038
1036
1030
1020

cyclones of the circumpolar westerlies impinge on southern Africa in winter months (Fig.2.15c). Heavy rains occur in the Cape with snowfalls in the Cape and Drakensberg mountains resulting in high run-off in Natal and Cape rivers. As a cyclone passes, the south Atlantic high pressure penetrates polewards and advects very cold air over the eastern subcontinent resulting in more snow in mountainous areas (Figs.2.15d). Because rain in sub-tropical areas comes predominantly from the south-west Indian Ocean, isohyets are aligned roughly north-south with annual rainfall receipts exceeding 800 mm over much of the east coast and Drakensberg mountains. Rainfall decreases towards the west, culminating in the Kalahari and Namib deserts of Botswana and Namibia. Thus, although river discharge to the sea is higher for the east coast than the west coast, in a global context run-off from the east coast is moderate (Milliman & Meade, 1983).

2.5 PHYSICAL OCEANOGRAPHY

From bathymetric evidence (section 2.3), as well as seismic and sedimentological evidence (chapters 8 & 9) it has been recognised that sediment distribution in the N.N.V. is profoundly influenced by geostrophic currents. Water circulation in the South West Indian Ocean, and in the Agulhas Current in particular, have been discussed by several authors using traditional hydrographic data (see reviews of Lutjeharms, 1971; 1972; Van Foreest; 1977). More recently, the use of satellite-tracked buoys (Stavropoulos & Duncan, 1974; Grundlingh, 1977; Grundlingh & Lutjeharms, 1979; Lutjeharms et al., 1981) satellite imagery (Harris & Van Foreest, 1977; Harris et al., 1978; Malan & Schumann, 1979; Lutjeharms & Valentine, 1981), and inertial jet modelling (Van Foreest, 1977) have confirmed the general flow-paths predicted by earlier work, and thrown light on many of the complexities (Fig.2.17).

The South Indian Ocean subtropical gyre comprises westerly flow of the South Equatorial Current, southerly flow of the Western Boundary Current system (including the Agulhas Current), easterly flow at the subtropical convergence, and northerly flow in a broad zone off Australia. Madagascar, the

FIGURE CAPTIONS

Fig.2.17. Circulation patterns in the South West Indian Ocean. Surface/near-surface currents = black arrows. (1) North Equatorial Current (part of the North Indian Ocean sub-tropical gyre). (2) South Equatorial Current. (3) East Madagascar Current. (4) Mozambique Current. (5) Agulhas Return Current. (6) West Wind drift at sub-tropical convergence (all sub-units of the South Indian Ocean sub-tropical gyre). Note the two flow-paths Route A/Route B of the Agulhas Current, as well as the re-cycling of Agulhas Return Current water in the Transkei and Mozambique Basins and west of the South-West Indian Ocean Ridge. Open arrows indicate cold under-currents of Antarctic Bottom Water in abyssal areas. 2000 m and 4000 m isobaths outline the main bathymetric features.

Fig.2.18. A compilation of water circulation patterns in the northernmost Natal Valley (after references in section 2.5). Hydrographic data are superimposed on to a 100 m interval isobath chart. Water movement is deduced from hydrographic profiles 1-6, inertial jet modelling, and satellite-tracked buoys. A simplified version showing the main flow-paths is given in Fig.9.1.

FIG 2.17

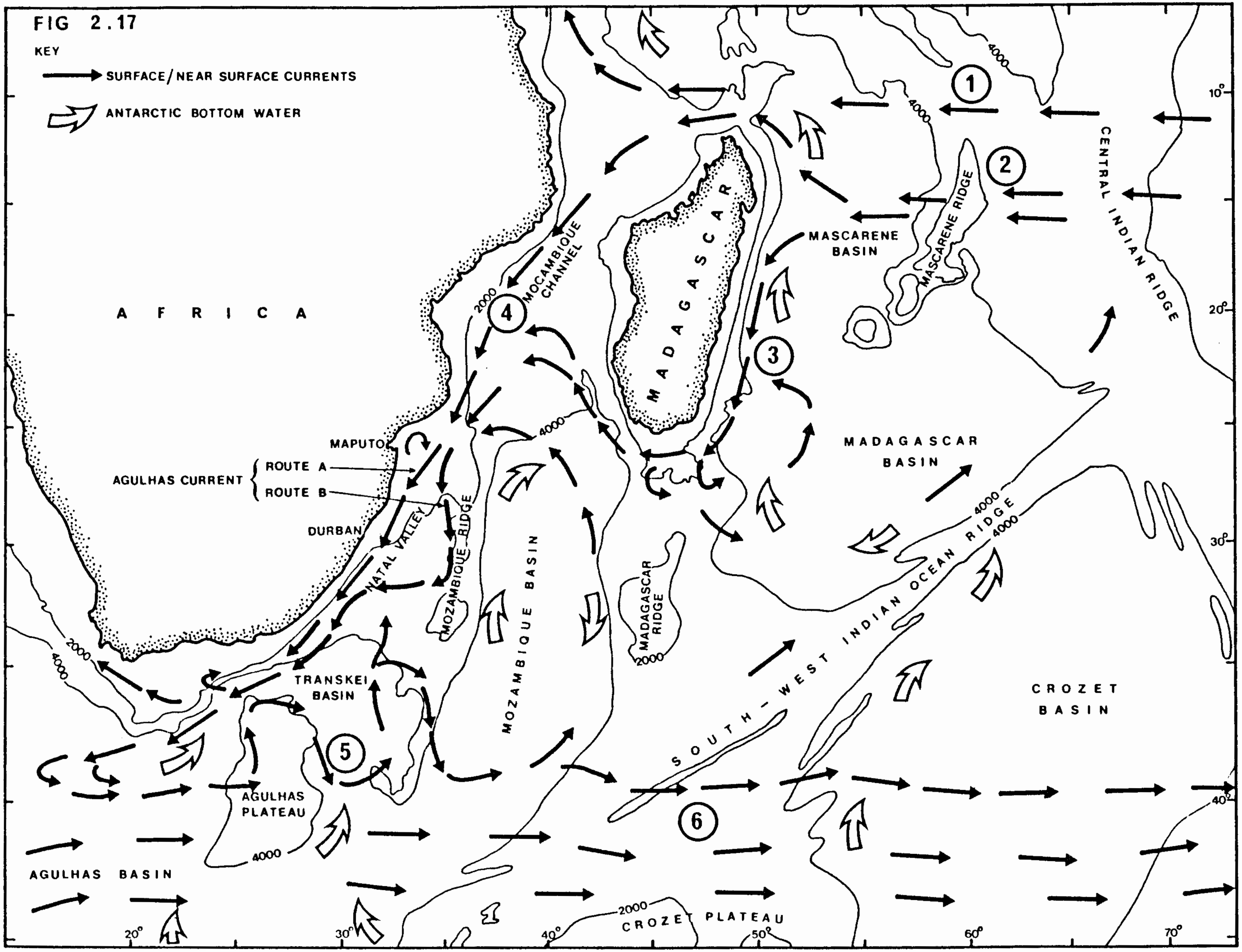
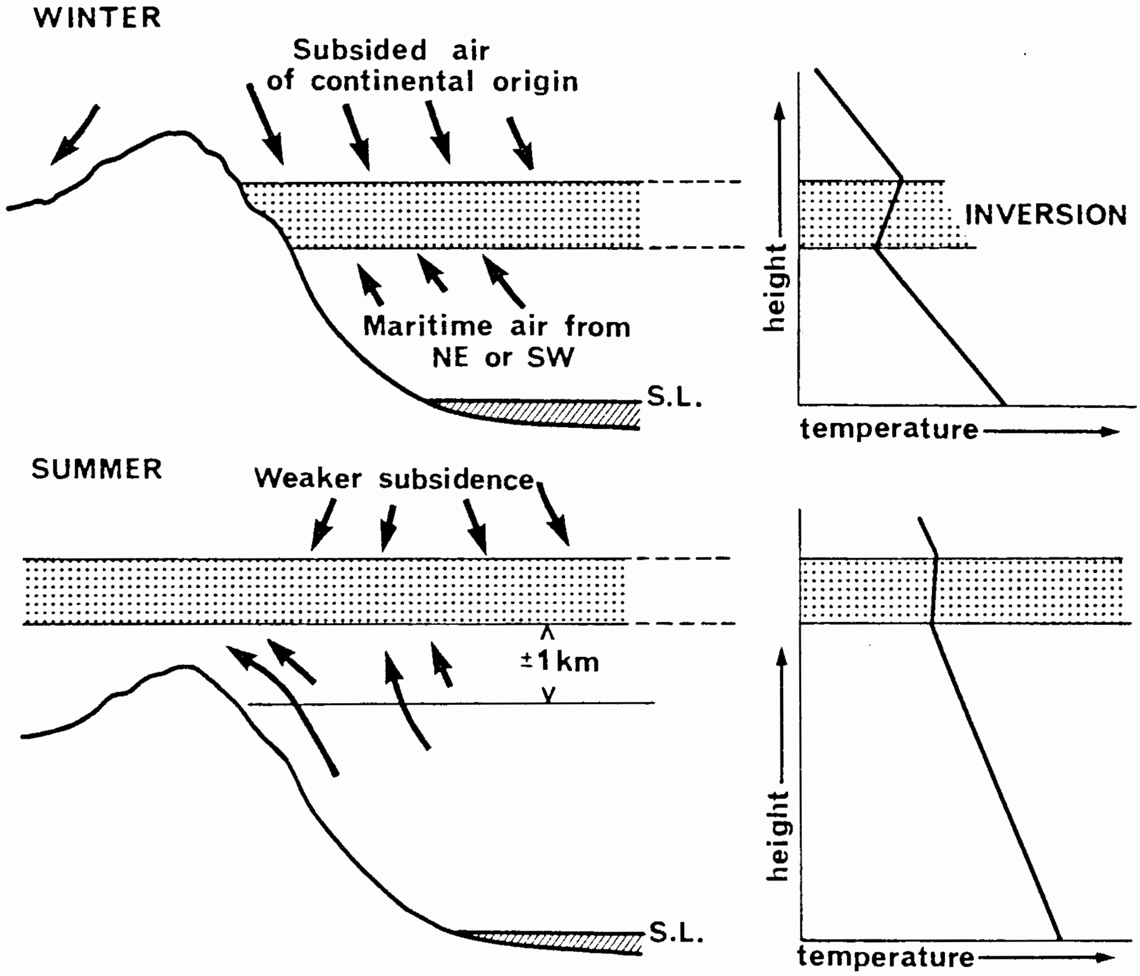

KEY
SURFACE/NEAR SURFACE CURRENTS
ANTARCTIC BOTTOM WATER
AFRICA
MADAGASCAR
MOCAMBIQUE CHANNEL
MASCARENE BASIN
MASCARENE RIDGE
CENTRAL INDIAN RIDGE
MADAGASCAR BASIN
MADAGASCAR RIDGE
MOZAMBIQUE BASIN
MOZAMBIQUE RIDGE
NATAL VALLEY
TRANSKEI BASIN
AGULHAS PLATEAU
AGULHAS BASIN
SOUTH-WEST INDIAN OCEAN RIDGE
CROZET BASIN
CROZET PLATEAU
MAPUTO
DURBAN
AGULHAS CURRENT
ROUTE A
ROUTE B
1
2
3
4
5
6
2000
4000
10°
20°
30°
40°
50°
60°
70°

FIG. 2.16


WINTER
Subsided air
of continental origin
INVERSION
height
Maritime air from
NE or SW
S.L.
temperature
SUMMER
Weaker subsidence
±1 km
height
S.L.
temperature

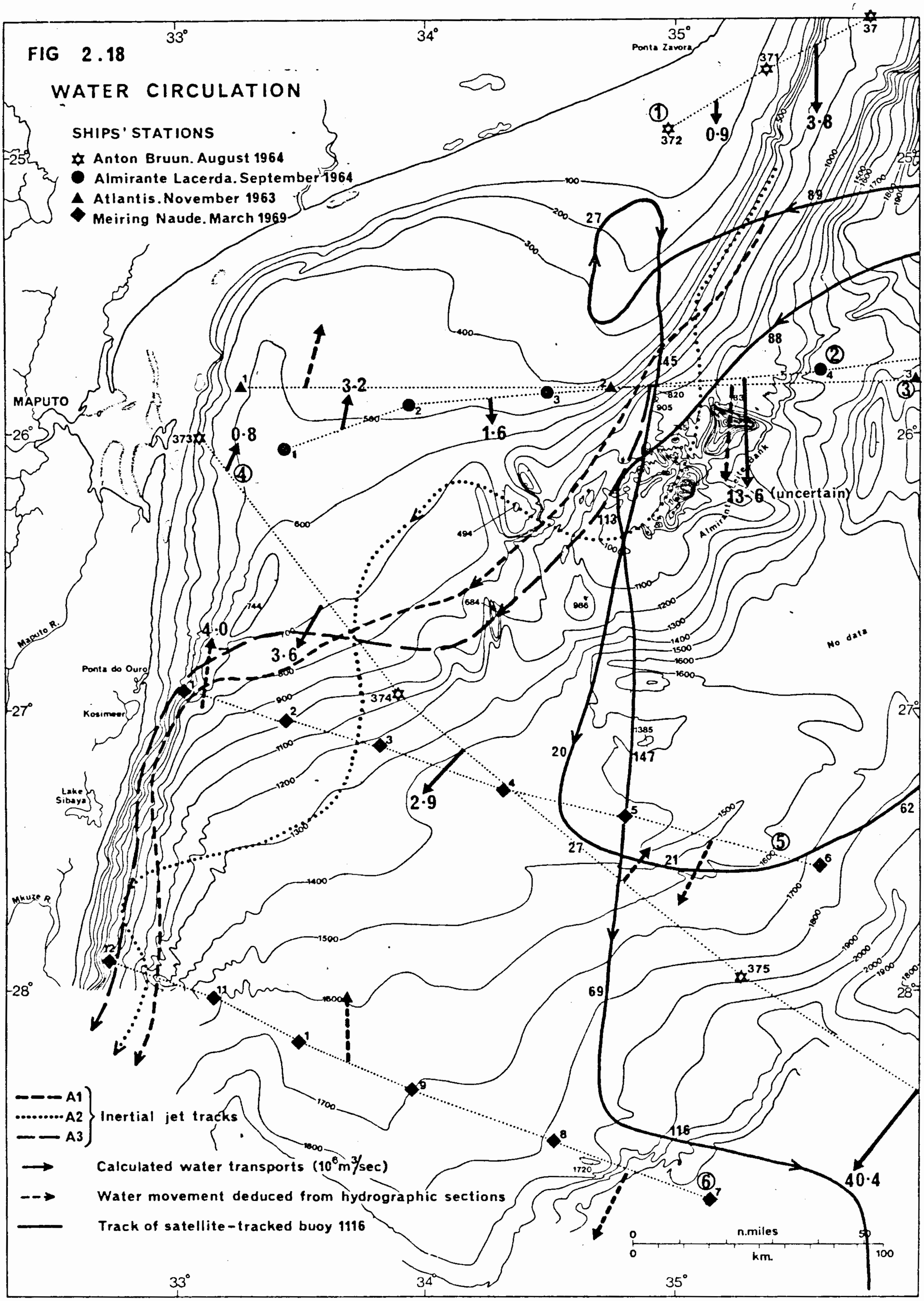


FIG 2.18
WATER CIRCULATION
SHIPS' STATIONS
Anton Bruun. August 1964
Almirante Lacerda. September 1964
Atlantis. November 1963
Meiring Naude. March 1969
A1
A2
A3
Inertial jet tracks
Calculated water transports (10⁶m³/sec)
Water movement deduced from hydrographic sections
Track of satellite-tracked buoy 1116
MAPUTO
Ponta Zavora
Ponta do Ouro
Kosimeer
Lake Sibaya
Maputo R.
Mkuze R.
Almirante Leite Bank
No data
13·6 (uncertain)
n.miles
km.

Mozambique Ridge and mainland Africa all act as western boundaries (Lutjeharms et al., 1981). Although cold undercurrents of Antarctic Bottom Water flow into the various abyssal basins of the Indian Ocean (Warren, 1974) it does not reach the shallow northernmost Natal Valley (Kolla et al., 1976).

2.5.1 Sources of the Agulhas Current

The N.N.V. lies in the source region of the Agulhas Current where three main water masses converge around 35°E, 25°S: a) Mozambique Current water which is supplied partly by the South Equatorial Current through the northern end of the Mozambique Channel and partly by the East Madagascar Current recycling in the Mozambique Channel and joining the southerly flow around 18°S (Duncan, 1970; Harris, 1970; 1972; Lutjeharms, 1971); b) East Madagascar Current water which meanders westwards from the southern tip of Madagascar towards the zone of convergence (Harris & Van Foreest, 1977; Harris et al., 1978; Lutjeharms et al., 1981); both sources a) and b) supply warm relatively saline 'tropical' water; c) discreet eddies of 'subtropical' water which are cooler and less saline, advect northwards then westwards from the Mozambique Basin and join the East Madagascar in-flow from the east (Grundlingh, 1977; Bang & Pearce, 1978).

2.5.2 Main Flow-paths

Southward and south westward flow from the zone of convergence takes one of two routes, both of which diverge from the coast (Harris & Van Foreest, 1977) (Figs.2.18 & 9.1): the inner flow-path (Route A) travels diagonally across the region from NE to SW, and impinges on the coast near 28°S, from where it flows southward near the coast; Route B, which may have a large anti-cyclonic eddy at its northern end (Fig.2.18), flows down the eastern Central Terrace and follows the Mozambique Ridge southwards. Around 32°S, Route B swings westwards and joins Route A near the African coast (Fig.2.17).

Note that sections 5 and 6, which were measured in March, 1969 (Harris & Van Foreest, 1977; 1978), show Route A further east than section 4 which was measured in August, 1964 (Harris, 1972). This indicates that the Route A flow-path varies or meanders and in Figure 9.1 the mean flow has been shown between these two extremes. Routes A and B may alternatively dominate circulation, varying in importance over periods of 3-4 weeks.

2.5.3 Counter-Current and Cyclonic Eddy

Off Maputo, an in-shore counter-current flows in a north-easterly direction and is associated with the cyclonic eddy south of Ponta Zavora (Orren, 1963; Lutjeharms, 1971; Grundlingh, 1977; Harris, 1978) (Fig.2-18). It is unlikely that a single well-defined clockwise circulation fills the whole bight in the coastline, but rather, several eddies of different scales probably occur simultaneously. In a similar but smaller coastal bay further south off Durban, the Natal Bight, this type of multiple clockwise circulation has been described by Pearce et al. (1978) and Malan & Schumann (1979).

From about 29°S to Ponta do Ouro, the counter-current is more intermittent than it is to the north (Harris, 1978). Its nature is uncertain, but Van Foreest (1977) postulates that Route A water may be forced back up the coast to form the counter-current when Route B flow impinges on the coast around 32°S and dominates circulation. An alternative suggestion (Pearce, 1977) is that the current reverses from south-west to north-east in response to the periodic passage of low pressure cells up the Natal coast.

2.5.4 Dynamic Up-welling

Dynamic up-welling is associated with the counter-current and particularly with the cyclonic eddy east of Maputo. Evidence for this lies in the up-bowing of the isopycnals shown in the density distributions of hydrographic sections measured in the area (eg. section 3 shown in Fig.9.2). This is further borne out by the high values of the nutrients nitrate, silicate and phosphate (Wyrtki et al., 1971) high phosphate and low oxygen (Orren, 1963) and high suspended iron values (Orren, personal communication, 1979) found in this region's surface waters.

2.5.5 Velocities and Water transports

Southerly flow may reach 3 knots in the main streams (see British Hydrographic Navigational Charts, and speeds reported by Grundlingh (1977) for Buoy 1116). Water transports are large: Harris (1970) estimates a southerly flow of 20 megatons/sec through section 1 which increases to 46 megatons/sec through section 4 (Harris, 1972), while 4 megatons/sec has been calculated for the counter-current (Harris, 1972, and Van Foreest, 1977).

2.5.6 Temporal Variations

Some of the features described here are intermittent while others are unstable but semi-permanent. On the basis of data collected during infrequent hydrographic campaigns, it was previously thought that seasonal variations occurred in the flow regime (eg. Darbyshire, 1964; Tripp, 1967; Duncan, 1970). Notwithstanding such evidence, recent satellite infra-red imagery and synoptic hydrographic data suggest that important changes occur in the system over periods of 3-4 weeks rather than seasonally (Harris & Van Foreest, 1977).

PART 2

PLATE TECTONIC RECONSTRUCTIONS

A wide variety of solutions have been proposed for the reconstruction of Gondwanaland (Fig.3.1). Reasonable agreement exists as to the form of the West Gondwanaland refit (South America and Africa) although in detail, various rotation parameters have been calculated (Table 3.1). Concerning East Gondwanaland and its match with West Gondwanaland, controversy has long raged (eg. Smith, 1976; Tarling & Kent, 1976; Harrison et al., 1979; Dalziel, 1980).

The Natal Valley occupies a central and crucial position within this controversy (eg. Veevers et al., 1980, Fig.2). In order to establish a tectonic framework for this area, a new plate tectonic reconstruction is presented here. Revision of the plate tectonic status of the Natal Valley has, in many instances, wider implications for the evolution of the southern oceans. Thus digressions from the immediate Natal Valley vicinity result from describing the new reconstructions and their regional importance.

The reconstructions were undertaken using programmes developed for the University of Cape Town Univac computer by C.J.H. Hartnady. The basic method is given in appendix 1.

Throughout this study, dating of events depends on comparison of magnetic polarity, radiometric and biostratigraphic time-scales. A burgeoning number of time-scales has appeared in recent literature (eg. Hiertzler et al., 1968; Larson & Hilde, 1975; Van Hinte, 1976; LaBrecque et al., 1977; Ness et al., 1980; Lowrie & Alvarez, 1981; Schlich, 1981; Harland et al., 1982; Lowrie, 1982; Odin, 1982). Various assumptions such as constant seafloor spreading rates and equal length biostratigraphic stages are inherent in various of the correlation schemes. I have used the time-scales of Larson & Hilde (1975) and LaBrecque et al. (1977) for the Late Jurassic-Cretaceous and Late Cretaceous-Tertiary respectively. This allows direct comparison with

Rabinowitz & LaBrecque (1979) some of whose conclusions are criticised, and is consistent with previous publications on this subject (appendix 2).

CHAPTER 3

A REVISED FALKLAND PLATEAU/NATAL VALLEY RECONSTRUCTION LEADING TO A NEW REFIT OF WEST GONDWANALAND

3.1 INTRODUCTION

The fit of South America and Africa proposed by Bullard et al. (1965) was widely accepted as one of the most accurate continental reconstructions for over a decade. More recently, its accuracy has been challenged (Table 1). The original Bullard reconstruction did not include the Falkland Plateau, the continental nature of which was only established later through seismic refraction work (Ewing et al., 1971) gravity studies (Rabinowitz et al., 1976) and the drilling of Precambrian granitic and gneissic rocks in DSDP hole 330 on the Maurice Ewing Bank, eastern Falkland Plateau (Barker, Dalziel et al., 1977). When the Falkland Plateau is rotated back against Africa using the Bullard parameters (Table 1) a gap of 75 km exists between the steep northern slope of the Falkland Plateau and the equally steep continental margin off the south and south-eastern coast of South Africa (see Figs.2.1 & 3.3). A revision of the fit around the South Atlantic (Rabinowitz & LaBrecque, 1979) closes the gap along these sheared margins by fitting a geophysical anomaly (combined isostatic gravity/magnetic anomaly) on the southern African margins to the equivalent anomaly on the Argentinian/Falkland Plateau margin. This reconstruction (Rabinowitz & LaBrecque, 1979) produces a close fit of South America and Africa, but exacerbates some of the overlap problems inherent in the earlier fit (Bullard et al., 1965). Specifically overlaps occur in the Natal Valley and the Gulf of Benin, and in solving these overlaps, while retaining the close fit of the Falkland Escarpment and the Agulhas margin, a new reconstruction of South America and south central Africa is achieved.

3.2 CONSTRAINTS USED IN CALCULATION OF ROTATION PARAMETERS

Two alternative dates for the opening of the South Atlantic have been proposed on the interpretation of magnetic data: anomaly M22 (Emery et al.,

FIGURE CAPTIONS

Fig.3.1. Two basic options for the reconstruction of Gondwanaland: a) after Norton & Sclater (1979) with West Antarctica's position after De Wit (1977); b) after Powell et al. (1980).

Fig.3.2. New palaeoposition of the Maurice Ewing Bank (Falkland Plateau) in relation to the Natal Valley. Falkland Plateau 2000 and 3000 m isobaths dashed. Natal Valley bathymetry at 500 m intervals. Basement highs delineated by seismic reflection profiling except south of 31°S where basement is schematically drawn from Scrutton (1976b). Northern extent of Cape Slope Anomaly, which marks the steep continental margin of the Agulhas Fracture Zone, from Du Plessis & Simpson (1974). Note how this margin and the South Tugela Ridge form a re-entrant into which the north-eastern tip of the Maurice Ewing Bank fits. The small 'peak' on the north-eastern apex of the Maurice Ewing Bank 3000 m isobath is an expression of the Falkland Fracture Zone which extends eastwards into the Argentine Basin. Insert shows locations of Figures 3.2-3.5.

Fig.3.3. Comparison of Falkland Plateau palaeopositions obtained using rotation parameters of 1 Rabinowitz LaBrecque (1979); 2 Bullard et al. (1965) and 3 this study. Rotation parameters are listed in Table 3.1. Falkland Plateau delineated by the 3000 m isobath.

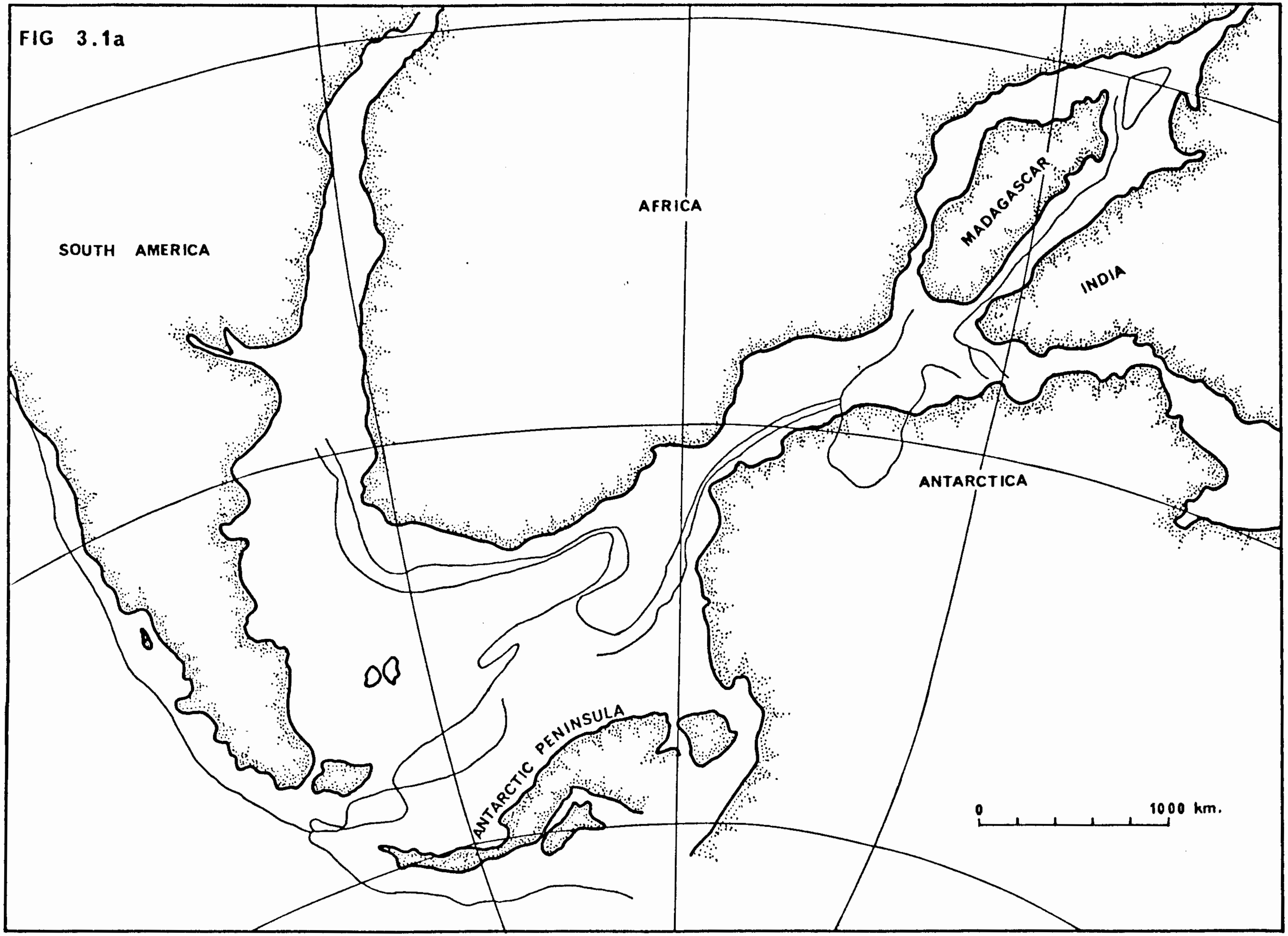


FIG 3.1a
SOUTH AMERICA
AFRICA
MADAGASCAR
INDIA
ANTARCTICA
ANTARCTIC PENINSULA
0
1000 km.

FIG 3.1b

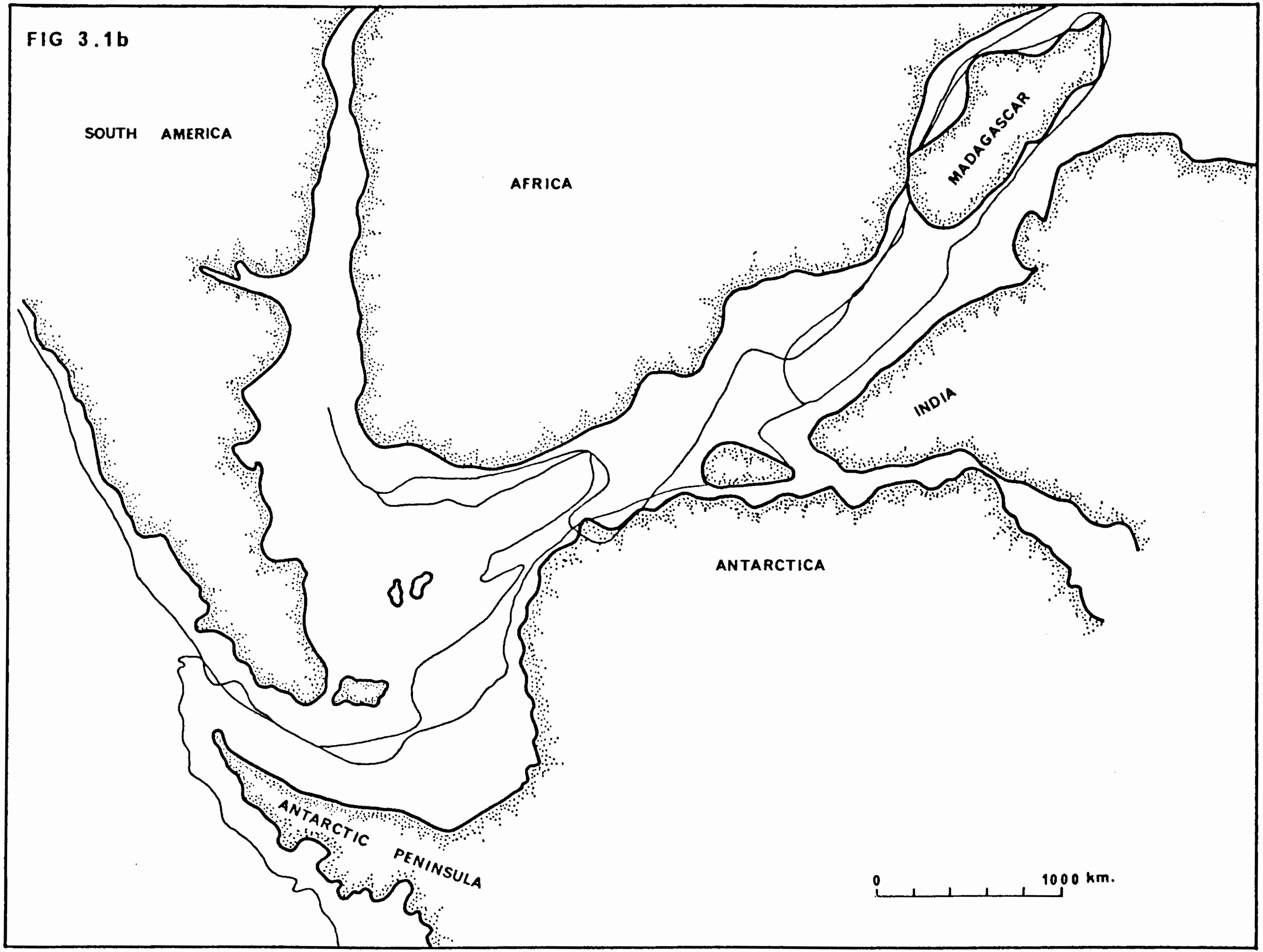
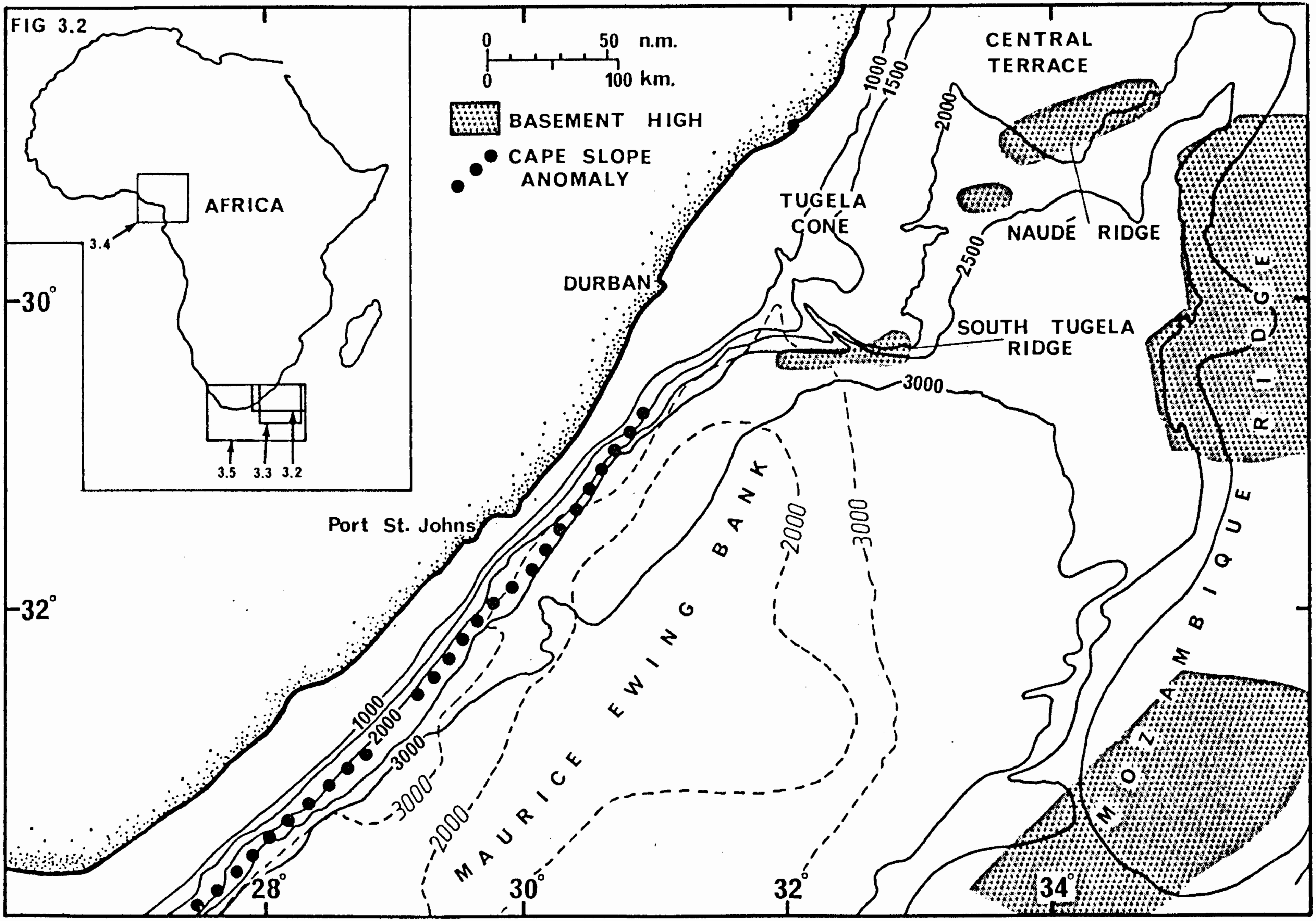

SOUTH AMERICA
AFRICA
MADAGASCAR
INDIA
ANTARCTICA
ANTARCTIC PENINSULA
0
1000 km.

FIG 3.2


AFRICA
3.4
3.5
3.3
3.2
30°
32°
0
50 n.m.
0
100 km.
BASEMENT HIGH
CAPE SLOPE ANOMALY
DURBAN
Port St. Johns
TUGELA CONE
CENTRAL TERRACE
NAUDÉ RIDGE
SOUTH TUGELA RIDGE
1000
1500
2000
2500
3000
MAURICE EWING BANK
MOZAMBIQUE RIDGE
28°
30°
32°
34°

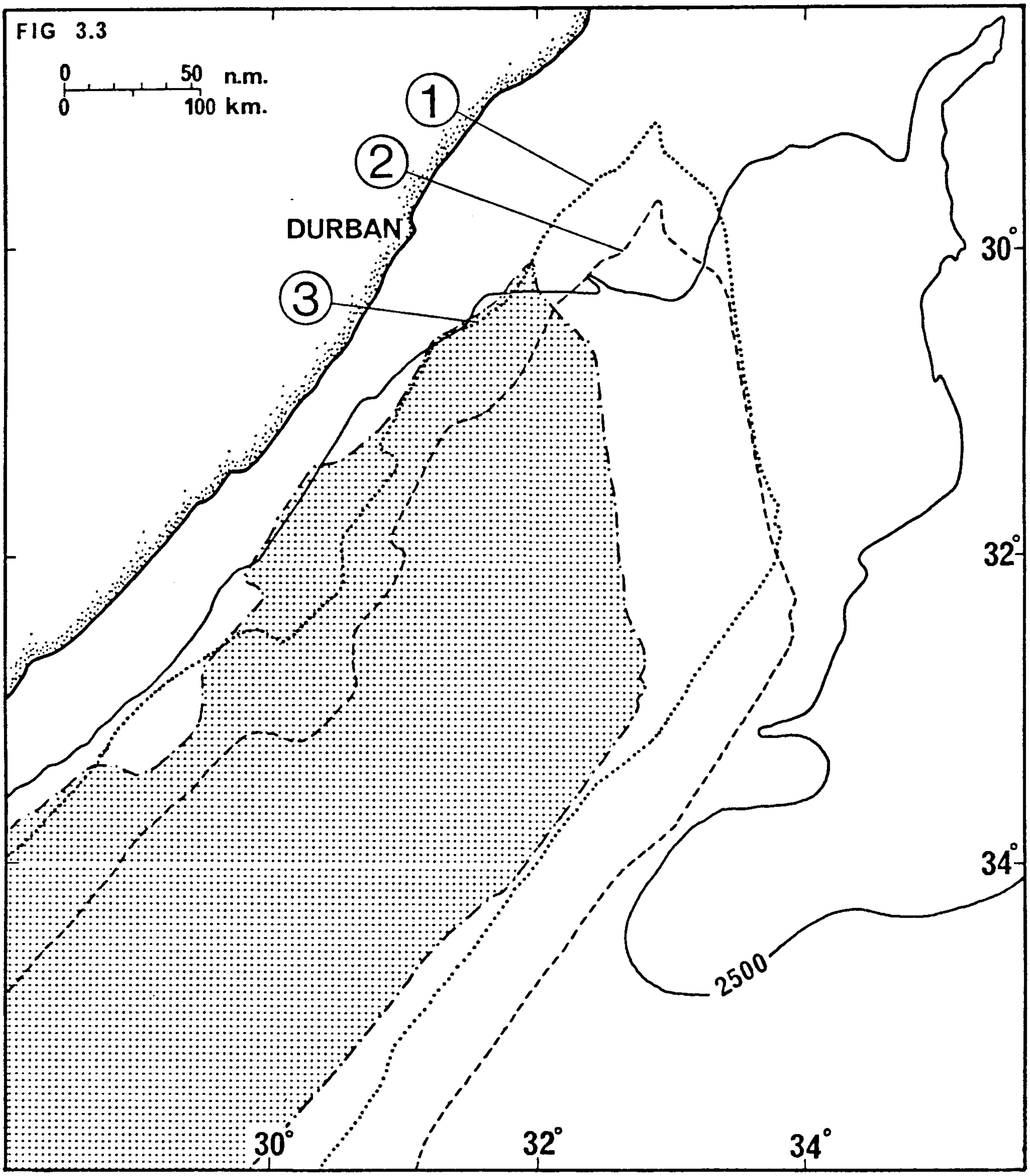


FIG 3.3
0 50 n.m.
0 100 km.
DURBAN
1
2
3
30°
32°
34°
2500
30°
32°
34°

1975; du Plessis, 1979) and anomaly M12/13 (Larson & Ladd, 1973; Rabinowitz, 1976). These are dated as 148 m.y. (basal Kimmeridgian) and 130 m.y. ago (basal Valanginian) respectively (Larson & Hilde, 1975). The projected age of basal sediments in DSDP hole 361 (Bolli, Ryan et al., 1978) appears to verify the Early Cretaceous date. During the initial stages of drifting between South America and Africa, the kinematics of relative rotation must have been severely constrained by the orientation and length of the intra-continental shear zone between the Falkland Plateau and the Agulhas margin of South Africa (cf. Le Pichon & Hayes, 1971; Scrutton, 1979). As there is little evidence for Early Cretaceous compression in South Africa (see chapter 7), rotations which cause substantial convergence across the Falkland Agulhas Fracture Zone (FAFZ) are considered untenable (Le Pichon, 1968; Le Pichon & Hayes, 1971; Delteil et al., 1974; Mascle & Sibuet, 1974; Sibuet & Mascle, 1978). Strictly conservative strike-slip motion is assumed to have occurred along the Agulhas margin (Fig.3.2), which is best defined by the Cape slope magnetic anomaly (du Plessis & Simpson, 1974).

By statistical fitting (method of Cruden & Charlesworth, 1972) of a small circle to a set of points on the Cape Slope anomaly, an early pole of rotation of 6°N, 2°W was obtained. This is within the limits of error ($\pm$5°) of the early pole of Rabinowitz & LaBrecque (1979)(3.5°N, 2.6°W). Rotation about either of these poles is appropriate, since neither causes misalignment, convergence or distension along the FAFZ (chapter 4). By trial and error, it was found that a clockwise rotation of 1.65° about the 6°N, 2°W pole maintained the sheared margin superimposition, while eliminating the overlap problems of the Rabinowitz-LaBrecque reconstruction (1979). The resultant total rotation parameters (Table 3.1) simultaneously solved the overlap in the Gulf of Benin.

3.2.1 Natal Valley/Falkland Plateau overlap

The African continental slope south of 30°20'S is steep and regular (Figs.2.1, 3.2) a character generated by the transform fault movement which

Table 3.1

Comparison of total reconstruction rotations, South America to south central Africa

	Source	Pole of Rotation	Rotation Angles
1.	Bullard et al (1965) (500 fm fit)	44.0°N 30.6°W	57.0°
	" " " (1000 fm fit)	44.1°N 30.3°W	56.1°
2.	Rabinowitz & LaBrecque (1979) (anomaly G)	45.5°N 32.2°W	57.5°
3.	Rickard & Belbin (1980)	45.93°N 30.12°W	56.68°
4.	Vink (1982)	47.0°N 33.8°W	58.0°
5.	Valencio et al., (1983)	52°N 29°W	53.5°
6.	This study	46.75°N 32.65°W	56.40°

took place along it. At 30°20'S, this slope swings eastwards to form the steep southern face of the Tugela Cone which is cored by an acoustic basement high - the south Tugela Ridge. Basement highs also underlie the Naudé Ridge and the Mozambique Ridge (chapter 8). The continent/ocean boundary (C.O.B.) is considered to underlie the southern edge of the Tugela Cone because (a) it is underlain by an acoustic basement high; (b) its southern face is steep and regular; (c) the Cape Slope anomaly which marks the sheared Agulhas margin terminates near the re-entrant formed by the south face of the Tugela Cone and the continental slope south of Durban (Fig.3.2); (d) the appropriate Mesozoic magnetic anomalies have been identified just south of the Tugela Cone (see further discussion in chapter 4).

On the Maurice Ewing Bank the COB, as delineated by seismic refraction work (Ewing et al., 1971), approximates to the 3000 m isobath. In the new reconstruction its north-eastern apex fits into the re-entrant in the Natal Valley isobaths caused by the south Tugela Ridge (Fig.3.2) while overlap problems of previous fits (Bullard et al., 1965; Rabinowitz & LaBrecque, 1979) are illustrated in Figure 3.3

3.2.2 Gulf of Benin Precambrian basement overlaps

In order to test if the refit proposed above (3.2.1) is valid along the length of the South Atlantic rift between South America and Africa, widely separated areas were considered. Two main criteria were evaluated: basement overlaps in West Africa (3.2.2) and predrift geotectonic features extending from one continent to the other (3.3).

In the reconstruction of South America and Africa, a well-known overlap occurs in the equatorial region where the north-eastern corner of Brazil overlaps the Niger Delta (Bullard et al., 1965). The magnetic signature of the Romanche fracture zone underlies the Niger Delta, proving that this latter feature is a post-drift outbuilding of deltaic sediments (Delteil et al., 1976). (In chapters 6 and 8, an analogy is drawn between the Niger Delta and coastal Mozambique). Closer inspection of the Rabinowitz &

FIGURE CAPTIONS

Fig.3.4.

a) The Rabinowitz-LaBrecque (1979) reconstruction for the Gulf of Benin area, compared to b) this study. Geotectonic divisions and features modified from maps in de Almeida et al. (1968) for South America and from the International Tectonic Map of Africa (ASGA-UNESCO, 1968). 1 - Archaean to Early Proterozoic basement of the Chaillu and N Gabon massifs in Africa and the Sao Francisco craton in S America. 2 - Tectonothermally reworked early Precambrian ("PD" after ASGA-UNESCO, 1968) in Cameroon and Central African Republic associated with Pernambuco-Alagoas Massif and Sergipano Fold Belt of NE Brazil. 3 - Dahomeyan Belt (Pan-African Cycle) of Africa and the North-eastern Fold System ("Caririan Belt"; Braziliano Cycle) median massifs and metasedimentary belts of NE Brazil. 4 - Late Proterozoic platform sedimentary cover rocks. 5 - South American coastline. 6 - African coastline. 7 - Major fault systems of Pan-African or Braziliano age. "A" represents Pernambuco lineament of NE Brazil and "B" represents the Foumban lineament of W Africa. 8 - Phanerozoic (mainly post-Jurassic) cover and S Atlantic Ocean. "C" represents Cretaceous Jucano Basin referred to in text. Note the area of Precambrian basement overlap (shaded black) in reconstruction "a" and its effective elimination in reconstruction "b". Note also the improved alignment of the Pernambuco and Foumban lineaments in reconstruction "b", and the possible correlation of the northern boundary of the Pernambuco-Alagoas Massif (S America) with the southwestern boundary of the Dahomeyan Belt (Africa).

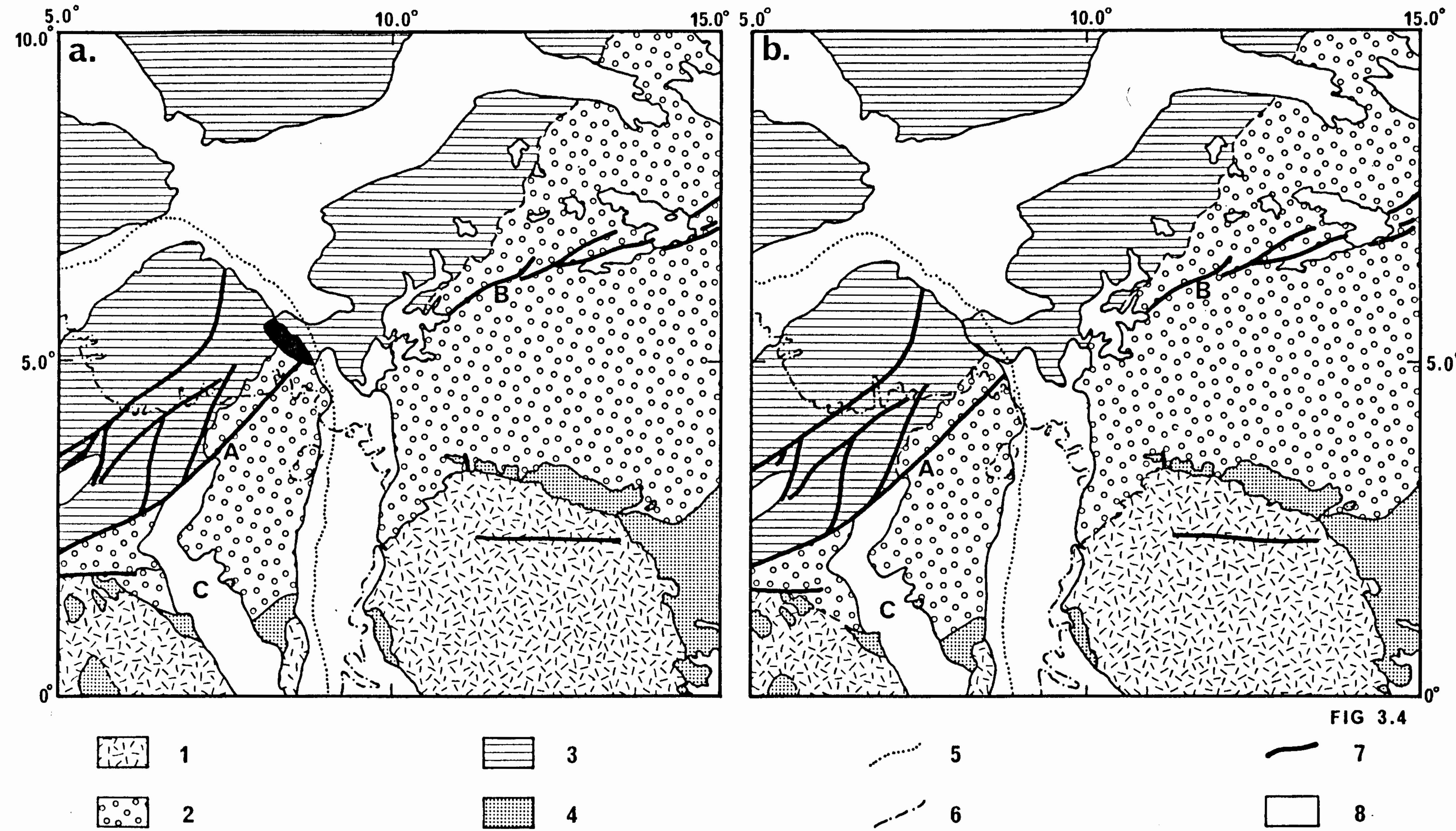


FIG 3.4

LaBrecque (1979) fit shows that an area of Precambrian basement exposed just east of the Niger Delta is overlapped by its South American equivalent (Fig.3.4a). While the large Niger Delta overlap is quite acceptable the ∿35 km overlap of Precambrian basement is not. Its existence indicates either an error in the reconstruction (Rabinowitz & LaBrecque, 1979) or a break-down of the first order assumption that the separated plates have behaved perfectly rigidly (see section 4.4). The same clockwise rotation adjustment which eliminates Natal Valley overlaps (Fig.3.3, Table 3.1) also resolves this Precambrian overlap (Fig.3.4b). Pre-drift reconstructions which produce a tighter fit than shown in Figures 3.3 and 3.4b are considered untenable.

3.3 MATCHING OF PRE-DRIFT STRUCTURAL AND STRATIGRAPHICAL FEATURES

The matching of on-land geotectonic features was one of the earliest tools used in continental reconstructions (du Toit, 1937). This technique has been used more recently to improve the fit between Australia and Antartica by matching the Dundas Trough (Australia) and the Bowers Group of Antarctica (Laird et al., 1977). Similar arguments have also been applied to the North Atlantic (Lefort, 1980). In the case of the South Atlantic, pre-drift structures on the rifted margins are unlikely to be sensitive enough markers to help choose between the five sets of rotation parameters listed in Table 3.1. However, along the offset margins (the equatorial fracture zones and the FAFZ), pre-drift structures aligned perpendicular to small circles of the rotation pole may provide markers with which to test continental reconstructions. This is particularly so in the case of the FAFZ since it is distant from the early rotation pole and there is in this region, up to 450 km difference implied by the various rotation parameters in Table 3.1.

3.3.1 Outeniqua and Falkland Plateau Basins

During the rifting stage which preceded true seafloor spreading between South America and Africa, E-S-E-trending basins and swells were formed in the

FIGURE CAPTIONS

Fig.3.5. Matching of features across the Agulhas Falkland Fracture Zone using new rotation parameters. Outeniqua Basin shown by isopachytes in km after Dingle (1979). Basement highs A.A. - Agulhas Arch. P.A.A. - Port Alfred Arch. Smaller, intervening basement highs from West to East - Infanta Arch, St Francis Arch, Gamtoos-Recife Arch. Sub-basins between these arches - Bredasdorp Basin, Pletmos Basin, Gamtoos Basin and Algoa Basin. South-western Cape Plutons: a. Saldanha Batholith; 2. Cape Granite; 3. Groot Haelkraal Granite; 4. George Granite. The northern front of the Cape Fold Belt is taken as the limit of strongly overfolded strata (from the Geological Map of South Africa, Dept. of Mines 1970). Falkland Plateau Basin shown by isopachytes in km from Ludwig et al. (1978). Note that no great accuracy is claimed by these authors, particularly for the eastern edge of the basin, and is "intended only as a guide" to the general features. Edges of "basement high" taken from profiles SR1 = seismic reflection profiles (Barker, 1977) and seismic refraction profiles (SR2, SR3)(Ludwig et al., 1979).

South African 2000 m and 3000 m isobaths dotted. Falkland Plateau 2000 m and 3000 m isobaths solid line. On the eastern Falkland Plateau the 2750 m isobath shows the bathymetric indentations considered to mark a structural change by Lonardi & Ewing (1971). Note how the suspected continuation of the northern front of the Cape Fold Belt, and the eastern and western margins of the Falkland Plateau Basin line up with their counterparts in South Africa. DSDP site 330 falls geographically within the realm of the Cape Fold Belt which may explain the elevated position of the gneisses there and the correspondence of their isotopic ages with Cape Pluton and Cape Fold Belt data rather than with the Namaqua/Natal basement ages south of Durban.

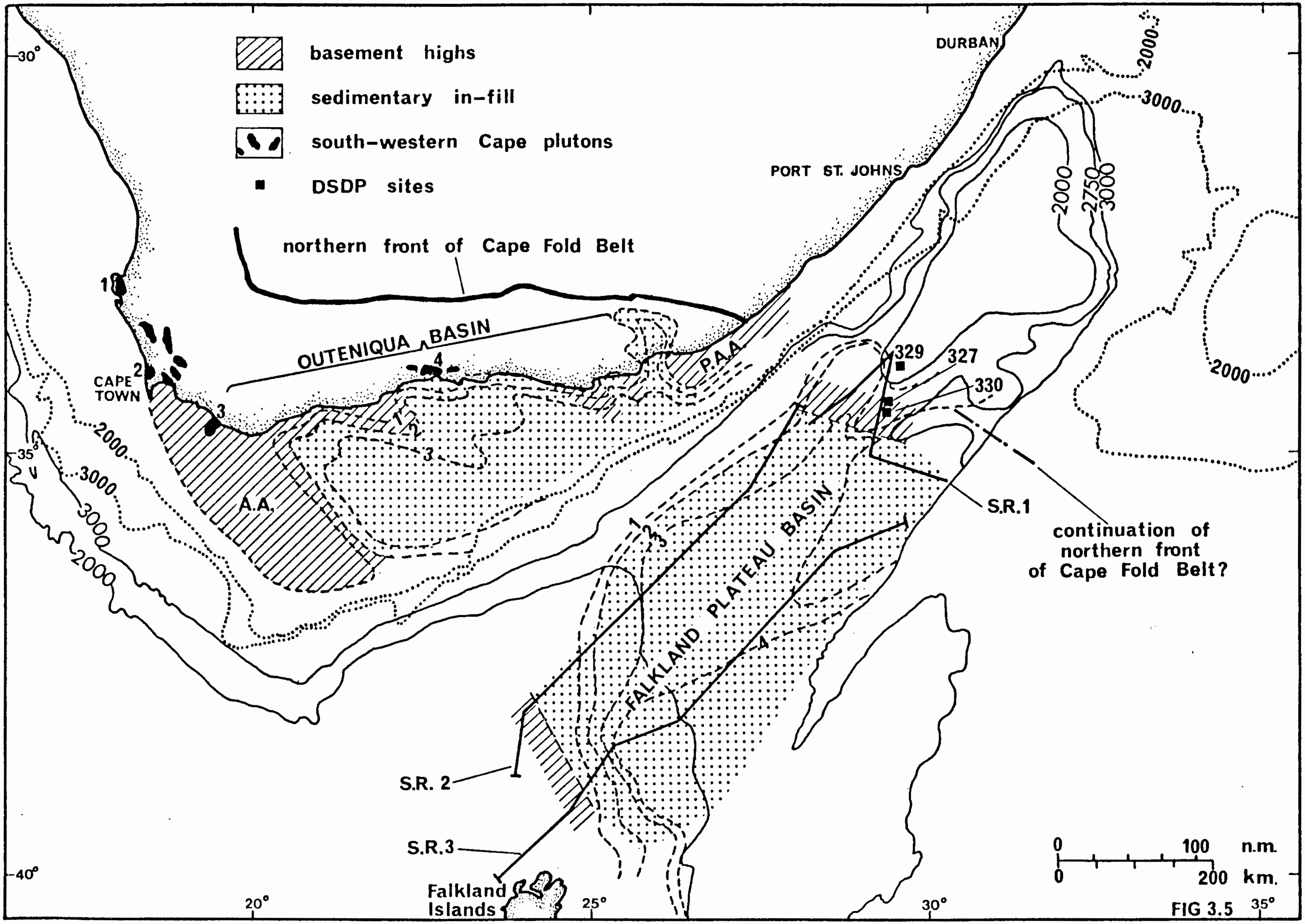

basement highs
sedimentary in-fill
south-western Cape plutons
DSDP sites
northern front of Cape Fold Belt
DURBAN
PORT ST. JOHNS
CAPE TOWN
OUTENIQUA BASIN
P.A.A.
A.A.
FALKLAND PLATEAU BASIN
329
327
330
S.R.1
S.R. 2
S.R.3
continuation of northern front of Cape Fold Belt?
Falkland Islands
2000
2750
3000
0 100 n.m.
0 200 km.
30°
35°
40°
20°
25°

FIG 3.5

Cape Province and on the Agulhas Bank, the offshore extension of South Africa (Dingle, 1973c; 1978; 1979; du Toit, 1979; McLachlan & McMillan, 1979). These structures curve more southeasterly offshore where the intermontaine basins are infilled forming the Outeniqua Basin (Fig.3.5). Palaeozoic sandstones, granites and Precambrian metasediments are exposed on the Agulhas Arch which forms the western limit of the Outeniqua Basin (Dingle et al., 1971). Cape Group sandstones and Karoo rocks underlie the eastern margin - the Port Alfred Arch (Dingle, 1978; du Toit, 1979). Intervening basement highs reduce the sedimentary infill thickness by about 500 m (Dingle, 1973c).

Oldest sediments in the Outeniqua Basin are assigned to the Uitenhage Group which can be divided into at least three formations - a continental Enon Formation, paralic Kirkwood Formation and marine Infanta Formation (Fig.3.6). The Enon Formation, a high-energy alluvial fan and piedmont deposit, passes laterally and upwards into the Kirkwood Formation, a finer-grained fluviatile deposit with limited local lignite bands. This, in turn passes laterally and vertically into marine silts, shales and sapropels of the Infanta Formation. The oldest rocks are at least Portlandian (Dingle & Klinger, 1972; Klinger et al., 1972; McLachlan & McMillan, 1979) and may be partly correlated with the mid-Jurassic volcanic and conglomeratic Zuurberg Group (Dingle & Scrutton, 1974; Dingle, 1978). The Hauterivian hiatus (top of column A, Fig.3.6) is related by du Toit (1979) to the on-set of drifting following the model of Falvey (1974).

The western end of the Falkland Plateau Basin is defined by seismic refraction work showing a basement high of seismic velocity 5.4 km/s (Ludwig et al., 1978; 1979). (This is within the seismic velocity range of basement on the Agulhas Bank (Ludwig et al., 1968)). Seismic refraction data are sparser to the east of the basinal area, but seismic reflection profiles (Barker, 1977) show basement (which correlates with granitic metasedimentary gneiss) rising from the basin towards the Maurice Ewing Bank (Fig.3.5). In

FIGURE CAPTIONS

Fig.3.6. Comparison of stratigraphy of Outeniqua and Falkland Plateau Basins. Column A from Du Toit (1979) and McLachlan & McMillan (1979). Column B from Dingle (1978). Disagreement exists as to whether the sequence extends back to the mid-Jurassic. The base of the sequence is not older than Bajocian (Dingle, 1978). Column C (Barker, Dalziel et al., 1977) is based on DSDP hole 330 just east of, but in seismic continuity with the Falkland Plateau Basin.

Fig.3.7. a) A cross-section of the southern Karoo basin (after Truswell, (1970). 1 Table Mountain sandstone. 2 Bokkeveld 3 Witteberg 4 Dwyka Tillite 5 Upper Dwyka Shale 6 Lower Ecca 7 Middle Ecca 8 Upper Ecca 9 Northern Ecca Facies.

b) Location of possible correlations across the Falkland Agulhas Fracture Zone. Basin on the Falkland Plateau opposite the Port St. Johns Basin is delineated after Ludwig et al. (1978). Port St. Johns Basin after Ludwig et al. (1968) and Simpson & Dingle (1973).

FIG 3.6

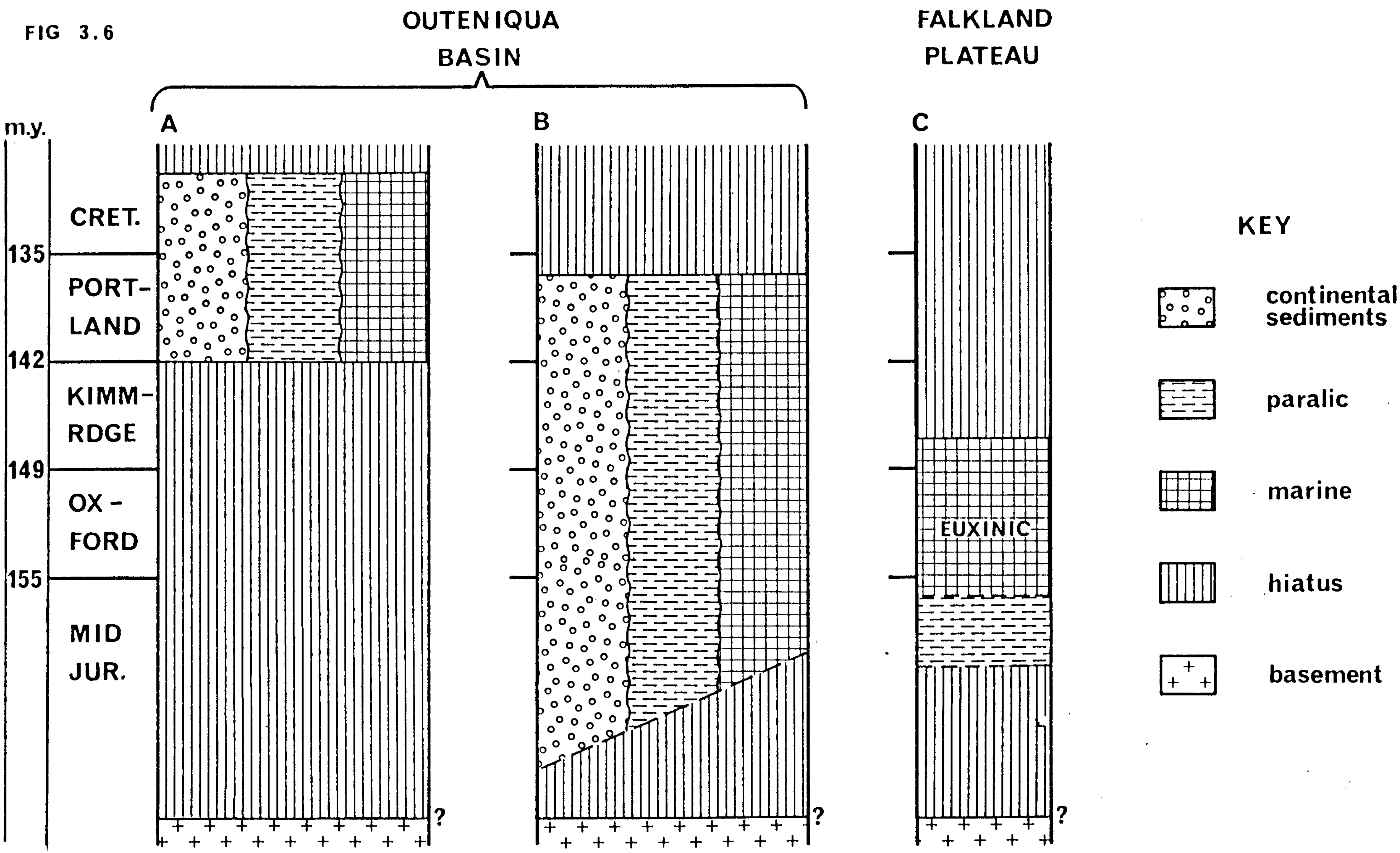
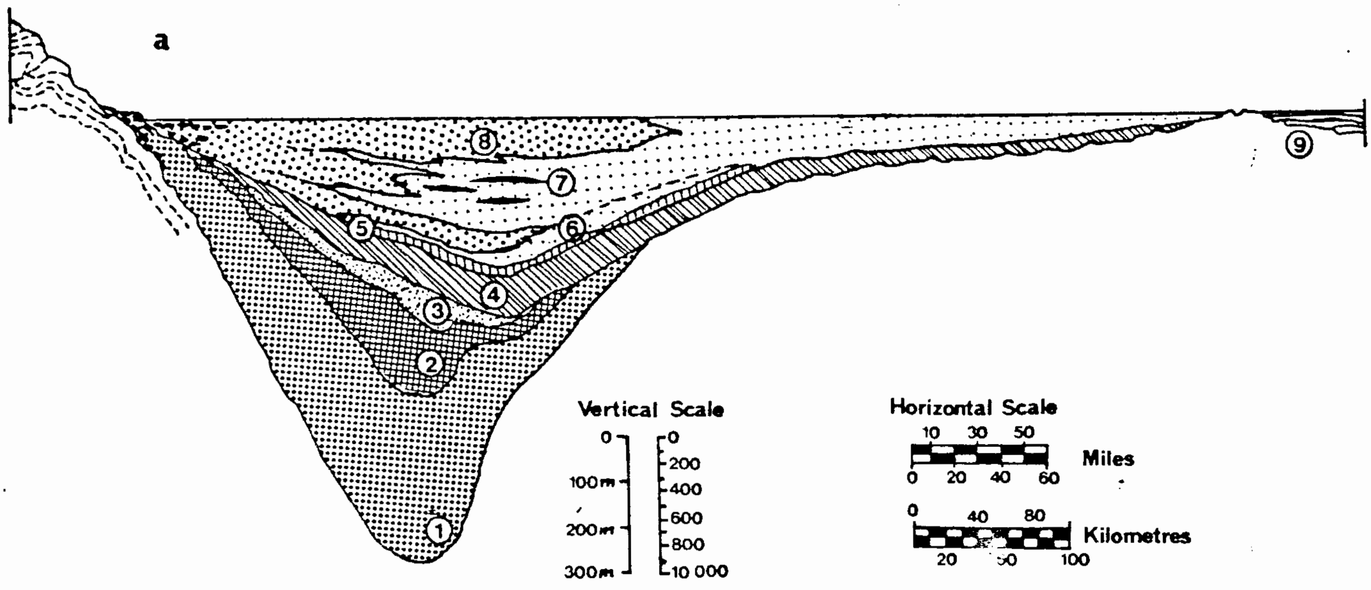
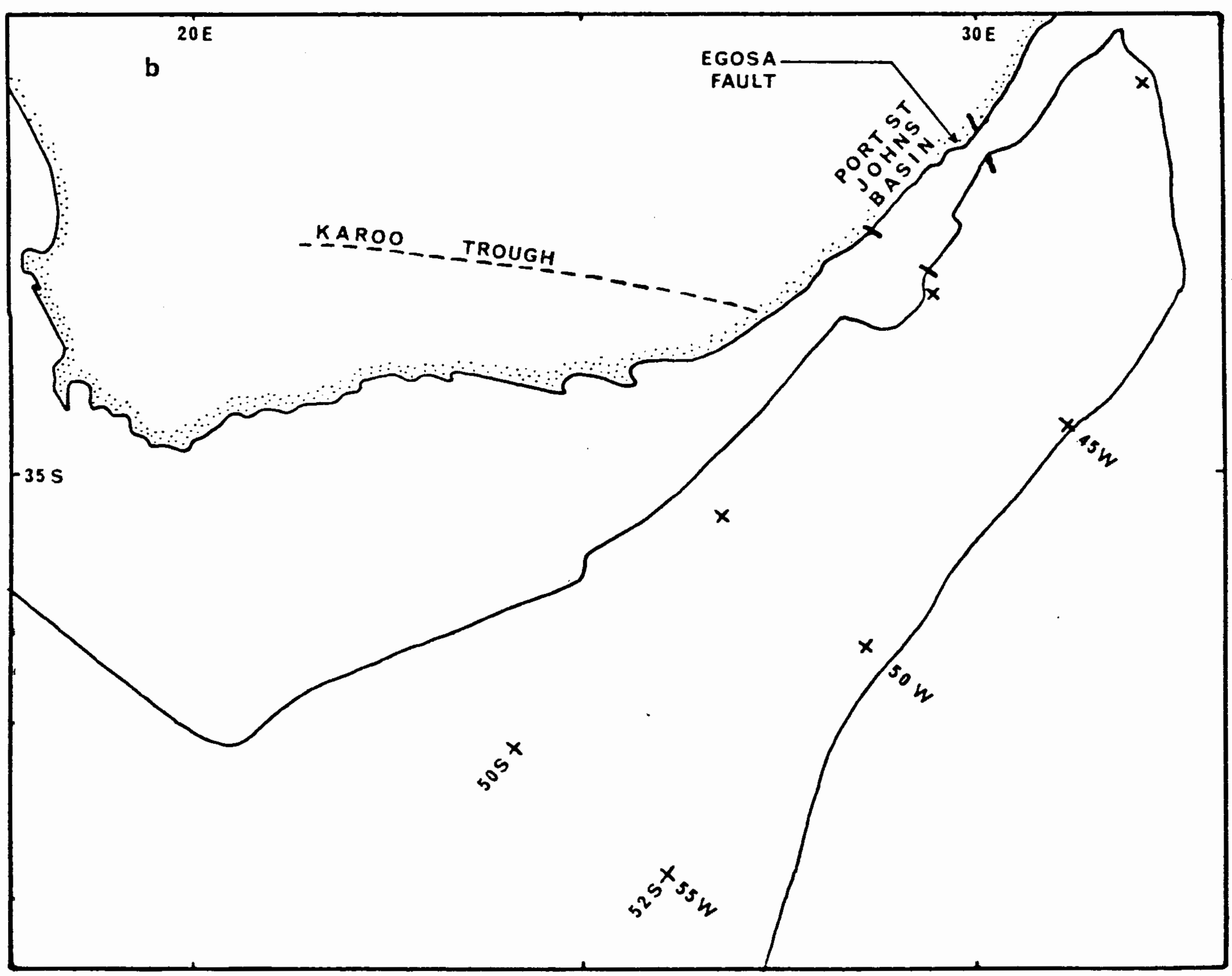

OUTENIQUA
BASIN
FALKLAND
PLATEAU
A
B
C
m.y.
135
142
149
155
CRET.
PORT-
LAND
KIMM-
RDGE
OX -
FORD
MID
JUR.
EUXINIC
?
KEY
continental
sediments
paralic
marine
hiatus
basement

FIG 3.7


a
8
7
6
5
4
3
2
1
9
Vertical Scale
0
100m
200m
300m
0
200
400
600
800
10 000
Horizontal Scale
10 30 50
Miles
0 20 40 60
0 40 80
Kilometres
20 100


b
20E
30E
EGOSA
FAULT
PORT ST
JOHNS
BASIN
KAROO TROUGH
35 S
45W
50 W
50S
52S
55W

DSDP hole 330 near the eastern margin of the basin, a basal lignitic silty sandstone is followed by clean sandstones which are overlain by marine silts and clays (Barker, Dalziel et al., 1977). This succession has been interpreted as a fluviatile sequence overlain by a beach, succeeded by marine continental shelf deposits (Thomson, 1977). The marine beds are Oxfordian, possibly mid-Jurassic near the base. The overlying beds are Oxfordian sapropelic claystones. The Kimmeridgian-Neocomian hiatus (top of column c, Fig.3.6) spans the time of initial break-up of Africa and the Falkland Plateau (Barker, Dalziel et al., 1977, p.999).

An objection to the matching of the basins has been raised on the basis of the Indo-Pacific affinities of the Upper Jurassic/Lower Cretaceous macrofossils of DSDP sites 327 and 330 (Jones & Plafker, 1977) as opposed to the Tethyan faunas of South Africa (Dingle, 1978). However, Hallam (1977) considers that a marine link existed between the Andean (Indo-Pacific) and Tethyan macrofaunas from at least the Portlandian, while Dingle (personal communication, 1980) notes an excellent correlation in the benthic ostracoda of the Early Cretaceous of the Falkland Plateau and Outeniqua Basins. Moreover, the basins have been previously matched on the basis of the similar stratigraphies and basin dimensions (Barker, Dalziel et al., 1977; Thomson, 1977) while Scrutton (1976a) tentatively correlated the Falkland Island basement high with the Agulhas Arch using gravity data. The similarity in structure and stratigraphy of the Outeniqua and Falkland Plateau Basins suggests that together they formed a unified feature in pre-drift times (Fig.3.5). This match mitigates against a more northeasterly palaeoposition for the Falkland Plateau.

3.3.2 Tectonic boundary locus of the Cape Fold Belt province

Prior to the Mesozoic rifting of West Gondwanaland, the last major compressive tectonic episode to affect southern Africa and its continental surroundings was the Permo-Triassic Cape Fold Belt orogeny which is considered to extend across Antarctica (du Toit, 1937; Barron et al., 1978; Dalziel, 1980). It must also cross the Falkland Plateau (du Toit, 1979; Newton, 1980).

Palaeozoic strata of the Cape and lower Karoo Supergroups are strongly overfolded, and locally overthrust towards the north along lines which curve gradually from easterly through to more southeasterly trends as the Agulhas continental margin between Port Elizabeth and East London is approached. The northern tectonic front of the Cape Fold Belt province, marked by a fairly sudden transition to the gently-folded or flat lying beds of the main Karoo basin, is therefore sharply truncated almost at right angles by the sheared continental margin (Figs.2.2, 3.5).

If the equivalent tectonic structure could be accurately recognised beneath the Mesozoic-Cenozoic cover of the Falkland Plateau, it would provide a most important constraint on an accurate palaeocontinental reconstruction. As yet there is no published marine geophysical data on pre-Mesozoic Falkland basement structure which allows such a boundary line to be drawn with confidence. However, the existence of the relatively elevated Maurice Ewing Bank as a distinct eastern topographic province of the Falkland Plateau may be controlled by differences in the deep basement structure across its western boundary. Lonardi & Ewing (1971) originally divided the northern margin of the Falkland Plateau into two major crestal topographic units "separated by a north-south offset of their crest at 46°30'W, clearly defined by the 1500 fm isobath" (op. cit. p.112). Along the southern margin of the plateau, the same isobath defines a deep embayment, delimiting the Maurice Ewing Bank on its southwestern side. A line drawn across the plateau from the northern isobath offset locus through the axis of the embayment on the southern side (Fig.3.5) may be a close approximation to the Cape Fold Belt province boundary extension, if the hypothesis of basement structure control is correct.

It is significant that this boundary line, as located indirectly from the topographic pattern, includes DSDP hole 330 within the extension of the Cape Fold Belt province (Fig.3.5). Apart from the folding of Palaeozoic strate another major feature of this province is the uplift of large inliers of Precambrian basement. DSDP hole 330 bottomed in Precambrian granitic and metasedimentary gneissic rocks only a few metres below fossiliferous Jurassic sediment (Barker, Dalziel et al., 1977). Although these rocks show petrographic and geochemical affinities (Tarney, 1977; Simpson, 1977) to 1018 m.y. old Precambrian rocks exposed in the Marble Delta area 115 km south of Durban, their isotopic age patterns closely resemble those obtained from the basement inliers of the Cape Fold Belt.

The oldest Rb/Sr isochron for the Maurice Ewing Bank rocks (Beckinsale et al., 1977) indicates an isotope homogenisation event at 554 $\pm$66 m.y. (recalculated using ^{87}Rb decay constant of $\lambda = 1.42 \times 10^{-11}$ yr^{-1}) which allows for a much older age for the original sediments before metamorphism. This has been correlated (Beckinsale et al., 1977) with the 541 $\pm$8 m.y. Rb/Sr age of the Cape Granites near Cape town (Allsopp & Kolbe, 1965) which are associated with deformation of the Late Precambrian Malmesbury sediments. Similar radiometric dates have been obtained for the Saldanha Batholith farther north (Schoch et al., 1975; Schoch & Burger, 1976) (Fig.3.5). Farther east the George Granite has been isotopically dated (Rb/Sr) at 532 $\pm$8 m.y. (Krynauw & Gresse, 1980). In this area, too, the event involves deformation of surrounding metasediments. A later phase of deformation isotopically dated as 275 $\pm$1 m.y. has been linked with the Cape Fold Belt episode (Krynauw & Gresse, 1980). The Groot Haelkraal Granite (Fig.3.5), although less metaphosed than the George Granite, suffered an argon loss event dated at 248 $\pm$2 m.y. (Gentle et al., 1978). It is considered that all of these southwestern Cape plutons were formed around 550 m.y. and were affected to a greater or lesser degree during the Permo-Triassic phase of the Cape Fold Belt orogeny (Gentle et al., 1978). A radiometric date of 287 m.y. interpreted as an argon loss event and tentatively correlated with the Cape Fold Belt orogeny (Beckinsale et al., 1977) was also obtained on the Maurice Ewing basement rocks. In Figure 3.5, the palaeoposition of DSDP 330 gneisses falls geographically within the realm of the Cape fold Belt and this may explain their uplift and the correlation of their isotopic ages with Cape Pluton and Cape Fold Belt events. In contrast, the isotopic age characteristics of the Falkland Plateau gneisses might be expected to correlate with the 900-1100 m.y. Namaqua-Natal mobile belt outcropping in southern Natal if their palaeoposition lay 140 km northeast (Rabinowitz & LaBrecque reconstruction 1979) and beyond the limits of the Cape Fold Belt (section 2.1).

3.3.3 Further potential correlations across the Falkland Agulhas Fracture Zone (FAFZ)

Several features of South African geology are truncated by the sheared Agulhas margin. One of the most prominent is the Karoo Basin (Fig.2.2). In cross-section (Fig.3.7a), up to 3000 m of Dwyka Formation shales and Ecca Group sandstones have been deposited in a large syncline. A similar trough accommodated the Cape Group sandstones and shales, but the locus of sedimentation and subsidence migrated 50 km north by Karoo times (Truswell, 1970). Maximum thicknesses of Karoo rocks occur near the coast 60 km south of East London (Fig.3.7b). Palaeocurrent directions indicate provenance from the south and south-east indicating that the basin should continue on to the Falkland Plateau and beyond. If the preferred reconstruction is correct this trough should be located at 47°W on the Falkland Plateau's northern edge, extending south-eastwards towards 46°W on its southern flank.

A second potential feature to be correlated is the Port St. Johns Basin lying between the Port Alfred Arch and the Port Shepstone High (Ludwig et al., 1968; Simpson & Dingle, 1973). This feature has been outlined by seismic refraction only and its exact boundaries are not well determined, but lie between 31.1°S and 32.3°S. On the Falkland Plateau's northern flank a small basin lying approximately between 42.7°W and 44.7°W has also been outlined by seismic refraction (Ludwig et al., 1978). These two features appear to match in the revised reconstruction (Fig.3.7b).

Near Port ST. Johns, the Egosa and related faults appear to offset the continental margin. This fault is pre-drift and is possibly syn-rift (chapter 7). It is opposed by a small bathymetric offset on the Maurice Ewing Bank which may be controlled by the same feature.

3.3.4 Pernambuco/Foumban fault system

Although it has been mentioned that the rifted sections of the South Atlantic are less sensitive to testing the five reconstructions cited in Table 3.1, the fit between Africa and South America in the equatorial region is tested by correlation of the Pernambuco fault system of Brazil (Fig.3.4) with the Foumban lineament of Cameroun (De Almeida, 1968). The Pernambuco dextral transcurrent fault system is a 700 km long E-W trending structure which partly forms the boundary between the Caririan or Northeastern Fold Region of the Braziliano Orogenic cycle (1000-500 m.y.) and the Pernambuco Alagoas Massif (DeAlmeida et al., 1976). This fault system however, appears to mark the northern limit of the Mesozoic/Cenozoic Jucano Basin which extends southwards as a straight trough to link with the Cretaceous Sergipe-Gabon Basin of the Atlantic coast (Wardlaw & Nicholls, 1972). It is thus possible that Cretaceous re-activation caused some left lateral displacement of the Pernambuco-Alagoas Massif relative to the Caririan Belt and the Sao Francisco craton during the opening of the Jucano Rift. similarly, along the Foumban lineament, small Cretaceous sedimentary basins have been locally tectonised (Vincent, 1968). Burke & Dewey (1974) have therefore questioned the Late Precambrian age of the "Ka Borogop mylonite belt" (Foumban lineament). However, Gorini & Bryan (1976) consider that the fault zone is Precambrian and was re-activated in the Cretaceous. The correlation and alignment of these two Late Precambrian structures (Fig.3.4) can be used as a subordinate constraint on West Gondwanaland reconstructions.

The reconstruction proposed here (Fig.3.4b) not only eliminates an unacceptable basement overlap, but it displaces the Pernambuco lineament relatively southwards and thus brings it into better (though not perfect) alignment with the Foumban lineament. More detailed mapping of Precambrian basement structures and isotopic age provinces in this critical area is needed to confirm this conclusion and evaluate the quantitative significance of possible post-drift deformation.

3.4 DISCUSSION OF THE REVISED FIT

In this section some of the more immediate implications of the revised reconstruction are considered. A detailed comparison with the other refits (Table 3.1) is deferred until seafloor spreading anomalies in the Natal Valley are described and used to construct palaeopositions for the Falkland Plateau (chapter 4, 4.1 - 4.3). These results also bear upon the refit of West Gondwanaland.

3.4.1 Second-order movements

3.4.1.1 Two-plate behaviour of Africa during West Gondwanaland break-up

The equatorial fracture zones provide another area where cross-cutting pre-drift features may provide a test of palaeo-reconstructions. However, alignment (Hurley et al., 1967) of the western tectonic front of the Dahomeyan orogenic belt against the West African craton and the equivalent front of the Caririan folded region against the San Luis craton in South America (500 km west of Fig.3.4) is not exact. To improve it requires an eastward relative displacement of about 200 km for the Dahomeyan Front. The implication is that north-western Africa has moved relative to south central Africa, with the finite displacement between the proposed north-western and south central African plates being of the order of 200 km. Burke & Dewey (1974) suggested this movement occurred sometime after the Early Cretaceous. The discovery of Precambrian rocks under the Benue Trough (Avbovko, 1980) suggests this movement may not have occurred as opening of a "narrow Benue Ocean" (Burke & Dewey, 1974) but by intracontinental crustal stretching or shearing. Left lateral strike slip motion in a generally east-west direction along a sequence of strike-slip faults seems a plausible mechanism (Benkhelil, 1982; Neev et al., 1982). Thus the rotation parameters for the northern part of the West Gondwanaland fit may be significantly different from the southern fit parameters. Recently, Pindell & Dewey (1982) have in fact used a two-plate model for Africa in studying the evolution of the Central Atlantic and the Caribbean. To produce a reconstruction which is everywhere accurate, both major plate movements and possible second-order movements constituting a break-down of the assumption of rigid plate behaviour must be accounted for.

3.4.1.2 Natal Valley Gap

The Falkland Plateau's palaeoposition in the Natal Valley has been noted previously (Scrutton, 1973; Barker, Dalziel et al., 1977; Simpson, 1977; Rabinowitz & LaBrecque, 1979). However the status of the crust north of the Falkland Plateau in its reconstructed position remains debatable. Intermediate thickness crust, as determined by gravity models and seismic refraction lies under the northernmost Natal Valley and the Mozambique Ridge (Darracott, 1974; Scrutton, 1976b; Chetty & Green, 1977). Despite geochemical studies on DSDP 249 basalt samples from the Mozambique Ridge (Erlank & Reid, 1974; Thompson et al., 1982) agreement has not been reached as to the oceanic or continental nature of this feature. Moreover, doubt exists over the nature of crust lying between the Naudé Ridge, the Mozambique Ridge and the Falkland Plateau in its reconstructed position (Fig.3.2). Seafloor in this region lies at abyssal depths, while basement lies between -5 and -7 km depth (Ludwig et al., 1968; Chetty & Green, 1977; Dingle et al., 1978; chapter 8). Six possible explanations are discussed below, with explanation (f) being preferred:

a) The region consists of thinned continental crust which has subsided to greater depths than the Tugela Cone. However, seismic refraction velocities in the area between the Falkland Plateau and the Mozambique

Ridge (Chetty & Green, 1977) and on the eastern flank of the Tugela Cone (Ludwig et al., 1968) are more typical of oceanic crust (see Fig.8.3 for refraction site locations).

b) Although seismic refraction results (Ewing et al., 1971) suggest the 3000 m isobath approximates the C.O.B. on the south-eastern tip of the Falkland Plateau, a spur of continental crust extending east of this area could fill the gap south of the Naudé Ridge (Fig.3.2). Such a spur exists just north of South Georgia Island. However it appears to be too large and of the wrong orientation, while South Georgia island may have been placed south of Patagonia at this time (eg. Winn, 1978). Moreover seafloor spreading anomalies have been mapped in this area (Fig.4.11).

c) The northern part of the Agulhas Plateau lies in the seafloor spreading compartment between the Natal Valley and the Falkland Plateau. If this region was continental, it may originally have been located at the apex of the Falkland Plateau, fitting south of the Naudé Ridge. However, the northern area of the Agulhas Plateau is considered oceanic in character (Barrett, 1977; Tucholke et al., 1981) while ridge-jumps occurring in this spreading compartment occurred at the wrong time to have rifted the northern part of the Agulhas Plateau from the Falkland Plateau (du Plessis, 1977; Barker, 1979; LaBrecque & Hayes, 1979; chapters 4 and 5).

d) Movement along both the north and south sides of the Maurice Ewing Bank is considered to have taken place along fracture zones, and leaky transforms may have caused some movement perpendicular to the direction of the transform faults. However, the gap north of 33°S (Fig.3.2) cannot be explained by this mechanism.

e) The southeastern edge of the Maurice Ewing Bank is straight and in its reconstructed position it is sub-parallel to the western Mozambique Ridge between 29° and 32°S. Similarly, the southern margin of the Maurice Ewing Bank is parallel to the western edge of the Mozambique Ridge south of 32°S. If part of the Mozambique Ridge south of Naudé Ridge is rotated back towards the Falkland Plateau in a west-south-westerly direction, these two sub-parallel sections could be juxtaposed and the gap eliminated. Late Jurassic or Early Cretaceous back-arc spreading may have occurred in the Falkland Trough (south of Falkland Plateau) contemporaneous with back-arc spreading in southern Patagonia (De Wit, 1977; De Wit & Stern, 1981). This movement, precursor to South Atlantic oceanic crustal accretion, may have extended to the Mozambique Ridge. This possibility is negated by the preferred explanation ((f) below), but did point to the possibility that the Mozambique Ridge has not always occupied its present position relative to Africa.

f) Modelling of the separation of East Antarctica and Africa, and matching of the correct conjugate seafloor spreading sets (chapter 6) necessitates that the Mozambique Ridge is either oceanic in character or did not occupy its present position relative to Africa prior to anomaly M2 times (113 m.y. ago - Larson & Hilde, 1975). In this model, the Natal Valley crust east of the Tugela Cone, including the area north of the Naudé Ridge is generated at mid-ocean ridge sections between Mozambique and Antarctica.

3.5 SUMMARY

A new set of parameters for the total reconstruction of South America and south central Africa is calculated: Euler pole 46.75°N, 32.65°W; rotation 56.40° counter clockwise of South America relative to Africa. This fit is considered pre-drift rather than pre-rift. The fit is obtained by fitting the geophysically and bathymetrically defined north-eastern apex of the Falkland Plateau into the re-entrant angle defined on the South African margin by the steep south-east facing sheared Agulhas margin and the southern face of the Tugela Cone.

Simultaneously, known Precambrian outcrops in northeastern Brazil and in the Gulf of Benin area of West Africa are juxtaposed rather than overlapped. Reconstructions producing a tighter fit of these cratonic areas are considered untenable. The fit is also constrained by alignment of at least three tectonic features crossing from one continent to another: (1) the geophysically defined eastern and western boundaries of the submarine Jurassic Outeniqua Basin of South Africa and the Falkland Plateau Basin; (2) the Late Precambrian transcurrent fault and mylonite belts of Pernambuco (Brazil) and Foumban (West Africa); (3) the northern tectonic front of the Triassic Cape fold Belt and the major morphological feature on the Falkland Plateau with which it is closely lined up. Isotopic ages of Falkland Plateau gneisses correspond to Cape Pluton and Cape Fold Belt ages, suggesting their palaeoposition lies within the realm of the Cape Fold Belt. The fit implies that $\sim$200 km of relative movement has occurred between north-west Africa and south-central Africa.

CHAPTER 4

MESOZOIC SEAFLOOR SPREADING MAGNETIC ANOMALIES IN THE NATAL VALLEY AND THEIR USE IN COMPUTING CRETACEOUS PALAEOPOSITIONS FOR THE FALKLAND PLATEAU AND SOUTH AFRICA

4.1 INTRODUCTION

Veevers et al., (1980, Fig.2) highlighted the paucity of Mesozoic anomalies in the Indian Ocean. In the Natal Valley, the revised plate tectonic reconstruction and the premise that the southern face of the Tugela Cone marks the C.O.B. (chapter 3) prompted a search for magnetic anomalies in the Transkei Basin and the Natal Valley. An analogous situation occurred in the Cape Basin where the work of Talwani and Eldholm (1973) prompted identification of the Cape Sequence Mesozoic anomalies (Larson & Ladd, 1973).

Recognition of Mesozoic seafloor spreading anomalies in the Georgia Basin, off the tip of the Maurice Ewing Bank (Barker, 1979; LaBrecque & Hayes, 1979) showed that the Falkland Plateau separated from the Natal Valley approximately at the same time as seafloor spreading began in the Argentine and Cape Basins of the South Atlantic (Larson & Ladd, 1973; Rabinowitz & LaBrecque, 1979). However, understanding of Gondwanaland break-up and the complicated seafloor spreading regime south of the Falkland Agulhas Fracture Zone (FAFZ) has been hindered by the lack of firmly identified magnetic anomalies in the Natal Valley (eg. Tucholke et al., 1981). Previously, du Plessis (1977, 1979) had the appropriate data but did not recognise anomalies immediately south of the Tugela Cone. This was probably because he calculated the original offset along the FAFZ to be only 1130 km, thus placing the C.O.B. in the Natal Valley south of Port St. Johns (compare Fig.4.1). This result was achieved indirectly from his estimate of the present-day offset along the FAFZ (given as 280 km and a ridge-jump of 850 km considered to be the only asymmetry in spreading). R.R.S. Shackleton's 1975 cruise (Barker, 1979) and a cruise by R.V. Vema in 1977 (E.S.W. Simpson, personal communication, 1981) failed to obtain unequivocal data.

Magnetic data from R.V Thomas B. Davie cruises (du Plessis, 1977; 1979) have been re-examined and Mesozoic seafloor spreading anomalies recognised in the Natal Valley south of Durban. These identifications are compared with synthetic profiles and various magnetic sequences associated with opening of the South Atlantic. These results have implications for recognition of the continent-ocean-boundary (C.O.B.) and interpretation of the physiography of the Natal Valley.

4.2 NATAL VALLEY MAGNETIC ANOMALIES

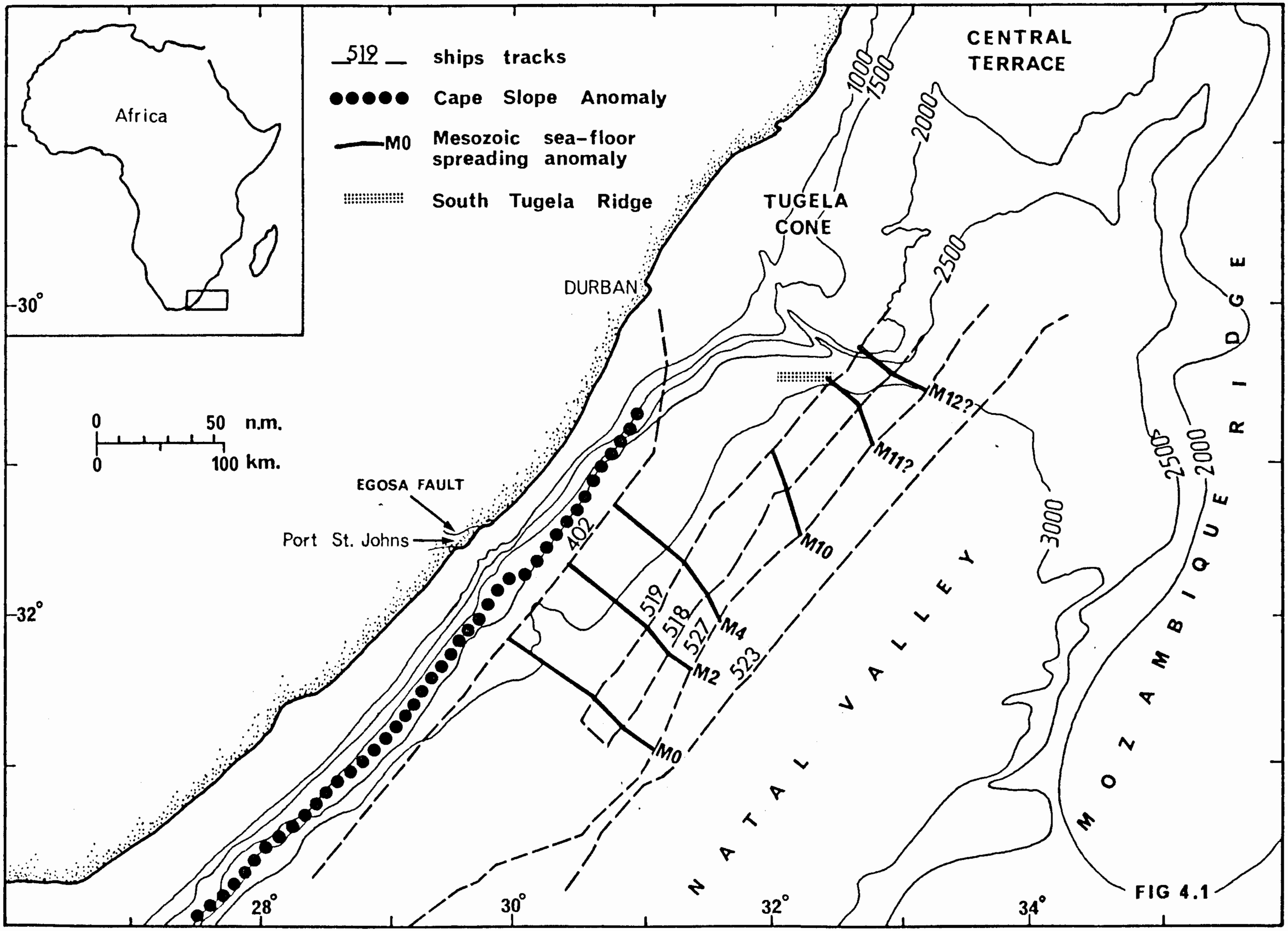
On tracks oriented NE-SW, Mesozoic anomalies M0-M10 are firmly identified, while anomalies M10N-M12 are tentatively recognised (Fig.4.1). The anomalies strike NW-SE and are perpendicular to the Cape slope anomaly (du Plessis & Simpson, 1974) which marks the sheared Agulhas margin. Anomaly M4 provided the key to the identification, appearing prominently in all profiles as a broad positive anomaly 250 gammas in amplitude, except in 402, where it is slightly subdued near the Agulhas margin (Figs.4.1, 4.2). On profiles 518, 519 and 402, anomalies M0-M2 are well correlated as a prominent triple peak. On profile 527, the M0-M2 sequence is poorly formed, but resembles the R.R.S. Shackleton profile (Barker, 1979) obtained in the conjugate area of the Georgia Basin. Profiles 402 and 527 further resemble Georgia Basin data in that magnetic activity occurs in crust younger than M0. Anomalies M4-M10 on profile 527 correlate very well with the synthetic sequence, whereas on the other profiles this part of the sequence is less well exhibited. Note that profile 527 is slightly compressed relative to the other profiles, and anomalies are positioned slightly to the south (Fig.4.1). It is uncertain whether this implies a slower spreading rate, a small offset, or just navigational inaccuracies, since Decca and dead reckoning were used on the cruise. Note that profile 523 (Fig.4.1) does not show the same sequences, suggesting that the tectonic regime is disturbed by a fracture zone or a change in the spreading orientation in this region. A possible model for the disturbance in spreading regime is given in chapter 6 (Figs.6.6 - 6.10).

FIGURE CAPTIONS

Fig.4.1. Ships tracks and magnetic lineations identified as the Mesozoic seafloor spreading sequence M0-M10 and possibly M10N-M12. The Cape Slope Anomaly is after Du Plessis & Simpson (1974).

Fig.4.2. Natal Valley magnetic profile 402, 519, 518 and 527 compared to a synthetic anomaly generated using a half spreading rate of 1.5 cm yr^{-1} and the time-scale of Larson & Hilde (1975). Also shown are Barker's (1979) Shackleton profile from the Georgia Basin, Cape Basin profiles 2204 and 2206, and a synthetic profile (Larson & Ladd, 1973; Rabinowitz & LaBrecque, 1979). Note that since the original publication of this work (Martin et al., 1982b) the implied COB in the Cape Basin (between anomalies M10 and the peak of the broad positive anomaly deforming anomalies M10N-M12) has been corroborated by multi-channel seismic profiling near profile 2204 (Austin & Uchupi, 1982).

Fig.4.3. Profiles from the Argentine Basin and a synthetic anomaly generated at a spreading rate of 1.3 cm yr^{-1} after Rabinowitz and LaBrecque (1979). Identifications from M0-M4 and anomaly G as originally reported. M10 identifications tentative in order to complete reconstruction Figure 4.5. Note that landward of M10, the profile is deformed by a broad positive anomaly.



519 ships tracks
Cape Slope Anomaly
M0 Mesozoic sea-floor spreading anomaly
South Tugela Ridge
Africa
30°
32°
28°
30°
32°
34°
0 50 n.m.
0 100 km.
EGOSA FAULT
Port St. Johns
DURBAN
CENTRAL TERRACE
TUGELA CONE
1000
1500
2000
2500
3000
402
519
518
527
523
M0
M2
M4
M10
M11?
M12??
NATAL VALLEY
MOZAMBIQUE RIDGE

FIG 4.1

FIG 4.2

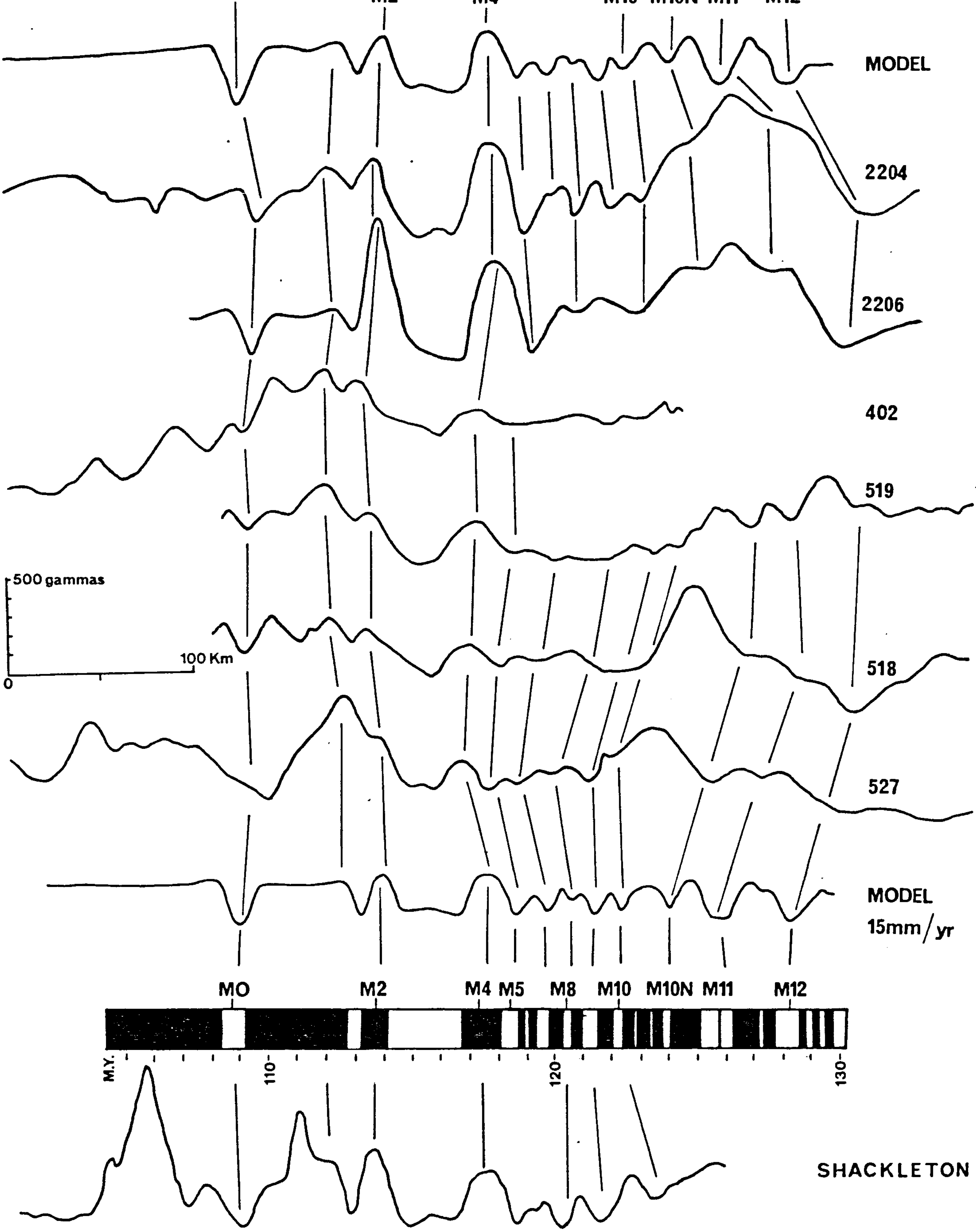
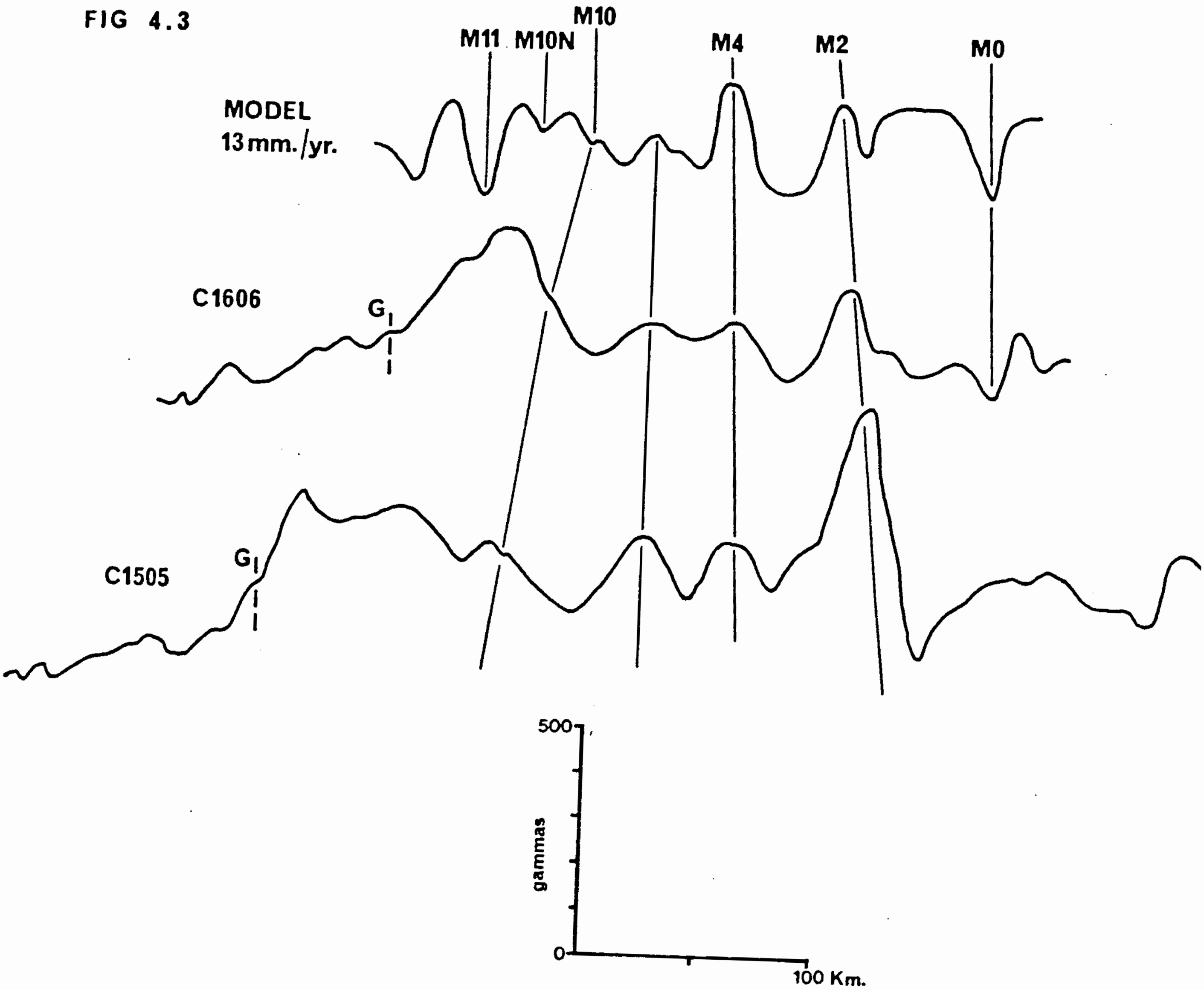

MO
M2
M4
M10
M10N
M11
M12
MODEL
2204
2206
402
519
500 gammas
100 Km
0
518
527
MODEL
15mm/yr
MO
M2
M4
M5
M8
M10
M10N
M11
M12
M.Y.
110
120
130
SHACKLETON

FIG 4.3


M11
M10N
M10
M4
M2
M0
MODEL
13 mm./yr.
C1606
G1
C1505
G1
500
gammas
0
100 Km.

Using the Larson and Hilde (1975) time-scale (Fig.4.2, synthetic sequences), the half-rate of seafloor spreading in the Natal Valley is 1.3-1.4 cm yr^{-1} for M0-M4, and 1.6 cm yr^{-1} for M4-M10, and is similar to spreading rates in the southernmost Cape Basin. A slight asymmetry of spreading favouring the South American side occurred, since the Shackleton profile shows half-rates of 1.5 cm yr^{-1} and 1.7 cm yr^{-1} in the M0-M4 and M4-M10 sequences respectively. Anomalies M10N, M11 and M12 are possibly represented on the north-eastern ends of profiles 518, 519 and 527 but a major positive anomaly deforms this section of the sequence. A similar situation has been noted in the Cape Basin (Larson & Ladd, 1973) where disruption by the rifting process is considered a likely cause for poorly lineated magnetic signatures at the landward end of the sequence. There is some uncertainty as to the exact correlation of M10 in the Cape and Georgia Basin (compare Larson & Ladd, 1973, Fig.2; Barker, 1979, Fig.4 and Rabinowitz & LaBrecque, 1979, Fig.7). I have followed Barker (1979) and placed M10 immediately seaward of a broad positive anomaly and landward of four small anomalies which are older than M4. (Fig.4.2, compare profile 527 from the Natal Valley with the Shackleton profile). For consistency I have placed M10 immediately seaward of a large positive anomaly on the Cape Basin profiles Vema 2204 and 2206 shown in Figure 4.2. This scheme matches the model better than previous, slightly different schemes: Larson & Ladd's model (1973) places M10N in this position while Rabinowitz and LaBrecque's model (1979) places M9 there. Irrespective of the exact correlation with the time-scale, inter-profile comparison shows that the scheme given in Figure 4.2 is consistent. M10 is the last undeformed seafloor spreading anomaly on the seaward side of magnetic signatures associated with rifting (c.f. Rabinowitz & LaBrecque, 1979, p.5996).

The identification of anomalies M0-M10 (and possibly of anomalies M10N-M12) in the Natal Valley stretching south-west of the Tugela Cone is strong corroborative evidence that the southern face of the Tugela Cone delineates the continent-ocean-boundary (c.f. chapter 3). Specifically the new rotation parameters (Table 4.1) juxtapose the tip of the Falkland Plateau with the South Tugela Ridge. This implies that anomalies identified as M10N-M12 are due to splintered continental basement and possible intrusion of volcanic material during the rifting phase. The combination of volcanic material, possibly emplaced at different times and thus of differing magnetisation, non-magnetised continental basement, and varied basement topography may explain the poorer inter-profile correlation in this region and the deformation of anomaly shapes.

A second major implication of these M-sequence anomaly identifications is that they are offset ∿1300 km to the northeast from their equivalents at the southernmost end of the Cape Basin. This is an improved estimate of the original offset along the FAFZ. This has previously been estimated as 1130 km (du Plessis, 1977; 1979) and 1400 km (Barker, 1979; LaBrecque & Hayes, 1979). The probable root of du Plessis' error has been discussed above. The estimate of 1400 km was made from the Falkland Plateau side and appears to be the distance from anomaly G on the Argentine margin to the very tip of the Falkland Plateau. This estimate is incorrect because these two end-points are not equivalent, since the tip of the Falkland Plateau is just landward of anomaly M10 (Fig.4.11) while anomaly G is colinear with anomaly M12 and appears to correlate with rifted continental crust (section 4.4).

4.3 CRETACEOUS FALKLAND PLATEAU PALAEOPOSITIONS RELATIVE TO SOUTHERN AFRICA

4.3.1 Introduction

Identification of seafloor spreading magnetic anomalies in relation to the known geomagnetic reversal time-scale is a subjective exercise. In several ocean basins, the same marine magnetic traverses have been keyed to geomagnetic reversal time-scales differently by different authors. For example, controversy in this regard exists in the South Atlantic (Dickson et al., 1968; Larson & Ladd, 1973; Emery et al., 1975; Rabinowitz, 1976; du Plessis, 1979; Van der Linden, 1980). In the Weddell Sea, various interpretations of the magnetic anomalies have been proposed by the same authors because of a lack of corroborative evidence with which to constrain the models (LaBrecque, 1977; Barker & Jahn, 1980; LaBrecque & Barker, 1981).

In the South Atlantic, DSDP hole 361 appears to bear out the interpretation of Cape Sequence anomalies as M0-M10. I have used the matching data from the Natal Valley and the Georgia Basin to compute Early Cretaceous palaeopositions for the Falkland Plateau relative to Southern Africa. These reconstructions bear upon the pre-drift positions of the continents, early formation of the South Atlantic and South-west Indian Oceans and the dating of significant events occurring in the Cretaceous Normal Polarity Epoch. Reconstructions for several anomalies were computed in order to a) test whether anomaly identifications in the Natal Valley are consistent with Cape Basin identifications and b) to check for any extension or contraction of the offset along the FAFZ which may indicate continental stretching or ridge-jumps during early spreading. The resultant palaeopositions provide a useful frame-work within which to study Early Cretaceous stratigraphy and sedimentation patterns, particularly since euxinic strata of potential economic interest were deposited in the developing ocean basins and margins (Barker, Dalziel et al., 1977; Du Toit, 1979; McLachlan & McMillan, 1979; Tissot et al., 1980. section 8.4).

4.3.2 Computation

Assuming rigid plate tectonics (eg. LePichon et al., 1973, but see section 4.4.2) palaeopositions of the Falkland Plateau at various stages in the evolution of the South Atlantic and South-West Indian Ocean have been computed (Figs. 4.5-4.8). The finite anticlock-wise rotations of South America which produce these reconstructions (Table 4.1) were obtained by trial and error clock-wise rotations from the full fit position (Fig.4.4) about the new early pole of rotation, 6°N 2°W (chapter 3). This pole was obtained by assuming strictly conservative strike-slip motion along the eastern part of the Agulhas sheared margin (Fig.4.4) and by statistical fitting of a small circle to the Cape Slope Anomaly which delineates this margin.

FIGURE CAPTIONS

Figs.4.4-4.10. Computed reconstructions. In all cases Africa outlined by 3000 m isobath. Argentinian margin and Falkland Plateau shaded and outlined by 3000 m isobath. A full list of rotation parameters, time-scales used and ages is given in Table 1. All reconstructions mercator projection relative to Africa. Cape slope magnetic anomaly marks sheared Agulhas Margin. Similarly the Falkland Escarpment magnetic anomaly marks the sheared margin continent/ocean boundary. For these sheared margins, the peaks (troughs) of the anomalies were plotted, whereas the modelled C.O.B. is $\sim$5 km landward of the anomaly peak (du Plessis & Simpson, 1974). This and navigational inaccuracy explains any overlaps of these data sets.

Fig.4.4. - predrift: On the rifted margins, this reconstruction juxtaposes a broad prominent positive anomaly found landward of M10 (Figs. 2 & 3). Agulhas Fracture Zone (shown in present day position) separates oceanic crust of the two spreading compartments.

Figs.4.5-4.8. were obtained by successive clock-wise rotations of S America from the pre-drift reconstruction (Fig.4.4) about an early pole of 6°N 2°W. By trial and error, equivalent Georgia Basin and Natal Valley anomalies were superposed, simultaneously fitting Argentine and Cape Basin anomalies. Fig.4.5-M10. "Tentative" identifications of LaBrecque & Hayes (1979) shown with question marks do not line up well with the Shackleton profile (Fig.4-11) and do not superpose Natal Valley anomalies. Inset shows small offset at Egosa and related faults (Fig.4.1) and equivalent offset in the Falkland Escarpment forming a small basin as strike-slip occurs along sheared margins.
Fig.6 = M4. Fig.7 = M2. Fig.8 = M0. Fig.9. This reconstruction (using parameters of Rabinowitz & LaBrecque 1979) shows that the amount of post M0 crust produced by this stage (shown striped) necessitates a re-dating of the pole change age.
Fig.10 = chron 34 reconstruction (after Bergh & Barrett, 1980).

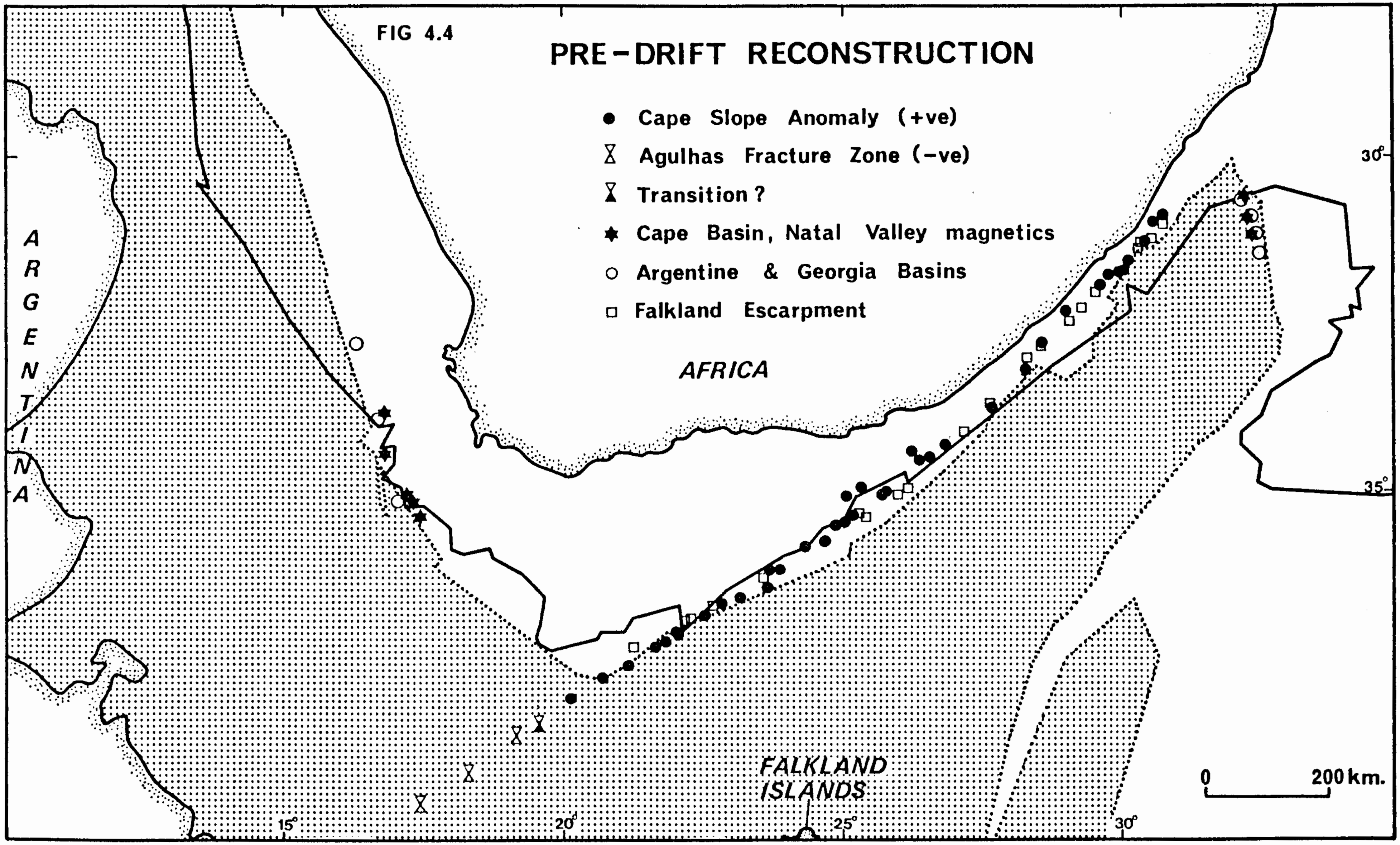


FIG 4.4
PRE-DRIFT RECONSTRUCTION
Cape Slope Anomaly (+ve)
Agulhas Fracture Zone (-ve)
Transition ?
Cape Basin, Natal Valley magnetics
Argentine & Georgia Basins
Falkland Escarpment
AFRICA
ARGENTINA
FALKLAND ISLANDS
0
200 km.
15°
20°
25°
30°
35°

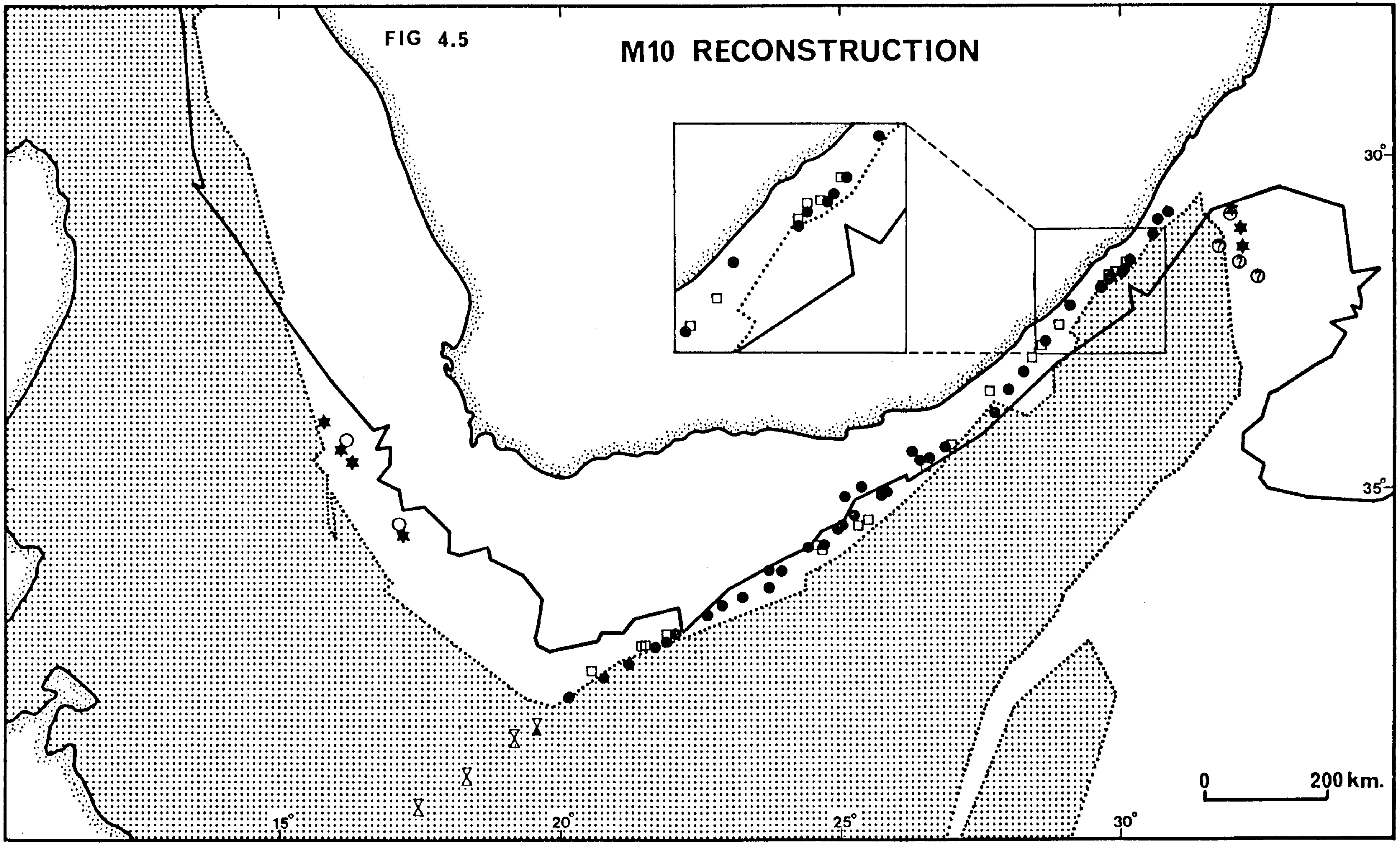


FIG 4.5
M10 RECONSTRUCTION
30°
35°
15°
20°
25°
30°
0
200 km.

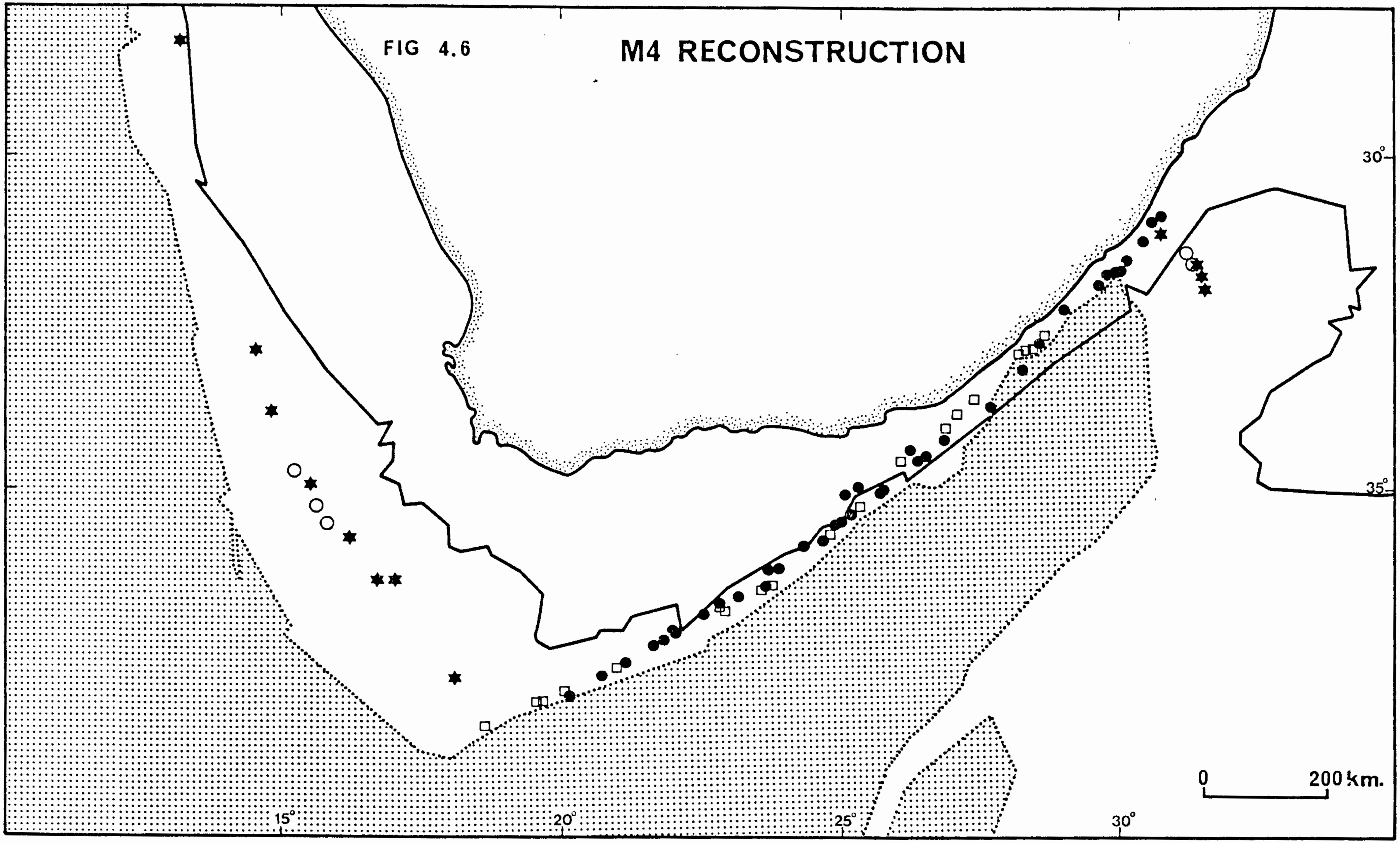


FIG 4.6
M4 RECONSTRUCTION
30°
35°
15°
20°
25°
30°
0
200 km.

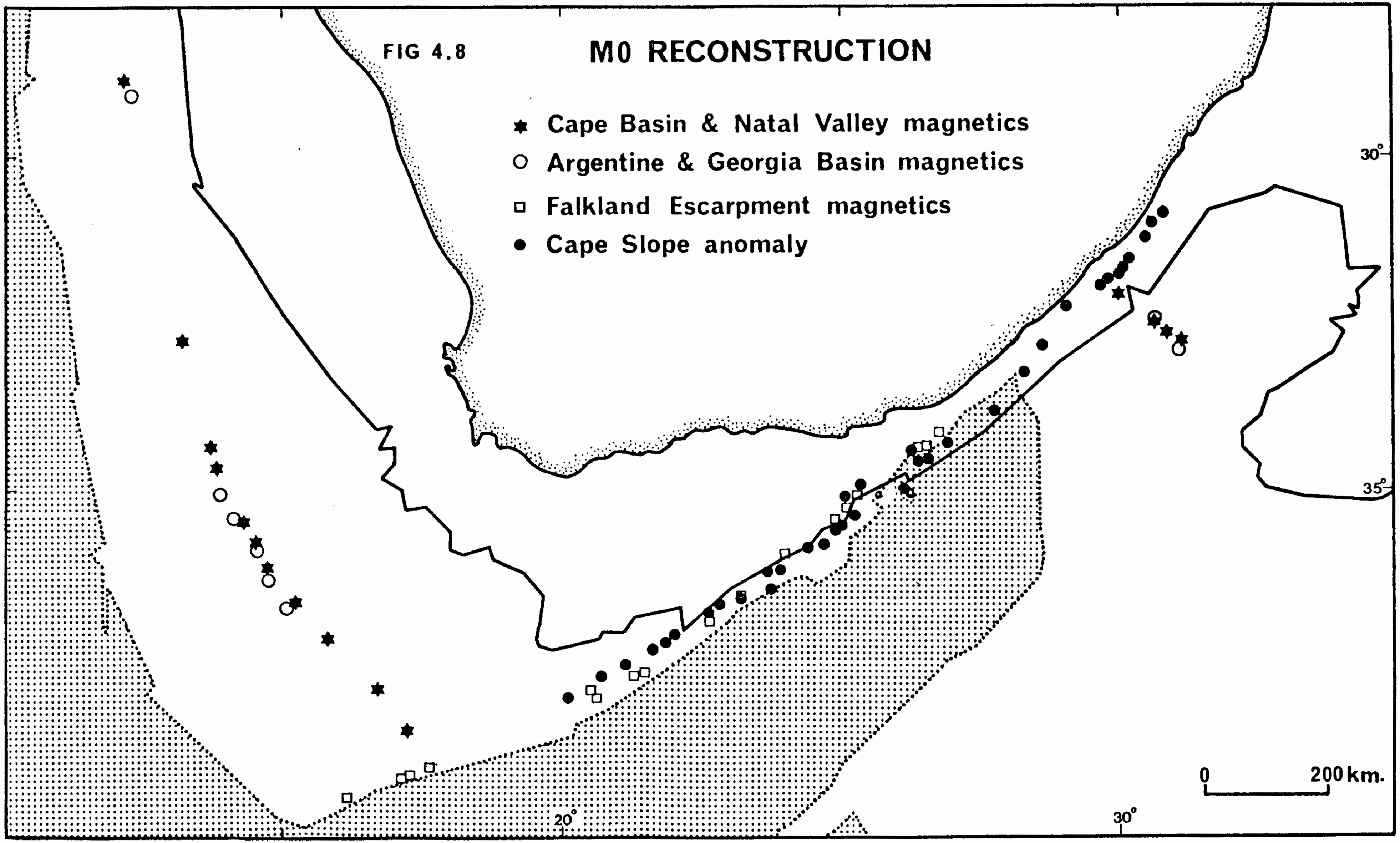


FIG 4.8
M0 RECONSTRUCTION
Cape Basin & Natal Valley magnetics
Argentine & Georgia Basin magnetics
Falkland Escarpment magnetics
Cape Slope anomaly
30°
35°
20°
30°
0
200 km.

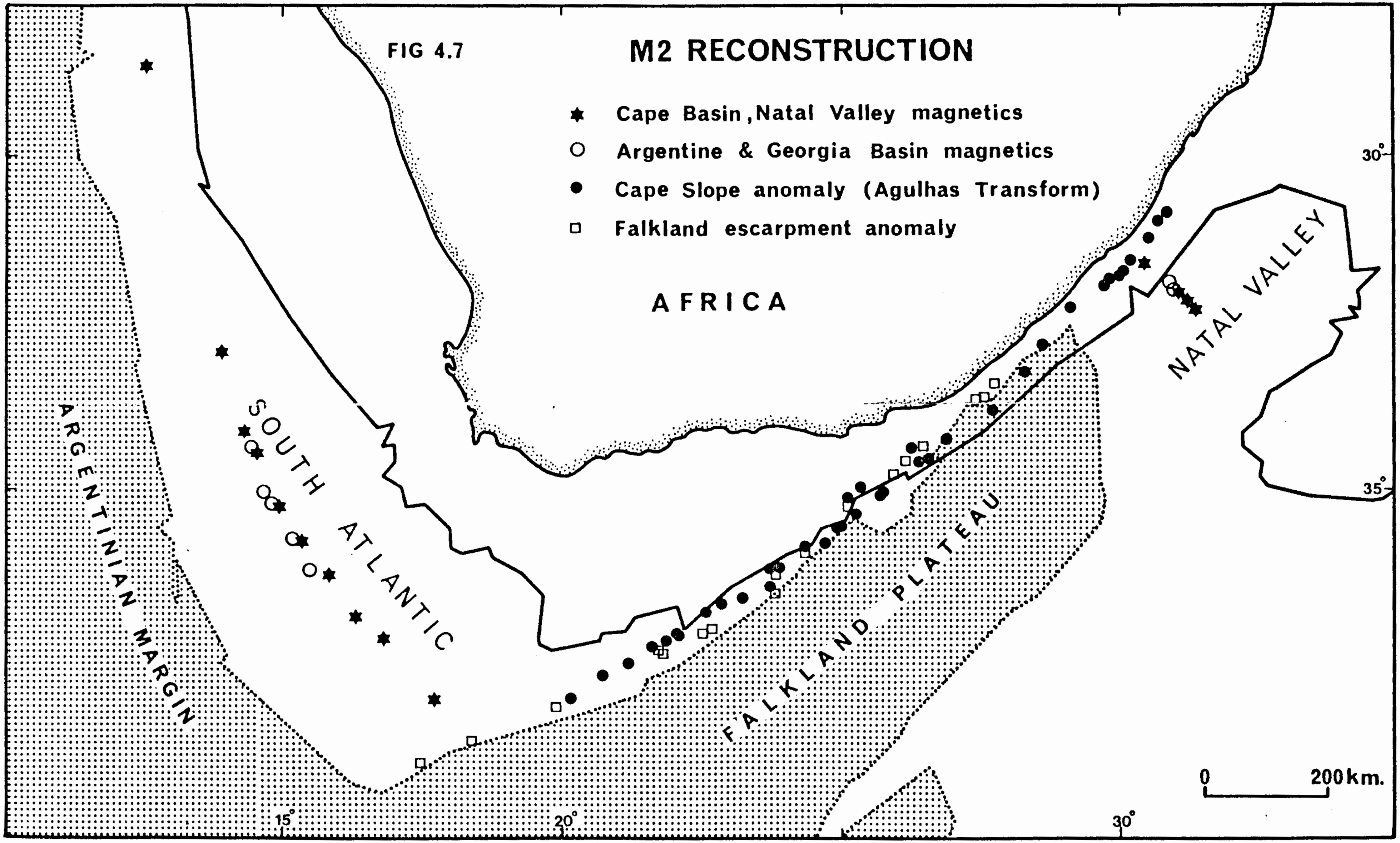
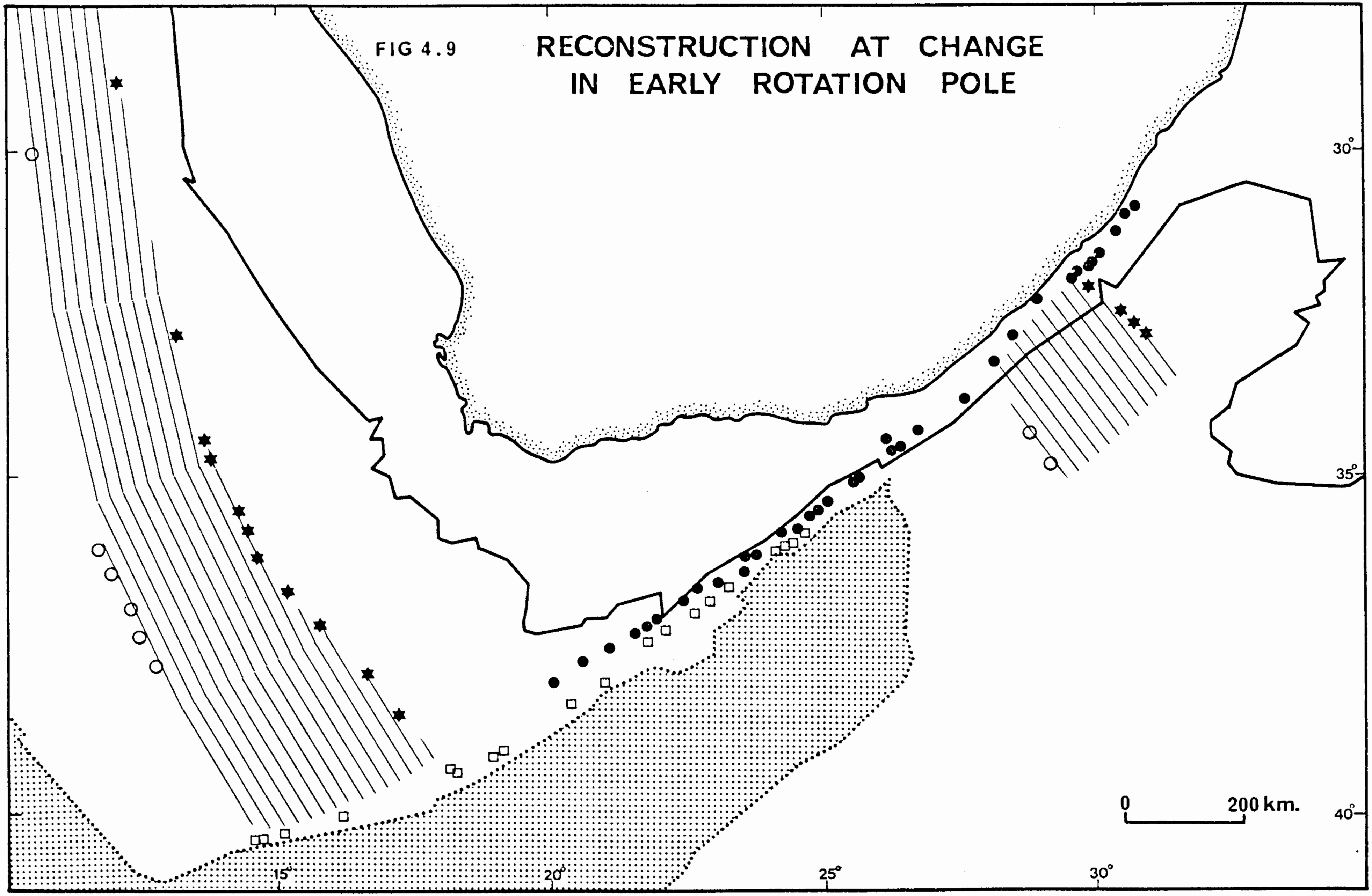


FIG 4.7
M2 RECONSTRUCTION
Cape Basin, Natal Valley magnetics
Argentine & Georgia Basin magnetics
Cape Slope anomaly (Agulhas Transform)
Falkland escarpment anomaly
AFRICA
NATAL VALLEY
SOUTH ATLANTIC
FALKLAND PLATEAU
ARGENTINIAN MARGIN
0
200km.
30°
35°
15°
20°
30°

FIG 4.9


RECONSTRUCTION AT CHANGE IN EARLY ROTATION POLE
30°
35°
40°
15°
20°
25°
30°
0
200 km.

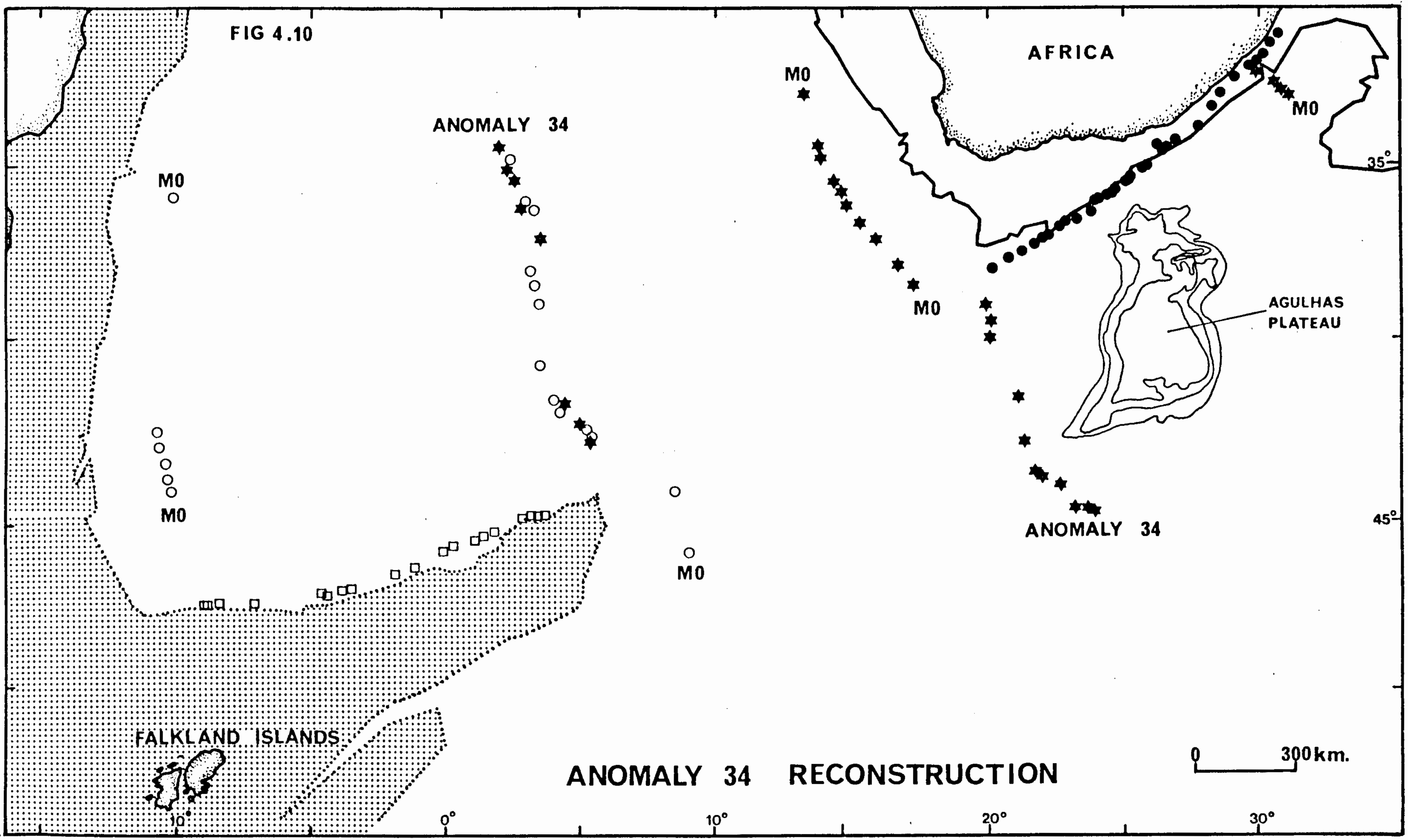


FIG 4.10
AFRICA
AGULHAS
PLATEAU
ANOMALY 34
ANOMALY 34
M0
M0
M0
M0
M0
M0
FALKLAND ISLANDS
35°
45°
0°
10°
20°
30°
10°
0
300km.
ANOMALY 34 RECONSTRUCTION

Note that off the Argentine margin, identification of magnetic anomalies between M4 and anomaly G were not previously made (LaBrecque & Rabinowitz, 1977; Rabinowitz & LaBrecque, 1979). Tentative identifications of anomaly M10 have been made here (Fig.4.3) in order to complete Figure 4.5. The positions of Mesozoic anomalies in the Cape & Argentine Basins (Fig.4.3; LaBrecque & Rabinowitz, 1977; Rabinowitz & LaBrecque, 1979) were also plotted in Figures 4.4-4.9.

4.3.3 Results

Palaeopositions for M10, M4, M2 and M0 times are shown in Figures 4.5-4.8. Cape and Argentine Basin anomalies were effectively juxtaposed by the rotations which fitted equivalent anomalies south of the FAFZ. This is evident from Table 4.1 where the rotations used here in the M0 and M2 reconstructions are compared with earlier M0 and M2 fits obtained using South Atlantic data (Larson & Ladd, 1973; Rabinowitz & LaBrecque, 1979). This indicates that the subjective identification of Natal Valley anomalies is consistent with the South Atlantic interpretation: seafloor spreading in the Cape and Argentine Basins began and proceeded simultaneously with that in the Georgia Basin and Natal Valley. Use of the new early pole of rotation - 6°N, 2°W - or that of Rabinowitz & LaBrecque (1979) - 3.5°N, 2.6°W - produces essentially similar results: a) the sheared Agulhas and Falkland margins slide past each other as the continents move apart (compare Figs.4.5-4.8 with Fig.4.9); thus, using Indian Ocean data to calculate an early pole in conjunction with Natal Valley (Indian Ocean) magnetic anomalies, we can outline the early opening of the South Atlantic Ocean; b) early South Atlantic opening proceeded in a scissors-like fashion causing compression along the Venezuelan and Guinea margins, and explaining the late opening of a connection between the North and South Atlantic Oceans in this region (Rabinowitz & LaBrecque, 1979). Within the limits of navigational accuracy and bearing in mind that the wavelengths of these anomalies are up to 60 km, Figures 4.5-4.8 indicate that the offset in

Table 4.1

Finite poles of rotation, Cretaceous reconstructions of South America and Africa

Description	Time (m.y. B.P.)	Pole of Rotation		Rotation Angle	Source
Pre-drift	pre 130	45.5°N	32.2°W	57.5	(1)
Pre-drift		44.0°N	30.6°W	57.0	(2)
Pre-drift	pre 122-127	46.75°N	32.65°W	56.40	(3)
M10	122	47.53°N	32.94°W	55.74	(4)
M4	117	49.33°N	33.67°W	54.30	(4)
M2	113	50.50°N	34.16°W	53.42	(4)
M2	113	46.27°N	31.08°W	52.91	(5)
M0	108	51.78°N	34.74°W	52.51	(4)
M0	108	52.19°N	34.48°W	52.61	(1)
Pole change		55.18°N	35.7°W	50.90	(1)
34	80	64.84°N	38.01	33.9	(6)

(1) - Rabinowitz and LaBrecque (1979)

(2) - Bullard et al., (1965)

(3) - Chapter 3

(4) - This section

(5) - Larson and Ladd (1973)

(6) - Bergh and Barrett (1980)

Time-scale from M0 - pre-drift from Larson and Hilde (1975). Anomaly 34 from LaBrecque et al., (1977).

spreading ridges of ~1300 km along the FAFZ remained constant from the time of M10 to M0 (122-108 m.y. ago - Larson & Hilde, 1975). The lack of reversal isochrons in the Cretaceous Normal Polarity Epoch prevents accurate palaeo-reconstructions for this period and the next well-constrained fit is obtained by superposing conjugate sets of anomaly 34 (80 m.y. on the time-scale of LaBrecque et al., 1977). Anomaly 34 identifications in the Cape Basin and south of the Agulhas Fracture Zone (du Plessis, 1977; Barker, 1979; LaBrecque & Hayes, 1979; Bergh & Barrett, 1980) indicate that by this time the offset along the FAFZ was ~1270 km - only slightly less than the original offset (Fig.4.10).

South of the FAFZ, 1180 km of oceanic crust exists between anomaly M0 in the Natal Valley and anomaly 34 in the Agulhas Basin. This suggests a total of 2360 km of crust was produced in the Cretaceous Normal Polarity Epoch. This agrees very well with the value of 2350 km predicted from a consideration of spreading rates (Barker, 1979). Thus two lines of evidence prove that only slight asymmetry (maximum 30 km) or a small ridge-jump adjustment in the spreading regime occurred between 108 and 80 m.y. ago (between M0 and anomaly 34).

4.3.4 Ridge-jumps in the Falkland Plateau Natal Valley (F.P.N.V.) spreading compartment

The present-day offset along the FAFZ is appreciably shorter than the original offset indicating that phases of asymmetric spreading or westward ridge-jumps have occurred. Du Plessis (1977) discovered a fossil ridge centred on anomaly 26, and estimated the associated ridge-jump to be 850 km. Assuming an original offset of only 1130 km, he considered this jump sufficient to reduce the offset to the present-day value given as 280 km. Barker (1979) suggested that two ridge-jumps between anomalies 28 and 27 accounted for a westward move in the spreading centre of ~800 km. These jumps were considered to have caused elevated crust at the site of the new spreading centre in the

Meteor Rise area (Barker's elevations W & X). Continued seafloor spreading split the elevated feature in half and the western half (Barker's elevation X or the Islas Orcadas Rise of LaBrecque & Hayes, 1979) now lies in the Georgia Basin (Figs.5.2 & 5.3). Estimating the original offset as 1400 km and the present-day one as 200 km, Barker calculated that the ridge must have jumped a total of 1200 km to the west. Only 800 km is accounted for by the above jump, pointing to the possibility of a 400 km ridge-jump in the Cretaceous Quiet zone. Such a jump causing an elevated area of oceanic crust has been invoked as the origin of the Agulhas Plateau (Barker, 1979; Tucholke et al., 1981). (However, see chapter 5). Interpreting the magnetic data differently, LaBrecque & Hayes (1979) placed anomaly 33 farther from anomaly 34 south-west of the Agulhas Plateau. This implies asymmetric spreading between anomalies 34 and 31, reducing the offset along the FAFZ by 280 km. In conjunction with a discreet ridge-jump of 825 km, this led to a total reduction in offset of 1105 km between anomalies 34 and 25. This interpretation is based on a larger data-set and is corroborated by complementary surveys of Bergh & Barrett (1980). LaBrecque & Hayes estimated the original offset as 1400 km and noted that 1105 km of westward movement of the ridge is only 55 km less than needed to produce the present-day offset given as 240 km.

Clearly, existing estimates of original and present-day offsets vary. Data presented here show that the original offset was 1300 km (Figs.4.4-4.8) while the offset was 1270 km at anomaly 34 times. Accepting the anomaly 34, 33 identifications of LaBrecque & Hayes (1979) because they are corroborated by Bergh & Barrett (1980), a total asymmetry of 1105 km occurred in the Late Cretaceous and Palaeocene. One may therefore predict that the present-day offset should be 165 km (1270-1105 km). Since "prediction is the legitimate child of forward looking quantitative analysis" - Ross Gunn, it was decided to check the present-day offset. Previous estimates varied because of the paucity of data close to the FAFZ in the southern Atlantic near the mid-ocean ridge. Recently obtained magnetic profiles near to the FAFZ show that the present-day offset is in fact only 170 km (Martin et al., 1982c).

The total offset reduction along the FAFZ can therefore by accounted by 1105 km of asymmetry occurring between anomalies 34 and 25. This also rules out any ridge-jumps in the Cretaceous Normal Polarity Epoch (section 4.3.3).

4.3.5 Dating of events in the Cretaceous Normal Polarity Epoch

Dating M0 and anomaly 34 as 108 and 80 m.y. BP respectively, and having ruled out the possibility of a ridge-jump in this period (sections 4.3.3 and 4.3.4), the 1180 km of crust between M0 and anomaly 34 implies an average half-spreading rate for the Cretaceous Normal Polarity Epoch in this region of 4.2 cm yr^{-1}. Four important events in this time (Tables 4.2 & 5.1) are dated with greater confidence using the new anomaly identifications presented here (section 4.2):

1) The change in early pole of rotation which was mis-dated by Rabinowitz & LaBrecque (1979).

2) A reconstruction based on juxtaposing salt boundaries in the Brazil and Angolan Basins which mark the effective permanent flooding of these areas.

3) Final separation of the Falkland Plateau and the southern tip of Africa, Allowing the establishment of deep sea-ways between the Indian and Atlantic Oceans.

4) Formation of the northern, oceanic part of the Agulhas Plateau (Tucholke et al., 1981).

This last point will be dealt with fully in chapter 5.

Using the same time-scales as used here (M0 = 108 m.y. and anomaly 34 = 80 m.y.) Rabinowitz & LaBrecque (1979) proposed that a change in the early opening pole for South America and Africa occurred at 107 m.y. They assumed (pp.5977 and 5984) that a lineated magnetic anomaly ∿50 km north of and subparallel to the Falkland Escarpment, represents the continuation of the Agulhas Fracture Zone. Their pole change position is given as the point

Table 4.2

Dating of important events in the Cretaceous
Normal Polarity Epoch

EVENT		AMOUNT OF POST M0 CRUST PRODUCED (km)	DATE CALCULATED (m.y.)
Pole change		∿250	105
Salt Boundary		∿360	103.7
Falkland Plateau clears tip of Africa		∿820	98.3
Agulhas Plateau	Older Limit	∿900	97.3
	Peak of Volcanism?	∿1265	93.0
	Younger Limit	∿1450	90.7

Dates calculated using the average half-spreading rate of 4.2 cm yr^{-1} for the FPNV spreading compartment during the Cretaceous Normal Polarity Epoch. The amounts of post M0 crust allow the events to be re-dated using any time-scale.

where the lineated magnetic anomaly intersects the western end of the Falkland Escarpment. A reconstruction using their pole change position (Fig.4.9, Tables 4.1, 4.2) shows that the conjugate M0 anomalies are 250 km apart in the Natal Valley. If this reconstruction represents the situation 107 m.y. BP, then a half-spreading rate of 12.5 cm yr^{-1} in the period 108-107 m.y. is required to explain 250 km of post M0 crust. Although a short burst of spreading at 12.5 cm yr^{-1} is not impossible it seems highly unlikely. The half-spreading rate in the period M4-M0 was 1.3 - 1.4 cm yr^{-1} (section 4.2) while the spreading rate calculated between anomalies 34 and 33 on the African plate south of the FAFZ is ~5.5 cm yr^{-1} (Bergh & Barrett, 1980; and calculated from LaBrecque & Hayes, 1979). If the spreading rates for the M4-M0 and anomaly 34-33 intervals are extrapolated towards the pole-change reconstruction this event is dated as 99-100 m.y. BP. Alternatively if the average spreading rate of 4.2 cm yr^{-1} calculated for the Normal Polarity Epoch is used, the pole change is dated as 105 m.y. (Table 4.2). Similarly the '106 m.y.' reconstruction of Rabinowitz & LaBrecque (1979) which juxtaposes the limits of salt deposition in the Angola and Brazil Basins is dated by the two alternative criteria above as 96-97 m.y. or 103.7 m.y. If salt deposition was replaced in the Late Albian (Pautot et al., 1973) then flooding may have been caused by the opening of a connection between North and South Atlantic oceans in early Late Albian time (Kennedy & Cooper, 1975). However, the salt is considered to be no older than early Late Aptian (Bolli, Ryan et al., 1978) with floodwaters coming from the south (Scheibnerova, 1978). Of course the datings of Table 4.2 are based on interpolation between the two end points of the Cretaceous Normal Polarity Epoch and therefore depend a) on the assumption of an average spreading rate and b) on the time-scales used. If the magnetic reversal/biostratigraphic correlation of Lowrie (1982) is modified to incorporate the biostratigraphic/radiometric dating of Odin (1982) the salt boundary reconstruction is dated at the Aptian/Albian boundary. As discussed in the introduction to part 3, detailed discussion of new time-scales is beyond the scope of this study.

4.3.6 Correspondence of magnetic and micropalaeontological dating relating to break-up

The pre-drift reconstruction presented here (chapter 3) is considered the tightest fit possible because it juxtaposes Precambrian outcrops in north-eastern Brazil with equivalent rocks in the Gulf of Benin (Fig.3.4b). This fit juxtaposes broad positive magnetic anomalies just landward of M10 (Fig.4.4). Comparison of the Larson & Hilde (1975) time-scale with this full fit position dates initial separation of the continents at 124-127 m.y. (Figs.4.2, 4.3, 4.11). Anomaly M10 is considered the earliest undeformed seafloor spreading anomaly both south of the FAFZ in the Natal Valley and Georgia Basin, and immediately north of the FAFZ in the southernmost Cape Basin. Considering reconstructions shown in Figures 4.4 and 4.5 as limits within which the pre-drift fit must lie, initial separation is dated as 122-127 m.y., with true seafloor spreading beginning 122 m.y. ago. This is in reasonable accord with sedimentological and micropalaeontological work (Du Toit, 1979; McLachlan & McMillan, 1979). Synthesising published work, including oil exploration data, these authors conclude that the rifting stage preceding drift or final separation is bounded by horizons D and C in the Pletmos basin of the Agulhas Bank. The onset of rifting marked by horizon D is dated by Du Toit (1979) as Middle to Upper Jurassic (specifically 180-145 m.y.) or by McLachlan & McMillan (1979) as Portlandian (∿140 m.y.). The block and step faulting of the rift stage stopped at horizon 'C' which is underlain by strata not younger than Hauterivian and dated as 121 m.y. An earlier widespread unconformity, horizon B in the Pletmos Basin is dated as Mid-Valanginian (128 m.y.). The preferred date for the onset of drifting is thus 121 m.y. but is possibly marked by the older unconformity dated at 128 m.y. (A possible flaw in this argument is that it is based on the break-up model of Falvey (1974) in which a "break-up" unconformity is tectonically induced. It has since been shown that in the type area for this model (Australia) the 'break-up' unconformity may in fact be due to a global sea-level fluctuation (Cande & Mutter, 1982)).

Near Port St. Johns, the Egosa and related faults offset the coastline and margin. A 300 m scarp marks the faulted contact between the Early Palaeozoic Table Mountain Group and the Ecca Group. On the Maurice Ewing Bank a small bathymetric offset occurs at 42.5°W (Fig.4.11) and this and the Egosa Fault are lined up on the pre-drift reconstruction. (A larger offset of the Falkland Escarpment at 46.4°W is discussed in chapter 3). Small outliers of Late Valanginian conglomerates (124-126 m.y.) are considered the earliest sediments deposited in half-graben delineated by the Egosa and related faults near Port St. Johns (McLachlan et al., 1976; Klinger & Kennedy, 1979). Open marine fossils indicate that the conglomerate was dumped directly into the sea, while some brecciation and distorted lenticles demonstrate contemporaneous movement along the boundary fault. The Egosa Fault appears to have been a pre-drift (possibly syn-rift) fault offsetting the FAFZ while it was still an intra-continental shear zone. As movement along the shear zone occurred, tension across the small offset allowed the formation of a small parallelogram shaped basin, with the Egosa Fault outlining a half-graben (compare inset of Fig.4.5 to chapter 7, Fig.7.4c). The micropalaeontological stratigraphy is consistent with the tectonic framework based on interpretation of magnetic anomalies dated on the time-scale of Larson & Hilde (1975): initial marine sedimentation in half-graben at Port St. Johns dates from at least 124-126 m.y. while onset of movement may be dated magnetically as 124-127 m.y. or 122.127 m.y. A regional unconformity marked by horizon B in the Pletmos Basin (128 m.y.) may mark this event. Final separation is marked by horizon C (121 m.y.) while undeformed magnetic anomalies date from 122 m.y. (M10).

Seismic profiling and drilling data from the Agulhas Bank (Du Toit, 1977) indicate that the Agulhas marginal fracture ridge (Scrutton & Du Plessis, 1973) was gradually uplifted in Hauterivian-Albian times. This coincides with

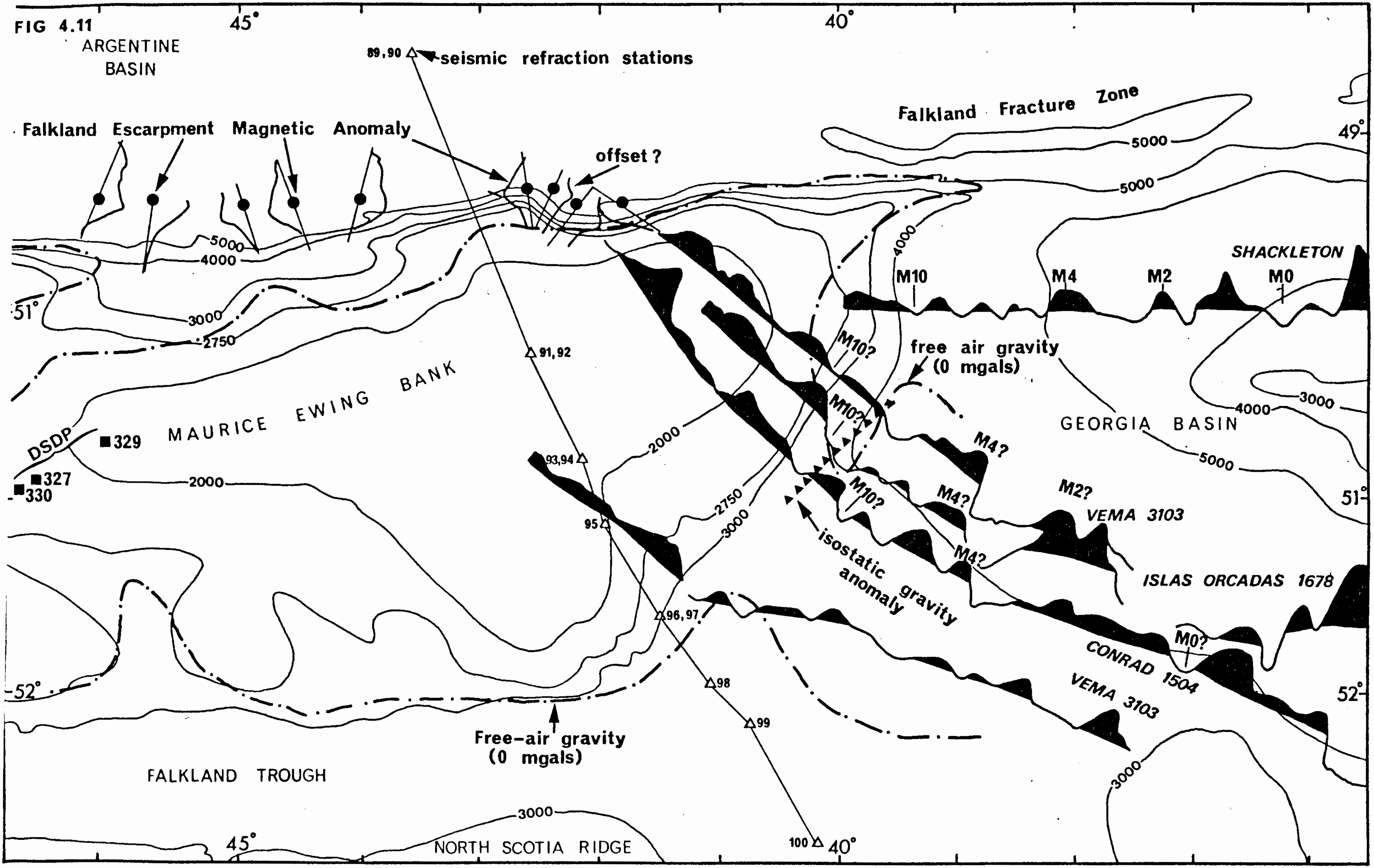
FIGURE CAPTIONS

Fig.4.11. Magnetic, gravity, seismic refraction, and bathymetric data for the Maurice Ewing Bank and environs (after Ewing et al., 1971; Rabinowitz, 1977; LaBrecque & Hayes, 1979; Barker, 1979). Positive free-air gravity anomalies of the Maurice Ewing Bank are separated from the Georgia Basin negative gravity area by a seaward slope in the anomaly, here indicated by the 0 mgal contour. Seismic refraction data suggest the COB lies south-east of station 97.

Fig.4.12. Three interpretations of the COB in the Cape Basin around 30°S - a region that may have been a 'locked zone' (sensu Courtillot, 1982). The vertical dashed line shows the COB inferred from the pre-drift reconstruction of chapter 3. a) magnetic and gravity profiles of Rabinowitz & LaBrecque (1979) showing anomaly G. b i) schematic representation of a COB with minimum continental stretching, b ii) and iii) based on gravity models, suggest an attenuated continental crust zone 200 km in width (after Emery et al., 1975; Scrutton, 1978).

Fig.4.13. Propagating rifts as a natural consequence of rotations of continents about a pole. 1) pre-rift. 2) & 3) propagation of rift proceeds without the necessity of increasing continental crustal stretching. 4) rifts propagating in opposite directions.

FIG 4.11



ARGENTINE BASIN
seismic refraction stations
89,90
91,92
93,94
95
96,97
98
99
100
Falkland Escarpment Magnetic Anomaly
offset ?
Falkland Fracture Zone
SHACKLETON
M10
M4
M2
M0
M10?
M4?
M2?
M0?
free air gravity (0 mgals)
GEORGIA BASIN
VEMA 3103
ISLAS ORCADAS 1678
CONRAD 1504
isostatic gravity anomaly
MAURICE EWING BANK
DSDP
329
327
330
Free-air gravity (0 mgals)
FALKLAND TROUGH
NORTH SCOTIA RIDGE
5000
4000
3000
2750
2000
45°
40°
49°
51°
52°

FIG 4.12

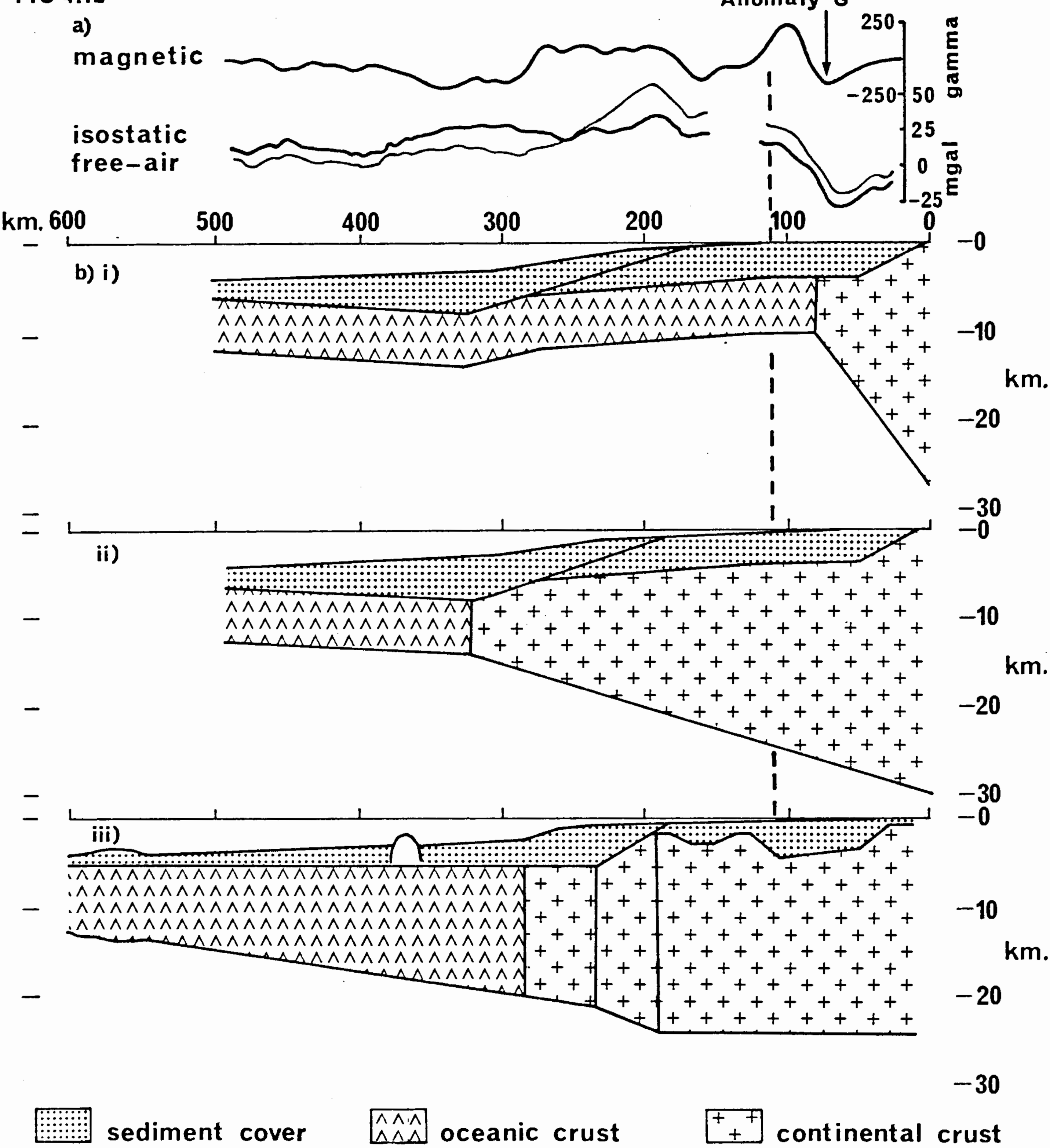
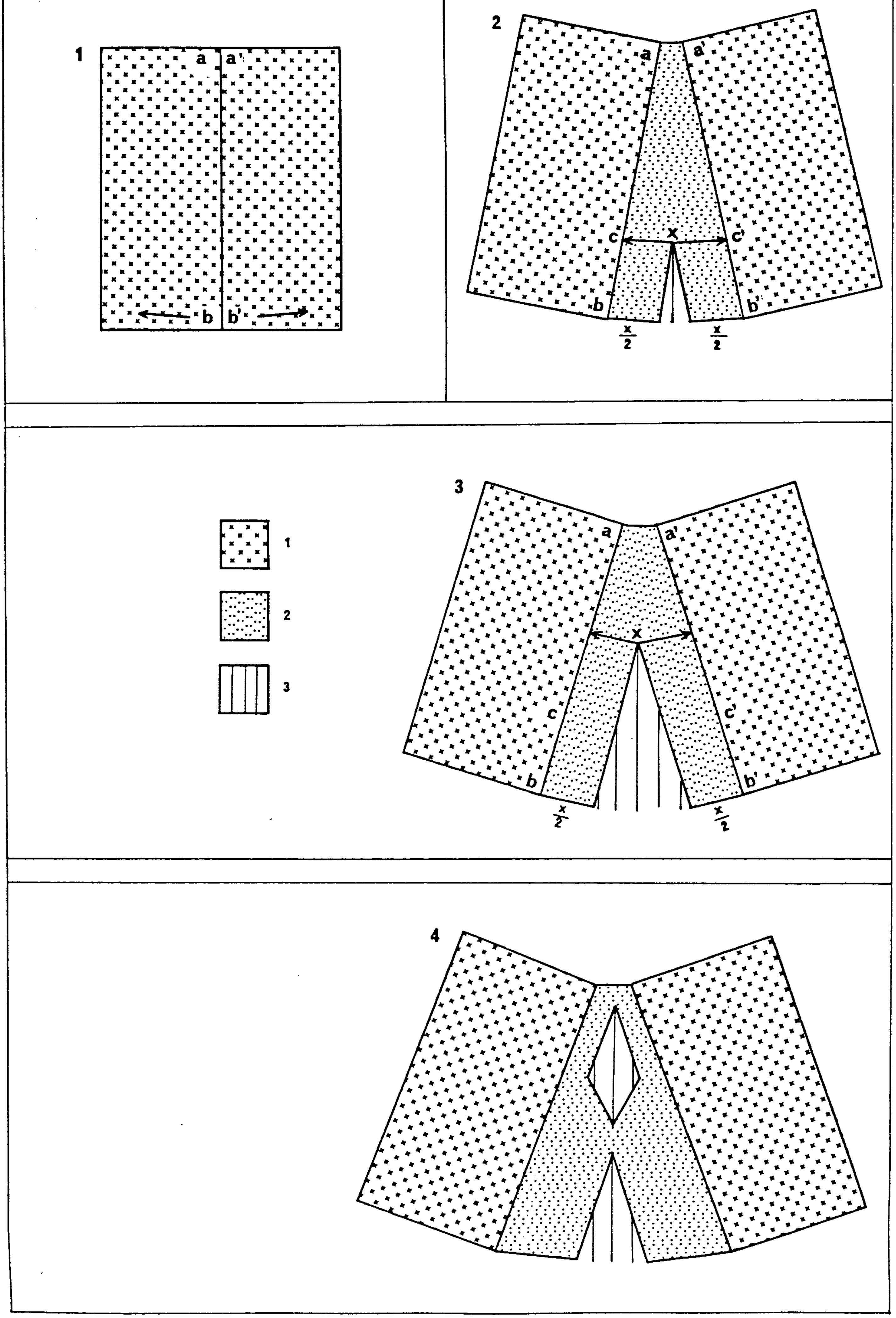

a)
magnetic
isostatic
free-air
Anomaly G
250
-250
gamma
50
25
0
-25
mgal
km. 600
500
400
300
200
100
0
b) i)
ii)
iii)
-0
-10
-20
-30
km.
sediment cover
oceanic crust
continental crust

FIG. 4.13



the period of intra-continental shearing (122-98 m.y.) during which time slivers of crust may be expected to be raised within a strike-slip zone.

4.4 CRITIQUE OF ALTERNATIVE METHODS FOR PRODUCING PALAEOCONTINENTAL RECONSTRUCTIONS

Having a) presented a revised reconstruction of West Gondwanaland (chapter 3); b) described a new interpretation of magnetic profiles in the Natal Valley leading to identification of Mesozoic seafloor spreading anomalies (sections 4.1 and 4.2); and c) used these anomaly identifications to produce Cretaceous palaeopositions of South America and south central Africa (section 4.3), it is appropriate to discuss the various refits of West Gondwanaland which have been proposed (Table 3.1).

It is important to distinguish between reconstructions which recreate the positions of the continents at the onset of seafloor accretion (pre-drift refits) and reconstructions which attempt to replace the continents to pre-rift relative positions.

Initial quantitative attempts at West Gondwanaland reconstruction were based on fitting selected isobaths and minimising the areas of gaps or overlaps between them (Bullard et al., 1965). During the evolution of continental margins over tens of millions of years, variable marginal subsidence or sedimentary outbuilding may substantially modify the original rift boundaries. Assuming that a chosen isobath marks the continent edge is therefore questionable, and reconstructions based solely on quantitative fitting of isobaths are subject to inherent inaccuracy. This point is strikingly illustrated in the case of the Mozambique coastal plain (chapter 6) and its West African analogue, the Niger Delta.

The palaeo-magnetic method (Valencio et al., 1983) is based on obtaining a virtual magnetic pole from both South American and African Jurassic rocks and closing the apparent polar wander curves of the two continents assuming a finite rotation. The rotation proposed (Table 3.1) is very similar to that needed to produce an M2 fit (Table 4.1). The refit parameters of Valencio et al., (1983) are not consistent with seafloor spreading magnetic anomaly identifications associated with the South Atlantic (Rabinowitz & LaBrecque, 1979; Barker, 1979; section 4.2). This may be due to convergence between South America and Africa between the Jurassic and the Early Cretaceous (Valencio et al., 1983).

Rabinowitz & LaBrecque (1979) produced a refit by accurately matching geophysically defined continent-ocean-boundary (C.O.B.) effects. Matching of geophysical C.O.B. 'markers' is limited in that model solutions for potential field data are non-unique. For example, the identification of anomaly 'G' as a C.O.B. marker (Talwani & Eldholm, 1973; Rabinowitz, 1976) has been criticised because alternative models based on gravity data suggest continental crust extends farther sea-ward (Fig.4.12); Emery et al., 1975; Scrutton, 1978; Karner & Watts, 1982). Off the west coast of Australia, where acoustic basement under the margin is not deeply buried because of low sedimentation rates, an alternative geophysical C.O.B. effect has been postulated (Roots et al., 1979). There, the C.O.B. is considered to be marked by a seaward gradient in the Free Air gravity anomaly and a positive magnetic anomaly aligned with a change in the character of acoustic basement. In contrast anomaly 'G' is a landward gradient of the isostatic gravity anomaly aligned with a negative magnetic anomaly measured on margins where fast sedimentation accumulation has blanketed the margin.

An alternative method of continental reconstruction is based on the recognition and precise matching of truncated large-scale tectonic structures which are transverse or significantly oblique to the younger rifted margins. This method is similar to a technique (Ramsey, 1969; p.68-72) for the measurement of strain and displacement across orogenically deformed zones and depends on the identification of linear features common to both sides of the deformed or rifted zone. The problem with this method lies in the recognition of tectonic features which extend beyond the region deformed by rifting, and in establishing whether these were distinctly rectinilear or uniformly curvilinear. It may be argued that continental rifts or shear zones may naturally tend to propagate through "notches" or "kinks" on older cross-cutting tectonic structures. Here, it is assumed that this is not the case and that the pre-drift tectonic structures discussed above were aligned linear features prior to their disruption by separation of the continents. If this is valid, then an accurate fit of the continents can be obtained, independent of the controversy surrounding the positioning of the C.O.B. on sediment-draped margins. Secondly, a fit resulting from matching of pre-drift tectonic features may provide independent constraints on the position of the C.O.B.

Reconstructions which match the C.O.B. as it is in the present-day differ from refits aimed at modelling the continental crustal configuration prior to rift, because the continental margin is significantly modified in the rifting process. Models for rifting involve crustal stretching and thinning or subsidence, which is accomplished through uplift and erosion (Dewey & Bird, 1970), metamorphism at the base of the crust (Falvey, 1974), intrusion of basaltic crust (Burke & Whiteman, 1973; Van der Linden, 1975), necking of continental crust (McKenzie, 1978) or through oceanward creep due to gravitational instabilities across young passive margins (Bott, 1973; 1979). Most of these models envisage stretching occurring in the rifting stage, whereas in the Bott model, oceanward creep is most pronounced once oceanic crust begins accreting. Crustal stretching and second order movements constitute a break-down of the rigid-plate hypothesis (chapter 3; Vink, 1982). Thus pre-rift reconstructions are tighter than reconstructions which juxtapose the present-day expression of the C.O.B. (Fig.4.12). The situation is complicated by the recognition that rifting may proceed via a propagating crack (Turcotte & Oxburgh, 1973; Hay, 1977; Hey et al., 1980; Courtillot, 1982; Courtillot et al., 1980; Vink, 1982). Two approaches to a quantitative pre-rift have emerged: a) LePichon & Sibuet (1981) have quantified crustal stretching and provided equations with which to reconstruct the original configuration; in order to do this, multi-channel seismic data similar in quality to their data is required around the continental margin; b) Vink (1982) considers this task untenable and instead he has estimated the crustal extension, considering that in the South Atlantic, extension has increased from south to north.

4.4.1 Comparison with Rabinowitz-LaBrecque fit - a pre-drift reconstruction

Rabinowitz & LaBrecque (1979) infer from their reconstruction of conjugate 'G' anomalies that the rifted continental blocks of South America and Africa behaved rigidly during and after the rifting episode. Overlaps produced by their rotation parameters necessitate non-rigid behaviour such as crustal stretching or second-order intra-plate movements. Post-drifting opening of the Jucano basin south of the Pernambuco lineament (Fig.3.4) <u>may</u> have pushed part of the Pernambuco Alagoas Massif eastwards thus extending eastern Brazil. Unless movement also occurred north of the Pernambuco lineament (c.f. Rand & Mabesoone, 1982), the basement overlap (Fig.3.4a) would not be affected by this. Alternatively post-drifting opening of the Benue trough (Burke & Dewey, 1974) <u>may</u> have moved the Chaillu and N. Gabon massifs south-easterly. It is questionable whether closing the Benue trough would resolve the overlap (Fig.3.4a) since the direction of movement is likely to have been north-west south-east (Burke & Dewey, 1974) or even east-west along strike-slip faults (Benkhelil, 1982; Neev et al., 1982). A third possibility is that crustal stretching along the rifted Atlantic margins has caused the overlap. In rifting models the stretched crust usually subsides whereas the overlap discussed here involves Precambrian rocks exposed onland. The Danakil Horst in the Red Sea is an example of thinned continental crust which is still upstanding but this is associated with a very young ocean where the continental margin may yet subside. In the Cape Basin (Fig.4.2) the earliest negative anomaly may be interpreted as anomaly M12 (Rabinowitz & LaBrecque, 1979 identification rather than that of Larson & Ladd, 1973), but has been considered to be of M13 age by extrapolation of spreading rates (Rabinowitz & LaBrecque, 1979). Off the tip of the Falkland Plateau magnetic anomalies as far back as M10 have been recongised (sections 4.2, 4.3, Fig.4.11). The overlap in the Natal Valley region produced by fitting the 'G' anomaly in the South Atlantic (Fig.3.3) similarly requires explanation by crustal stretching. If stretching occurred prior to final rifting, then extension would be taken up by both South America and south central Africa, since they formed a single West Gondwanaland plate at this stage. Only stretching in the Falkland Plateau/Natal Valley region after break-up in the Cape and Argentine Basins can explain the overlap. Therefore, to accommodate the Rabinowitz-LaBrecque model, 140 km of stretch must occur between M13 (130 m.y.) and M10 (122 m.y.). This gives a rate of stretch of 1.75 cm yr^{-1} which is greater than the subsequent seafloor spreading rate (section 4.2).

Some extension undoubtedly occurred on the Falkland Plateau in order to initiate the various basins, such as the Magellanes and Malvinas Basins (Ludwig et al., 1979) and the Falkland Plateau Basin (Fig.3.5). Like the Falkland Plateau Basin, the other two contain marine rocks dating back to the Oxfordian (Olea & Davis, 1977) and were thus established prior to South Atlantic seafloor spreading.

Another possibility is that the Falkland Plateau Basin was initiated through a limited amount of seafloor spreading and is thus underlain by oceanic crust (Rabinowitz, quoted in Ludwig et al., 1979; Ludwig, Krasheninnikov et al., 1980). The detailed reconstructions of section 4.3 (Figs.4.4-4.10) show that the FAFZ did not change in length in the period M10-M0. This shows that if seafloor spreading did occur in the Falkland Plateau Basin it occurred prior to M10 times and thus prior to South Atlantic spreading. This is in agreement with the stratigraphy (Fig.3.6) which, as noted above and in chapter 3, points to the initiation of these basins in Mid-Late Jurassic times.

A consideration of the magnetic signatures of the Agulhas sheared margin also disagrees with the anomaly 'G' interpretation of the C.O.B. The positive Cape Slope Anomaly is considered to mark the sheared margin between continental and oceanic crust. The Agulhas Fracture Zone south-west of Africa separates oceanic crust of the Cape Basin from oceanic crust in the southern spreading compartment and is marked by a negative magnetic anomaly (Figs.4.4 & 4.5). One crossing shows a very subdued anomaly. This change-over from positive to negative anomalies gives an indication of the extent of continental crust along the shear zone and suggests continental crust of the African Plate extends to 20°E at the tip of the Agulhas Bank. This is in agreement with the C.O.B. proposed below for the rifted margins in the southern South Atlantic.

In the Natal Valley, Mesozoic magnetic anomalies extending south-west from a region associated with rifting, confirm that the southern face of the Tugela Cone marks the C.O.B. This mitigates against the pre-drift palaeoposition of the Falkland Plateau implied by the rotation parameters of Rabinowitz & LaBrecque (1979)(Fig.3.3). Near the tip of the Maurice Ewing Bank, a continental crustal signature extends to seismic refraction station 97 at the south-eastern corner of the Falkland Plateau (Fig.4.11) whereas beyond this at station 99 an oceanic signature was recorded (Ewing et al., 1971; seismic refraction results correlate with seismic reflection and drilling results of Barker, Dalziel et al., 1977). The C.O.B. marker of Roots et al. (1979) and even that of Rabinowitz & LaBrecque (1979)(Fig.4.11) suggest that continental crust lies just landward of M10. (The model of Roots et al. may be appropriate because the tip of the Falkland Plateau resembles Roots' study area, West Australia, in being relatively thinly blanketed in sediment. This is due to a) its removal through seafloor spreading from contiguous supplies of terrestrial detritus and b) erosion and non-deposition due to action of deep-sea currents (Ciesielski & Wise, 1977)). Note also that the identification of Barker (1979) is considered most reliable since the identifications of LaBrecque & Hayes (1979) are described as 'tentative' and do not superpose Natal Valley anomalies in palaeoreconstructions (Fig.4.5).

The revised pre-drift reconstruction implies that in the southernmost Cape Basin the C.O.B. lies at least 70 km seaward of anomaly G. This is corroborated by recent multi-channel seismic profiling (Gerrard & Smith, 1980; Uchupi & Austin, 1981) which delineates the C.O.B. of south-western Africa. In particular anomaly G is related to pinch-out of a pre-drift volcanic wedge (Gerrard & Smith, 1980). The new reconstruction juxtaposes the broad positive anomaly found landward of M10 (Figs.4.2, 4.3, 4.4). I suggest therefore that the rifted margins immediately north and south of the FAFZ show a consistent pattern: the C.O.B. lies between the positions of anomaly M10 (Fig.4.5) and the broad positive magnetic anomaly just landward of that (Fig.4.4). Since the reconstruction of Figure 4.4 is considered the tightest possible (chapter 3), the best pre-drift fit lies between Figures 4.4 and 4.5 (Martin et al., 1982b). This premise has recently been substantiated by Austin & Uchupi (1982) who interpret a transition from oceanic crust to rifted continental crust occurring between anomaly M10 and the peak of the broad positive anomaly just landward of it. This agreement between the proposed pre-drift fit and the delineation of the C.O.B. by seismic profiling suggests that post-drift continental crustal stretch has not been appreciable. Thus the stretching process proposed by Bott (1973, 1979) has not occurred. LePichon & Sibuet (1981) came to a similar conclusion through their modelling of the rifting process.

Further to the north off the Orange River, the refit presented here places the implied C.O.B. only 40 km seaward of anomaly G. Both gravity modelling (Fig.4.12) and multi-channel seismic work (Austin & Uchupi, 1982) places the C.O.B. farther seaward. This area probably marks a 'locked zone' in the scheme of Courtillot (1982) where separation has been accommodated by continued crustal stretching rather than oceanic accretion (see section 4.4.2 below).

4.4.2 Discussion on Vink fit

The recognition that continental rifts propagate implies that continent-ocean-boundaries are not isochrons. Thus the pre-drift fit proposed here (chapter 3) based on a) alignment of pre-break-up geotectonic features and b) the premise that the South Tugela Ridge marks the C.O.B. in the Natal Valley and which c) correctly predicted the position of the C.O.B. in the southernmost Cape Basin, is relegated to a reconstruction which is pre-drift for the areas immediately around the FAFZ. Two models for propagating rifts have been proposed. In the first, rigid plate tectonics is preserved in that as one rift propagates, a second rift recedes (Hey, 1977; Hey & Vogt, 1977). In the case of continental rifting, the equivalent of the receding rift is a scenario where the relative rotation pole migrates ahead of the tip of the rift with a wave of compression migrating ahead of that (Hey et al., 1980). In the second model, rigid plate tectonics is violated in that crustal stretching ahead of the propagating rift increases the size of the plate. Second order movement of minor plates (chapter 3) may also constitute a break-down of the premise of rigid plate tectonics only if considered on a broad scale. If each microplate is considered at the appropriate scale then movement may be modelled in terms of rigid plates. One may envisage a hierarchy of scales on which movement occurs (c.f. Courtillot, 1980). In the case of stretching of continental crust, the plates on either side of the rift increase in size via stretching over a diffuse zone. However, if crustal stretching includes basaltic injection (Burke & Whiteman, 1973; Van der Linden, 1975) there is a fine distinction between plates increasing in size at an accreting plate boundary by seafloor spreading (which occurs over a finite width of the ridge - Luyendyk & Macdonald, 1977) and plates increasing in size at a rift via crustal stretching incorporating basaltic intrusion.

Vink (1982) uses a 'flat earth' model in an attempt to demonstrate that crustal stretching should gradually increase with increased propagation of a rift. In this model the amount of separation between the two continents is equal throughout the length of their eventual margins. Crustal extension ahead of the rift is required to increase so as to compensate for seafloor spreading behind the rift.

If we retain Euler's theorem, accepting that increases in plate size due to crustal extension are equivalent to plate growth at oceanic accreting margins, then a different scenario results (Fig.4.13). Consider a continent which is about to break into two plates (Fig.4.13). As movement occurs, point b at the southern end of the split has moved farther from b' than a has moved from a' (Fig.4.13). The separation of the two plates increases from a to b. Let us assume that the continent is homogeneous along the length a b. In the model of LePichon & Sibuet (1981) the lithosphere asthenosphere boundary rises as lithosphere thinning proceeds until a critical value is reached: partial melting increases due to higher temperatures at the lithosphere asthenosphere boundary and the asthenosphere breaks through to the surface. Let this critical separation distance, x = the separation between c and c'. Therefore between a and c, separation is accommodated by lithosphere extension while between c and b, oceanic accretion occurs. As separation between the plates increases (Fig.4.13) the critical separation value x moves north and oceanic accretion gradually extends northwards. Thus propagating rifts are a natural result of motions about a rotation pole. As in Vink's model seafloor spreading isochrons abut the continent ocean boundary which is not an isochron (Fig.4.13). The difference is that crustal extension is equal along the length of the continental margin. Note that the rotation pole need not be positioned at the propagating rift tip (Hey et al., 1980) but may be ahead of it. Also the propagating rift is modelled to react to a stress situation, which implies that in the seafloor spreading process 'slab-pull' or some other driving mechanism is more important than 'ridge push' (c.f. Hey et al., 1980).

If we then drop one of the assumptions above, that the continent is homogeneous along the length a b, we approach reality. In contrast to the generalised 'Wilson cycle' model, rifts cross older geotectonic features rather than opening along them (eg. chapter 3; Torquato & Cordani, 1981). Thus we expect different parts of the continent to react differently in the rifting process, resulting in the scenario of Figure 4.13.4, where oceanic crustal accretion (asthenosphere break-through) occurs at different areas of the rift zone. Thus oceanic accretion may initiate at different areas and rifts may accrete in different directions. This is similar to the scenario of Courtillot (1982), while rifts propagating in opposite directions have been outlined in the Pacific (Johnson et al., 1983).

4.4.2.2 The South Atlantic case

In the S. Atlantic the configuration of magnetic anomaly isochrons (although those north of $\sim$32°S are tentative) compared with the configuration of the C.O.B. as delineated by multi-channel seismics (Gerrard & Smith, 1980; Austin & Uchupi, 1982) shows that in the Cape Basin, a rift did propagate northwards. It need not have been the only propagating rift involved in South Atlantic opening (Fig.4.13.4).

In the model of Hey (1977), as one rift propagates another recedes, and a pole of rotation migrates ahead of the rift tip. Section 4.4.2.1 suggests the pole may be ahead of the rift tip. Moving poles of rotation have been documented particularly well in the south-west Pacific (Smith, 1981). In the South Atlantic the data have allowed early and late poles to be differentiated, but have not outlined a smooth locus of pole positions. The early pole of rotation (Rabinowitz & LaBrecque, 1979; chapter 3) suggests that opening occurred in a scissors-like fashion. Either some compression is implied along the Guinea and Venezuelan margins (Rabinowitz & LaBrecque, 1979) or

Africa was already in a two-plate state (chapter 3, Pindell & Dewey, 1982). In the latter case, compression may have been experienced along the Pelusium Line (Neev et al., 1982) stretching from the Gulf of Benue towards Israel - a likely area for the boundary between the West African plate and the south central plate. Thus during early propagation of the South Atlantic rift, it is possible that compression or even closure of a rift was experienced either between Guinea & Venezuela or along the Pelusium Line. The implications are a) that rigid plate tectonics may be preserved (c.f. Hey et al., 1980) and b) no continental crustal extension is expected north-west of the pole position during early opening (122-105 m.y. - section 4.3); if massive crustal stretching occurred in Ghana and Guyana (Vink, 1982, Fig.11) it occurred after the change in the early rotation pole.

Vink (1982) considers that it is an impossible task to estimate crustal extension all around the continental margin in order to model its pre-rift shape. It is for this reason that he presents a reconstruction of the continents using their present-day shape. He assumes a) that the South Atlantic rift propagated from the environs of Cape Town towards the Gulf of Benue, and b) because of this,crustal extension should increase from south to north. As demonstrated above, neither of these assumptions need be valid. Thirdly, by reconstructing the continents in their present shape, he misaligns various pre-drift tectonic features.

4.5 SUMMARY

Mesozoic seafloor spreading anomalies M0-M10 have been identified in the Natal Valley south-west of an area associated with rifting (possibly distorted anomalies M10N-M12). The anomalies are perpendicular to the sheared Agulhas margin, striking NW-SE. Half spreading rates of 1.3 - 1.4 cm yr^{-1} and 1.6 cm yr^{-1} are calculated for M0-M4 and M4-M10 respectively. These identifications confirm that a) seafloor spreading north and south of the Falkland Agulhas Fracture Zone (FAFZ) began simultaneously; b) the southern face of the

Tugela Cone marks the C.O.B. in the Natal Valley. The Natal Valley anomalies are off-set ∿1300 km by the FAFZ from their equivalents in the southern Cape Basin. I have computed successive Falkland Plateau palaeopositions using Natal Valley and Georgia Basin anomalies. These show that the offset in spreading ridges at the FAFZ remained ∿1300 km long from M10 to M0 time. By anomaly 34 time, the offset was ∿1270 km. Therefore no major ridge jumps had occurred by then. Dating M0 as 108 m.y. BP and anomaly 34 as 80 m.y. BP, the average half spreading rate immediately south of the FAFZ for the Cretaceous Normal Polarity Epoch is 4.2 cm/yr. Using this, I date (a) the change in early pole of rotation at 105 m.y.; (b) a reconstruction which juxtaposes salt boundaries in the Brazil and Angolan basins at 103.7 m.y.; (c) final separation of the Falkland Plateau from southern Africa at 98.3 m.y.

Comparison of magnetic data with implied C.O.B. positions in the southernmost Cape and Argentine Basins, and the Georgia Basin suggests continental separation began 122-127 m.y. BP with undeformed magnetic anomalies dating from 122 m.y. (M10). This scheme is consistent with micropalaeontological and sedimentological data around southern Africa.

The Rabinowitz-LaBrecque (1979) reconstruction (considered pre-drift) is rejected because a) it necessitates untenable stretching of the Falkland Plateau in post-M10 times and b) the C.O.B. on the southernmost Atlantic margin of southern Africa implied by the refit presented here is corroborated by recent multi-channel seismic work. The Vink (1982) model for estimating a pre-rift reconstruction is rejected because two inherent assumptions in his argument - namely that the South Atlantic rift propagated northwards in its entirety and that crustal extension increases along a propagating rift - are shown to be not necessarily valid.

CHAPTER 5

EVOLUTION OF THE AGULHAS PLATEAU AND ITS RELATIONSHIP TO OTHER ASEISMIC BATHYMETRIC HIGHS

5.1 INTRODUCTION

The northern part of the Agulhas Plateau lies in the same seafloor spreading compartment as the Natal Valley and the Falkland Plateau (F.P.N.V. compartment). Its oceanic character presents no obstacle to the reconstructions of section 4.3. The southern continental part of the Agulhas Plateau lies in the same spreading compartment as the Burdwood Bank (C.A.P.B.B. spreading compartment). The new identifications of seafloor spreading anomalies (section 4.2) and the model for the evolution of the south-west Indian Ocean (section 4.3) have implications for the evolution of the Agulhas Plateau, as well as other aseismic rises in the southern ocean.

The Agulhas Plateau is a large aseismic bathymetric high lying between 250 and 950 km south of South Africa. It is separated from the African margin by the Agulhas Passage and is flanked by the Agulhas and Transkei Basins (Fig.5.1). Its central region is 2000-3000 m deep, rising about 2500 m above the surrounding abyssal areas. A northern triangular shaped area of rough relief is offset to the north-east from a southern rectangular shaped portion of smoother relief.

Regarding the nature of its underlying crust, Scrutton (1973) suggested that it comprises oceanic crust of an abandoned spreading centre. Through seismic refraction and magnetic modelling, Barrett (1977) showed that the northern region is underlain by thickened oceanic volcanic basement. Barker (1979) suggested the excess volcanism of the northern Agulhas Plateau represents the site to which an accreting ridge jumped in the Cretaceous Quiet zone. Recently Tucholke et al. (1981) proposed a continental origin for the southern area of smooth basement. A layer of seismic velocity 5.8 - 6.4 km/sec (minimum thickness 4.3 - 7.7 km) represents a 'granitic layer' indicating continental crust. This interpretation is supported by dredge-hauls of quartzo-feldspathic gneiss of lower amphibolite to granulite facies which give isotopic dates of 498 $\pm$17 and 1105 $\pm$36 m.y. (Allen & Tucholke, 1981). Continental crust is flanked to the west by a narrow south-west trending strip of thickened oceanic material (Fig.5.1), and to the north-west the crust is intermediate between oceanic and continental in character. Patches of south-east trending irregular basement may represent thickened oceanic material intruding continental basement.

5.2 ORIGIN OF THE NORTHERN OCEANIC AREA

Three lines of evidence indicate that no appreciable ridge-jump occurred during the Cretaceous Normal Polarity Epoch in the seafloor spreading compartment between the Falkland Plateau and the Natal Valley (F.P.N.V.): a) the offset along the FAFZ remained 1300 km long from M10-M0 times and was 1270 km by chron 34; b) the 1180 km of crust between M0 in the Natal Valley and anomaly 34 in the Agulhas Basin is very close to half of the expected 2350 km; c) the asymmetry of spreading occurring between anomalies 34 and 25 described by LaBrecque & Hayes (1979) is sufficient to account for the reduction to the present offset of only 170 km (section 4.3).

It is therefore unlikely that the northern oceanic part of the Agulhas Plateau which is up to 175 km wide, was caused by a ridge-jump. Using the half-spreading rate of 4.2 cm yr^{-1} calculated for the Cretaceous quiet zone (section 4.3) the Agulhas Plateau, lying between 450 and 725 km from M0, is dated at 97.3-90.7 m.y. (Table 5.1). The central area is dated as 93.0 m.y., in general agreement with micropalaeontological data which suggests basement of the Agulhas Plateau is Cenomanian (>95 m.y.)(Saito et al., 1974).

Tucholke et al. (1981) proposed that a triple junction formed on the Agulhas Plateau and migrated towards the triple junction indicated by the chevron shape of the anomaly 34 crust to the south-west (Fig.5.1). This suggests that the oceanic and thickened oceanic crust of the Agulhas Plateau is strongly diachronous.

FIGURE CAPTIONS

Fig.5.1. Crustal types of the Agulhas Plateau (after Tucholke et al., 1981). Magnetic seafloor spreading anomalies in the environs are included (references in text).

Fig.5.2. Locations of Islas Orcadas Rise, Meteor Rise and the Agulhas Plateau in the present-day South Atlantic and South-west Indian Ocean.

Fig.5.3. A reconstruction for anomaly 25 times (after LaBrecque & Hayes, 1979) showing Islas Orcadas Rise and Meteor Rise juxtaposed.

Fig.5.4. A reconstruction of the Agulhas Plateau and Islas Orcadas Rise 93 m.y. ago. The oceanic parts of these features are diachronous (Table 5.1), and thus they partly overlap in this refit. Islas Orcadas Rise and Falkland Plateau delineated by the 3000 m isobath. Agulhas Plateau by 3000 and 4000 m isobaths. Reconstruction parameters: pole 59.77°N, 36.63°W: 41.05.

Fig.5.5. Pre-drift, M0 and pole-change reconstructions of the Falkland Plateau relative to the Agulhas Plateau shown in its present position relative to Africa. The Falkland Islands and Argentina are shown in the pole-change position. These reconstructions suggest that the continental part of Agulhas Plateau rifted from Burdwood Bank (south of the Falkland Plateau) between 108 and 105 m.y. ago.

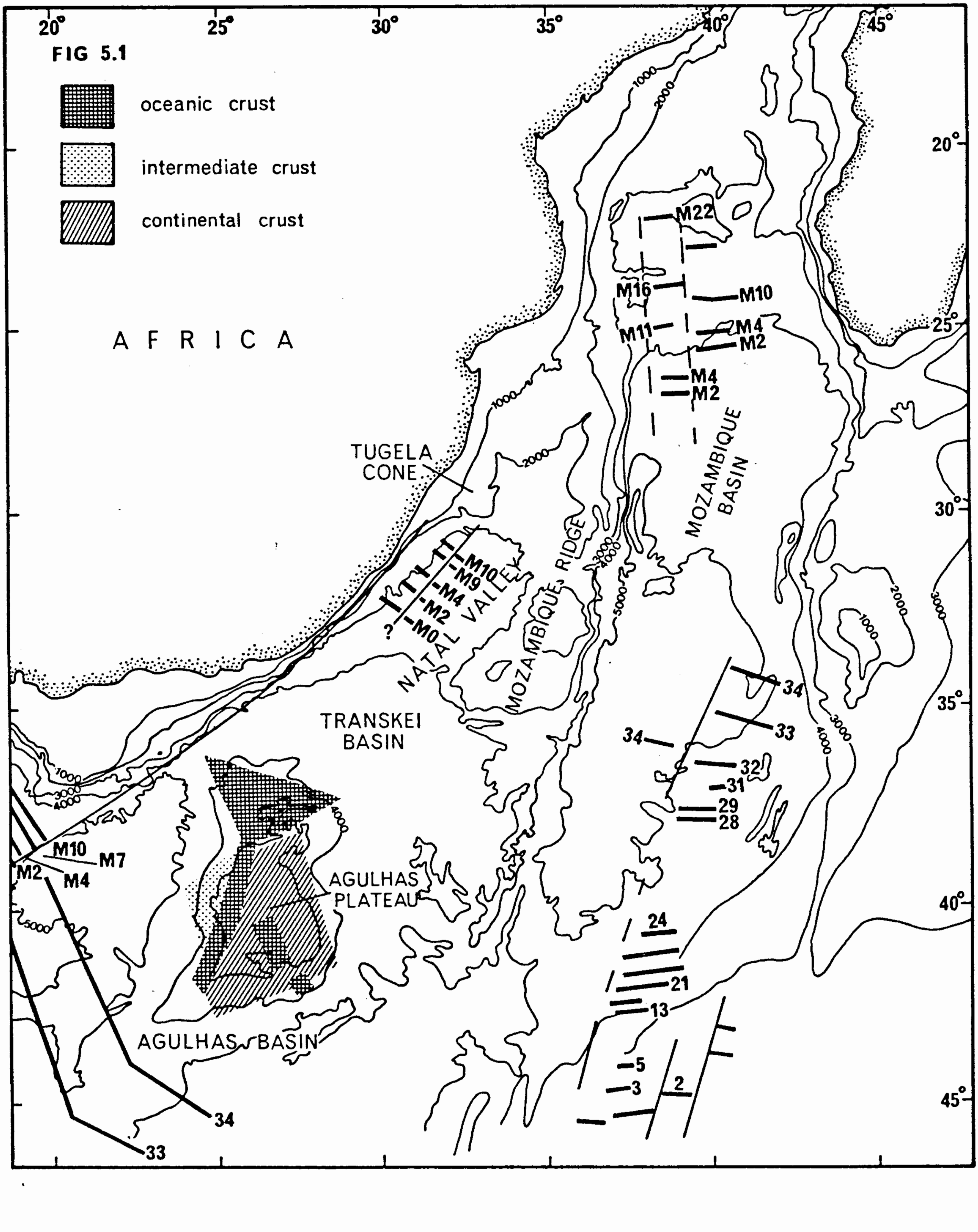


FIG 5.1
oceanic crust
intermediate crust
continental crust
AFRICA
TUGELA CONE
NATAL VALLEY
MOZAMBIQUE RIDGE
MOZAMBIQUE BASIN
TRANSKEI BASIN
AGULHAS PLATEAU
AGULHAS BASIN
20°
25°
30°
35°
40°
45°
M22
M16
M10
M11
M4
M2
M9
M0
M7
34
33
32
31
29
28
24
21
13
5
3
2
1000
2000
3000
4000
5000

FIG 5.2

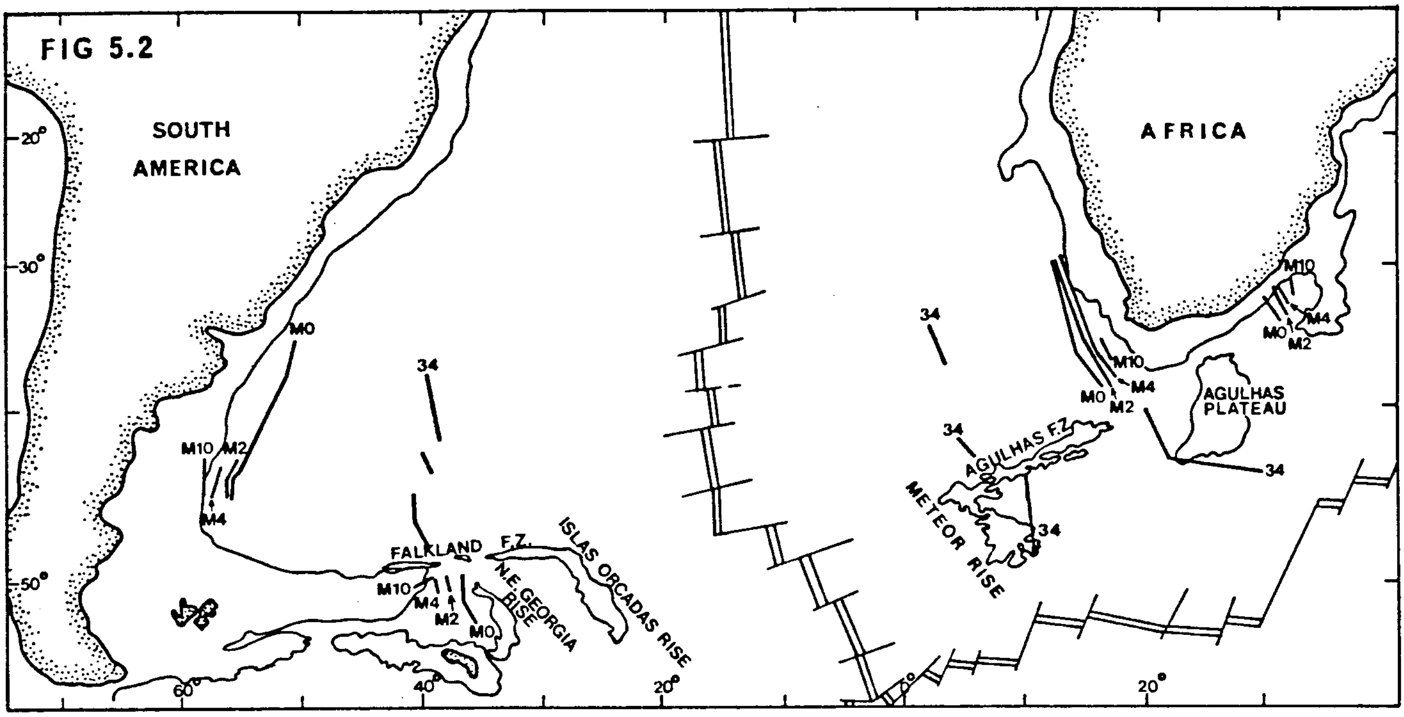
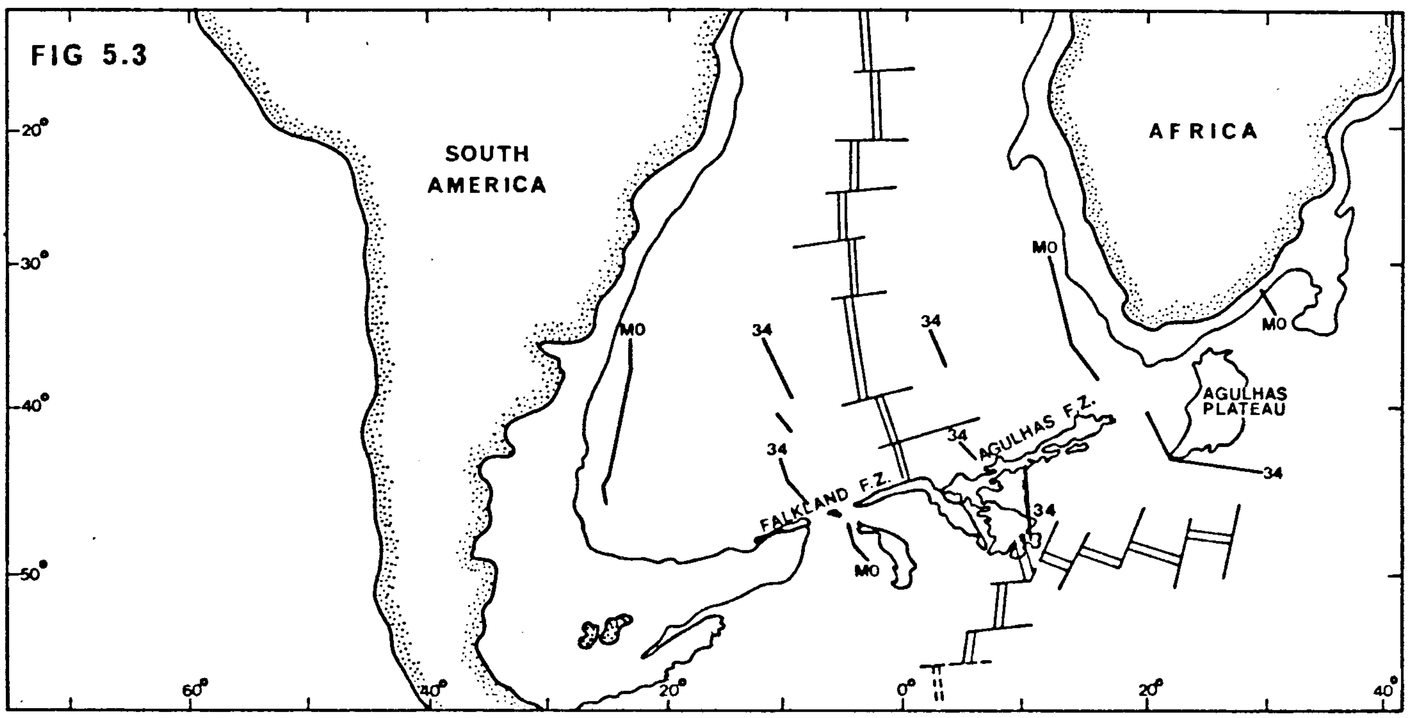

SOUTH AMERICA
AFRICA
M0
M10
M2
M4
34
FALKLAND F.Z.
N.E. GEORGIA RISE
ISLAS ORCADAS RISE
AGULHAS F.Z.
METEOR RISE
AGULHAS PLATEAU
20°
30°
50°
60°
40°
20°
0°
20°

FIG 5.3


SOUTH AMERICA
AFRICA
M0
34
FALKLAND F.Z.
AGULHAS F.Z.
AGULHAS PLATEAU
20°
30°
40°
50°
60°
40°
20°
0°
20°
40°

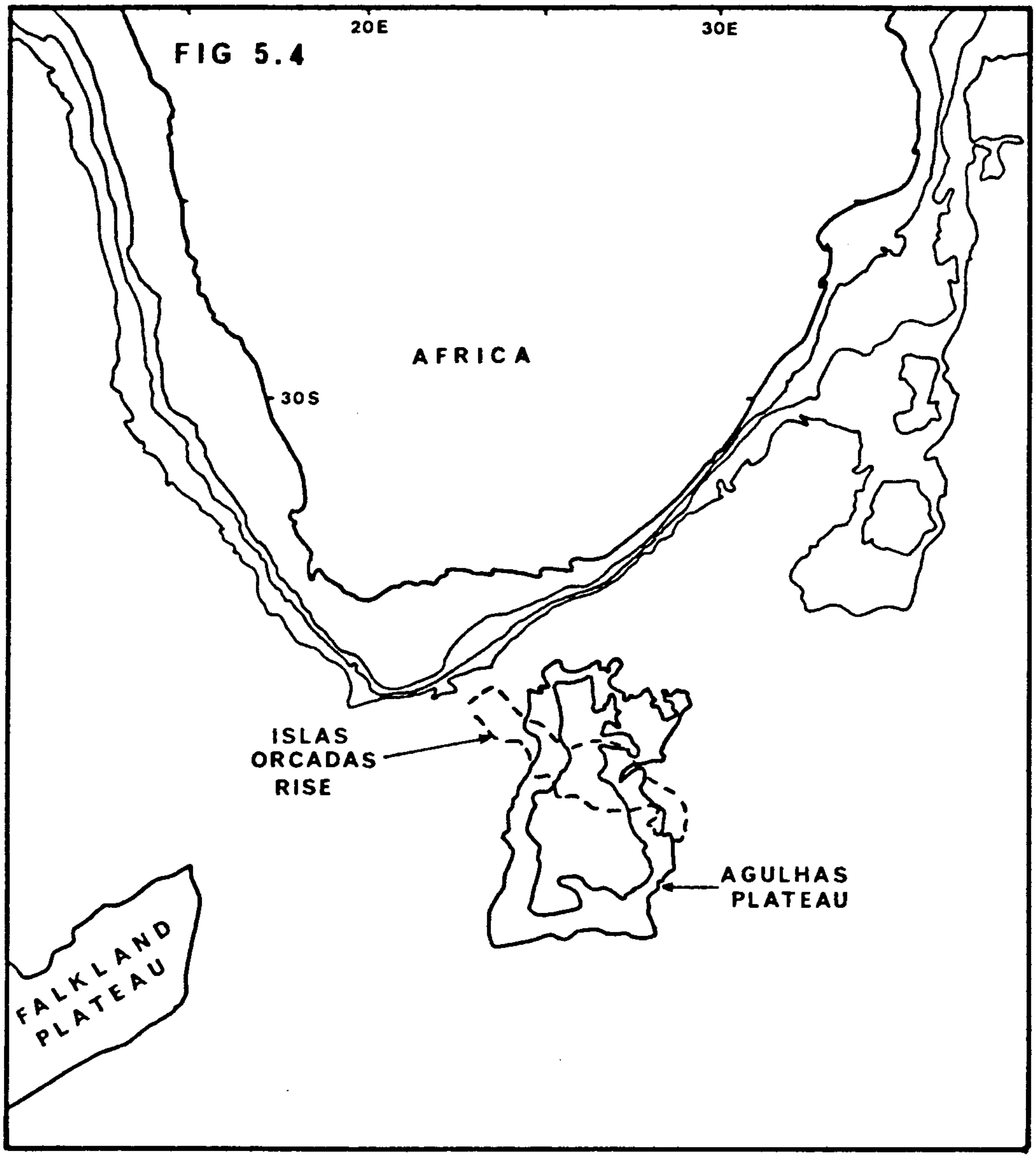

20E
30E
AFRICA
30S
ISLAS
ORCADAS
RISE
AGULHAS
PLATEAU
FALKLAND
PLATEAU

FIG 5.4

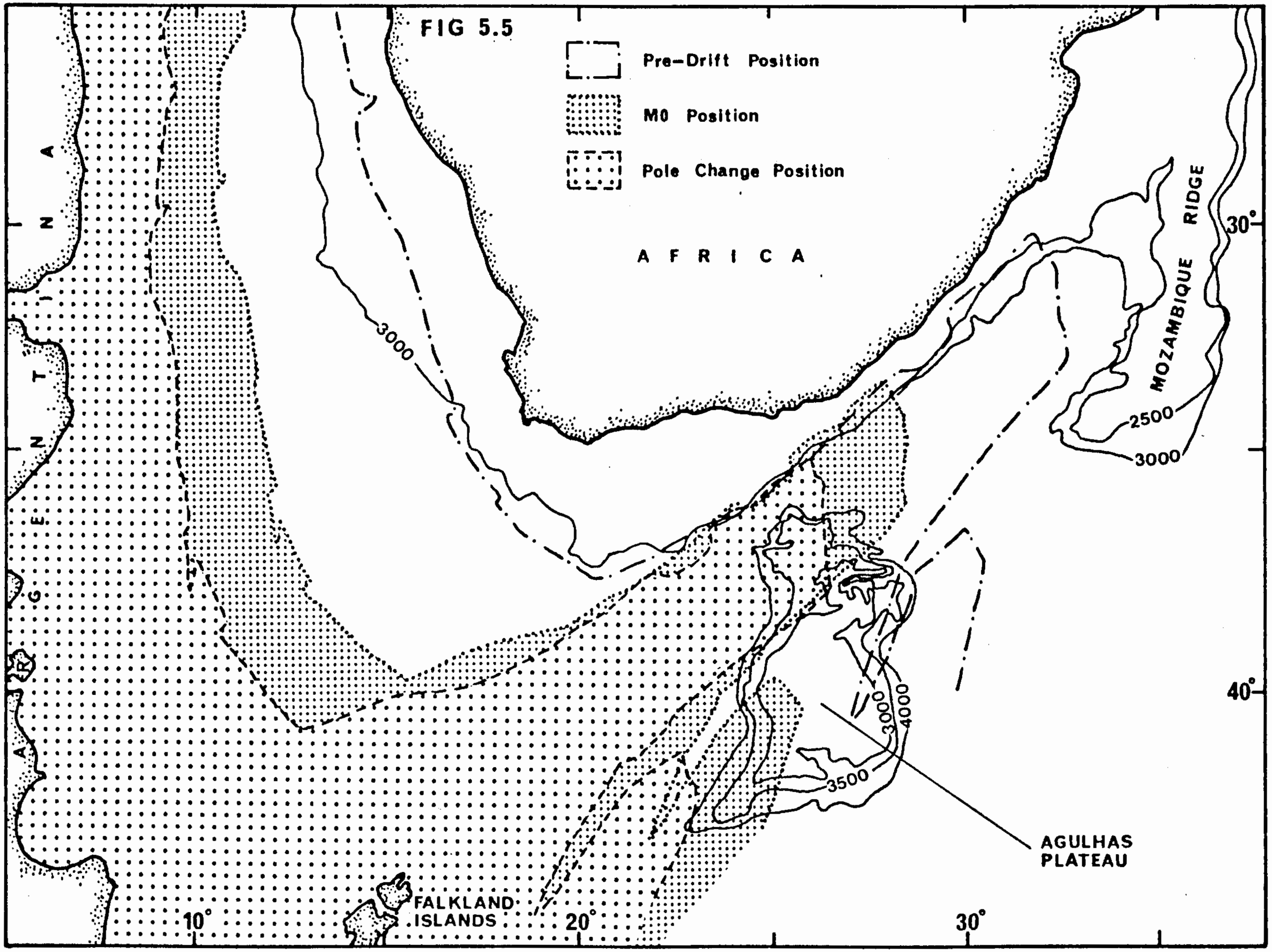


FIG 5.5
Pre-Drift Position
M0 Position
Pole Change Position
A F R I C A
MOZAMBIQUE RIDGE
2500
3000
3000
AGULHAS PLATEAU
4000
3000
3500
FALKLAND ISLANDS
10°
20°
30°
30°
40°

Considering the present-day southern Atlantic ocean (Fig.5.2) the Islas Orcadas Rise, lying in the Georgia Basin appears to be analogous to the Agulhas Plateau. The Islas Orcadas Rise trends 30°W and is not parallel to M0 crust, suggesting that it may also be diachronous. At latitude 50°S it centres on crust 590 km from M0 and is 80 km wide. It lies on crust dated at between 94.9 and 93.0 m.y. and in this respect is very similar to the oceanic part of the Agulhas Plateau. This is seen in a reconstruction for 93 m.y. (the centre of the Agulhas Plateau volcanism?) where the Islas Orcadas Rise and the Agulhas Plateau are overlapped (Fig.5.4). Note that the Maud Rise off the tip of Antarctica is also juxtaposed with the Agulhas Plateau in the 93 m.y. reconstruction (chapter 6).

In the well documented ridge-jump between anomalies 25 and 29 (Du Plessis, 1977; Barker, 1979; LaBrecque & Hayes, 1979) an old spreading centre was abandoned in the Agulhas Basin and a new one formed on a bathymetric high (Fig.5.3). Seafloor spreading from anomaly 25 to the present-day split and separated the bathymetric high into the Islas Orcadas Rise and the Meteor Rise (compare Figs.5.2 & 5.3). Since the Islas Orcadas and Meteor Rises are parts of the same feature, the Meteor Rise and the Agulhas Plateau have, by implication, a common origin. This is evident from Figures 5.2 and 5.3 where the Meteor Rise and Agulhas Plateau are symmetrically placed with respect to the anomaly 34 isochrons on either side of the abandoned ridge in the Agulhas Basin. At their northern ends, both Meteor Rise and Agulhas Plateau are 450 km distant from anomaly 34, while Meteor Rise is centred on crust 520 km from anomaly 34 (Table 5.1). It appears that one phase of excessive volcanism 97-90 m.y. ago produced an elevated feature on the accreting ridge, encompassing Islas Orcadas Rise, Meteor Rise and the Agulhas Plateau. If this feature was coincident with a triple junction which migrated to the south with time, its movement is not only traced by the south-western spur of the Agulhas Plateau (Bergh & Barrett, 1980; Tucholke et al., 1981) but also by

Table 5.1

Ages of the Agulhas Plateau, Islas Orcadas Rise and Meteor Rise

Feature		distance from M0	age (m.y.)	distance from anom. 34
	Older limit	450	97.3	830
Agulhas Plateau	peak of volcanism?	630	93.0	550
	younger limit	725	90.7	455
	older limit	550	94.9	
Islas Orcadas	peak of volcanism?	590	94.0	
Rise 50°S	younger limit	630	93.0	
	older limit		94.0	590
Meteor Rise	peak of volcanism?		92.4	520
	younger limit		90.7	450

Note that the Islas Orcadas and Meteor Rises were involved in a second phase of excess volcanism between anomalies 29 and 25 (65-58 m.y.). This probably increased the plan areas of these features, and thus affects the calculation of original ages.

the south-eastern end of the Meteor Rise (Fig.5.3). The symmetry exhibited by Meteor Rise, Islas Orcadas Rise and Agulhas Plateau mitigates against substantial subduction of oceanic crust below the Georgia Rise as suggested by LaBrecque & Hayes (1979).

Today the crust of the Islas Orcadas Rise, Meteor Rise and Agulhas Plateau is $\sim$2000 m above crust of the surrounding ocean basins (Barrett, 1977; LaBrecque & Hayes, 1979). Assuming subsidence according to the model of Parsons & Sclater (1977) the elevated crust of the single large bathymetric high (Fig.5.4) was originally $\sim$500 m deep with isolated areas exposed subaerially. This shallow depth, in conjunction with sea-level low-stands and possible subaerial exposure may have contributed to the hiatus on the Agulhas Plateau between Cenomanian clays (>95 m.y.) and Maastrichtian chalk (65-70 m.y.). Continued seafloor spreading and southerly migration of a triple junction between 93 and 80 m.y. (Figs.5.3 & 5.4) split the elevated feature in three - the combined Islas Orcadas Rise and Meteor Rise on the South American plate, the Maud Rise on the Antarctic plate (chapter 6) and the Agulhas Plateau on the African plate. When the South America-Africa ridge jumped around anomaly 25 - 29 (58-65 m.y.) the new spreading ridge was centred on the elevated crust produced 30 m.y. earlier. By this time it had subsided to about 2500 m depth. Thus the western part of the original feature was in turn split in half, and spreading from anomaly 25 to the present, separated the Islas Orcadas and Meteor Rises. The Islas Orcadas Rise remained on the South American plate while the Meteor Rise was incorporated into the African plate. During the second splitting phase the elevated crust was reheated and uplifted and then gradually subsided to its present depth as it moved away from the spreading ridge. This was probably accompanied by a second phase of excess volcanism. The bathymetric highs of the Meteor and Islas Orcadas Rises thus were not necessarily entirely caused by the ridge-jump occurring between 65 and 58 m.y. ago (Barker hypothesis, 1979). Rather these features have a composite history: they may, like the Agulhas Plateau incorporate an area of continental crust; volcanism occurred in two stages, firstly between 95 and 80 m.y. and secondly between 65 and 58 m.y.

5.3 MOVEMENT OF THE SOUTHERN CONTINENTAL AREA: AN ALTERNATIVE BREAK-UP MODEL

No seafloor spreading anomalies have been identified between the Agulhas Plateau and the Mozambique Ridge to constrain break-up in this region. Tucholke et al. (1981) suggested that as seafloor spreading proceeded in the Natal Valley and the south Atlantic 127-108 m.v. ago, a spreading centre formed oceanic crust south-west of the Agulhas Plateau. This they linked with the proposed seafloor spreading under the Falkland Plateau Basin (Ludwig, Krasheninnikov et al., 1980). When the Natal Valley spreading centre reached the southern end of the Mozambique Ridge it extended southeastwards separating the Agulhas Plateau and the Mozambique Ridge. At this stage in their model, spreading centres existed on both sides of the Agulhas Plateau. These jumped to the Agulhas Plateau intruding it and extending it. From chapter 4 we know that a) for the F.P.N.V. compartment at least, this model is invalidated because no ridge-jump occurred at this time; thus only half of the proposed straight ridge-section encompassing the F.P.N.V. and C.A.P.B.B. compartments could have been involved in a jump; b) any seafloor spreading in the Falkland Plateau Basin stopped prior to M10. An alternative model is suggested by the relationships shown in Figure 5.5.

I have assumed that the Burdwood Bank has been rigidly attached to the Falkland Plateau throughout the opening of the South Atlantic. At the South American margin of Gondwanaland, subduction of oceanic crust formed an active volcanic arc in the Middle and Late Jurassic (Dalziel et al., 1974; De Wit, 1977; De Wit & Stern, 1981). At this time South Georgia Island was adjacent to the southern side of Burdwood Bank (Winn, 1978). Silicic volcanism on

South Georgia and in Patagonia has been dated at 180-200 and 157-163 m.y. respectively (Tanner & Rex, 1979; De Wit, 1977; Herve et al., 1981). A back arc basin opened on the continental side of the arc after the silicic volcanic episode and is dated by the oldest marine sediments. These are at least basal Barremian (118 m.y.)(Dalziel et al., 1974) and range back to the Oxfordian (149 m.y.)(Olea & Davis, 1977). In South Georgia mafic rocks of the back-arc are cut by 127 m.y. tonalites and initial emplacement of mafic material is dated at ∿140 m.y. Silicic rocks of the Tobifera Formation are encountered east of Tierra del Fuego, and Burdwood Bank is considered part of this remnant arc (Ludwig et al., 1979). Burdwood Bank is thus a pre-drift feature, since South Atlantic opening began 122-127 m.y. ago (chapter 4). It is uncertain how much of Burdwood Bank and the North Scotia Ridge is due to subduction of the Falkland Trough below North Scotia Ridge. I have followed LaBrecque & Hayes (1979) in outlining Burdwood Bank with the 3000 m isobath.

In its pre-drift position Burdwood Bank lay near Antarctica (chapter 6) with the gap between filled by Agulhas Plateau. As seafloor spreading began in the Natal Valley, the Agulhas Plateau, attached to Burdwood Bank rifted away from Antarctica. Spreading ridges in the F.P.N.V. and C.A.P.B.B. compartments were subparallel and offset by 350 km. Between 108 and 105 m.y. ago (M0 and pole change reconstructions) Agulhas Plateau reached its present position relative to Africa (Fig.5.5). The C.A.P.B.B. spreading ridge then jumped ∿500 km to the west, rifting apart Agulhas Plateau and Burdwood Bank. South-east trending areas of thickened oceanic crust were emplaced in conjunction with south-east oriented faulting (Tucholke et al., 1981, Figs.2 and 6). This event may have been associated with the change in opening pole of the South Atlantic (Fig.5.5). F.P.N.V. and C.A.P.B.B. spreading ridges were at that stage, offset by ∿850 km. Through continued seafloor spreading the F.P.N.V. ridge reached the north-east corner of the continental Agulhas Plateau, and excess volcanism ensued. A slight (∿30 km) asymmetry favouring the African side may have occurred. South-west trending faults and east-west bathymetric trends resulted from the interaction of ridge and active transform fault (c.f. Searle, 1979; Schouten et al., 1980). The offset of the northern and southern parts of the Agulhas Plateau is associated with the transform fault and westward rafting of the western half of the thickened oceanic crust. Before anomaly 34 times the C.A.P.B.B. ridge jumped back, perhaps to the western edge of the Agulhas Plateau around 90-95 m.y. as the ridge in the F.P.N.V. compartment extruded excess material. At this later stage a triple junction was formed at a time when global plate re-organization occurred (Johnson et al., 1980).

5.4 WHY WAS THERE EXCESS VOLCANISM ON THE ASEISMIC RISES?

An episode of excess volcanism occurring between 97 and 90 m.y. (Table 5.1) built up the Meteor Rise, Islas Orcadas Rise, Agulhas Plateau (oceanic areas) and possibly even Maud Rise (Fig.5.4). Three possible causes are suggested.

a) The tip of the Falkland Plateau cleared southern Africa 98.3 m.y. ago (Table 4.2). This marked the release of one of the constraining forces on the direction of motion (c.f. LePichon & Hayes, 1971; Scrutton, 1979), and may be a contributary cause to the global re-organization of plate motion occurring around 90-95 m.y. Agulhas Plateau volcanism may have been a local expression of this.

b) In the alternative break-up model given above (section 5.3)it is suggested that Burdwood Bank and the continental Agulhas Plateau rifted apart between 108 and 105 m.y. As the accreting ridge in the F.P.N.V. spreading compartment migrated to the south-west relative to Africa it encountered an area of high heat-flow where a continental fragment had been split off 8-12 m.y. earlier. The result was possibly instability and increased extrusion on the accreting ridge.

c) Considering relative plate motions in terms of the hot-spot reference frame, the Bouvet hot-spot (Duncan, 1981) or possibly a Meteor Rise hot-spot

(Crough et al., 1980) was located in the Karoo of South Africa 90 m.y. ago. (However, Morgan (1981) provides an alternative model). This hot-spot appears to have been very active at this time since a large number of kimberlite pipes erupted around Kimberley between 95 and 78 m.y. ago (Crough et al., 1980). Vogt (1976) has outlined a theory where partial melts flow away from hot-spots along the length of accreting ridges in "sub-axial pipes". This theory was developed mainly for hot-spots occurring on or near accreting ridges. It is possible that pipe-flow from the Bouvet hot-spot, located around Kimberley, led partial melts to the accreting ridge between the FAFZ and the Agulhas Plateau 500-700 km to the south. Recently, Angevine and Turcotte (1983) suggested that the Agulhas Plateau topography is partially compensated by a low-density mantle layer similar to that postulated for the Walvis Ridge. As this latter feature marks a hot-spot trace, it is possible that the Agulhas Plateau does too. Later on, around 65 m.y. ago the Bouvet hot-spot may have approached the FAFZ again, causing the ridge-jump to the Islas Orcadas/Meteor Rise complex (C.J.H. Hartnady, personal communication, 1983).

5.5 SUMMARY

The northern oceanic part of the Agulhas Plateau is not the site to which an accreting ridge jumped. Oceanic crust of this area is between 97.3 and 90.7 m.y. old. Calculated ages of the Meteor and Islas Orcadas Rises are similar (Table 5.1). All three features originated through excess volcanism on the accreting ridge, possibly as a result of pipe-line flow from the Bouvet hot-spot which was located under the Karoo at this stage. Meteor Rise and Islas Orcadas Rise, forming a composite feature separated from the Agulhas Plateau between 93 and 80 m.y. ago. Between 65 and 58 m.y. the Islas Orcadas Rise and Meteor Rise were separated by a well-documented ridge-jump (Du Plessis, 1977; LaBrecque & Hayes, 1979; Barker, 1979) which may also be hot-spot related.

An alternative break-up model is suggested in which the continental part of the Agulhas Plateau had reached its present location relative to Africa by 108-105 m.y. ago.

CHAPTER 6

A REVISED RECONSTRUCTION OF EAST ANTARCTICA AND AFRICA: PLATE TECTONIC STATUS OF THE NORTHERN NATAL VALLEY AND MOZAMBIQUE RIDGE

Regarding reconstructions of South America and Africa, workers have concurred on the general form of the refit. Revision concerned the exact nature of the rotation parameters and approximately 450 km of difference implied by the different proposed rotations (chapter 3). In the case of East Antarctica and Africa, radically different reconstructions have been proposed (Fig.3.1): a) the Antarctic Peninsula is placed west of South America, while East Antarctica lies south of the Falkland Plateau; Madagascar is either i) placed next to Mozambique (Tarling, 1972; Barron et al., 1978; Harrison et al., 1979) or ii) is moved northwards towards Kenya (Veevers et al., 1971; 1980; Powell et al., 1980); b) East Antarctica and Mozambique are juxtaposed with Madagascar in the northerly position (Du Toit, 1937; Smith & Hallam, 1970; Norton & Sclater, 1979). Palaeomagnetic evidence (McElhinney et al., 1976) and the discovery of seafloor spreading anomalies in the Somali Basin (Segoufin & Patriat, 1980; Parson et al., 1981; Rabinowitz et al., 1983) requires a northerly position for Madagascar thus ruling out option (ai) above. Option (aii) leaves gaps in the region of the Mozambique Channel. Option b) causes the Antarctic Peninsula to overlap the Falkland Plateau, necessitating micro-plate re-organisation of West Antarctica (Barker, Dalziel et al., 1977; De Wit, 1977; Dalziel & Elliott, 1982). Tests of the global plate circuit indicate that at least a two-plate condition for Antarctica is required (Gordon & Cox, 1980).

The only available Mesozoic magnetic anomalies with which to constrain the refit of East Antarctica and Africa are those off Dronning Maud Land (Bergh, 1977) and in the Mozambique Channel (Segoufin, 1978; Simpson et al., 1979)(Fig.6.1). Reconstruction option (aii) implies that the Dronning Maud Land anomalies have equivalents near the southern Mozambique Ridge (Veevers et al., 1980) while option (b) matches the northern Mozambique Basin with the Dronning Maud Land crust: starting from the Smith & Hallam reconstruction, Segoufin & Patriat (1980) match the Antarctic anomalies with the most westerly identified Mesozoic crust in the Mozambique Basin; the Norton-Sclater fit (1979) places the Antarctic anomalies in the latitude now occupied by Madagascar. Establishing a correct refit of East Antarctica is crucial to determining the plate tectonic status of the northern Natal Valley and the Mozambique Ridge: either these marginal plateaux (chapter 2) are underlain by thinned continental crust, by thickened oceanic crust, or by a mixture of crustal types.

Using all available published seafloor spreading anomaly and fracture zone identifications (Fig.6.1) a sequence of progressively older reconstructions of East Antarctica and Africa has been computed (Figs.6.2-6.10). Being constrained by triple junction configurations and fracture zones, these allow successively older magnetic anomalies of the Antarctic plate to be matched correctly with the equivalent seafloor spreading compartment of the African plate (Hartnady & Martin, 1983). The result is a revised reconstruction of East Antarctica in Gondwanaland achieved entirely on the basis of marine geophysical data. Independent on-land geological criteria are used to test the reconstruction.

6.1 RECONSTRUCTIONS

6.1.1 Chron 29 = Cretaceous/Tertiary boundary = 65 m.y. ago

This reconstruction effectively eliminates Cenozoic crust generated along the South West Indian Ocean Ridge (SWIOR). Crucial to this reconstruction are the new anomaly identification SE of Madagascar and north of Crozet Island (Goslin et al., 1981). In both these areas, anomaly 29 crust forms a magnetic bight indicating a fossil triple junction. The reconstruction was achieved in two steps:

a. by using the pole of rotation and rotation rate of Sclater et al. (1981) for Antarctica/Africa relative motion;

FIGURE CAPTIONS

Fig.6.1. Compilation of fracture zone and magnetic anomaly identifications between Africa and Antarctica (references in text). Black dots are epicentres of recent earthquakes on the mid-ridge.

Figs.6.2-6.10. Computed reconstructions. In all cases black dots = half-rotated epicentres; black stars = anomalies on the African Plate; open squares = anomalies on the Antarctic plate; open circles = anomalies on the S. American plate.

Fig.6.2. = anomaly 29 - note the triple junction configuration east of the Madagascar Ridge. PEFZ = Prince Edward Fracture Zone.

Fig.6.3. = anomaly 34. Note the Mesozoic anomalies off Dronning Maud Land, Antarctica and in the Mozambique Basin.

Fig.6.4. = 93 m.y. Note juxtaposition of Maud Rise and Agulhas Plateau (c.f. Fig.5.4). F.P. = Falkland Plateau; A.P. = Agulhas Plateau; M.R. = Maud Rise; C.R. = Conrad Rise.

Fig.6.5. = 105 m.y. (c.f. Fig.5.5).

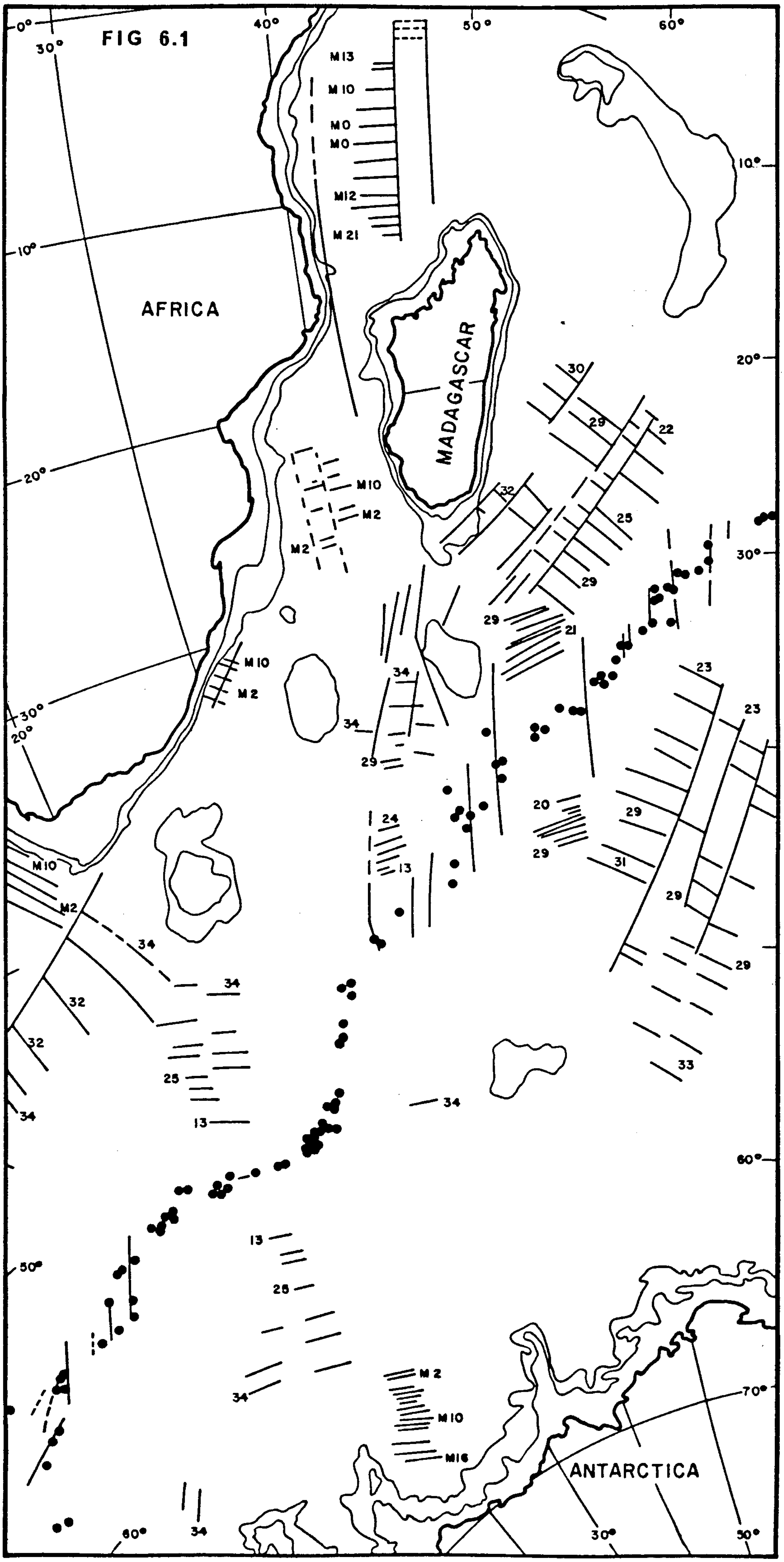

AFRICA
MADAGASCAR
ANTARCTICA
0°
30°
40°
50°
60°
10°
20°
30°
50°
60°
70°
M13
M10
M0
M0
M12
M21
M10
M2
M2
M10
M2
M10
M2
30
29
22
32
25
29
29
21
34
34
29
23
23
20
29
29
31
29
29
24
13
34
34
32
32
25
13
34
34
33
13
25
34
M2
M10
M16

FIG 6.1

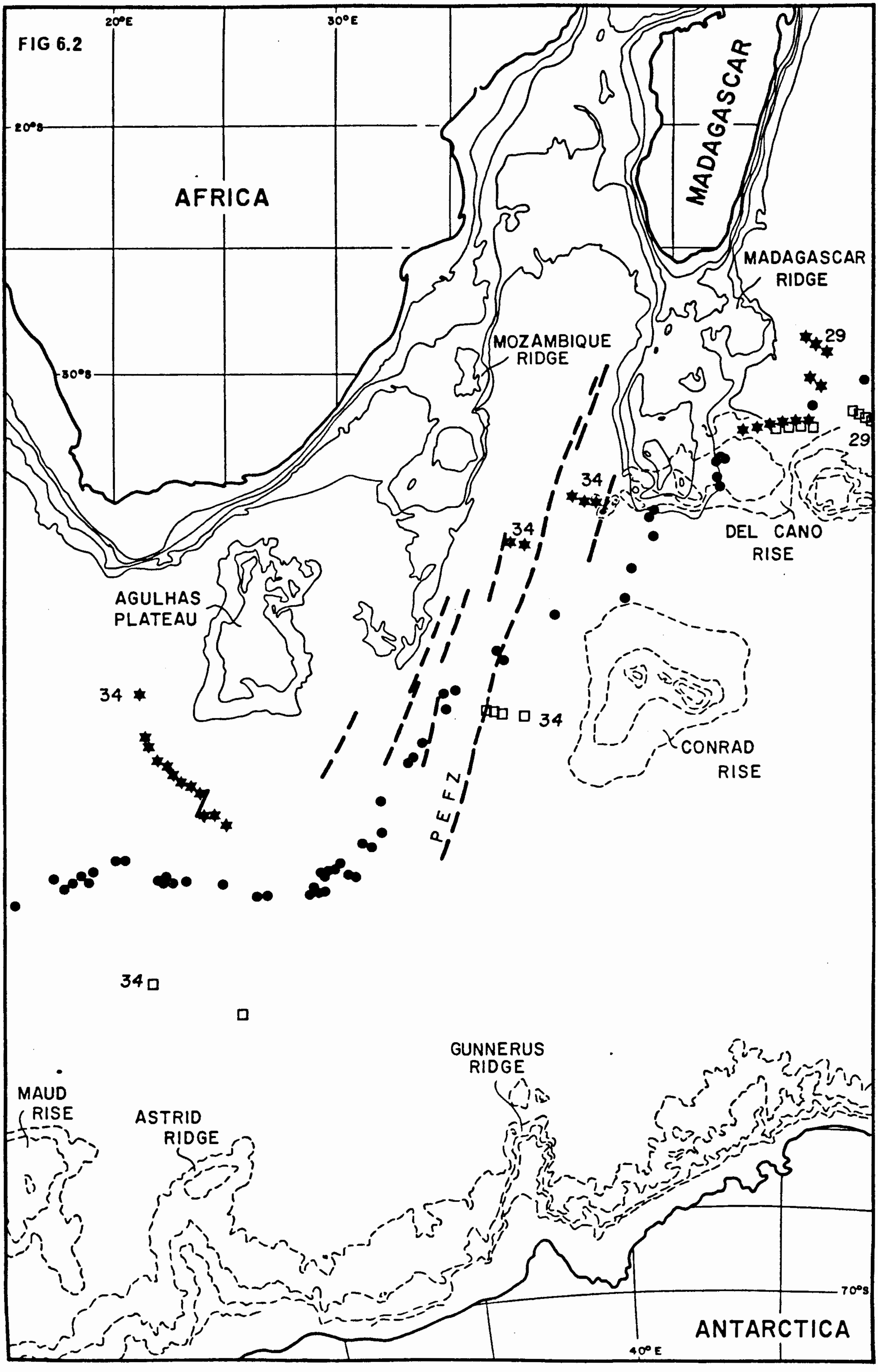


FIG 6.2
20°E
30°E
20°S
30°S
70°S
40°E
AFRICA
MADAGASCAR
MADAGASCAR RIDGE
MOZAMBIQUE RIDGE
AGULHAS PLATEAU
DEL CANO RISE
CONRAD RISE
PEFZ
29
34
GUNNERUS RIDGE
MAUD RISE
ASTRID RIDGE
ANTARCTICA

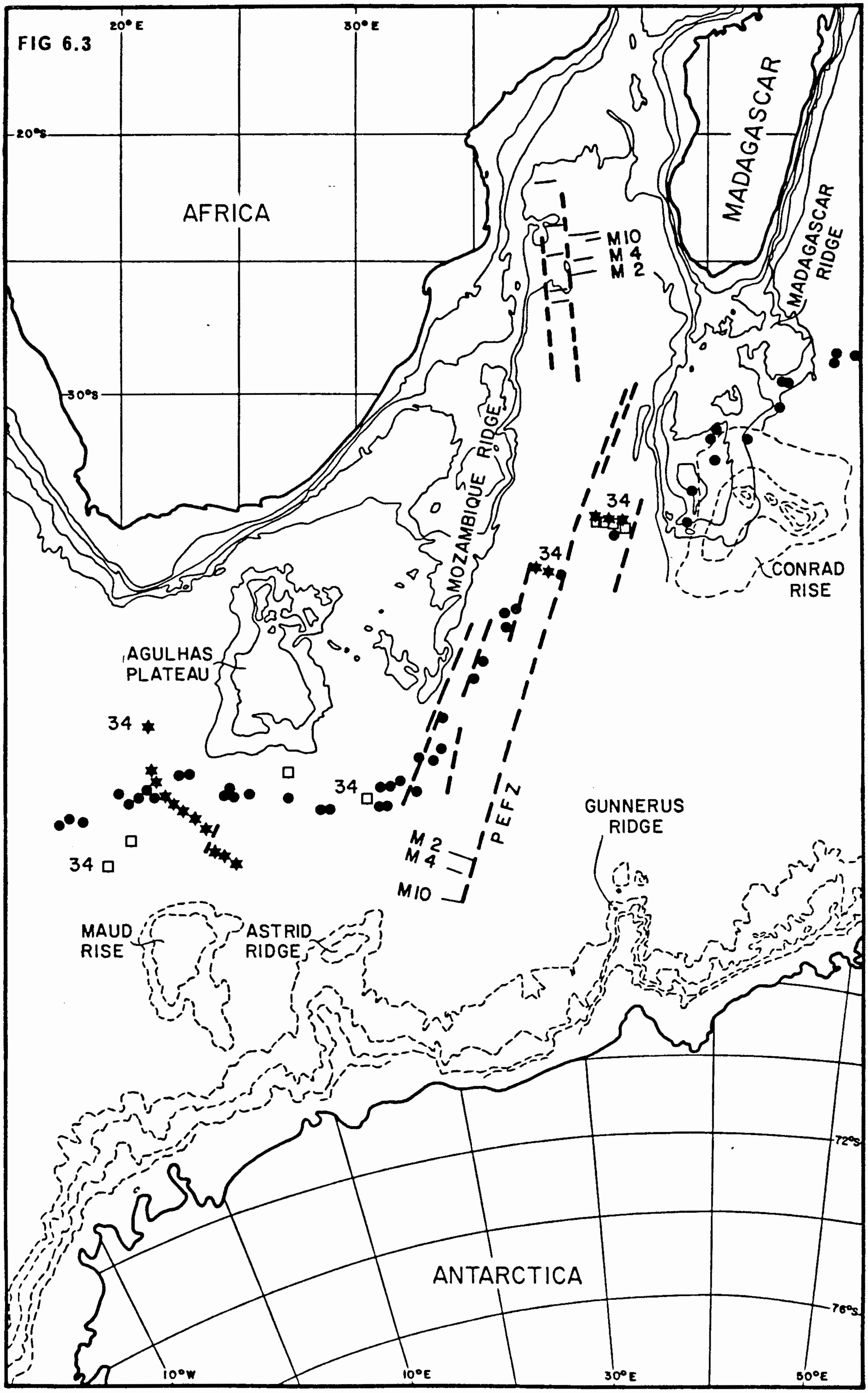


FIG 6.3
20° E
30° E
20°S
30°S
AFRICA
MADAGASCAR
MADAGASCAR RIDGE
M10
M 4
M 2
MOZAMBIQUE RIDGE
34
CONRAD RISE
AGULHAS PLATEAU
PEFZ
GUNNERUS RIDGE
MAUD RISE
ASTRID RIDGE
72°S
76°S
ANTARCTICA
10°W
10°E
30°E
50°E

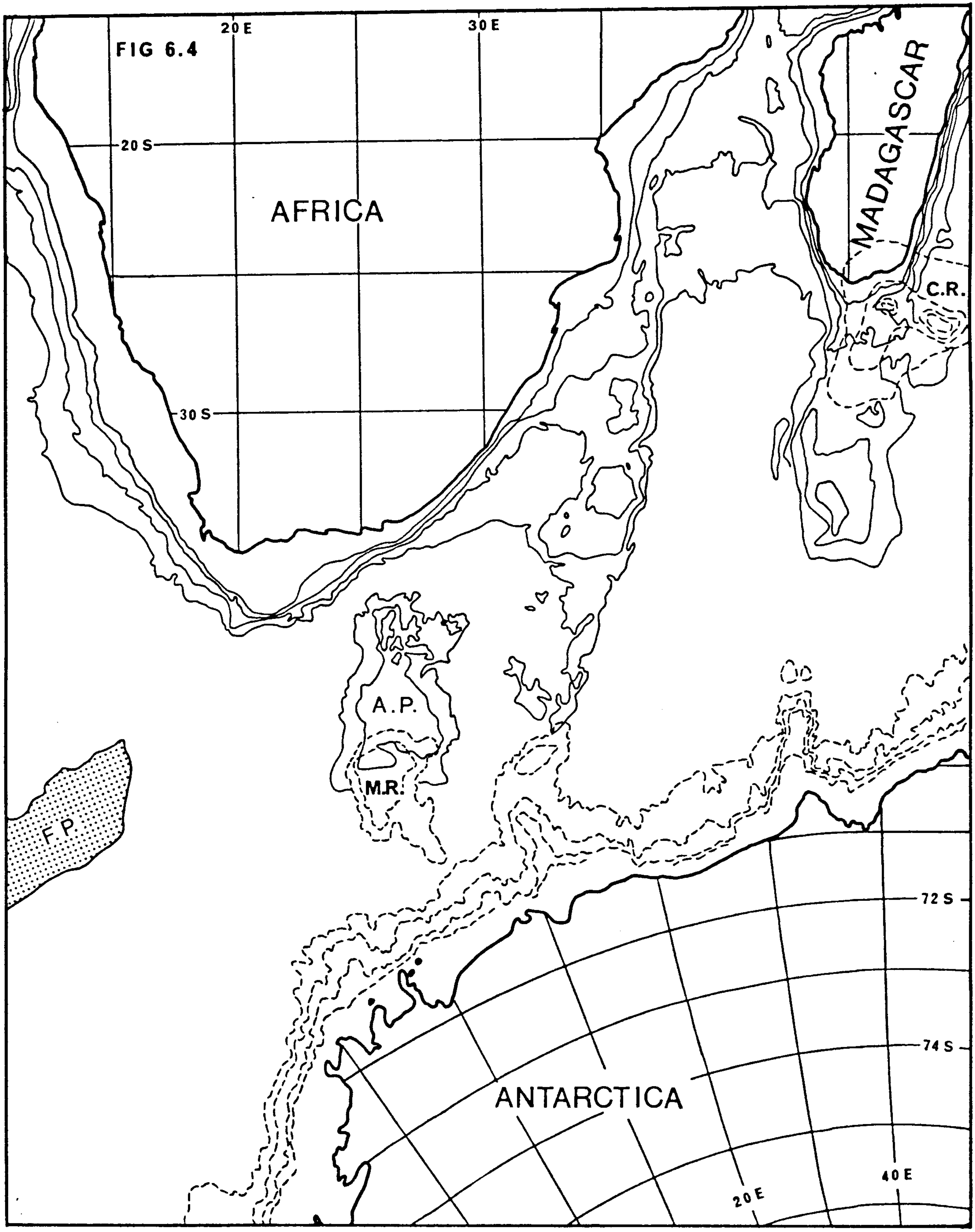


FIG 6.4
20 E
30 E
20 S
30 S
AFRICA
MADAGASCAR
C.R.
A.P.
M.R.
F.P.
72 S
74 S
ANTARCTICA
20 E
40 E

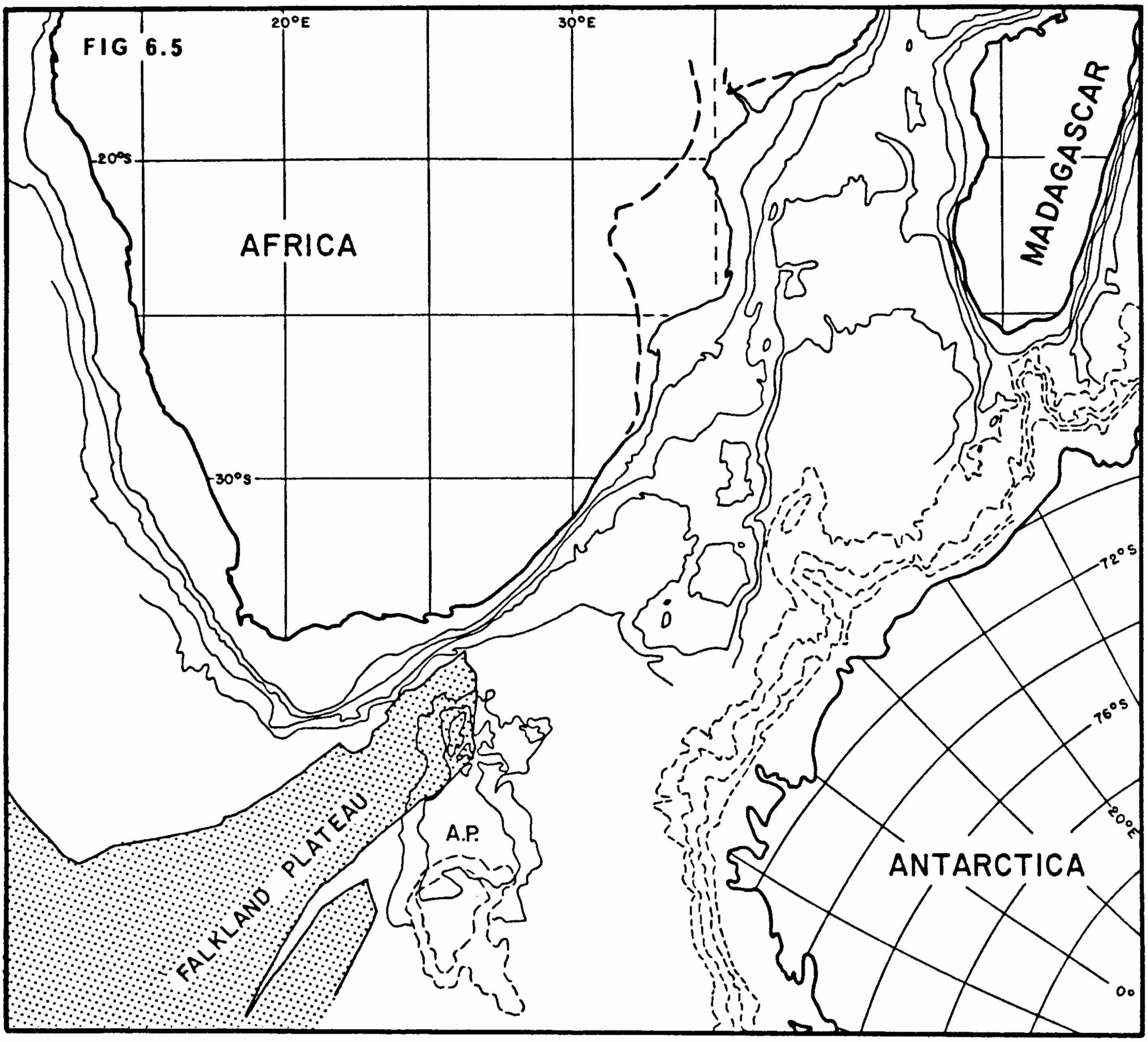


FIG 6.5
20°E
30°E
20°S
30°S
AFRICA
MADAGASCAR
FALKLAND PLATEAU
A.P.
ANTARCTICA
72°S
76°S
20°E
0°

b. by applying a small corrective rotation which moved the anomaly 29 crust of the Antarctic plate approximately 80 km west relative to the anomalies on the African plate (Table 6.1).

This refit is considered accurate because it recreates the triple junction configuration SE of Madagascar and juxtaposes the Del Cano Rise with the Madagascar Ridge (Goslin et al., 1981). These conjugate features are considered oceanic in nature (Goslin et al., 1981; Sinha et al., 1981; Duncan, 1981) and have been related to the Prince Edward hot-spot. Moreover, the present-day configuration of the fracture zones taken from GEBCO sheet 509 (Fig.6.2) indicate that spreading on SWIOR has continued in a similar orientation for a long geological time. The use of the instantaneous pole with only a small correction is thus legitimate. For this reconstruction we have also applied a half rotation to the earthquake epicentres which delineate the present day spreading ridge. This gives an indication of the Cretaceous/Tertiary boundary mid-ocean ridge: note that the half-rotated epicentres are located on the palaeo-triple-junction. The Prince Edward Fracture Zone is plotted from the profiles of Bergh & Norton (1976). This, and the fracture zones to the west indicate that the orientation of the recent magnetic lineations on the spreading ridge NW of Prince Edward Island is caused by frequent offset rather than by oblique spreading. The most important point to be gleaned from Figure 6.2 is that, in its reconstructed position, the anomaly 34 crust west of the Conrad Rise lies <u>east</u> of the Prince Edward Fracture Zone. It should therefore be fitted with anomaly 34 crust lying east of the Prince Edward Fracture Zone in the Mozambique Basin. Fisher & Sclater (1983) have recently matched the spreading compartments in the same way, using independent reasoning, although they did not superpose magnetic anomalies.

6.1.2 Chron 34 = Late Santonian = 80 m.y.

This reconstruction is based on superimposition and parallel alignment

Table 6.1

Finite poles of rotation

Antarctica relative to Africa

description	age	rotation pole		angle
chron 29	65 m.y.	4.15°N	42.72° W	11.65°
chron 34	80 m.y.	17.07°N	48.40° W	19.74°
	93 m.y	10.37°N	45.20° W	26.56°
	105 m.y.	8.43°S	26.42° W	42.56°
M2	113 m.y.	7.22°S	28.14° W	43.82°
M10	122 m.y.	6.80°S	29.27° W	46.17°
M21	145 m.y.	1.67°S	35.99° W	53.43°

Agulhas Plateau relative to Africa

pre-drift	122 m.y.	6°N	2°W	7.25°

of chron 34 crust west of Conrad Rise with chron 34 crust east of the Prince Edward Fracture Zone and west of the Madagascar Ridge (Fig.6.3). This was achieved by applying two corrective rotations to the fit of Bergh & Barrett (1980)(Table 6.1). The exact parallel alignment of the two conjugate anomalies may be criticised since oblique spreading can occur (Atwater & Macdonald, 1977). However, obliquity normally occurs on slow-spreading ridges whereas at this stage the half-spreading rate for Africa and Antarctica was between 3.2 and 4.7 cm yr^{-1} (Bergh & Norton, 1976; LaBrecque & Hayes, 1979; Bergh & Barrett, 1980). This reconstruction does not superpose chron 34 crust north-east of Maud Rise with identified chron 34 crust south-west of the Agulhas Plateau, and in this respect we differ from Bergh & Barrett (1980). It places chron 34 crust north-west of Maud Rise in an appropriate position relative to the triple junction implied by the chevron-shaped anomaly 34 crust mapped south-west of the Agulhas Plateau. This refit implies that a fracture zone exists on the south-eastern flank of the Agulhas Plateau (Fig.6.3). Again half-rotations have been applied to earthquake epicentres and these lie on the palaeo-ridge in the Mozambique Basin. However the half-rotated epicentres south of Agulhas Plateau do not match the magnetic anomalies, suggesting that an adjustment to the spreading configuration has occurred between 80 m.y. and the present day. This may have occurred when a well-documented ridge-jump occurred south of the Agulhas/Falkland Fracture Zone between South America and Africa (chapter 5).

The main point demonstrated by Figure 6.3 is that Antarctic crust exhibiting Mesozoic or M-sequence magnetic anomalies lies west of the Prince Edward Fracture Zone. It should therefore be matched with crust in the Mozambique Basin lying west of the Prince Edward Fracture Zone or its extensions. The Prince Edward Fracture Zone (Bergh & Norton, 1976) and sub-parallel fractures in the Mozambique Basin (Goslin et al., 1981) are oriented north-north-east whereas the fracture zones associated with the M-sequence magnetic anomalies farther north are oriented 10°E of S and are sub-parallel to the Davie Fracture Zone west of Madagascar. A change in spreading direction is implied some time in the Cretaceous magnetic quiet zone, when the spreading ridge had migrated to approximately 28-29°S relative to Africa (Fig.6.3). This suggests that the northern equivalent of the Prince Edward Fracture Zone lies east of the easternmost Mesozoic crust identified on Figure 6.3. Thus the Mesozoic crust identified off Dronning Maud Land should be matched with the easternmost spreading compartment identified so far in the Mozambique Basin. This forms the basis for reconstructions involving Mesozoic oceanic crust (Figs.6.7-6.10).

The other points arising from this reconstruction for chron 34 concern the aseismic plateaux of the south-west Indian Ocean and the Africa-Antarctica Basin. The Conrad Rise is juxtaposed with the Madagascar Ridge, suggesting these two features have a common origin. Half-rotated epicentres suggest that these conjugate features were being rifted apart at this stage. It is not known however whether Conrad Rise is oceanic or continental in character. Symmetry is also displayed by

a. the Agulhas plateau and Maud Rise

b. the Mozambique Ridge and Astrid Ridge

c. the Madagascar Ridge and Gunnerus Ridge.

These features appear to form conjugate pairs.

6.1.3 The Cretaceous Normal Polarity Epoch

Although the lack of magnetic reversals in the Cretaceous magnetic quiet zone prevents accurate reconstructions for this period, two refits have been produced for this stage (Figs.6.4, 6.5, Table 6.1). The 93 m.y. reconstruction (c.f. chapter 5) shows Agulhas Plateau and Maud Rise juxtaposed (LaBrecque & Hayes, 1979; Bergh & Barrett, 1980). The possible relationship of the split-up of these features to hot-spot/pipe-line activity has already been discussed (chapter 5). The Conrad Rise, at this time, is placed just south of the tip of Madagascar. Duncan (1981) has shown that the Prince Edward hot-spot is located at the southern end of Madagascar between 90 and 70 m.y. The formation of Conrad Rise and its separation from the Madagascar Ridge is thus probably related to Prince Edward hot-spot activity.

Following the model of chapter 5, Maud Rise is considered either to a) lie on the African plate prior to 93 m.y.; or b) to have formed during the episode of excess volcanism around 93 m.y. In the reconstruction for 105 m.y. it is retained next to the Agulhas Plateau. Astrid Ridge is positioned close to the Mozambique Ridge while Gunnerus Ridge is placed south of Madagascar. Seafloor spreading stopped in the Somali Basin at this time (Segoufin & Patriat, 1980), and probably began farther south, separating Madagascar and Antarctica. Figures 6.2-6.5 indicate that Gunnerus Ridge, Conrad Rise and Del Cano Rise are all associated with the Madagascar Ridge, suggesting a complex inter-play of seafloor spreading, ridge-jumps and hot-spot activity.

6.1.4 M2 = Late Barremian = 113 m.y.

Because Magnetic anomaly M0 has not been mapped in the Mozambique Basin (Segoufin, 1978; Simpson et al., 1979), the next oldest well constrained reconstruction (Fig.6.6) is achieved by superimposition and parallel alignment of Antarctic M2 crust with the easternmost M2 crust so far identified in the Mozambique Basin. Half-rotated epicentres in the Mozambique Basin lie near M2 crust, suggesting that little re-adjustment of spreading has occurred there. The Falkland Plateau is replaced into the Natal Valley (chapter 4). Madagascar is rotated slightly northwards using Somali Basin anomalies (Segoufin & Patriat, 1980). The Agulhas Plateau complex probably reached its present position relative to Africa at or just after M0 time (chapter 5). Prior to this it was part of the South American plate, and is therefore displaced slightly to the north-east for the M2 reconstruction. The northern oceanic part of the Agulhas Plateau need not be included in this refit. The

FIGURE CAPTIONS

Fig.6.6. a) M2. Solid lines west of Antarctica = lineated magnetic anomalies in the eastern Weddell Sea. The numbers refer to anomaly identifications of LaBrecque & Barker (1981) which are invalidated by this study. b) Velocity triangle for the S. American African Antarctic triple junction between M10 and M0 times. This shows that the lineated anomalies in the Weddell Sea are aligned appropriately for their generation at this triple junction but that the total separation rate is 29 cm yr^{-1}, suggesting a half-spreading rate of $\sim$1.5 cm yr^{-1} as opposed to 0.6 cm yr^{-1} given by LaBrecque & Barker (1981).

Fig.6.7. M10. In this version, Mozambique Ridge is in its present position relative to Africa.

Fig.6.8. M10. Here, the southern Mozambique Ridge is assumed to have been part of the Antarctic plate prior to M2 times.

Fig.6.9. M21. Note the various inter-continental geotectonic features. The Haag Nunataks and the Whiteout Conglomerate, situated on the Ellsworth Block have been rotated relative to East Antarctica using the parameters of Watts & Bramall (1981).

Fig.6.10 M21. Here, the southern Mozambique Ridge is assumed to be continental and is placed next to Nuanetsi and Sabi volcanic outcrops.

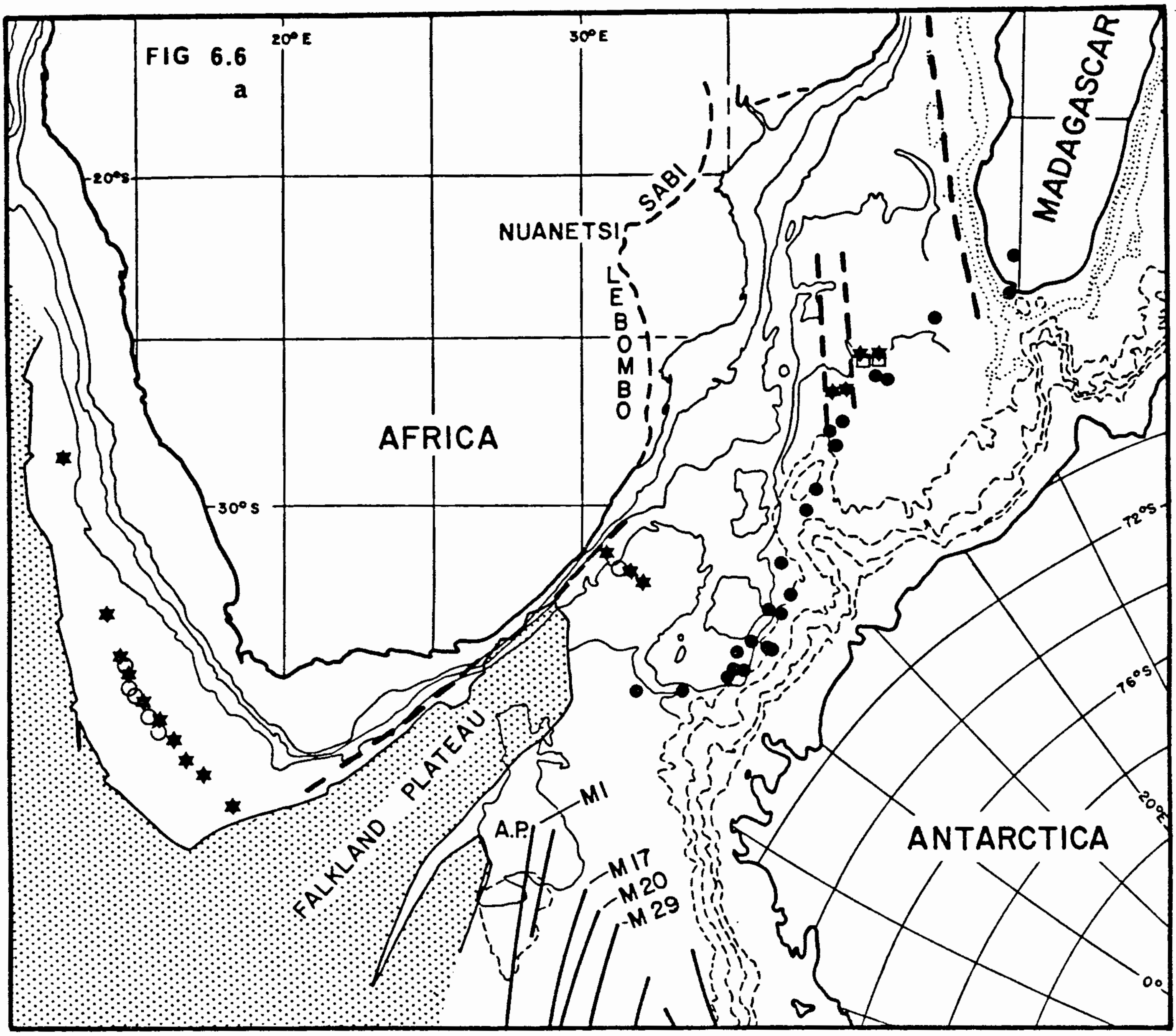
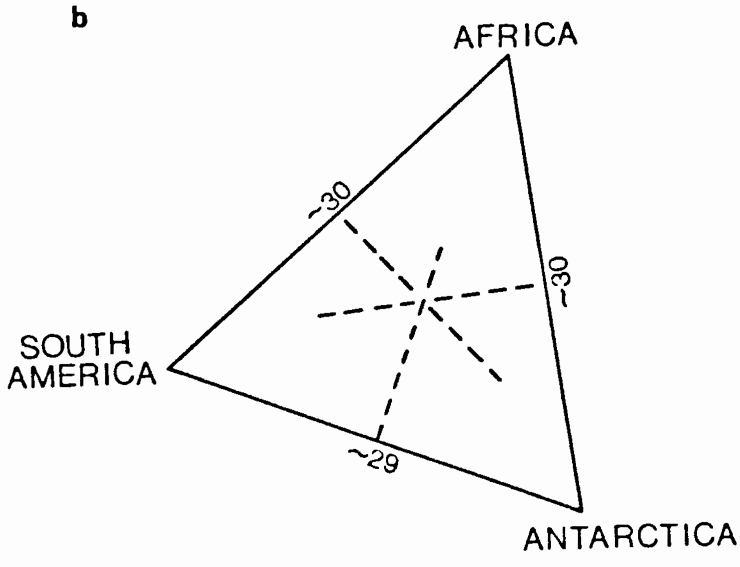


FIG 6.6
a
20° E
30° E
20° S
30° S
SABI
NUANETSI
LEBOMBO
AFRICA
MADAGASCAR
FALKLAND PLATEAU
A.P.
M1
M17
M20
M29
72° S
76° S
20° E
0°
ANTARCTICA


b
AFRICA
SOUTH AMERICA
ANTARCTICA
~30
~30
~29

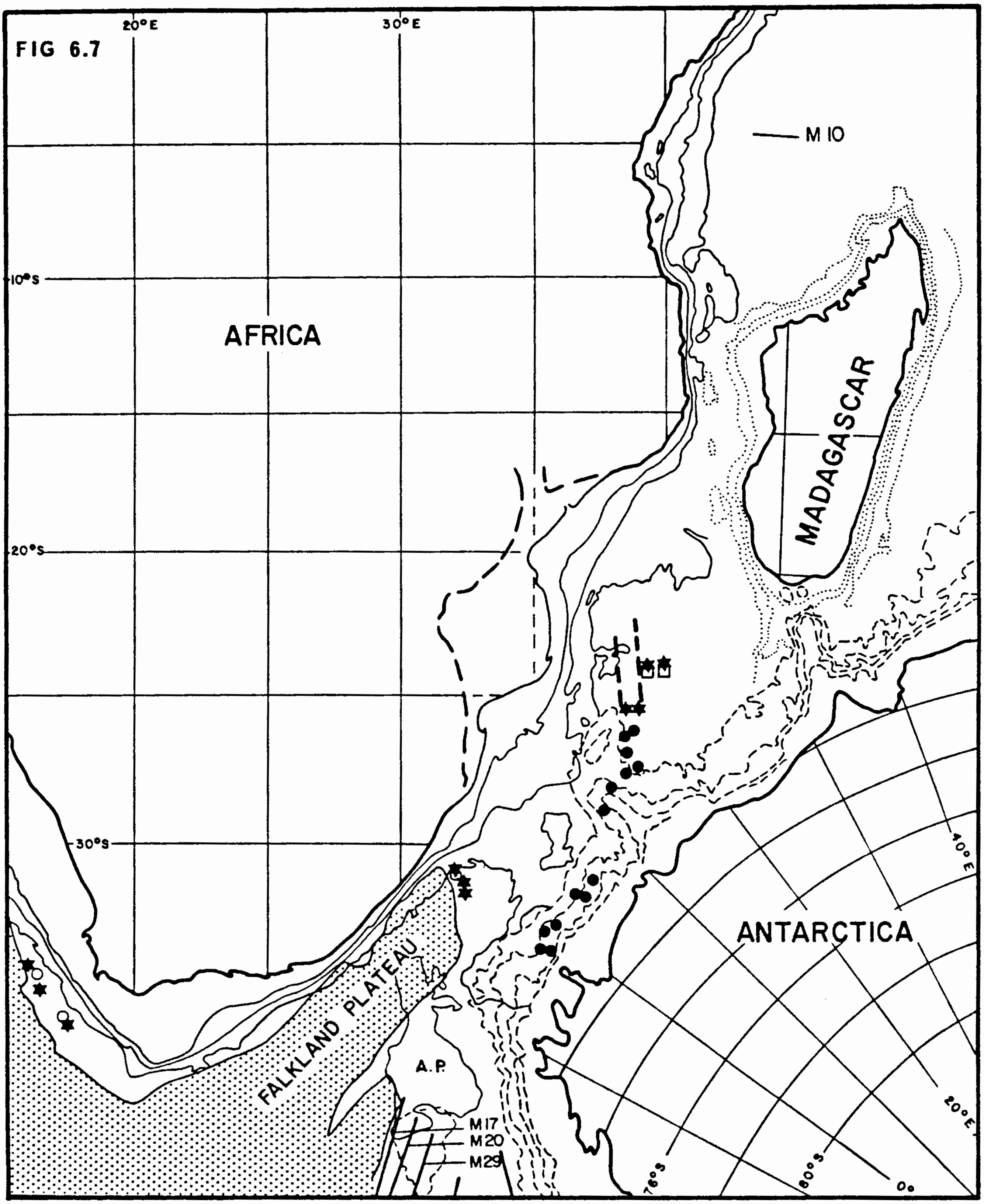


FIG 6.7
20°E
30°E
10°S
20°S
30°S
AFRICA
MADAGASCAR
M10
FALKLAND PLATEAU
A.P.
M17
M20
M29
ANTARCTICA
40°E
20°E
76°S
80°S
0°

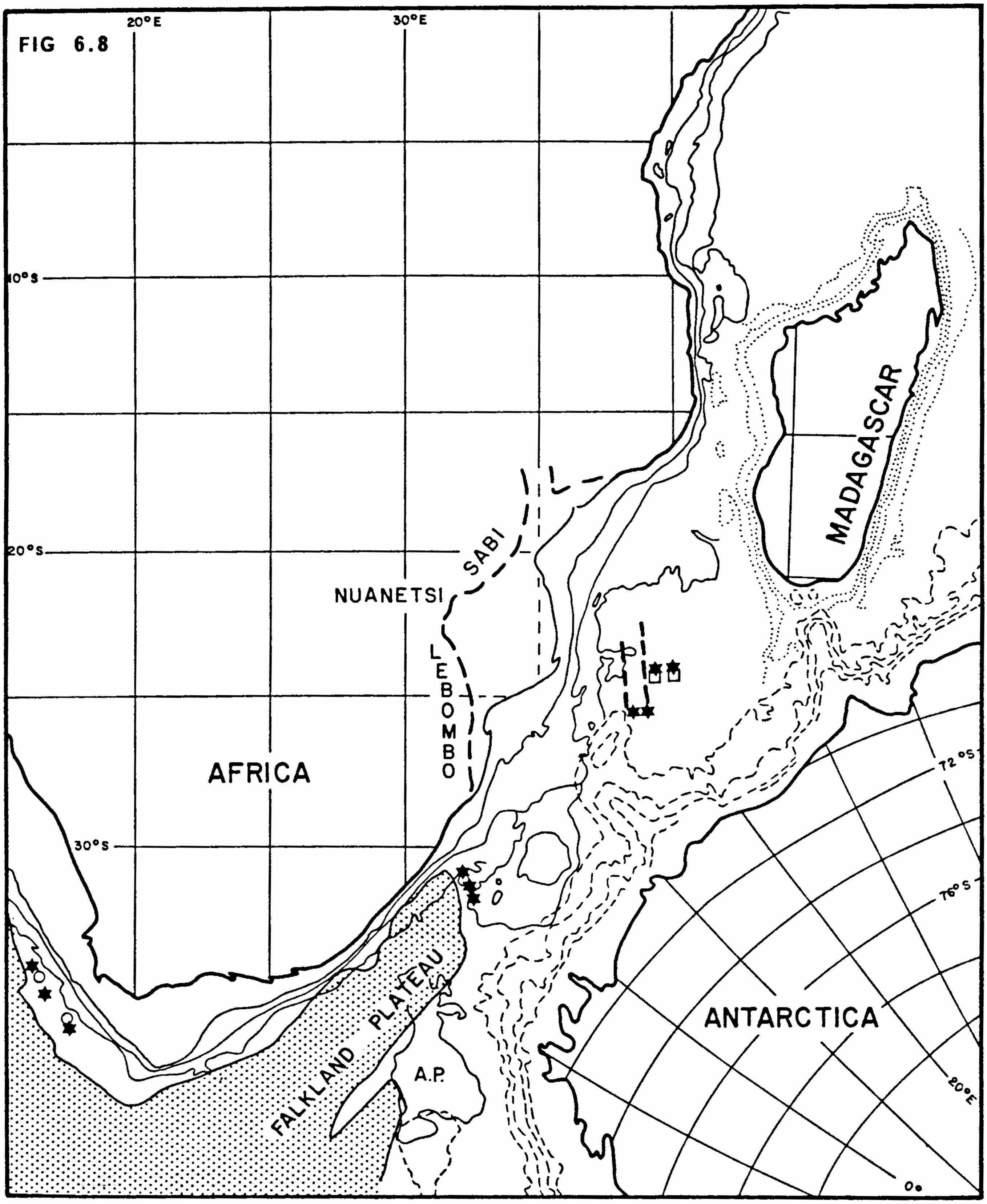
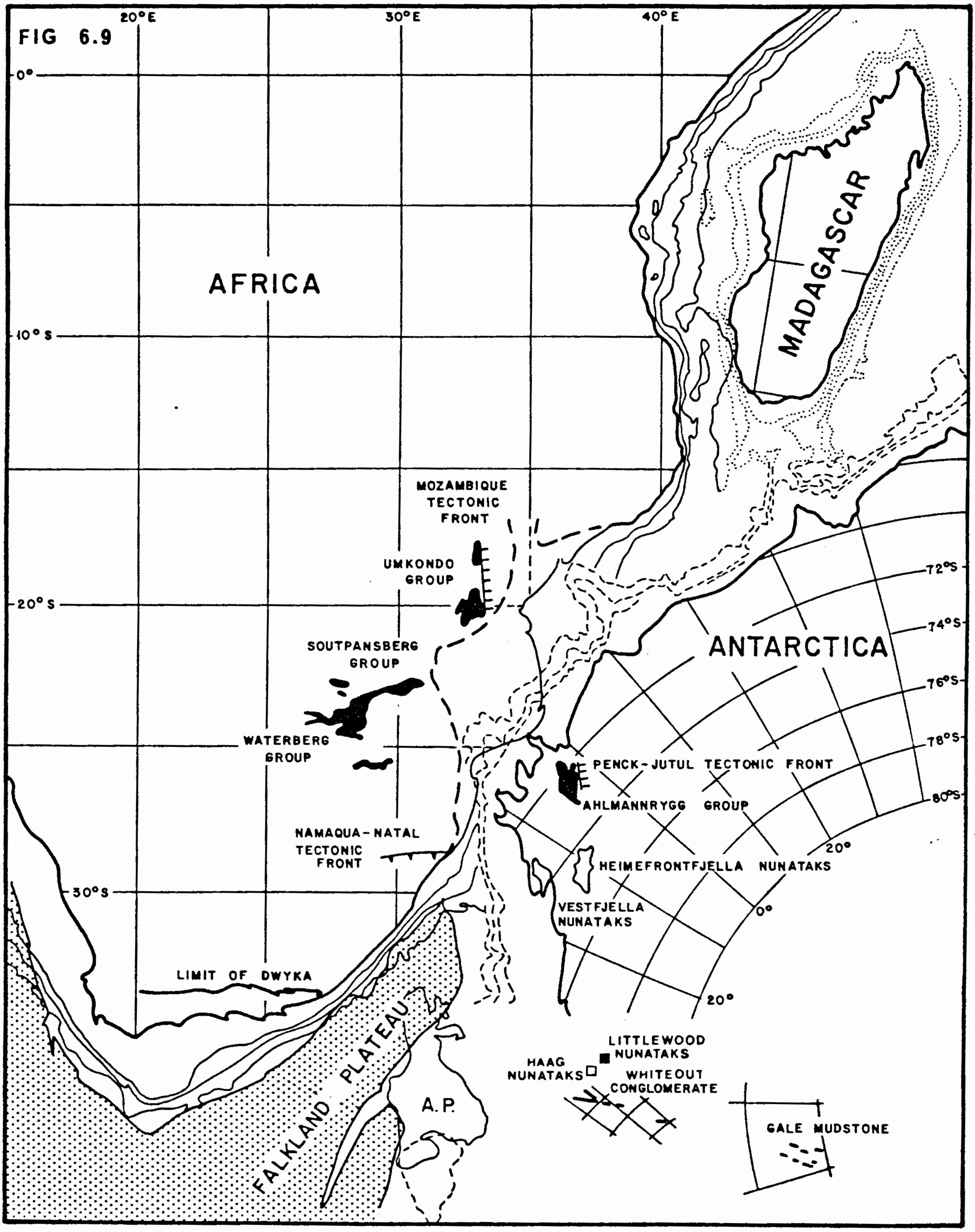


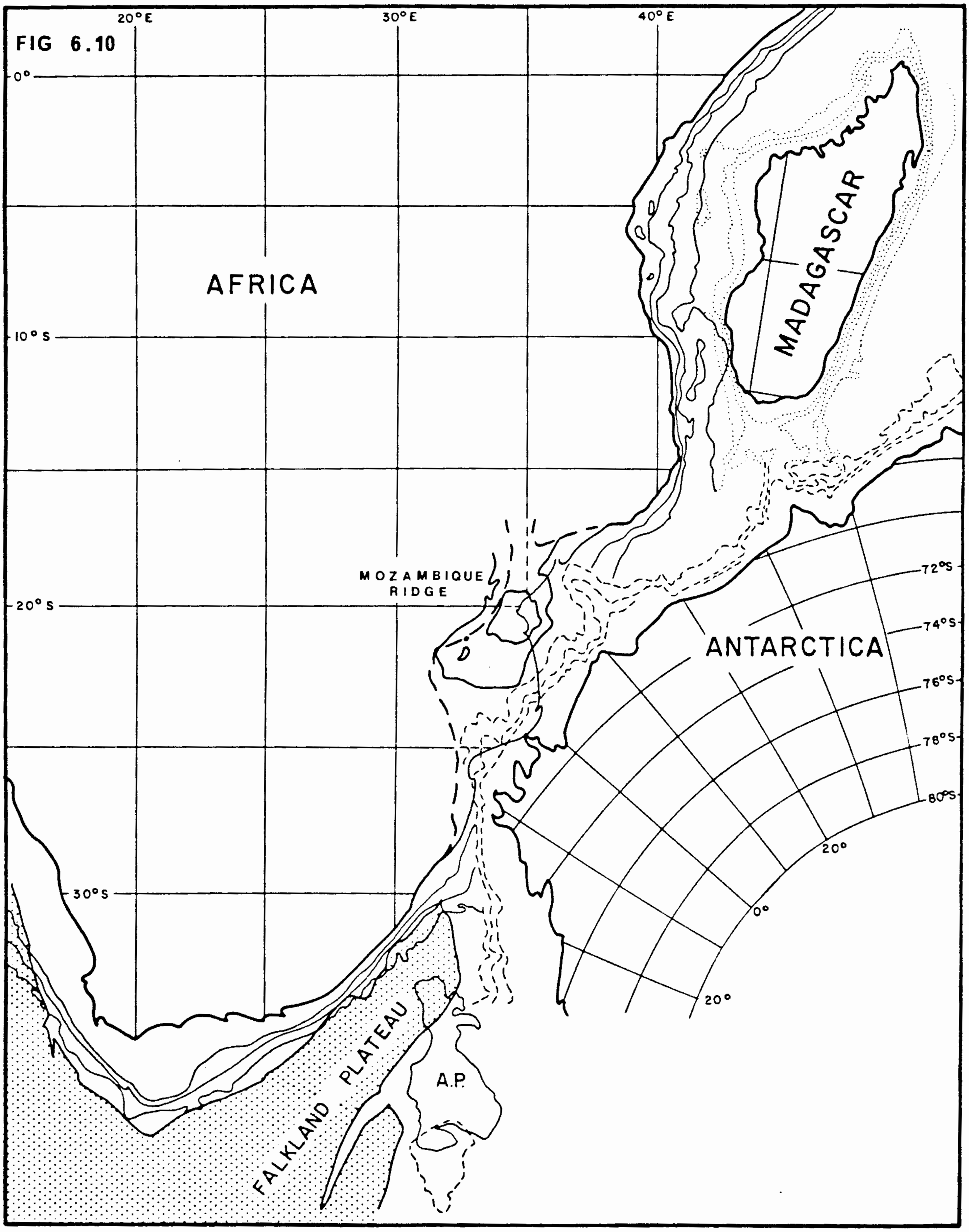
FIG 6.8
20°E
30°E
10°S
20°S
30°S
72°S
76°S
20°E
0°
AFRICA
MADAGASCAR
ANTARCTICA
SABI
NUANETSI
LEBOMBO
FALKLAND PLATEAU
A.P.

FIG 6.9



20° E
30° E
40° E
0°
10° S
20° S
30° S
AFRICA
MADAGASCAR
ANTARCTICA
MOZAMBIQUE TECTONIC FRONT
UMKONDO GROUP
SOUTPANSBERG GROUP
WATERBERG GROUP
NAMAQUA-NATAL TECTONIC FRONT
LIMIT OF DWYKA
FALKLAND PLATEAU
A.P.
72°S
74°S
76°S
78°S
80°S
PENCK-JUTUL TECTONIC FRONT
AHLMANNRYGG GROUP
HEIMEFRONTFJELLA NUNATAKS
VESTFJELLA NUNATAKS
20°
0°
20°
LITTLEWOOD NUNATAKS
HAAG NUNATAKS
WHITEOUT CONGLOMERATE
GALE MUDSTONE

FIG 6.10


20° E
30° E
40° E
0°
10° S
20° S
30° S
AFRICA
MADAGASCAR
MOZAMBIQUE RIDGE
ANTARCTICA
72°S
74°S
76°S
78°S
80°S
20°
0°
20°
FALKLAND PLATEAU
A.P.

southern continental part of the Agulhas Plateau was intruded by oceanic crust around 93 m.y. and thus formed a smaller feature prior to that. Although this refit for the Agulhas Plateau complex is not constrained by magnetic anomalies, experimentation with alternative tectonic schemes has not uncovered a more likely fit. For example, if the Maud Rise (and attached Agulhas Plateau) remained part of the Antarctic plate prior to 93 m.y. its palaeoposition would impinge on oceanic crust formed between the Falkland Plateau and the Natal Valley.

The Astrid Ridge is placed adjacent to the Mozambique Ridge. To date, neither seismic refraction (Chetty & Green, 1977), nor geochemical analyses of basalts (Erlank & Reid, 1974; Thompson et al., 1982) has unequivocally proven whether the Mozambique Ridge is oceanic or continental in nature. Older Antarctic palaeopositions (Figs.6.7-6.10) which overlap Antarctica and the Mozambique Ridge, imply that the Mozambique Ridge must either a) have formed during the early separation of Antarctica and Africa, suggesting an oceanic origin; within this oceanic option, it may either have i) formed entirely on the African plate in which case its formation spanned the period from 145 m.y. (Fig.6.9) to 80 m.y. (Fig.6.3); or ii) it formed between 145 m.y. and 113 m.y. (Fig.6.6) with its southern half having originally been part of the oceanic Antarctic plate; b) alternatively, if it is continental, it did not originally occupy its present-day position relative to Africa. Its neat fit against the Astrid Ridge leads to the tentative suggestion that it moved southwards to its present-day position while attached to Antarctica. The half-rotated epicentres suggest that a ridge adjustment occurred in the Mozambique Ridge area: in M2 times the Astrid Ridge and the Mozambique Ridge were rifting apart near an R-R-R triple junction situated between the tip of the Falkland Plateau and Antarctica. This supports options (aii) and b) above. Because the balance of geochemical evidence indicates that Mozambique Ridge basalts are similar to mid-ocean ridge basalts (MORB) option (aii) is preferred: the Mozambique Ridge was originally a piece of the oceanic part of the Antarctic plate; it formed between 145 and 113 m.y. ago (M21-M2) and was captured by the African plate in a ridge-jump episode around M2 times.

Included in Figure 6.6 are the lineated anomalies found in the Weddell Sea (Barker & Jahn, 1980; LaBrecque & Barker, 1981). From a consideration of the velocity triangle related to the movement of the three plates (Fig. 6.6b) these anomalies are oriented in an appropriate direction for their generation on the southern limb of the triple junction separating South America and East Antarctica. Age assignments for the Weddell Sea anomalies have varied (LaBrecque, 1977; Barker & Jahn, 1980; LaBrecque & Barker, 1981) and the latest version, dating these reversals as M29-M1, does not agree with our model. The velocity triangle (Fig.6.6b) suggests that South America and East Antarctica were separating at a half-rate of 1.5 cm yr^{-1} rather than at 0.6 cm yr^{-1} as suggested by LaBrecque & Barker (1981).

6.1.5 M10 = Early Hauterivian = 122 m.y.

This reconstruction marks the production of the earliest oceanic crust between South America and Africa (chapter 4). Madagascar is rotated farther north towards Kenya/Somalia, while Antarctica approaches the eastern tip of the Falkland Plateau and overlaps the present position of the Mozambique Ridge (Fig.6.7). If this latter feature is oceanic and has always been part of the African plate, its southern region formed after M10 times. If it was attached to the Astrid Ridge (options aii) and b) of section 6.1.5), at this stage it would have fitted between the Tugela Cone, the Falkland Plateau and Astrid Ridge (Fig.6.8). DSDP hole 249 (Simpson, Schlich et al., 1974) at latitude 30°S on the Mozambique Ridge bottomed in basalt of indeterminate age, overlain by Neocomian (possibly Valanginian) clays. The doubt over the exact age of the basalt means that age considerations alone do not rule out any of the possible origins of the Mozambique Ridge (Figs.6.7, 6.8). If the scenario shown in Figure 6.8 is correct, this could explain why seafloor spreading anomalies were not found in the eastern half of the Natal Valley (chapter 4, section 4.2). Either way, the initiation of a three plate system suggests a triple junction was formed at this time. This implies that the earliest anomaly in the Weddell Sea associated with the triple junction is ∿M10 in age. Prior to this, oceanic crust may have been produced in the Weddell Sea as East Antarctica separated from the West Gondwanaland plate, or as micro-plates of West Antarctica moved.

6.1.6 M21 = mid Kimmeridgian = 145 m.y.

In this reconstruction (Fig.6.9) (which is considered pre-drift) South America and the Falkland Plateau wrap around the southern tip of Africa, with the eastern end of the Falkland Plateau juxtaposed with the southern face of the Tugela Cone (chapter 3). Madagascar is placed next to Kenya/ Somalia (Segoufin & Patriat, 1980). The original palaeoposition of Antarctica is obtained by applying the M2-M21 rotation of Segoufin & Patriat to the M2 reconstruction for Antarctica shown in Figure 6.6. Note that this refit, which was based on accurately juxtaposing seafloor spreading anomalies and fracture zones, places the Antarctic Margin against the Lebombo Mountains, and sub-parallel to and just south of the Sabi volcanic outcrops. If alternative refit parameters for Madagascar (Scrutton et al., 1981) are applied to Antarctica, a slightly more northerly position results. This would be in accord with the identification of M22 (Segoufin, 1978) in the Mozambique Basin. Following from Figure 6.7, Agulhas Plateau and Maud Rise fit between Falkland Plateau, Burdwood Bank and Antarctica. If the southern Mozambique Ridge is continental, it may have fitted between the Astrid Ridge and Nuanetsi (Fig.6.10).

Several important implications follow from this pre-drift reconstruction. Firstly, the 190-137 m.y. old (Flores, 1973; Cleverly & Bristow, 1979) acid volcanic rocks of Sabi, Nuanetsi and the Lebombo mark the Jurassic continental margin of south-eastern Africa. This has previously been suggested on petrological grounds (Cox, 1970; Flores, 1970; Betton & Cox,

1979) while from the orientation of dyke swarms, Reeves (1978) considered that Nuanetsi represents a Mesozoic triple junction. Secondly, the coastal plain of Mozambique is underlain by volcanic rocks (Flores, 1973), part of which must be oceanic, having formed during the early separation of Africa and Antarctica. The amount of oceanic crust under coastal Mozambique depends on the exact fit adopted ie. Figure 6.9 or a slightly tighter fit based on the Madagascar reconstruction of Scrutton et al. (1981), or Figure 6.10. In forming a post-break-up outbuilding feature, coastal Mozambique is analogous to the Niger delta of West Africa. Thirdly, Antarctica initially moved southwards along a transform fault system which lay just east of and sub-parallel to the Lebombo mountains, the eastern face of the Tugela Cone, the eastern face of the Falkland Plateau and the eastern side of the Agulhas Plateau in their reconstructed positions (Fig.6.9). This interpretation for the Lebombo area is controversial since it has traditionally been considered a region of tension (Du Toit, 1929; King, 1972; chapter 7). Lastly, this reconstruction exacerbates the overlap of West Antarctica and the Falkland Plateau/South Africa region, if West and East Antarctica are retained in their present relative positions. Movement of microplates possibly in a back-arc situation is required.

6.2 COMPARISON OF PRE-BREAK-UP INTERCONTINENTAL GEOTECTONIC FEATURES

Three pre-break-up geological features common to Africa and Antarctica are used to test the revised reconstruction of East and West Gondwanaland.

6.2.1 Late Palaeozoic Gondwanide Orogeny

In South Africa, the northern tectonic front of the Cape Fold Belt coincides with the southern outcrop limit of the Permian Dwyka Tillite Formation (Fig.6.9). This has equivalents in the Whiteout Conglomerate of the Ellsworth Mountains (Craddock, 1969) and the Gale Mudstone of the Pensacola Mountains (Schmidt & Ford, 1969). The Ellsworth Mountains are known to be a "displaced terrane" (Schopf, 1969; Watts & Bramall, 1981) and have recently been quantitatively refitted with East Antarctica. The Cape Fold Belt front has previously been matched with a morphotectonic boundary on the Falkland Plateau (chapter 3) and these features align very well with the limits of the Gale and Whiteout Formations. The philosophy of aligning the Cape Fold Belt with the TransAntarctic mountains has already been debated (eg. Dalziel, 1980; Harrison et al., 1980). Alignment of the orogenic belts is here considered to lend support to the revised reconstruction (Fig.6.9) since palaeohydraulic analysis of the Triassic Molteno Formation indicates a linear highland in the locality of the aligned orogens (Turner, 1980).

6.2.2 The Mozambique Belt and the Sverdrupfjella front

The western limit of Pan-African reactivation of the Mozambique Belt is marked in Zimbabwe by the eastern limit of the early Proterozoic Umkondo Group. In Dronning Maud Land, Antarctica, the Sverdrupfjella front forms an equivalent feature along the Penck-Jutul rift. This separates the high-grade metamorphic rocks of the Sverdrupfjella to the east, and the older platform covered by the little-deformed and slightly metamorphosed Ahlmanrygg Group to the west (Eastin et al., 1970; Watters, 1972). In the revised reconstruction, the Sverdrupfjella and Mozambique tectonic fronts have similar N-S trends and are aligned relatively well. It has been suggested (albeit within an alternative model for Gondwanaland reconstruction) that in the early stages of break-up, some opening occurred in the Jurassic Zambezi Graben (Burke & Whiteman, 1973). Such second order movements may have contributed to the misalignment by pushing the Mozambique Belt relatively westward (c.f. discussion in chapters 3 & 4).

6.2.3 Ahlmanrygg Group/Soutpansberg Group/Waterberg Group cratonic terranes

The Ahlmannrygg Group is intruded by mafic sills dated at about 1700 m.y. (Allsopp & Neethling, 1970) and it is significant that ∿1740 m.y. old basic volcanics occur within the lithologically similar Soutpansberg Group in South Africa (Barton, 1979). The latter unit, together with the correlative

Waterberg and Umkondo Groups, appears in close spatial proximity to the Ahlmannrygg Group in the proposed reconstruction (Fig.6.9), suggesting that the Archaean to early Proterozoic cratonic terranes of Southern Africa, including the Kaapvaal and Zimbabwe blocks, do have a direct, if limited extension in East Antarctica. This contradicts the common assumptions (Truswell, 1970)

a. that the Kaapvaal block terminates beneath the Lebombo Monocline structure, and

b. that the Mesozoic break-up of Gondwanaland always follows pre-existing Pan-African tectonic trends (Cox, 1978).

6.2.4 Does the Namaqua-Natal Belt have an equivalent in East Antarctica?

The overthrust tectonic front of the E-W trending middle Proterozoic (∿950-1200 m.y.) Namaqua-Natal Belt of South Africa (Figs.2.1, 6.9) is a major tectonic feature (Mathews, 1972) which should find a continuation in Dronning Maud Land. If so, then it places a potentially crucial latitudinal constraint on the accuracy of the proposed reconstruction. Direct linear extrapolation of the front eastwards suggests that it should extend just north of the Vestfjella and Heimefronfjella nunatak groups (Fig.6.9). If the crystalline basement rocks of these areas (Juckes, 1972) return middle Proterozoic (equivalent of Namaqua-Natal Belt) rather than early Proterozoic or Archaean ages (equivalent of the Kaapvaal craton) then the illustrated reconstruction based on closure to the Jurassic M21 anomaly, will be confirmed (Fig.6.9). Alternatively, if this tectonic and geochronologic boundary runs south of the Heimefrontfjella, it may indicate that this reconstruction does not entirely close the rift between northern Mozambique and Dronning Maud Land. A slightly tighter fit may be required which straightens the southeastwards bend of the Permo-Triassic Gondwanide orogeny and brings the Ahlmannrygg Group into still closer proximity to its Soutpansberg and Umkondo Group correlatives. Future geochronological and structural investigations of the Heimefrontfjella nunataks would constitute an important test of the reconstruction.

In South Africa, the Namaqua-Natal Belt is considered to underlie the Karoo trough and the southern boundary of this age province is the northern front of the Cape Fold Belt (Figs.2.1, 6.9). In this latter area, dates clustering around 500 m.y. and 250 m.y. have been obtained (chapter 3). The Ellsworth Block includes the ∿1000 m.y. old Haag nunataks, which, when rotated relative to East Antarctica, are placed adjacent to the ∿1000 m.y. old Littlewood nunataks (Eastin & Faure, 1971; Clarkson & Brook, 1977). Both of these lie just north of the Whiteout conglomerate outcrops of the Ellsworth Mountains, and can be considered to lie within the correct age province. The Agulhas Plateau has returned ages of ∿1100 and 500 m.y. (Allen & Tucholke, 1981) and may be transitional between the mid-Proterozoic and Cape Fold Belt provinces.

A problem arises in considering age provinces and the correlation of pre-break-up geotectonic features in the case of the Falkland Islands. There, the Cape Meredith metamorphic complex has yielded ∿1000 m.y. dates, while the equivalent to the northern front of the Cape Fold Belt has been mapped (Newton, 1980; Dalziel & Elliott, 1982 and references therein). The Early Cretaceous palaeoposition of the islands is well constrained (chapters 3 & 4). Perhaps in an earlier phase of movement, the Falkland Islands moved south-westwards relative to the rest of Gondwanaland. This is a similar scenario to that envisaged by Watts & Bramall (1981) for the Ellsworth Block, and which is included in reconstructions presented here (Fig.6.9). However, the relationships of the Falkland Islands (Harrison et al., 1980) and West Antarctica to Gondwanaland remains enigmatic.

6.3 A COMPARISON OF STRATIGRAPHIC AND MAGNETIC DATING: BREAK-UP IN MOZAMBIQUE

As mentioned above, the exact pre-drift fit of East Antarctica/ Madagascar with Africa depends on whether a fit based on closure to M21 is adopted (Fig.6.9), or whether a slightly closer fit is preferred, commensurate with the scheme of Scrutton et al. (1981). Break-up is dated therefore at >145 m.y., possibly as old as 170 m.y. depending on the position of the C.O.B. in the Mozambique Channel and the Comores Basin. The Lebombo volcanic rocks contain some conglomeratic beds and are overlain by a basal continental conglomerate in Zululand and the Sena sandstone in Mozambique (Flores, 1973). The earliest overlying marine sediments are Neocomian (Barremian - Froster, 1975), that is at least 27 m.y. younger than the proposed break-up. Why did the sea fail to penetrate to this area, when it had already flooded the Outeniqua and Falkland Plateau Basins (chapter 3) and northern Mozambique, Madagascar and Kenya by the Jurassic (Flores, 1970)?

The crust underlying the northern Natal Valley, which is here considered oceanic, is intermediate in thickness (Darracott, 1974; Scrutton, 1976b). Massive volcanism in Late Karoo times preceded break-up in this area. Perhaps, during initial break-up, anomalously thick oceanic crust was emplaced, and the area was subaerially exposed as is the case today in Iceland and Afar (chapter 8).

6.4 SUMMARY

Using published magnetic anomaly and fracture zone identifications (Fig.6.1) a revised sequence of reconstructions from chron 29 (65 m.y.) to chron M21 (145 m.y.) has been computed for East Antarctica and Africa. These indicate that the Mesozoic crust identified off Dronning Maud Land, Antarctica (Bergh, 1977) is equivalent to the easternmost identified Mesozoic crust of the Mozambique Basin (Simpson et al., 1979). This leads to a revised palaeoposition for East Antarctica which overlaps the coastal plain of Mozambique and the northern Natal Valley, and places the Antarctic continental margin against the Jurassic acid volcanic outcrops of the Lebombo Mountains, and just south of similar rocks stretching from Nuanetsi beyond

Sabi (Figs.6.9, 6.10). This corroborates the premise (made on petrological grounds (Betton & Cox, 1979)) that these rocks mark the Jurassic continental margin of south-eastern Africa. This refit is supported by a) alignment of the northern tectonic front of the Cape Fold Belt with the southern outcrop limits of the Whiteout Conglomerate in the Ellsworth Mountains in their relocated position (Watts & Bramall, 1981) and the Gale Mudstone of the Pensacola Mountains; b) the alignment of the western limit of Pan-African reactivation of the Mozambique Belt with the western limit of the Sverdrupfjella metamorphic rocks in Dronning Maud Land. The rift implies that the Mozambique Ridge is either oceanic in nature or, if continental, it is allochthonous with respect to Africa.

During initial separation (M21-M10) East Antarctica and Madagascar moved southwards along the Davie Fracture zone and a fracture zone just east of and sub-parallel to the Lebombo Mountains, continuing along the eastern face of the Tugela cone, the eastern margin of the Falkland Plateau and the eastern flank of the Agulhas Plateau. Contemporaneously oceanic crust must have been generated south of South America. During this stage, sediments from the proto-Limpopo and Zambezi rivers began building outwards forming the coastal plain of Mozambique.

At M10 times, South America and the Falkland Plateau began to separate from Africa implying that a triple junction was initiated between the Falkland Plateau and Antarctica in the southern Natal Valley (Figs.6.7, 6.8). At or just after M2 times Astrid Ridge and Mozambique Ridge separated, possibly as the result of a ridge adjustment (Fig.6.6). Just after M0 times, seafloor spreading stopped in the Somali Basin (Segoufin & Patriat, 1980) and a new ridge was initiated between Madagascar and Antarctica. This was contemporaneous with a change in the rotation pole between South America and Africa (chapter 4) and possibly with the splitting of Agulhas Plateau from Burdwood Bank (chapter 5). Around 93 m.y. a spreading re-organisation caused a split between Maud Rise and the Agulhas Plateau. Gunnerus Ridge, Conrad Rise and Del Cano Rise are all tectonically linked with the Madagascar Ridge and are most likely associated with the Prince Edward hot-spot.

CHAPTER 7

STRIKE-SLIP CONTROL OF COASTAL NATAL FAULTS DURING GONDWANALAND BREAK-UP

7.1 INTRODUCTION

An extensive and complex network of faults has been mapped in coastal Natal (Kent, 1938; Gevers, 1941; Beater & Maud, 1960; Maud, 1961). The faulting is apparently Jurassic in age (Frankel, 1960) and has been associated with the break-up of Gondwanaland (eg. Maud, 1961) although this has been disputed (Brock, 1960). From a consideration of ice flow directions associated with the Dwyka diamictite and palaeocurrent directions in the Ecca Group, geologists have long recognised that a land-mass lay to the east of Natal during Karoo times (Du Toit, 1937; King, 1953; Crowell & Frakes, 1972). However, it was believed that this land-mass moved away from southern Africa in an easterly or south-easterly direction. Du Toit (1929) considered that east-west oriented tension prevailed in the Lebombo volcanic province. When this region was interpreted as a continental margin, Cox (1970) and Flores (1970), postulated that Madagascar had moved from this area in an easterly direction. Kent (1974a) interpreted the Jurassic geology of southern Kenya and Tanzania as indicating east-west oriented tension, while in Natal, Gondwanaland break-up was associated with south-east oriented tension (Maud, 1961; Hardie, 1962; King, 1972).

Recent plate tectonic reconstructions (chapters 3 - 6) indicate that much of south-eastern Africa's continental margin has developed along transform faults producing sheared margins. This dictates that rather than tensional forces, the dominant strain has been manifested along rotational couples. An apparent dilemma arises between tensional régimes interpreted from land-based geology and transform or strike-slip régimes dictated by marine geophysics.

In this chapter I attempt to resolve the dilemma by

(a) briefly describing the structural style of Jurassic-Cretaceous faulting in coastal Natal;

(b) outlining the areas of south-eastern Africa which represent sheared margins produced during Gondwanaland break-up;

(c) examining the tectonic styles attendant on strike-slip or wrench faulting;

(d) demonstrating that many features of the Natal coast are possibly explicable in terms of strike-slip tectonics.

7.2 FAULT PATTERN OF ZULULAND, COASTAL NATAL, AND NORTHERN TRANSKEI

The main concentration of faulting is exposed on land between 28° 30' and 31°S and between 30 and 32°E (Fig.7.1). Extending southwards from this faulted and fractured region is the N-S oriented Bongwan Gas fault system which has a downthrown slice between two curved faults (Gevers, 1941). South of this, several en échelon east-west oriented faults fracture the Transkei coast, particularly around Port St. Johns and to the south of Figure 7.1.

The trace of the Egosa fault near Port St. Johns shows an 's' shape and delineates a half graben. From between Bongwan and Port Shepstone, a north-north-west trending arc of faults extends towards Highflats and then swings north-eastward and eastward towards Durban. This sequence of faults exhibits several series of north and north-east oriented en échelon faults, diverging and converging faults, zig-zags, and splays of faults emanating from curved sections of master faults. Around Highflats, to the east of the curved fault, the base of the Natal Group is upthrown to an elevation of $\sim$1000 m. En échelon curved faults of similar orientation occur along the coast. Note that on some faults the down-throw reverses from east to west along strike.

From the Mid Illovo Eston area a second arc stretches north-north-west then north-east towards the Ngoye Horst. Maud (1961) considers the area to the east of this arc to represent a single block dipping to the east, exhibiting tilted block step-faulting. Again, the fault pattern displays diverging and converging faults, reversal of throw, wedges and splays. En échelon faults occur along the coast north of Durban.

A third arc begins south-east of Greytown and extends towards the north-east side of Nkwaleni Graben (Beater & Maud, 1960). South-east of Greytown

FIGURE CAPTIONS

Fig.7.1. Fault pattern of Zululand, coastal Natal and northern Transkei (after Maud, 1961; Geological Survey maps, 1970). Key: 1 downthrow on faults; 2 Molteno Formation; 3 Sabie River Basalt (Lebombo nomenclature after Cleverly & Bristow, 1979); 4 South Tugela Ridge; 5 sea-floor magnetic anomalies; 6 Cape Slope magnetic anomaly. Bathymetry after Goodlad (1978).

Fig.7.2. Sheared margins of south-east Africa, fracture zones and magnetic anomalies of the south-west Indian Ocean.

Fig.7.3. Strain ellipse for east-west oriented dextral couple. ψ shear angle; C vector of compression parallel to minor axis of ellipse $x - x_1$; E vector of tension parallel to long axis of ellipse $y - y_1$. Note the associated structural styles (after Wilcox et al., 1973; Harding, 1974).

Fig.7.4. Structural styles associated with strike-slip movement. a) en échelon folds; b) offset by continued wrenching; c) pull-apart graben d) graben and horsts caused by fault termination; e) folding and faulting on side-stepping fault; f) development of splay fault; g) and h) anastomosing faults leading to upthrust and down-sagged pods. (c-h after Crowell, 1974 a,b; Reading, 1980).

Fig.7.5. a) diverging and converging blocks causing tilted blocks (after Crowell 1974b); b) pod-shaped slice between San Gabriel and San Andreas faults, California. Saw-tooth ornament is edge of upthrust block, Ridge Basin is down-dropped block (after Wilcox et al., 1973); c) part of Fig.1 possibly exhibiting similar features - uplifted block south-east of Greytown, tensional normal faulting of Nkwaleni Graben (N.G.) and Ngoye Horst (N.H.).

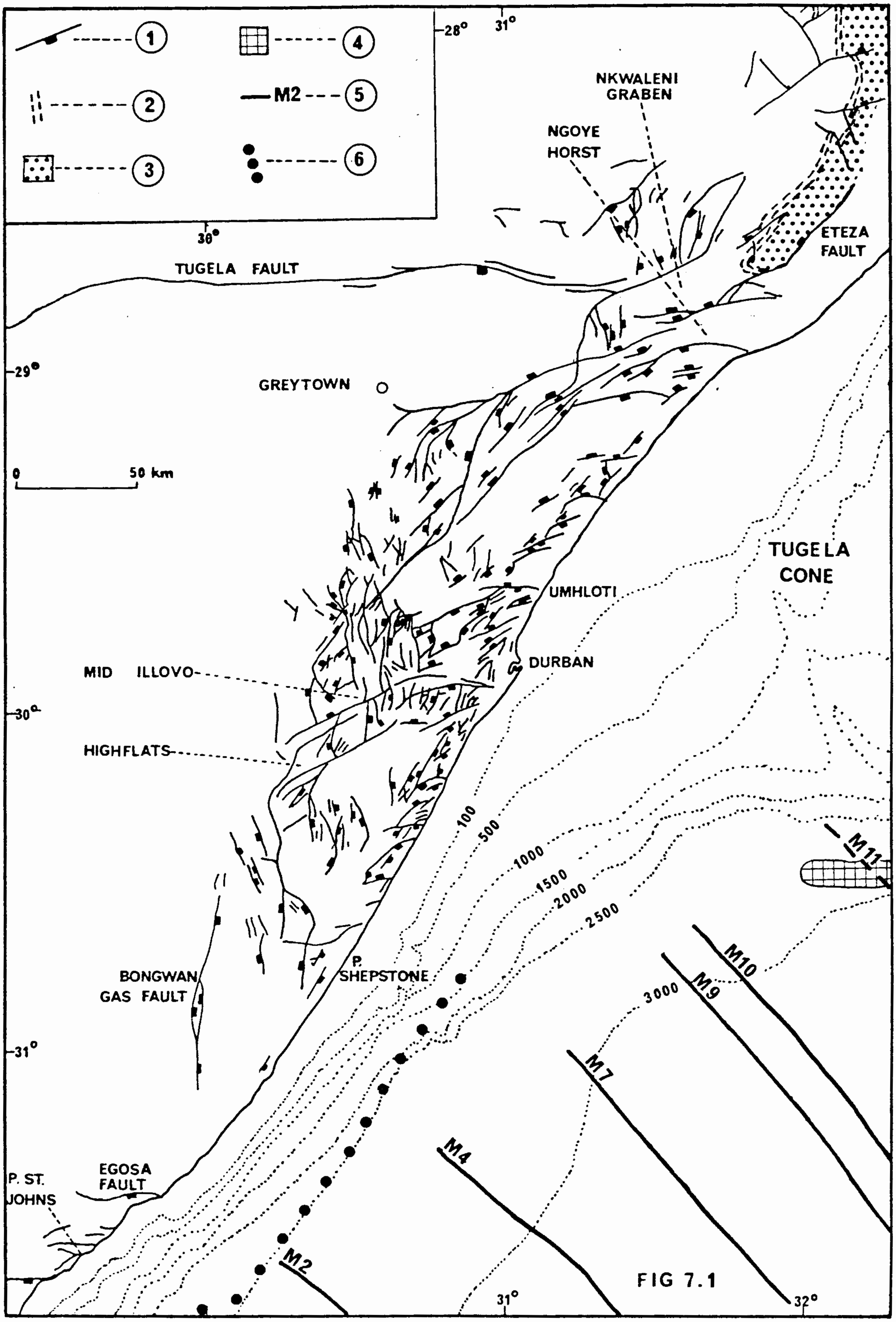

1
2
3
4
M2
5
6
28°
31°
30°
29°
30°
31°
31°
32°
NKWALENI
GRABEN
NGOYE
HORST
ETEZA
FAULT
TUGELA FAULT
GREYTOWN
0
50 km
TUGELA
CONE
UMHLOTI
DURBAN
MID ILLOVO
HIGHFLATS
100
500
1000
1500
2000
2500
3000
M11
M10
M9
M7
M4
M2
P. SHEPSTONE
BONGWAN
GAS FAULT
EGOSA
FAULT
P. ST.
JOHNS

FIG 7.1

FIG 7.2

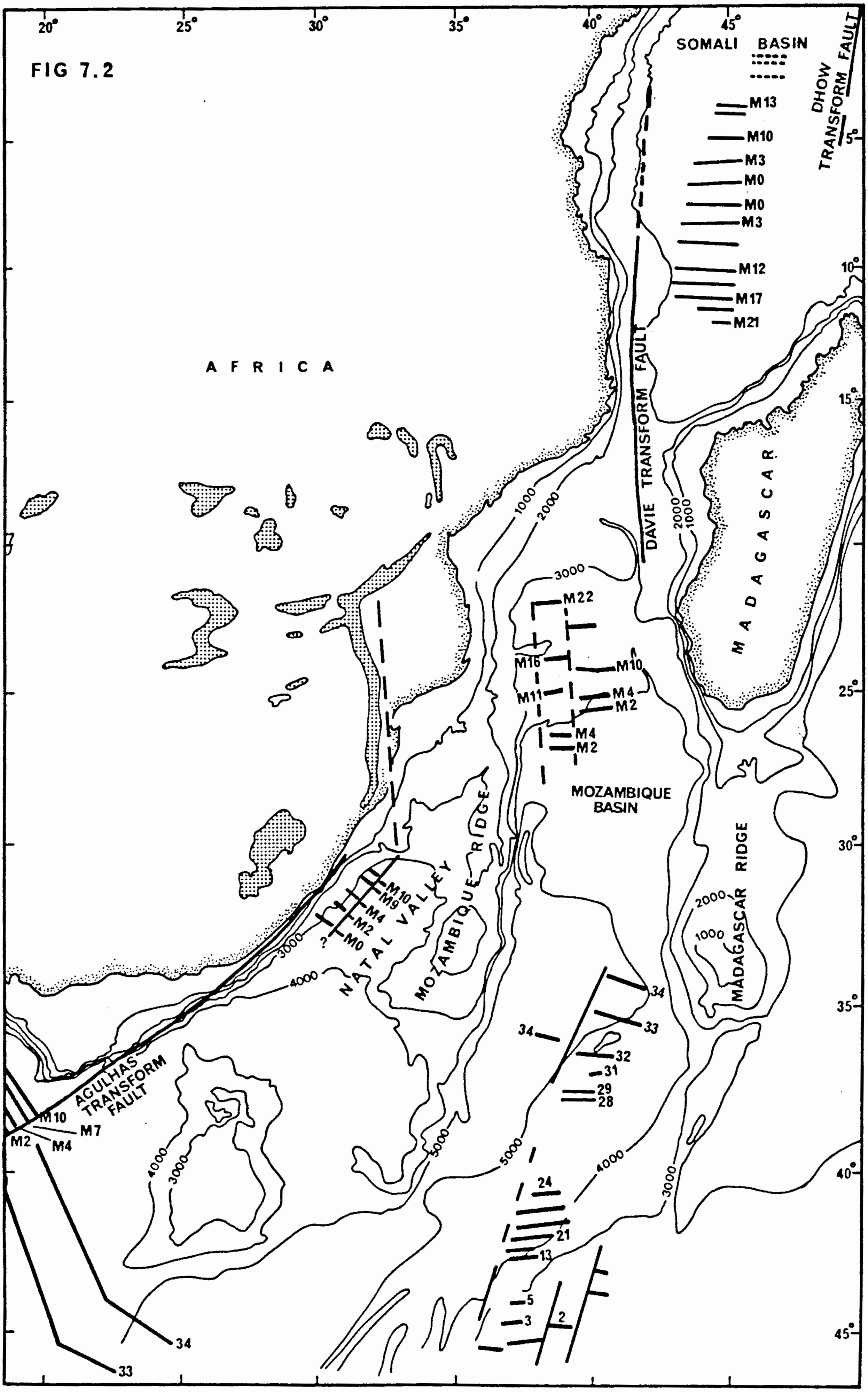

20°
25°
30°
35°
40°
45°
5°
10°
15°
25°
30°
35°
40°
45°
SOMALI BASIN
DHOW TRANSFORM FAULT
M13
M10
M3
M0
M0
M3
M12
M17
M21
AFRICA
DAVIE TRANSFORM FAULT
MADAGASCAR
1000
2000
2000
1000
3000
M22
M16
M10
M11
M4
M2
M4
M2
MOZAMBIQUE BASIN
MOZAMBIQUE RIDGE
NATAL VALLEY
MADAGASCAR RIDGE
M10
M9
M4
M2
M0
?
3000
4000
2000
1000
MADAGASCAR
34
33
34
32
31
29
28
AGULHAS TRANSFORM FAULT
M10
M7
M2
M4
4000
3000
5000
5000
4000
3000
24
21
13
5
3
2
34
33

FIG 7.3

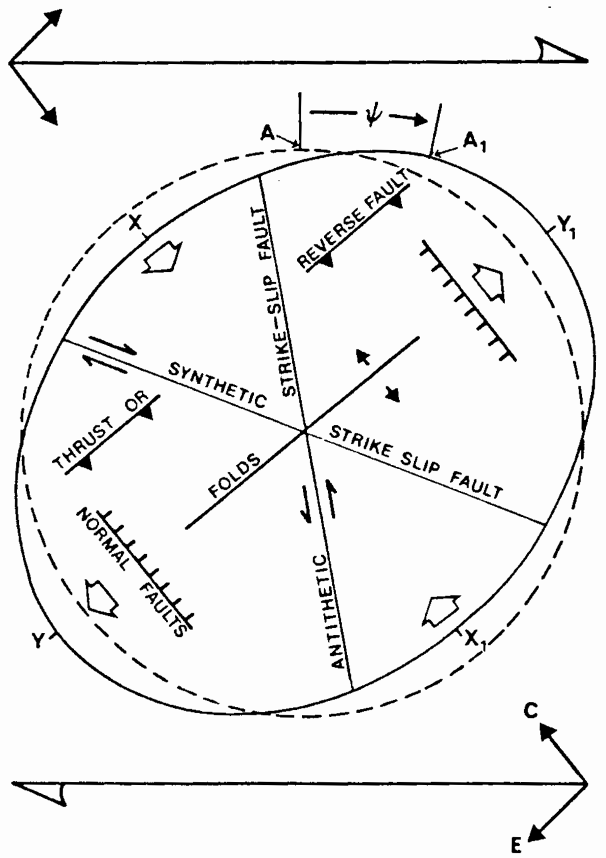
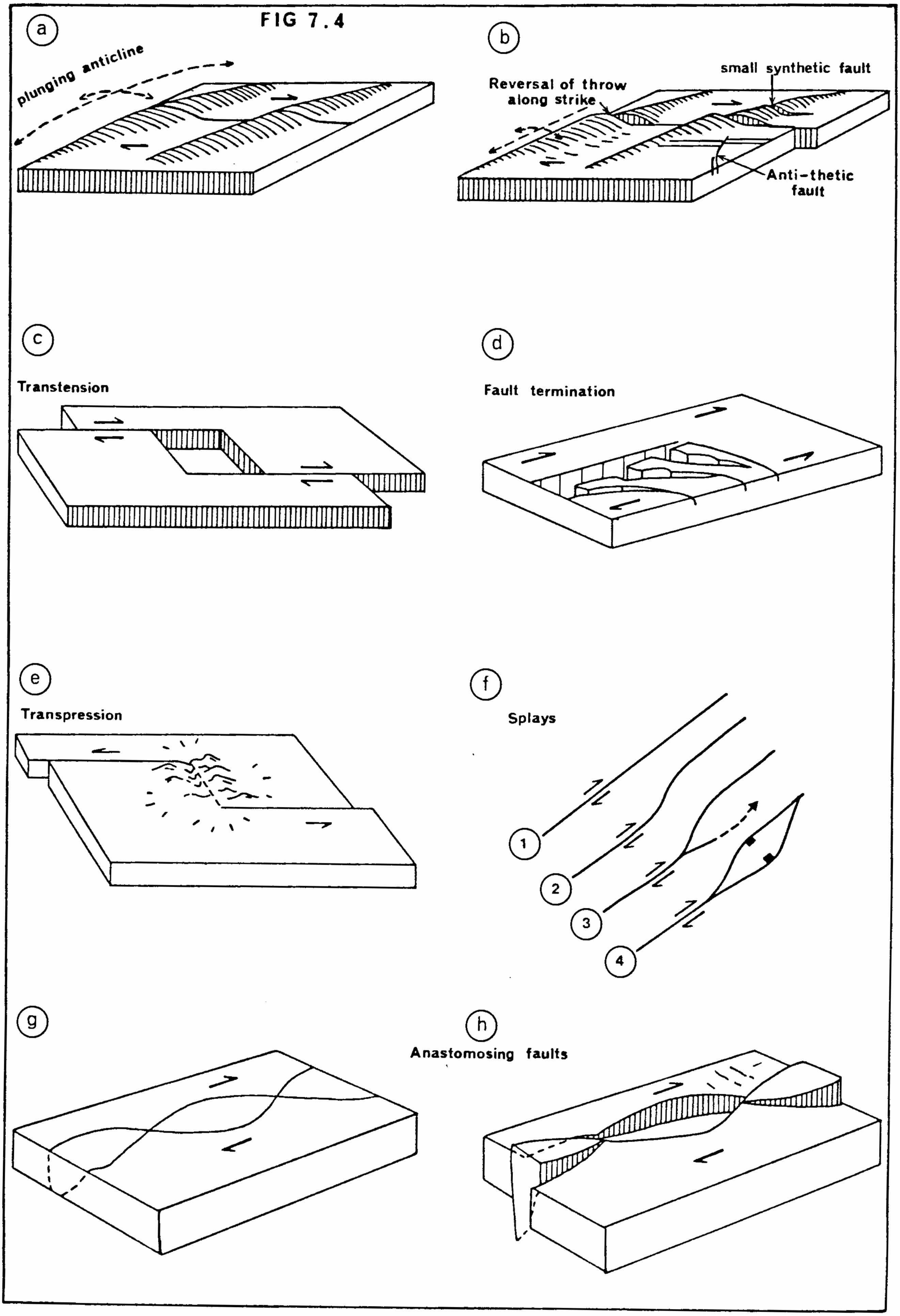

A
ψ
A1
X
Y1
REVERSE FAULT
STRIKE-SLIP FAULT
SYNTHETIC
THRUST OR
FOLDS
STRIKE SLIP FAULT
NORMAL FAULTS
ANTITHETIC
Y
X1
C
E

FIG 7.4


a
plunging anticline
b
Reversal of throw along strike
small synthetic fault
Anti-thetic fault
c
Transtension
d
Fault termination
e
Transpression
f
Splays
1
2
3
4
g
h
Anastomosing faults

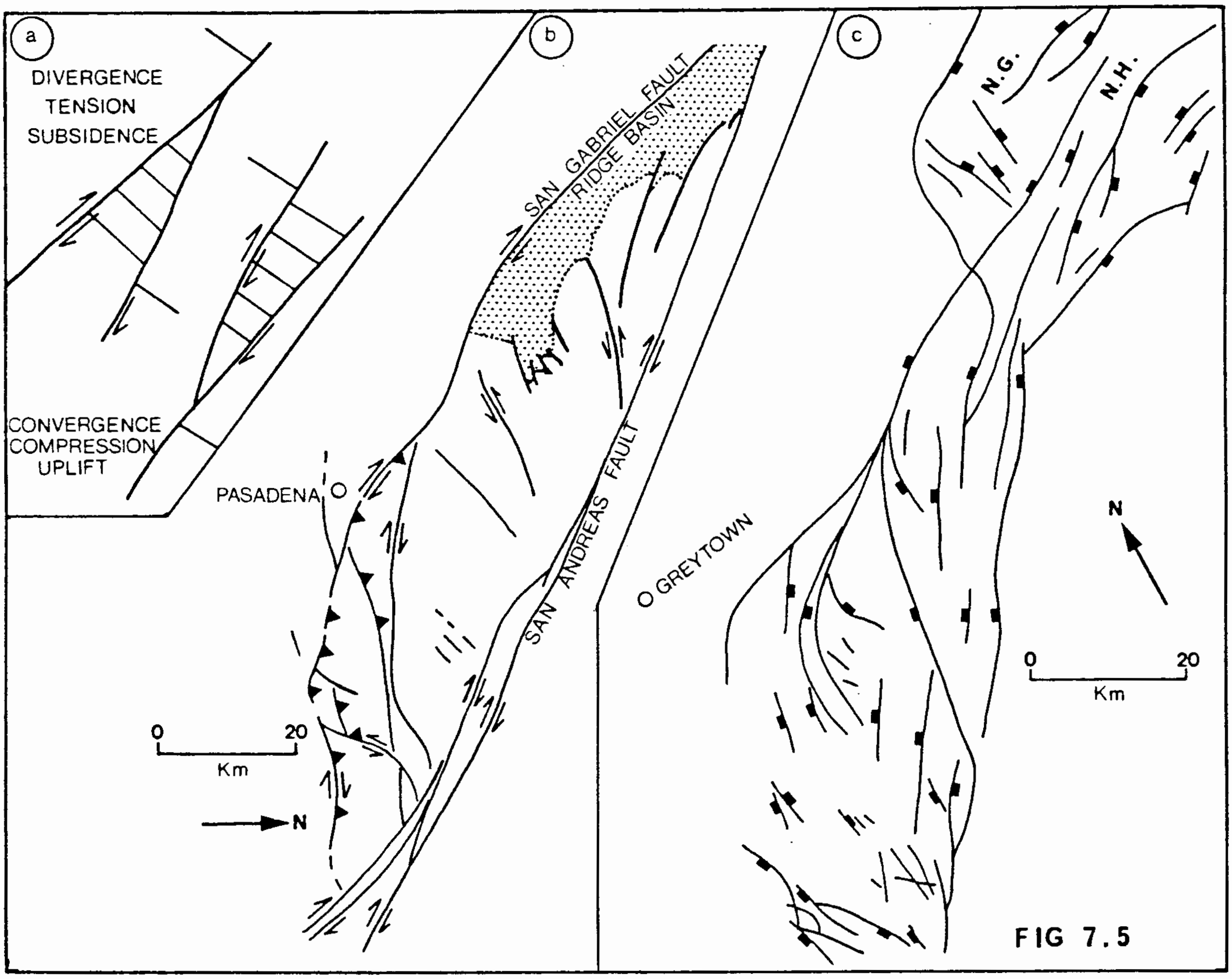

a
DIVERGENCE
TENSION
SUBSIDENCE
CONVERGENCE
COMPRESSION
UPLIFT
b
SAN GABRIEL FAULT
RIDGE BASIN
SAN ANDREAS FAULT
PASADENA
0
20
Km
N
c
N.G.
N.H.
GREYTOWN
N
0
20
Km

FIG 7.5

the eastern up-thrown side raises the base of the Natal Group ($\simeq$the Table Mountain Group) to its greatest elevation outside the Cape Fold Belt (1200 m). A curving fault, exhibiting an inverted 's' shape (sensu Wilcox et al., 1973) forms the northern margin of the Nkwaleni Graben. Throughout the area, curving faults are oriented west-north-west at their eastern extremity. West-north-west and north-west faults also occur north of Mid Illovo.

In the extreme north-east of Figure 7.1, east-north-east and north-north-west oriented faults intersect. East-north-east faults appear to offset the Molteno Formation and Lebombo Group volcanic rocks left laterally, while north-north-west oriented faults appear to cause right lateral offsets.

The Tugela fault is an east-west oriented post Karoo fault (Matthews, 1959) which parallels a Precambrian thrust zone between the Namaqua-Natal belt and the Kaapvaal Craton.

Folds have not been widely recognised, but are reported around Port Shepstone (Gevers, 1941). Du Toit noted folds near Durban (1954, p.570) while Kent (1938) described symmetrical folds near Umhloti (fold axes oriented 50°E of north). Elsewhere, drag folds are associated with motion on fault planes (M.J. McCarthy, personal communication, 1982).

Faulting is dated with reference to two key faults. In the north, the Eteza fault throws Lebombo volcanic rocks but apparently has not affected Coniacian and possibly Aptian conglomeratic lacustrine (fan deltaic) rocks overlaying the fault trace (Frankel, 1960; Kennedy & Klinger, 1975). Lebombo rhyolites are isotopically dated at 190 $\pm$5 m.y.(Cleverly, 1979) thus faulting occurred between $\sim$190 and 112-90 m.y. ago. In the south, Upper Valanginian (124-127 m.y.) conglomeratic marine sandstones have been laid down in faulted half graben at Mgazana and Mboyti (Egosa fault)(McLachlan et al., 1976; Klinger & Kennedy, 1979). The major fault has not grossly affected these Cretaceous rocks but some distortion of bedding and brecciation suggests some movement contemporaneous with deposition.

7.3 THE DISTRIBUTION OF SHEARED MARGINS IN SOUTH-EASTERN AFRICA

The plate tectonic reconstructions given in chapters 3 to 6 indicate that sheared margins predominate in south-eastern Africa (Fig.7.2). During the initial stages of fragmentation involving Antarctica/Madagascar/India and Africa/South America, break-up occurred along two long transform systems. The Davie Fracture Zone extended along the Tanzanian continental margin. A second transform system extended from the northern end of the Lebombo, along the east side of the Tugela Cone and along the south-eastern face of the Falkland Plateau in its reconstructed position (chapter 6). These transform faults lie along small circles about the pole of rotation which describes initial relative movement of East and West Gondwanaland. Rifted margins extend along the north-east oriented continental margins of Kenya and Somalia, and from Nuanetsi via Sabi along the north-east oriented continental margin of Mozambique. Motion began at least by M21 times (145 m.y.)(Segoufin & Patriat, 1980). Alternative interpretations suggest that seafloor spreading began in the Somali Basin as early as M25 times (153 m.y. ago)(Parson et al., 1981; Rabinowitz et al., 1983). Extrapolation of spreading rates to the Madagascar continental margin implies that break-up occurred as early as 175 m.y. ago (Rabinowitz et al., 1983). (This assumes that the Comores Basin is oceanic - see Maugé et al., 1982 for an alternative opinion). Similarly in the Mozambique Basin, alternative identifications of the continent-ocean boundary (Darracott, 1974; Lort et al., 1979) imply that break-up began 148-165 m.y. ago.

A third transform fault, the 1300 km long Agulhas margin, stretches from the southern face of the Tugela Cone along the south-eastern margin of South Africa. This transform fault lies along a small circle about the rotation pole describing the motion of the Falkland Plateau. A short rifted margin extends along the southern face of the Tugela Cone. This movement began 122-127 m.y. ago and by 98 m.y. the Falkland Plateau had cleared the tip of the Agulhas Bank.

7.4 STRIKE-SLIP OR WRENCH TECTONICS

7.4.1 Geometry

In large-scale plate tectonic models, transform faults are considered regions where plates passively move laterally past each other along a single vertical fault. In reality, strike-slip faults exhibit a variety of structural styles, since movement may occur along a complex zone of deformation which may be up to 500 km wide (eg. California borderland San Andreas system). Strike-slip or 'wrench' faults form in response to horizontal shear couples within the earth's crust, and these may be right lateral (dextral) or left lateral (sinistral). Examples of deformation accruing from dextral movements only are given here because the movements of East Antarctica/Madagascar and Falkland Plateau relative to South Africa are right lateral. Three basic styles are recognised (eg. Freund, 1965; Lowell, 1972; Wilcox et al., 1973; Crowell, 1974a & b; Reading, 1980).

a) the classic case of simple parallel wrenching in which crustal blocks move parallel with the strike-slip fault;

b) convergent wrenching in which blocks move obliquely towards each other;

c) divergent wrenching in which oblique movement is away from the fault.

Oblique movement leading to a combination of transcurrent and tensional faulting is termed transtensile, while a combination of transcurrent and compressional movement is termed transpressive (Harland, 1971).

A simple, dextral, parallel strike-slip fault results in a characteristic strain ellipse (Fig.7.3)(Wilcox et al., 1973; Harding, 1974). Point A on an undeformed circle moves to point A^1 defining the shear angle (ψ). Points moving closer together indicate compression, while points moving apart mark tension. Thus maximum compression and tension are parallel with the minor and major axes of the strain ellipse (X,X^1 and Y,Y^1 respectively on Fig.7.3). In a deep-seated shear couple, cover rocks may initially deform plastically, leading to en échelon folds oriented at 90° to the maximum compression. Compression is exerted over a finite distance, and folds die out some distance from the fault trace. Hence anticlines plunge away from the fault in the direction perpendicular to maximum compression (Fig.7.4a). The axes of en échelon folds are typically oriented at an angle of 30° $\pm$15° to the wrench strike (Wilcox et al., 1973). However, variations in local geology may influence this and fold axes sometimes parallel or even cross the wrench trend at low angles.

As deformation proceeds, two sets of intersecting vertical fractures form. One set forming at low angles to the wrench trend (10°-30°) is termed synthetic and is of the same sense as the main wrench. The second set oriented at high angles to the wrench trend (70°-90°) is of the opposite sense and is termed antithetic. Both of these fault sets (conjugate strike-slip faults) form en échelon. The main through-going fault then develops, mainly along the synthetic faults becoming a series of interconnected 'dog-legged' faults (this is possibly a propagating process - c.f. section 4.4). Through continued motion faults are rotated, particularly the antithetic ones. As a result they tend to be oriented at even higher angles to the wrench-strike, and the rotation imparts a characteristic reverse 's' shape in the case of dextral movement.

Oblique movement leading to transpression and transtension may be caused by non-parallel movement of crustal blocks or by changes in orientation of a master fault. Compression enhances folding and the formation of conjugate strike-slip faults, and may cause thrust faults and uplifted blocks. In a transtensile regime tension fractures lead to extensive block faulting producing graben and horsts oriented perpendicular to the axis of tension. This type of motion is often taken up by antithetic faults. These fractures thus exhibit negligible lateral displacements, and develop into high-angle normal faults with oblique slip.

7.4.2 Characteristic features

Because the main through-going fault develops along the synthetic faults at a low angle to the wrench trend, it tends to 'dog-leg' or 'zig-zag'.

Along strike therefore zones of transtension and transpression develop simultaneously. Transtension may lead to a pull-apart basin, particularly if one fault dies out and movement is taken up by an adjacent side-stepping fault (Fig.7.4.c). Volcanism, often manifested as basic dykes can occur in transtensile pull-apart basins (Freund, 1965; Wilcox et al., 1973; Crowell, 1974b, Fig.8). If motion is not continued on a parallel fault, tension may lead to horst and graben formation (Fig.7.4d). The end of a fault, or a side-stepping fault may equally cause transpression leading to folding and thrusting (Fig. 7.4e). In such circumstances one block is typically thrust above another leading to an uplifted block adjacent to a downwarped basin (Crowell, 1974a). Similar effects are caused by curving or bent faults. Transpression accentuates the formation of en échelon plunging folds. Offsetting of folds by subsequent faulting results in great variation in throw and reversals of vertical displacement along the wrench strike (Fig.7.4b). Characteristically, strike-slip systems comprise long straight master faults with subsidiary faults or splays branching off. As a master fault develops a curve, splays lead away from the change in strike. They may either continue in a new direction or rejoin the master fault forming a wedge (Crowell, 1974b)(Fig.7.4f). Alternatively a series of splays may form where a fault dies out (Freund, 1974). In a complicated strike-slip zone braided or anastomosing faults cause complex structural patterns such as uplifted and down-sagged wedges along strike from each other (Figs.7.4g,h). Variations and reversals of throw on faults may result from this process too. If curving faults converge, compression ensues, while tension exists in areas where faults diverge (Fig.7.5a). The leading edge of the compressed wedge is uplifted, while the trailing edge of a block subsides with divergence. This can lead to tilted fault blocks.

7.5 STRIKE-SLIP STYLES IN COASTAL NATAL

Faulting in coastal Natal was contemporaneous with break-up of Gondwanaland: break-up occurred in two phases: 175-145 m.y. and ∿125 m.y. ago;

faulting is dated as post 190 m.y. and pre 90-112 m.y. (Eteza fault); major movement on the Egosa and related faults is pre 124-127 m.y. with minor movements post-dating this. It appears (as suspected by Maud, 1961; Hardie, 1962) legitimate to interpret the fault pattern in terms of Gondwanaland break-up.

The magnetic signature of the Agulhas sheared margin has been traced to 30° 48'S (Duplessis & Simpson, 1974)(Fig.7.1) and by comparison with mapped magnetic anomalies, the sheared margin ends around 29° 50'S, 31° 25'E. On the Tugela Cone, tension ensued from an end-of-fault structure (c.f. Fig.7.4d) as well as from rifting tectonics (c.f. Goodlad et al., 1982), probably resulting in splintering of crust.

Although thousands of kilometres of lateral movement has occurred along the Agulhas transform fault, faulting in Natal is subsidiary to this and offsets may be slight. Similar structural styles are experienced in both large and small offsets (Harding, 1973; 1974). Transform faults follow small circles about rotation poles. Near Port Shepstone the Agulhas Transform Fault is oriented N 35°E, and at its termination trends N 25°E. Synthetic fractures should be oriented 70°-90° from this. In Natal we expect north-east oriented synthetic faults and east-west or west-north-west oriented antithetic faults (Figs.7.1 & 7.3). Note that en échelon synthetic faults may be inter-connected by faults at a low angle to the wrench-strike oriented sub-parallel to the preferred direction of thrust faults (Wilcox et al., 1973, Figs.7-10). In Natal these would be north-oriented. The dominant fault trends in Natal are compatible with this geometry.

Many of the features exhibited in coastal Natal are typical of strike-slip regimes.

1) Several faults display an en échelon configuration. This is well shown along the coastal sections between Port Shepstone and Durban and just north of Umhloti.

2) The large arcuate faults described in section 7.2 exhibit an inverse 's' shape. This is consistent with dextral or right lateral strike-slip.

3) Several faults show a reversal of vertical separation along strike (eg. 30 km south of Highflats, 40 km east of Highflats near the coast, 20 km west of Durban, west of Umhloti, south-east and east of Greytown, and 20 km south and south-west of the Ngoye Horst). This phenomenon has been documented by Beater & Maud (1960) and Maud (1961). As folding is of minor importance, reversal of throw is more likely to be caused by anastomosing faults rather than by offset folds (Figs.7.4a,b,g,h).

4) Faults are curved, dog-legged, or braided and form interconnecting grid patterns. Anastomosing faults may enclose downsagged or upthrust slices or wedges. The Bongwan Gas fault system includes a downthrown wedge where throw varies from the centre of the sag (300 m) to the tip (130 m). A component of strike-slip has been recognised on this fault (Gevers, 1941).

5) Where faults terminate, a series of subsidiary faults splay outwards. This is the case south-east of Greytown and at Highflats.

6) Large unbroken blocks or pods may be caught between curving faults. In the San Andreas fault zone of California, a large pod lies between the San Andreas and San Gabriel faults. This pod has moved south-easterly along the San Gabriel fault (towards the bottom left of Figure 7.5b - note the orientation of north). This has led to convergence causing uplift, reverse faults or high angle thrusts, and tilted blocks (Figs. 7.5a,b). Where the San Gabriel and San Andreas faults diverge, tension has resulted in the formation of the Ridge Basin. Note that at diverging and converging structures, the directions of offset are the same, and convergence or divergence is controlled by the relative movement of the blocks.

A section of Natal is compared in Figure 7.5c. Slight south-westerly movement of a pod has possibly elevated Natal Group rocks south-east of Greytown, and caused tilted block step faulting. Raised Natal Group rocks near Highflats may have a similar origin. Equally, divergence may have resulted in tension causing the Nkwaleni Graben and the Ngoye Horst. This latter feature is not an upthrust block, but is interpreted as a remnant high isolated by down-dropping blocks on either side (Beater & Maud, 1960). Tension appears to have been oriented north-south, and this is compatible with Figure 7.3.

7) Side-stepping faults leading to transtension cause pull-apart basins. The Egosa and related faults to the south outline half graben into which Upper Valanginian conglomerates were dumped directly into the sea. These faults appear to offset the continental margin (compare Figs.7.1 & 7.4c). The correspondence of micro-palaeontological dating (Upper Valanginian - 124-127 m.y. - McLachlan et al., 1976) and magnetic dating suggests that a graben opened as relative motion between Falkland Plateau and South Africa began ∿125 m.y.(chapter 4). Only the western half of this graben is preserved on the Transkei coast, with the eastern half having been rafted away on the Falkland Plateau. Again, tension was most likely oriented north-south (compare Fig.7.3).

8) Sedimentation in strike-slip basins often shows evidence for a nearby active source area indicated by restricted deposition of locally derived conglomerates and breccias (Reading, 1980). This is due to contemporaneous transpressional and transtensional regimes forming uplifted blocks along strike from down-dropped basins. In continental areas, lacustrine facies are typical in strike-slip basins. At Mgazana and Mboyti (Egosa fault) conglomerates include pebbles from the adjacent dolerites and Ecca shale, and amygdaloidal basalt possibly derived from Stormberg volcanic outcrops (McLachlan et al., 1976). At Eteza, the basal Cretaceous sediments are conglomeratic lacustrine rocks containing pebbles of

Lebombo basalt and rhyolite, dolerite, granite and Natal Group quartzites (Frankel, 1960). Such facies are not in themselves diagnostic of strike-slip regimes but only of fault-bounded basins.

Along the Agulhas Transform fault oceanic crust was emplaced against continental crust and the spreading centre migrated south-westward along the sheared margin. Oceanic crust emplaced at around 2500 m depth gradually subsides as it cools. South-east of Durban oceanic basement is now at a depth of 5500 - 7000 m (Chetty & Green, 1977). This subsidence causes the oceanic crust to decouple from the continental crust leading to faulting (Scrutton, 1979), and is the primary cause of steepness along sheared margins. The faulted nature of the Natal coastline (Kent, 1938; Maud, 1961) is thus borne out.

A discussion of the Lebombo Group volcanic rocks in a strike-slip context is beyond the scope of this chapter. However an extensive transtensile regime may have developed allowing widespread volcanism. In an extensional transform fault an elongate trough may develop. Oceanic crust emplaced against the transform fault gradually cools and subsides and the uplifted margin or rim of the trough may suffer collapse as a result. It then resembles a faulted monocline (Veevers & Powell, 1979). A strike-slip zone in Spitzbergen was also originally described as a faulted monocline (Lowell, 1972). Troughs along transform faults occur in the Gulf of Aquaba and the Dead Sea and in transtensional regimes (leaky transforms) volcanism occurs (Freund, 1965; Ben-Avraham et al., 1979). Extensional regimes may later become dominantly strike-slip as has happened in New Zealand and in the Rhine Graben (Norris et al., 1978; Sengor et al., 1978).

Some features are possibly consistent with dextral strike-slip movement along the Lebombo.

1) In the north of Figure 7.1 some east-north-east oriented faults appear to offset Lebombo volcanic rocks left laterally (Du Toit, 1929). Movement in a left lateral sense has been limited and has been rather of a dip-

slip nature on normal faults. This is similar to the situation arising when antithetic faults assume oblique-slip motion in a tensional regime. North-north-west faults appear to cause right lateral offset. These intersecting faults may be conjugate fractures of antithetic and synthetic nature respectively. The fault orientations resemble conjugate fault sets to the west of the dextral Davie Transform fault (Scrutton et al., 1981, Fig.2).

2) Conjugate fractures occur in Swaziland around 26° 30'S, 30° 50'E.

3) Many of the Rooi Rand dykes are oriented south-south-west or south-west (perpendicular to the proposed extension vector) and form en échelon (Du Toit, 1929; Cleverly, 1979, Fig.2). These dykes indicate gross extension of the crust which has been estimated to average 40% (Saggerson et al., 1983). However, their sudden termination around 28° 15'S is difficult to interpret in terms of a straight-forward tensional regime (J.W. Bristow, personal communication, 1980). This may be explicable in terms of closely contiguous transtensional and transpressive regimes in a strike-slip zone.

7.6 SUMMARY

A consideration of sea floor magnetic evidence dictates that transform faults controlled much of the Gondwanaland break-up along the south-eastern margin of Africa. The well documented Agulhas sheared margin runs from the south-western tip of the Agulhas Bank to the southern face of the Tugela Cone. A transform fault is postulated to lie along the east side of the Lebombo volcanic outcrops (chapter 6). A third transform fault, the Davie Fracture Zone lies along the east coast of northern Mozambique, Tanzania and southern Kenya (Segoufin & Patriat, 1980). These areas have traditionally been interpreted as regions of tension rather than of strike-slip (Du Toit, 1929; Maud, 1961; Kent, 1974a). Correspondence of the dating of the faults of Coastal Natal with magnetic dating of continental rifting suggests that the faults are associated with Gondwanaland break-up. The nature and orientation of faulting in Natal seems to be consistent with a dextral strike-slip model. The following features are diagnostic: en échelon fault configuration; reverse 's' plan of fault traces; reversal of fault throw along strike; faults which are curved, 'dog-legged', intersecting, splayed or anastomosing; pull-apart graben resulting from side-stepping faults; upraised tilted blocks possibly caused by convergence of curved faults; down-dropped basins due to diverging faults; conglomeratic lacustrine facies in fault bounded basins. The fault-controlled nature of the Natal coast south of Durban is substantiated.

Previously strike-slip motion has been recognised on the Bongwan Gas fault (Gevers, 1941), and an east-west oriented rotational couple was suggested by Kent (1938). A couple motion was considered and rejected by Maud (1961) and was recognised as prevailing on a plate tectonic scale by Scrutton et al. (1975). By noting reversal of throw on micro-faults, McCarthy (1979) reasoned that a rotational couple could explain Natal faulting.

The strike-slip interpretation is at variance with previous theories which consider the major controlling feature of coastal Natal to be a monocline (King, 1972) or faulting due to north-west/south-east oriented tension (Beater & Maud, 1960; Maud, 1961; Hardie, 1962). If strike-slip tectonism has influenced Natal geology and geomorphology, then existing geomorphological theories for the province (King, 1972) may need revision in this light.

A major problem in the strike-slip interpretation is the total lack of reported thrust faults of Jurassic-Cretaceous age in Natal (P.E. Matthews, personal communication, 1983). However three points must be raised in this regard. a) Offshore seismic profiling (chapter 8) suggests that thrust faults occurring at 28° 45'S, 33° 25'E have offset a mid-Cretaceous seismic reflector. b) As strike-slip motion proceeded along the Agulhas transform fault, a broad zone of faulting affected the Agulhas Bank (Du Toit, 1979). If such effects extended to the present southern Cape, some Jurassic-Cretaceous re-activation within the Permo-Triassic Cape fold belt may have ensued.

A large transpressive zone may be the cause of north-east oriented structures in the Cape Hangklip area near Cape Town. This scenario provides a possible mechanism for long-debated features (eg. Newton, 1980). Again however, in a well-mapped area, no thrust faults had been reported. During recent investigations (C.J.H. Hartnady, personal communication, 1982) a large thrust fault has been discovered in the Cape Peninsula. c) The situation in southern Africa, where hard evidence for strike-slip or transpressive motion is not immediately obvious, appears to be typical. Reading (1982) states that "in older tectonic regimes, where evidence for lateral motion is so difficult to obtain, vertical movements may be the only proveable fault motion". Given the model presented here, perhaps strike-slip-related faults will be recognised in Coastal Natal.

PART 3

THE NORTHERNMOST NATAL VALLEY SEDIMENTARY BASIN

Previous work done in the area is covered in chapter 1, regional setting is sketched in chapter 2, while chapters 3-7 describe the plate tectonic situation of the Natal Valley. In part 3 the basin configuration and in-fill of the Natal Valley is discussed. Originally, a preliminary account was given for the area north of 31°S (Dingle et al., 1978). Here, the main thrust is aimed at the northernmost area comprising the Central Terrace, Limpopo Cone, Almirante Leite Bank, Inharrime Terrace and continental shelf and slope. General relationships with the southern area are outlined. The description is primarily based on seismic profiler data augmented by stratigraphic data provided by core, grab and dredge samples raised from the seafloor (Figs.8.1 & 8.2). Over 4770 km of continuous seismic profiles are presented and considered in detail here. Attention has also been given to a considerable quantity of data peripheral to the main study area in order to complete geological maps, and elucidate the acoustic stratigraphy (Fig.8.3). The seismic profiling method is described in appendix 1.

CHAPTER 8

BASIN CONFIGURATION AND SEDIMENT IN-FILL

8.1 INTRODUCTION

The characteristics of the area are conveniently described in relation to five regional seismic reflecting horizons which divide the sedimentary succession into five units (section 8.2, Fig.8.4). The configuration of each reflector and sedimentary unit are described in descending geologic age (section 8.3). Depositional rates and styles, relationships to the plate tectonic model and implications for hydrocarbon potential are outlined in sections 8.4 - 8.6.

8.2 ACOUSTIC STRATIGRAPHY

The stratigraphic significance of regionally correlated acoustic reflectors is established through comparison with the succession on the coastal plain (Kennedy & Klinger, 1971; Frankel, 1972; Flores, 1973; Forster, 1975; Siesser & Miles, 1979) in SOEKOR drill-hole Jc-1 (Du Toit & Leith, 1974; McLachlan & McMillan, 1979) in DSDP hole 249 (Simpson, Schlich et al., 1974) and as determined from seafloor samples (Martin et al., 1982a). Seismic refraction stations in the area also provide correlation points (Ludwig et al., 1968; Chetty & Green, 1977; Lort et al., 1979).

8.2.1 Acoustic basement

The basalts and rhyolites of the Lebombo Mountains form the western edge of the sedimentary basin of the Natal Valley and the Mozambique and Zululand coastal plain. These volcanic rocks have returned isotopic ages between 190 and 137 m.y. (Flores, 1973; Cleverly & Bristow, 1979). These are overlain by continental conglomerates and the Sena sandstone (Flores, 1973). In Mozambique the overlying shallow-marine glauconitic sandstones contain a Barremian fauna (Forster, 1975). DSDP hole 249 on the Mozambique Ridge bottomed in basalt of indeterminate age overlain by Neocomian black claystones and volcanic siltstones (Simpson, Schlich et al., 1974). Originally, a Valanginian age was favoured for the basal sediments on the basis of the presence of the ostracod *Majungaella nematis* but more recently the range of this species has been ex-

FIGURE CAPTIONS

Fig.8.1. Track-chart of seismic profiles. Solid line = satellite navigation, dashed line= Decca chain and dead reckoning.

Fig.8.2. Sample positions. D = dredge hauls; open square = gravity cores (successful); filled square = gravity cores (unsuccessful); open circle = piston cores; closed circle = grabs. Unsuccessful gravity cores which appear to have hit consolidated older material are used to delineate the limits of past-'L' (most recent) sediments (Fig.8.46).

Fig.8.3. Location of drill-sites, sonobuoys and refraction sites used in erecting an acoutic stratigraphy: Jc1 borehole (Du Toit & Leith, 1974); DSDP site 249 (Simpson, Schlich et al., 1974); stars = sonobuoys (Lort et al., 1979); dashed lines = refraction sites (Ludwig et al., 1968; Chetty & Green, 1977).

Fig.8.4. Stratigraphic columns in the northern Natal Valley region, showing the correlation of reflectors McDuff and Angus with the JC-1 borehole and DSDP borehole 249. Note a horizon in the Tugela Cone area correlates with the Early Palaeocene break in the JC-1 borehole. Horizon Jimmy and a reflector in the Angus-Jimmy sequence are dated micropalaeontologically (Fig.8.5) while 'L' has not been precisely dated yet. Pre-McDuff = sedimentary unit 1; McDuff-Angus = unit 2; Angus-Jimmy = unit 3; Jimmy - 'L' = unit 4 and post-'L' = unit 5.

Fig.8.5. Some age diagnostic foraminifera from key samples. Sample positions schematically shown in relation to seismic reflectors. Species ranges given as longest range determined from literature (after Martin et al., 1982a; Salmon (in press)).

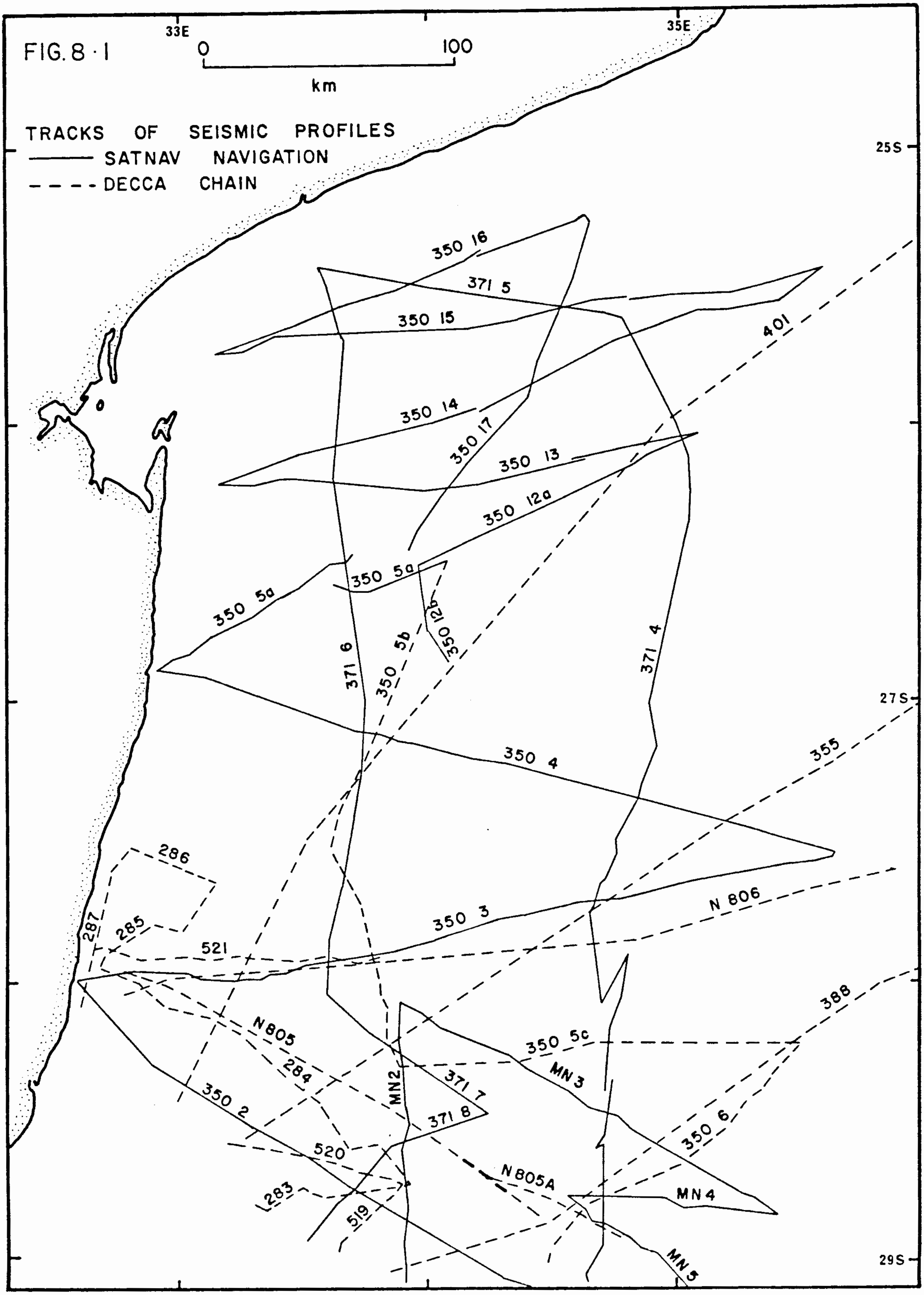


FIG. 8 · 1
33E
35E
0
100
km
TRACKS OF SEISMIC PROFILES
SATNAV NAVIGATION
DECCA CHAIN
25S
27S
29S
350 16
371 5
350 15
401
350 14
350 17
350 13
350 12a
350 5a
350 5a
350 12b
350 5b
371 6
371 4
350 4
355
286
287
285
521
350 3
N 806
388
N805
350 5c
284
MN2
371 7
MN3
350 2
371 8
350 6
520
N805A
283
519
MN4
MN5

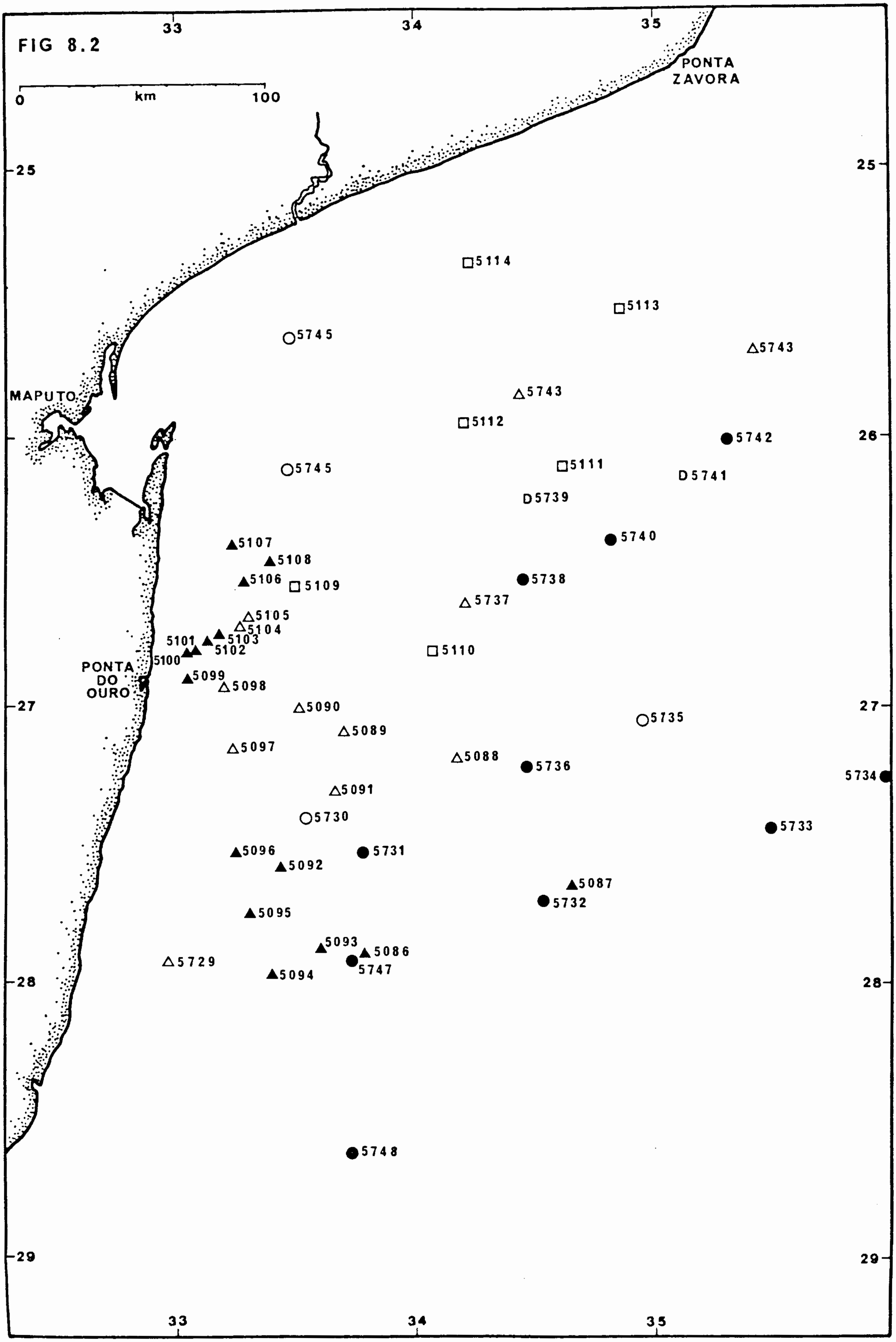


FIG 8.2
0 km 100
33
34
35
25
26
27
28
29
PONTA ZAVORA
MAPUTO
PONTA DO OURO
5114
5113
5745
5743
5743
5112
5742
5745
5111
5741
5739
5107
5740
5108
5106
5109
5738
5737
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5086
5729
5747
5094
5748

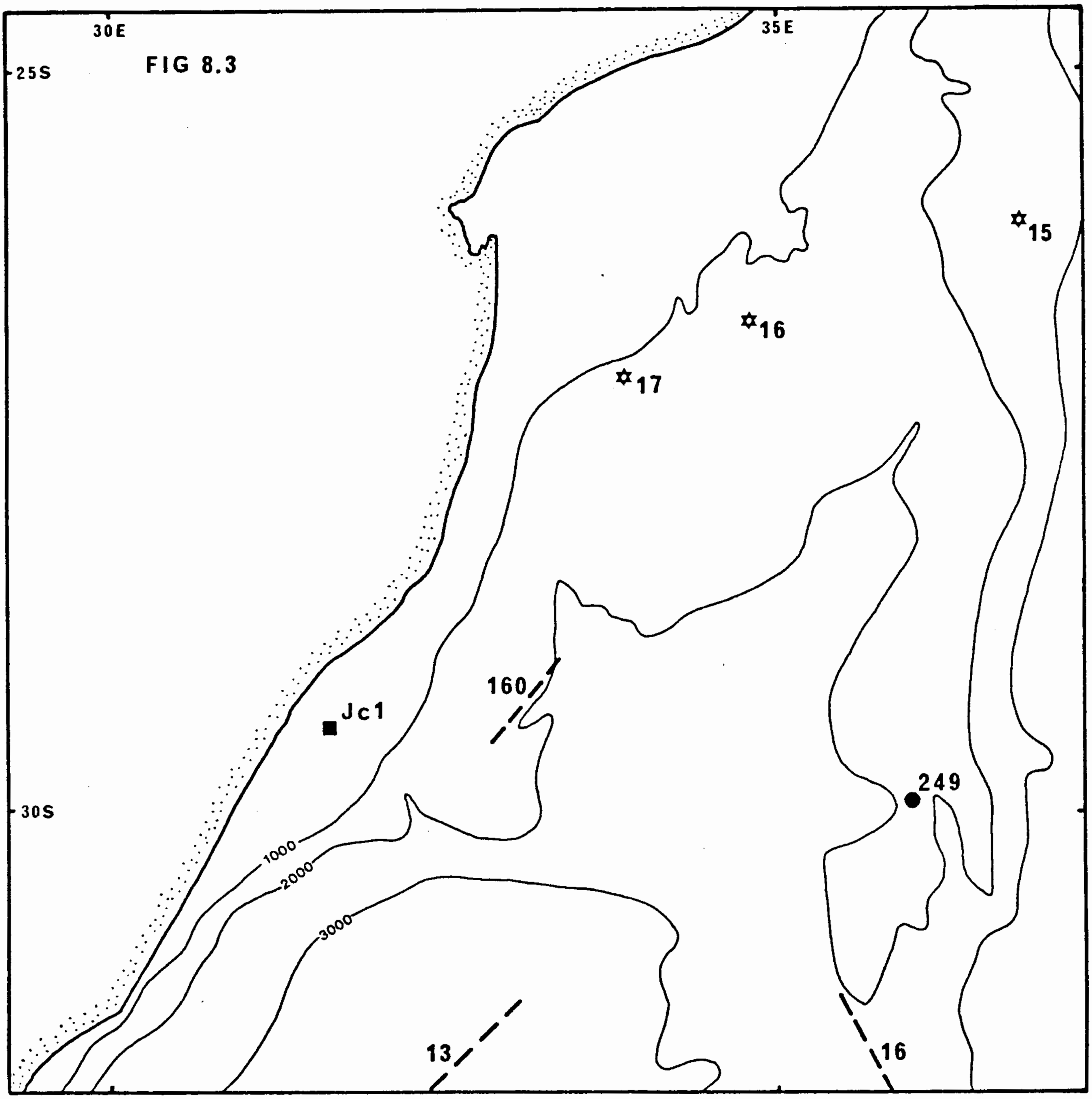
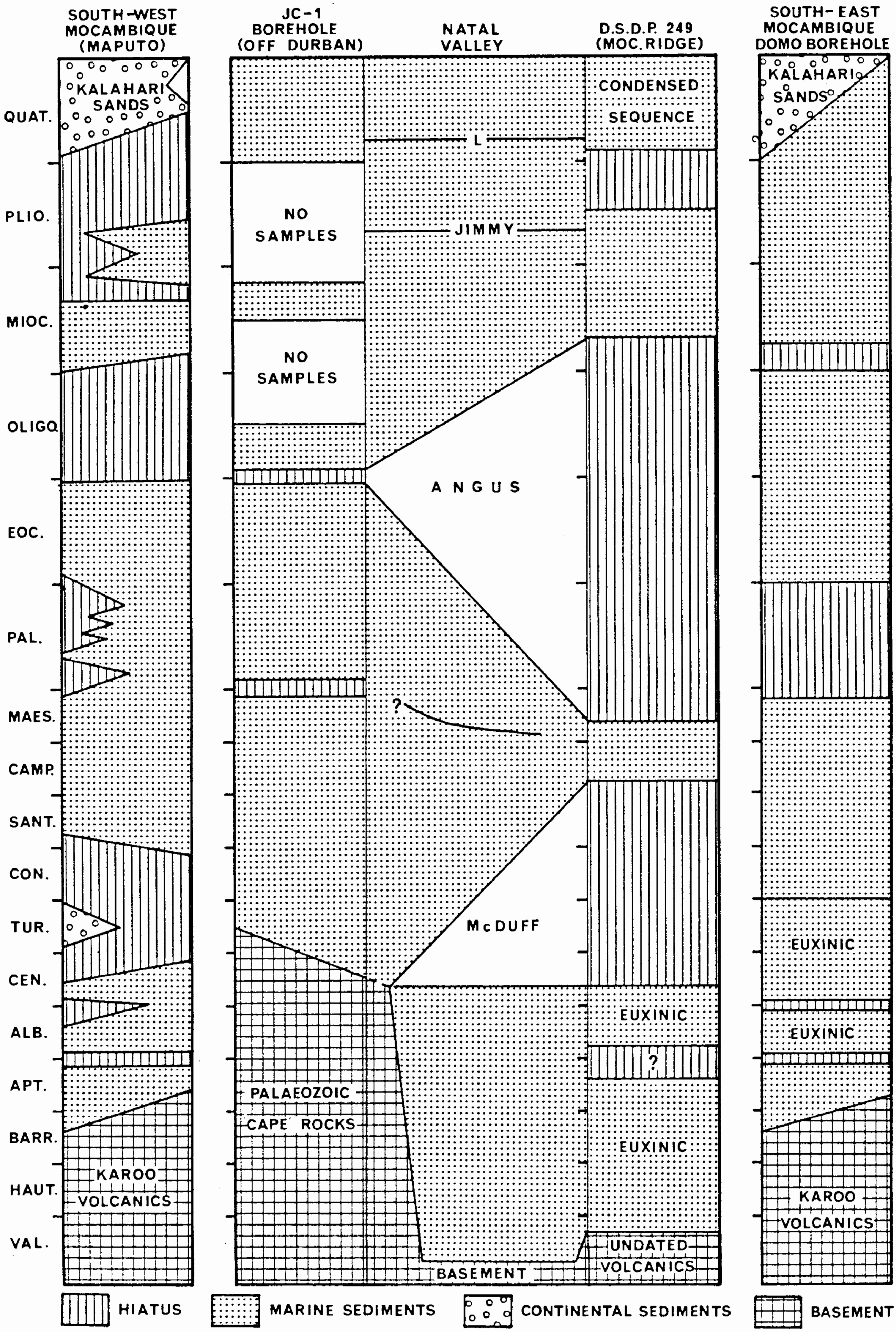

30E
35E
25S
FIG 8.3
30S
15
16
17
Jc1
160
249
1000
2000
3000
13
16

FIG 8.4


SOUTH-WEST MOCAMBIQUE (MAPUTO)
JC-1 BOREHOLE (OFF DURBAN)
NATAL VALLEY
D.S.D.P. 249 (MOC. RIDGE)
SOUTH-EAST MOCAMBIQUE DOMO BOREHOLE
QUAT.
PLIO.
MIOC.
OLIGO.
EOC.
PAL.
MAES.
CAMP.
SANT.
CON.
TUR.
CEN.
ALB.
APT.
BARR.
HAUT.
VAL.
KALAHARI SANDS
KAROO VOLCANICS
NO SAMPLES
NO SAMPLES
PALAEOZOIC CAPE ROCKS
L
JIMMY
ANGUS
?
McDUFF
BASEMENT
CONDENSED SEQUENCE
EUXINIC
?
EUXINIC
UNDATED VOLCANICS
KALAHARI SANDS
EUXINIC
EUXINIC
KAROO VOLCANICS
HIATUS
MARINE SEDIMENTS
CONTINENTAL SEDIMENTS
BASEMENT

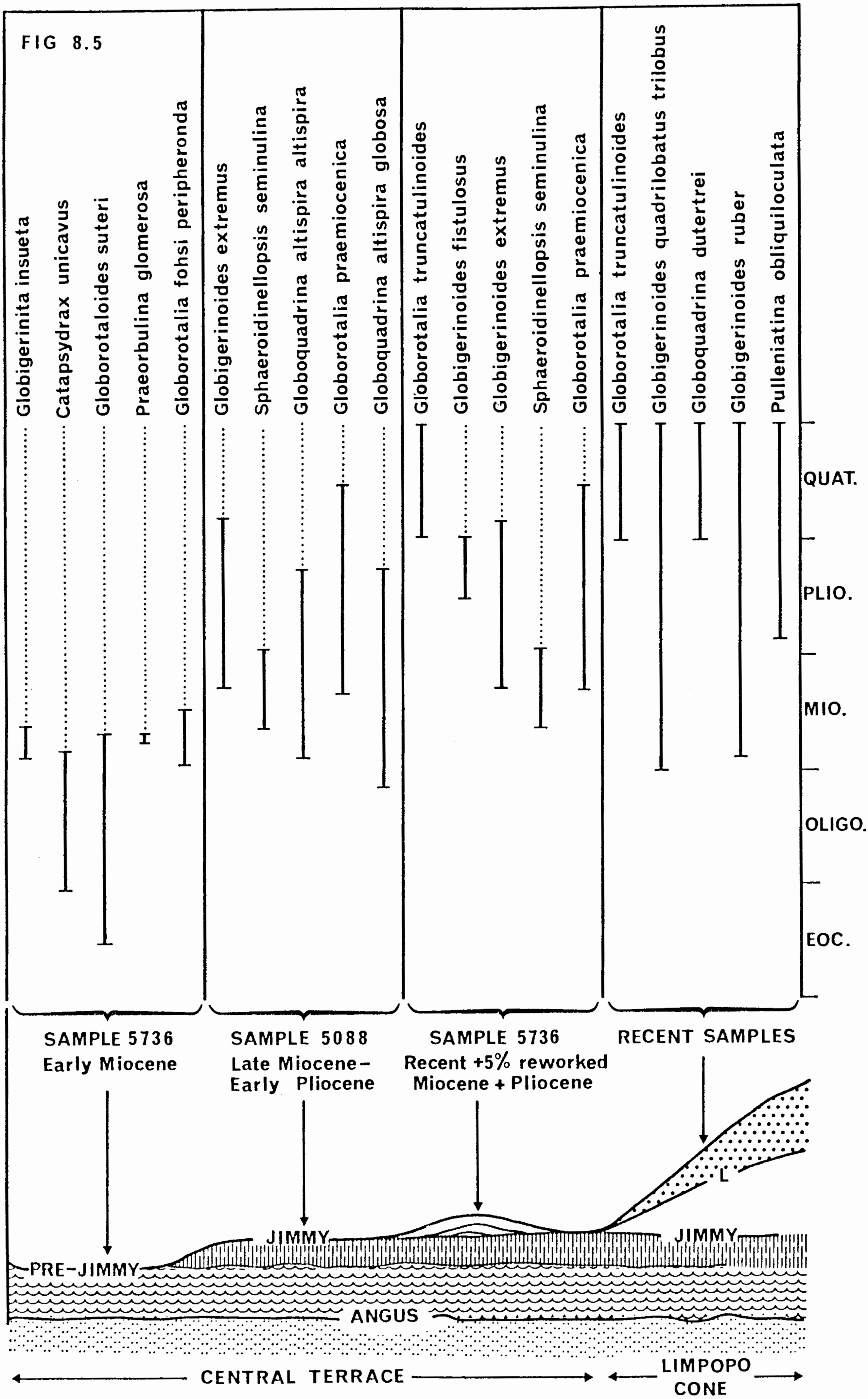


FIG 8.5
Globigerinita insueta
Catapsydrax unicavus
Globorotaloides suteri
Praeorbulina glomerosa
Globorotalia fohsi peripheronda
Globigerinoides extremus
Sphaeroidinellopsis seminulina
Globoquadrina altispira altispira
Globorotalia praemiocenica
Globoquadrina altispira globosa
Globorotalia truncatulinoides
Globigerinoides fistulosus
Globigerinoides extremus
Sphaeroidinellopsis seminulina
Globorotalia praemiocenica
Globorotalia truncatulinoides
Globigerinoides quadrilobatus trilobus
Globoquadrina dutertrei
Globigerinoides ruber
Pulleniatina obliquiloculata
QUAT.
PLIO.
MIO.
OLIGO.
EOC.
SAMPLE 5736
Early Miocene
SAMPLE 5088
Late Miocene – Early Pliocene
SAMPLE 5736
Recent +5% reworked Miocene + Pliocene
RECENT SAMPLES
L
JIMMY
JIMMY
PRE-JIMMY
ANGUS
CENTRAL TERRACE
LIMPOPO CONE

tended (Portlandian-Cenomanian - R.V. Dingle, personal communication, 1983). Basal sediments are either Valanginian or Hauterivian. The Jc-1 borehole bottomed in Palaeozoic Cape Supergroup and Karoo rocks overlain by Cenomanian/Turonian claystones. Seismic evidence shows an extensive older sequence down-dip. Thus volcanic basement is at least pre-Barremian and may be as old as Earliest Cretaceous or Late Jurassic. In the upper reaches of the Tugela Cone, even older rocks of the Cape and Karoo Systems form basement.

8.2.2 McDuff

Reflector McDuff tops a very reflective sedimentary unit (stratigraphic unit 1 - Fig.8.4), which lies on acoustic basement. The type locality for this reflector is profile 520 (Fig.8.22). McDuff is tentatively correlated with the major pre-Cenomanian/Turonian reflector which abuts basement at 2.7 sec two-way time, 7 km east of Jc-1 on the profile shown by Du Toit & Leith (1974). Although McDuff cannot be directly equated with the sedimentary pockets on the Mozambique Ridge, a reflector of similar character marks the hiatus between Cenomanian ash-rich, euxinic silts and Campanian, open-marine, clay-rich nanno-chalk in DSDP site 249 (Simpson, Schlich et al., 1974). An equivalent hiatus occurs on the Zululand and South Mozambique coastal plains, although no such break occurs in south-eastern Mozambique (Kennedy & Klinger, 1971; Flores, 1973; Forster, 1975; Dingle, 1978). The change from euxinic to open-marine sedimentation on the Mozambique Ridge has been linked to a similar change in south-eastern Mozambique (Girdley et al., 1974). McDuff lies close to the interface between sedimentary units having seismic velocities of 4 and 2 km/sec identified by Ludwig et al. (1968)at refraction site 160. At DSDP site 249, compressional wave velocities at 0.5 kbar pressure of 4.43 and 2.15 km/sec were measured on recrystallised limestone and glauconitic mudstone respectively within the Cenomanian-Neocomian succession (Christensen et al., 1974). Sonobuoy stations 16 and 17 of Lort et al. (1979) lie very close to seismic profiles TBD 371 4 and TBD 350 4 respectively (Figs.8.1&8.3). Near station 16, where only one velocity of 4 km/sec was measured, McDuff outcrops on an up-faulted block. Near station 17, McDuff lies near the base of a unit of seismic velocity 3.08 km/sec and near to the unit having velocity of 4.4 km/sec. Thus at all four localities where velocity data are available, pre-McDuff sediments have velocities greater than 3.0 km/sec except for the low-velocity mudstone reported by Christensen et al. (1974). This is in agreement with the velocity data for Early Cretaceous rocks exposed on-shore (Ludwig et al., 1968), supporting the dating of McDuff as a mid-Cretaceous horizon. In its position relative to basement and its age, McDuff is analogous to horizon AII in the South Atlantic (Emery et al., 1975; Bolli, Ryan et al., 1978; Austin & Uchupi, 1982).

8.2.3 Angus

Reflector Angus lies at the top of stratigraphic unit 2 (Fig.8.4) and marks an unconformity separating well stratified reflective sediment below from a less reflective unit above. Again the type locality is profile 520. Angus correlates well with a reflector marking the basal Oligocene hiatus in the Jc-1 borehole separating sub-aqueous deltaic sands from shallow marine clays. This has been equated with the hiatus in DSDP site 249 between Maastrichtian, clay-rich nanno-chalk and Lower Miocene, foram-rich nanno-ooze (Du Toit & Leith, 1974; Dingle et al., 1978). An Oligocene unconformity occurs on the Zululand and South Mozambique coastal plain. In the Mozambique Channel and eastern and south-eastern Mozambique, the Oligocene is complete (Flores, 1973; Simpson, Schlich et al., 1974, site 242; Forster, 1975), although in the latter area, a Lower Miocene break is evident. Micropalaeontological studies of seafloor samples allow direct dating of reflectors higher in the sedimentary column (Fig.8.5). The oldest material recovered (Early Miocene) is stratigraphically younger than Angus, thus supporting its dating as Oligocene (Figs.8.4 & 8.5).

In the original analysis of Jc-1 borehole (Du Toit & Leith, 1974) no hiatus was recorded at the Cretaceous/Tertiary boundary. Later work detected such a break which correlates with a corresponding reflector. A prominent reflector does occur in the McDuff to Angus sequence, and has been correlated to the Cretaceous/Tertiary boundary hiatus locally in the Tugela Cone area (Martin et al., 1982a). A longer but comparable hiatus occurs on the Zululand coastal plain (Fig.8.4).

8.2.4 Jimmy

Stratigraphic relationships within the Neogene sediment column are schematically depicted in Figure 8.5. A reflector within the Angus - Jimmy sequence lies on yellowish gray Early Miocene foraminiferal muddy sand. Reflector Jimmy lies on a similar lithology dated as Late Miocene/Early Pliocene. Sample 5088, placed on seismic profile 350 4 (Fig.8.13) may be considered the type locality, although profiles 371 4 and 371 6 helped clarify relationships with both the Early Miocene reflector and Angus. On the Zululand coastal plain both a short Late Miocene/Early Pliocene and a Pliocene hiatus occurred (Siesser & Dingle, 1981). By contrast, in south-eastern Mozambique, the Late Miocene-Recent sequence appears complete (Forster, 1975). In DSDP site 249, much of the Pliocene is missing while the overlying Quaternary is a condensed sequence (Simpson & Schlich et al., 1974).

8.2.5 Horizon 'L'

Reflector 'L' lies at the top of a sequence of reflectors on the central Limpopo Cone and marks a major depocentre shift. Profiles 350 15 and 371 6 exhibit this most clearly. Altough it outcrops, samples of this horizon have not been collected, thus precluding accurate dating. Up to 0.73 sec of sediment lie between Jimmy and 'L' while in other areas post 'L' sediments are up to 0.44 sec thick. In Figure 8.4, horizon 'L' has been arbitrarily placed within the post-Jimmy sequence.

8.3 STRUCTURE AND SEDIMENT DISTRIBUTION

Tracings of eighteen generally east-west-trending profiles are shown as Figures 8.6 - 8.23, while nine profiles trending generally north-south are shown as Figures 8.24 - 8.32. Figure 8.33 shows part of a multi-channel seismic profile provided by SOEKOR. Five structure contour charts have been compiled for the major reflecting horizons, while seven isopachyte charts help to elucidate basin in-fill (Figs.8.34-8.45). All five major horizons - basement, McDuff, Angus, Jimmy and 'L' outcrop within the area as shown in the geological map (Fig.8.46).

8.3.1 Basement configuration

The available single-channel seismic profiles detected acoustic basement only where basement highs are exposed or are thinly covered by sediment. Basement was detected on profiles 350 13, 350 12, 371 8, 519, MN2 and 371 4 (Figs. 8.10, 8.11, 8.29, 8.30, 8.31 and 8.32). Multi-channel seismic profiles (eg. Fig.8.33) provide the most useful record of basement on a regional scale. Because of the sparsity of data on basement, I have a) included data from profile 350 2; b) included spot values taken from profile MD2-21 of Lort et al. (1979) and Figure 4 of Beck & Lehner (1974); c) contoured the basement structure contour map and the isopachyte map of total sediment at intervals of 250 m sec (Figs.8.34 & 8.45).

Basement topography is very uneven and is characterised by relief of up to 1 sec. Topography is most pronounced in the region of two major basement highs - the Almirante Leite Bank and the Naudé Ridge where relief exceeds 1.8 and 1.5 sec respectively (Figs.2.13, 2.14, 8.34). Small scale relief is also indicated by the hyperbolic echoes seen on the multi-channel profiles (Fig.8.33 and appendix 1). This configuration is considered typical of oceanic basement (Roots et al., 1979; Austin & Uchupi, 1982).

The Almirante Leite Bank is a complex basement high centred at 26° 10'S 34° 50'E. Relationships on profiles 350 13 (1000 hrs) and 371 4 (0730 hrs) suggest that it is an upfaulted block (Figs.8.10 & 8.32). The faults throw 700-800 m, while fault planes dip at 27-32°. Basement outcrops over a wide

FIGURE CAPTIONS

Figs.8.6-8.32. Tracings of continuous seismic profiles obtained using air-guns and single-channel arrays (see appendix 1). Tracks are shown in Fig.8.1. Figs.8.6-8.23 are east-west-oriented profiles (west to left) arranged from the most northerly (8.6) to the most southerly (8.23). Figs. 8.24-8.32 = north-south-oriented profiles (north to right) arranged from the most westerly (8.24) to the most easterly (8.32). Fig.8.32 (profile 371 4) is shown in two parts because the profile is very long. Vertical scales in seconds two-way reflection time. Horizontal scales in km. Vertical exaggeration is given in each case. Numbers at top refer to time of the day during collection of continuous profiles. Physiographic sub-provinces (c.f. section 2.3) are given . Regional acoustic reflectors are marked: B = basement; M^{C} = McDuff; A = Angus; J = Jimmy, L = reflector L. Note that tracks 350 2, 350 6, MN4 and MN5 shown in Fig.8.1 were used to complete Fig.8.46 (Geology) but are not shown here as they are being considered in detail by Mr S.W. Goodlad in a neighbouring study. Profiles 401, 355 and 388 were only available as photographic copies and again were used only to help draw Fig.8.46 (see appendix 1).

Fig.8.33. Tracing of part of multi-channel profile N805 (shown in heavier dashed line in Figure 8.1) at the scale at which it was made available (see appendix 1). Only basement and reflector McDuff could reliably be traced and correlated on these profiles.

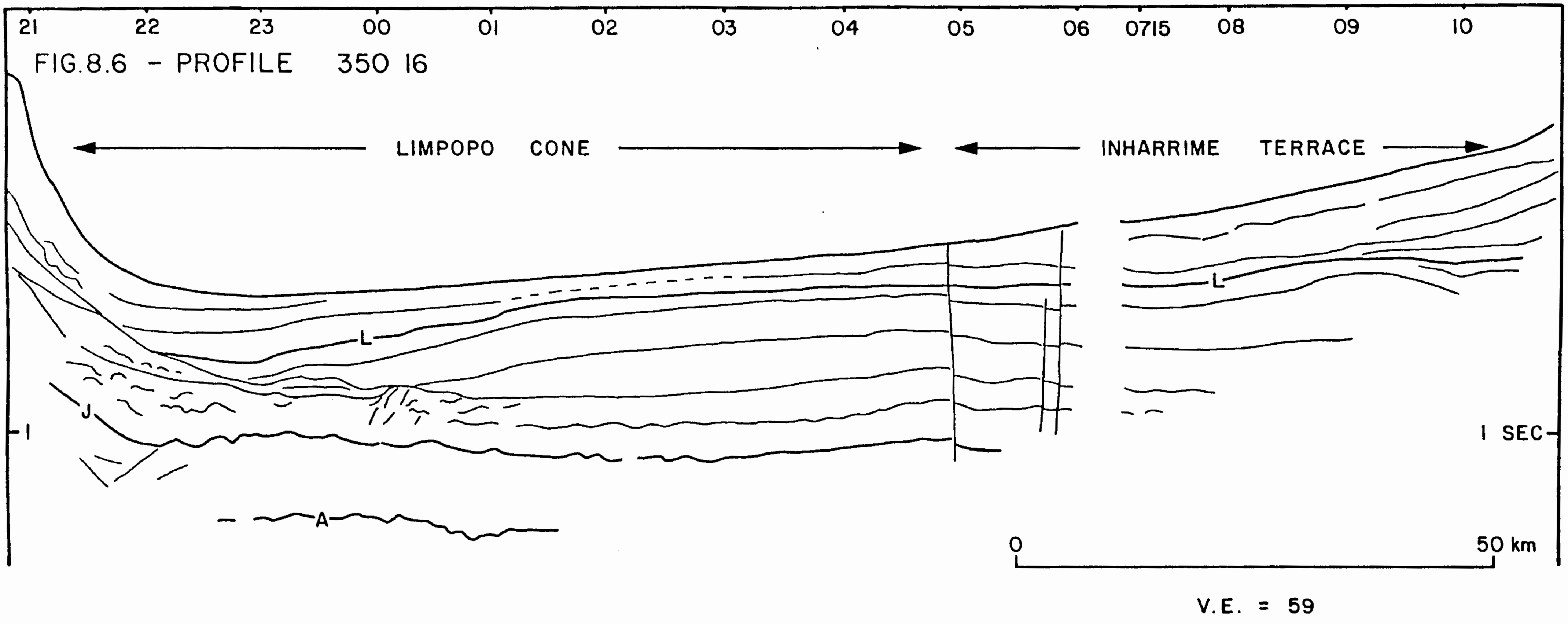


FIG.8.6 - PROFILE 350 16

FIG. 8·7 PROFILE 371 5

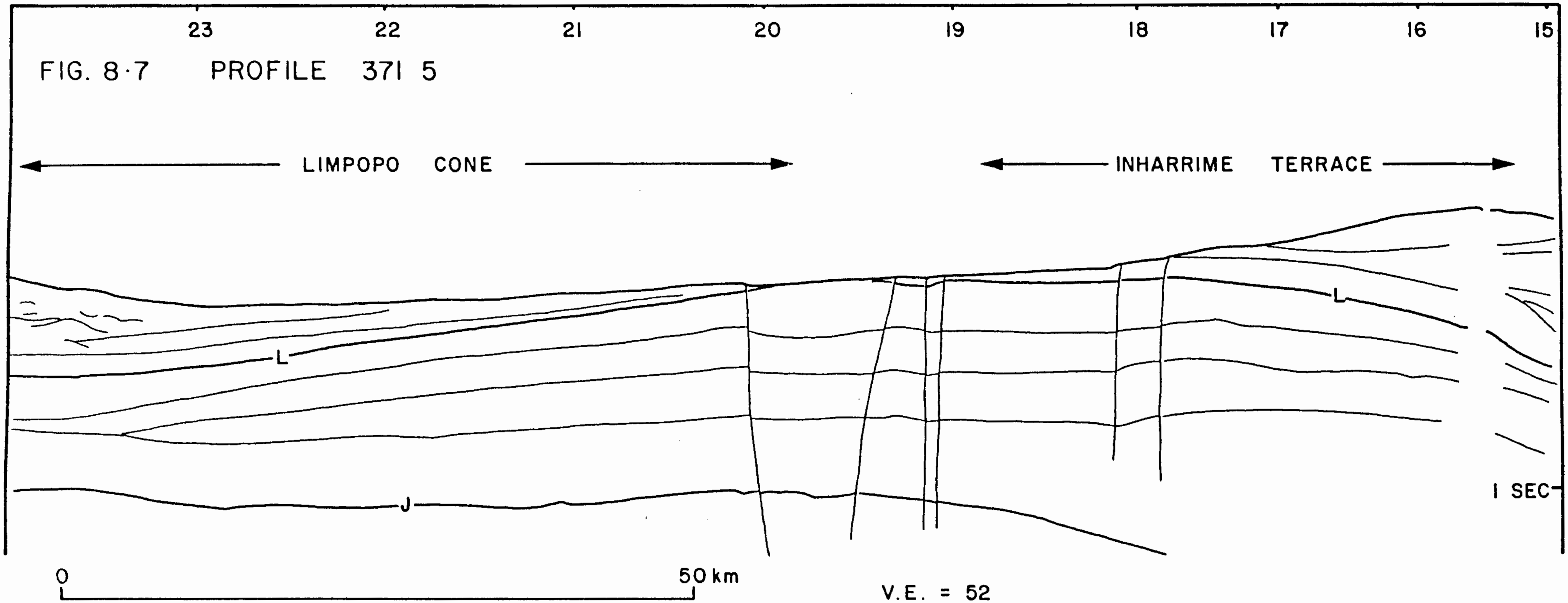
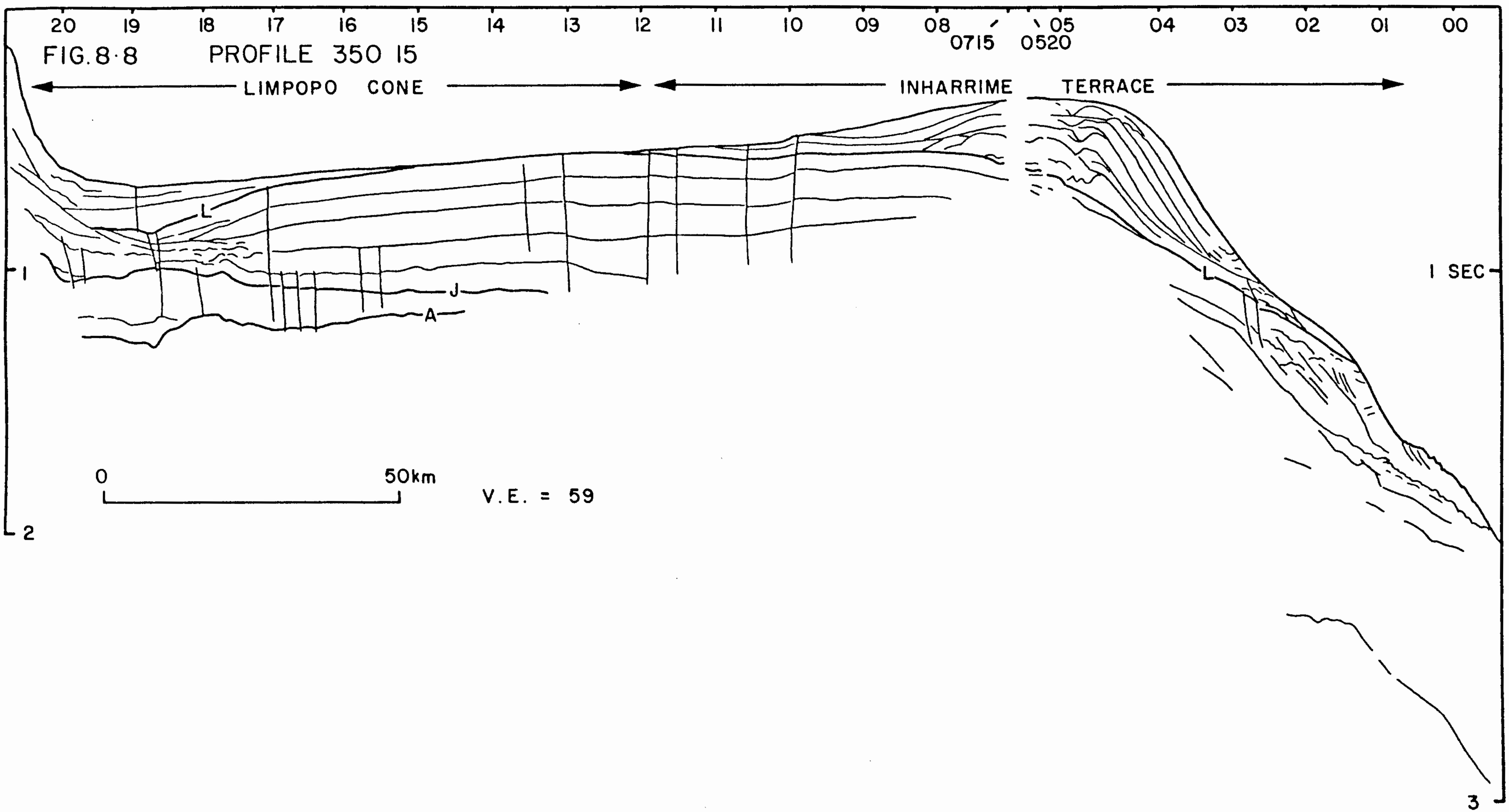


FIG. 8·8


PROFILE 350 15
20
19
18
17
16
15
14
13
12
11
10
09
08
0715
05
0520
04
03
02
01
00
LIMPOPO CONE
INHARRIME TERRACE
L
J
A
1
1 SEC
2
3
0
50km
V.E. = 59

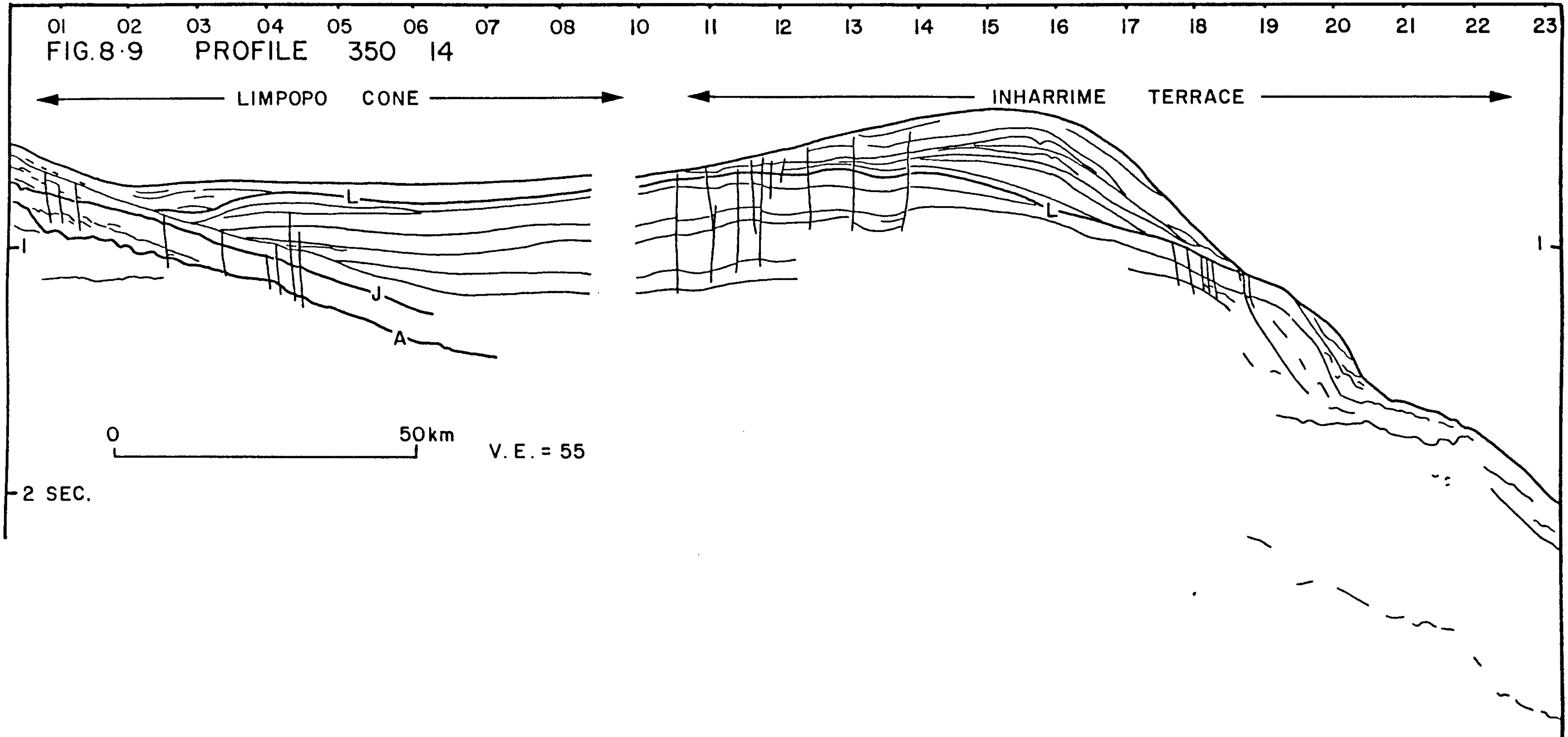


FIG. 8·9 PROFILE 350 14

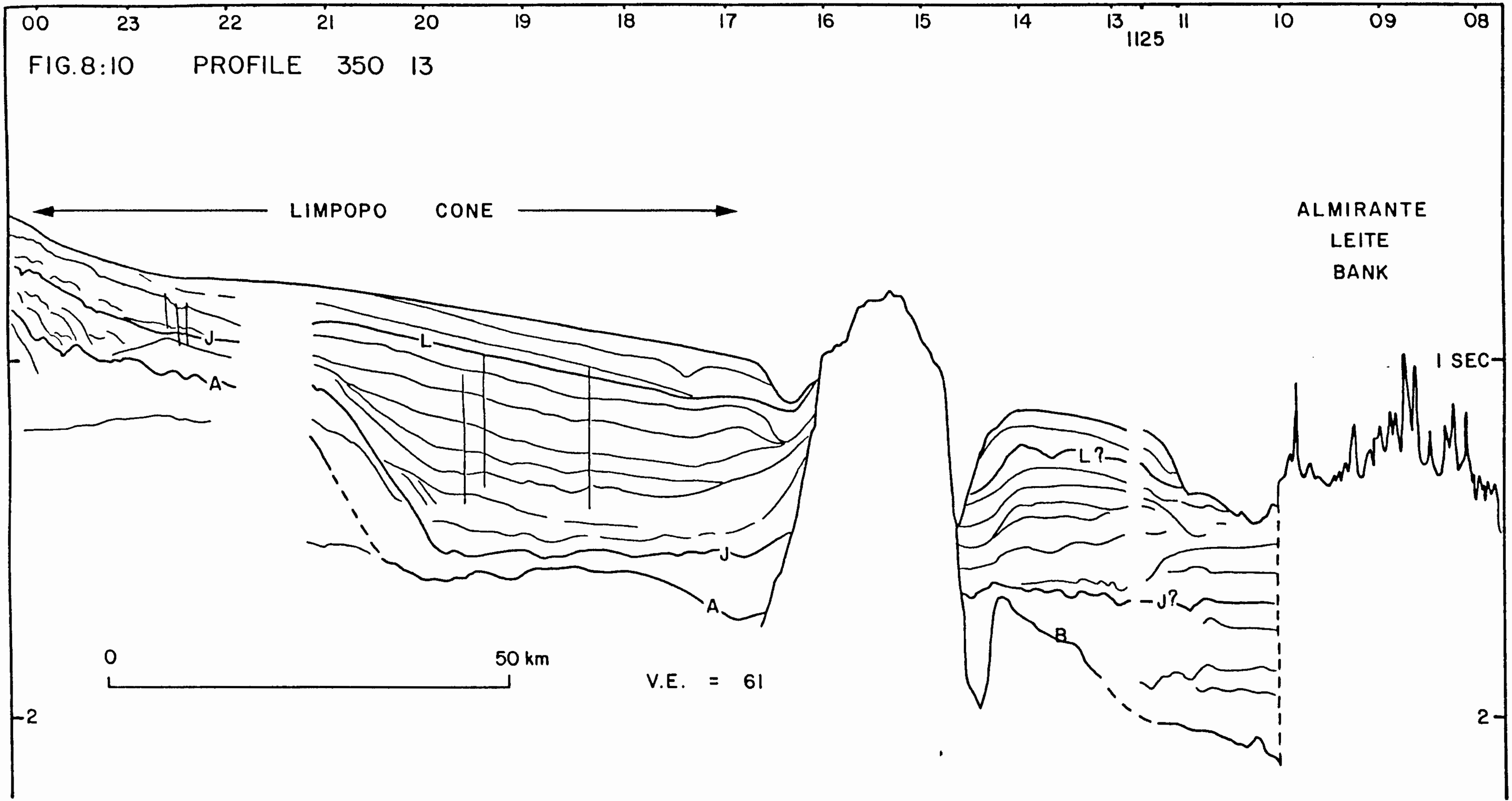
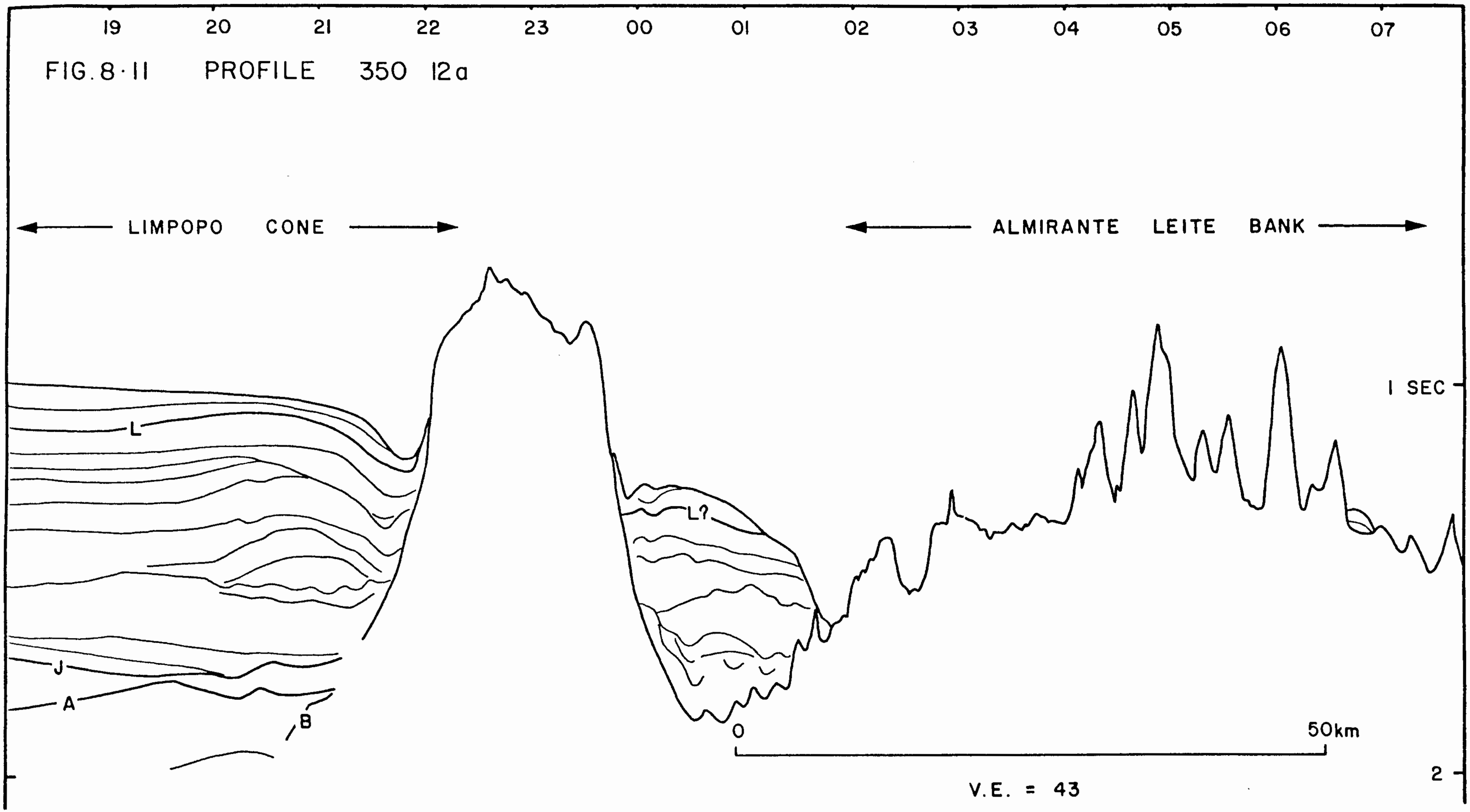


FIG.8:10 PROFILE 350 13


19
20
21
22
23
00
01
02
03
04
05
06
07
FIG. 8·11 PROFILE 350 12a
LIMPOPO CONE
ALMIRANTE LEITE BANK
1 SEC
2
L
L?
J
A
B
0
50km
V.E. = 43

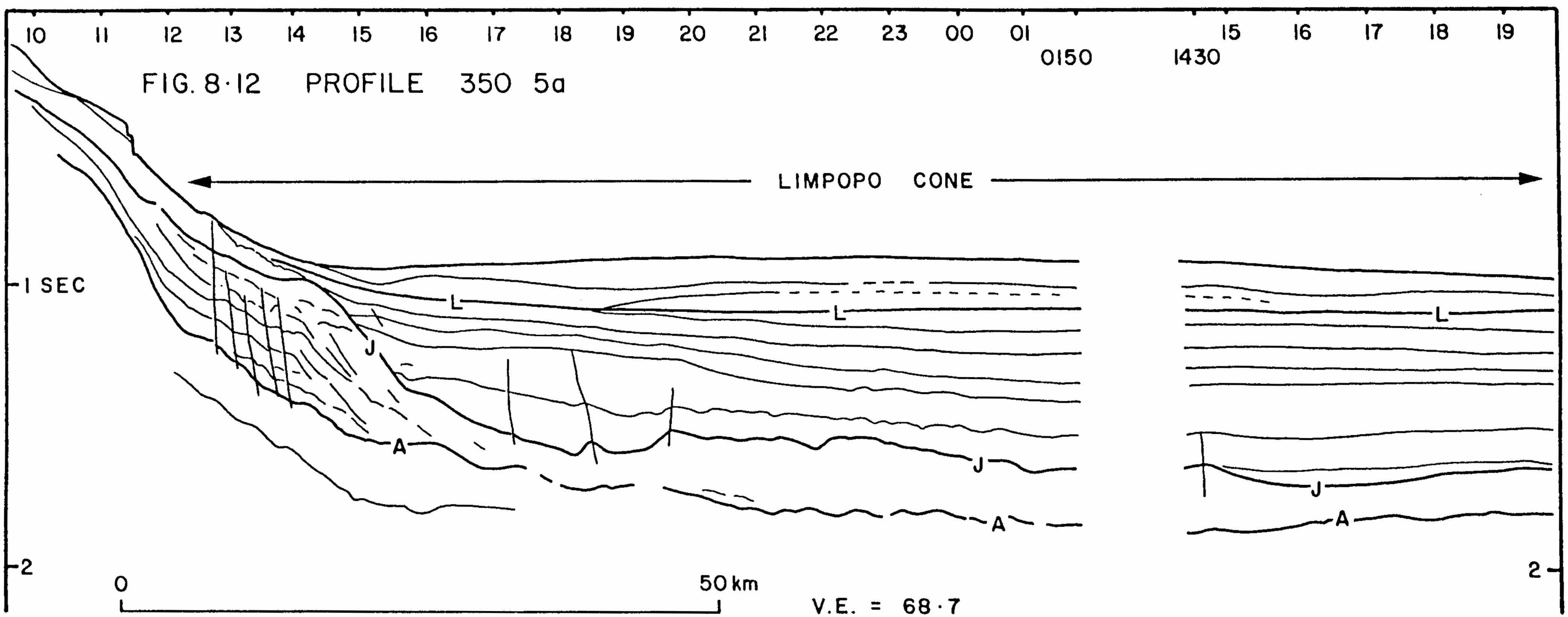


FIG. 8·12 PROFILE 350 5a

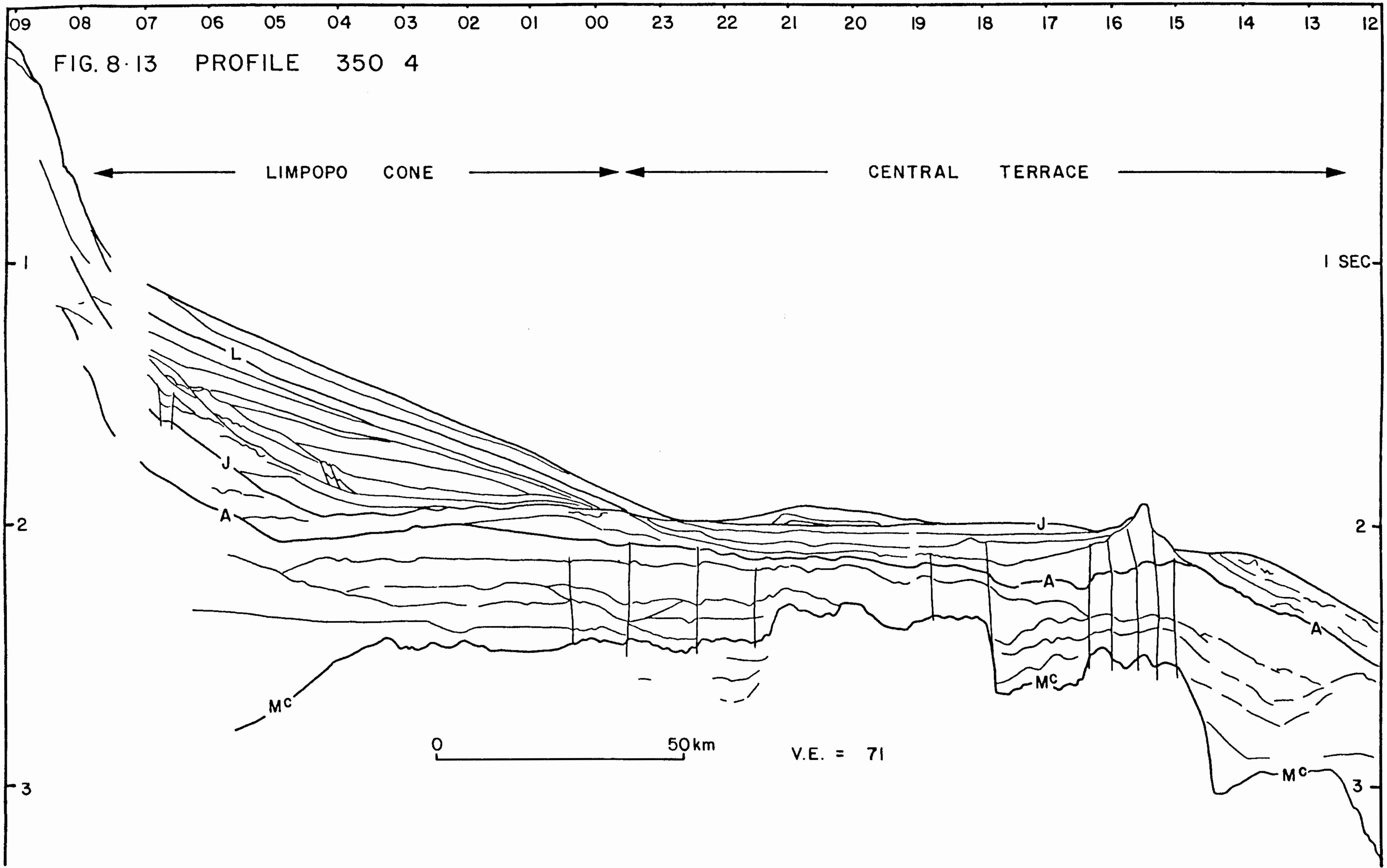


FIG. 8·13 PROFILE 350 4

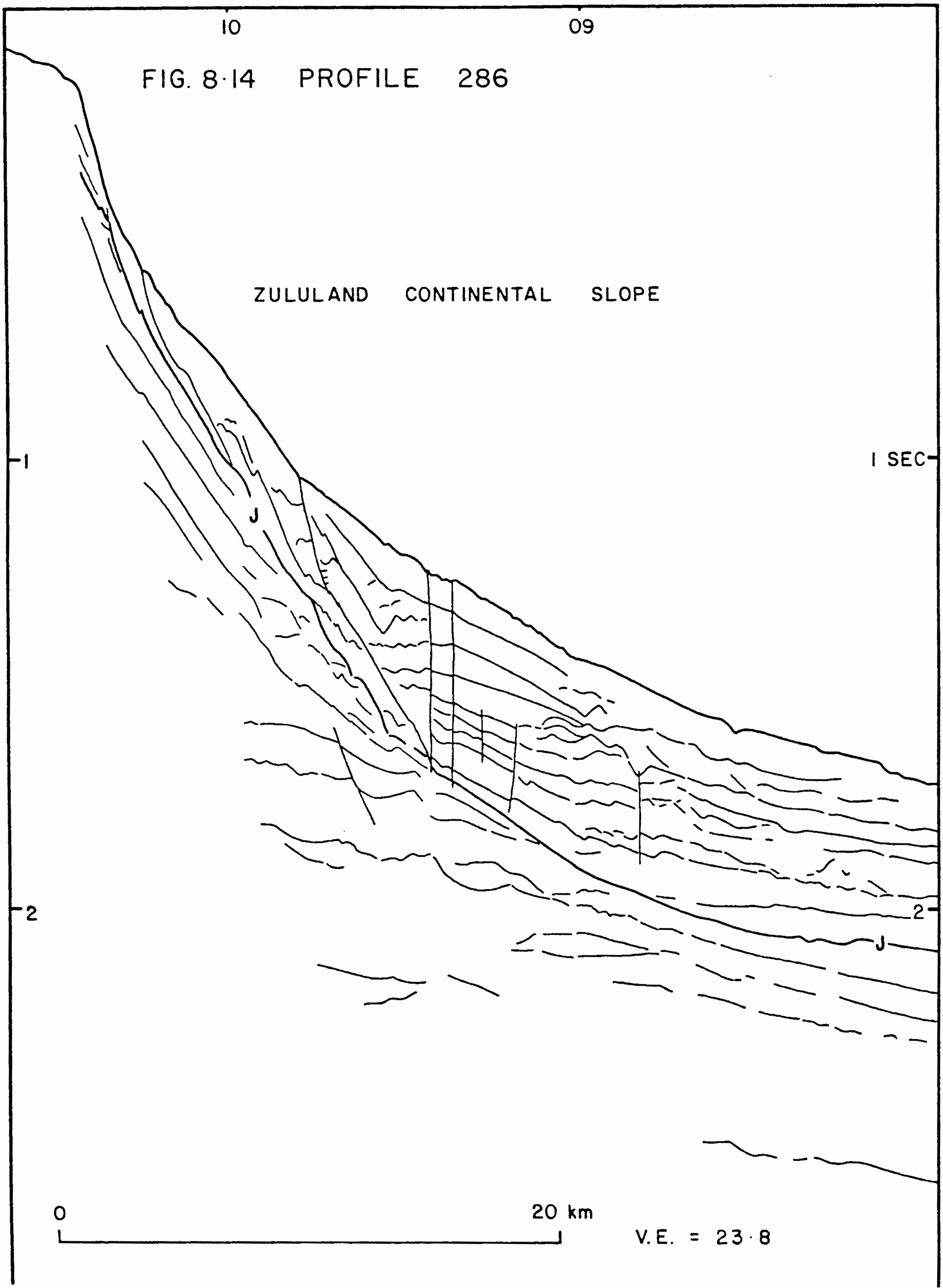


FIG. 8·14 PROFILE 286

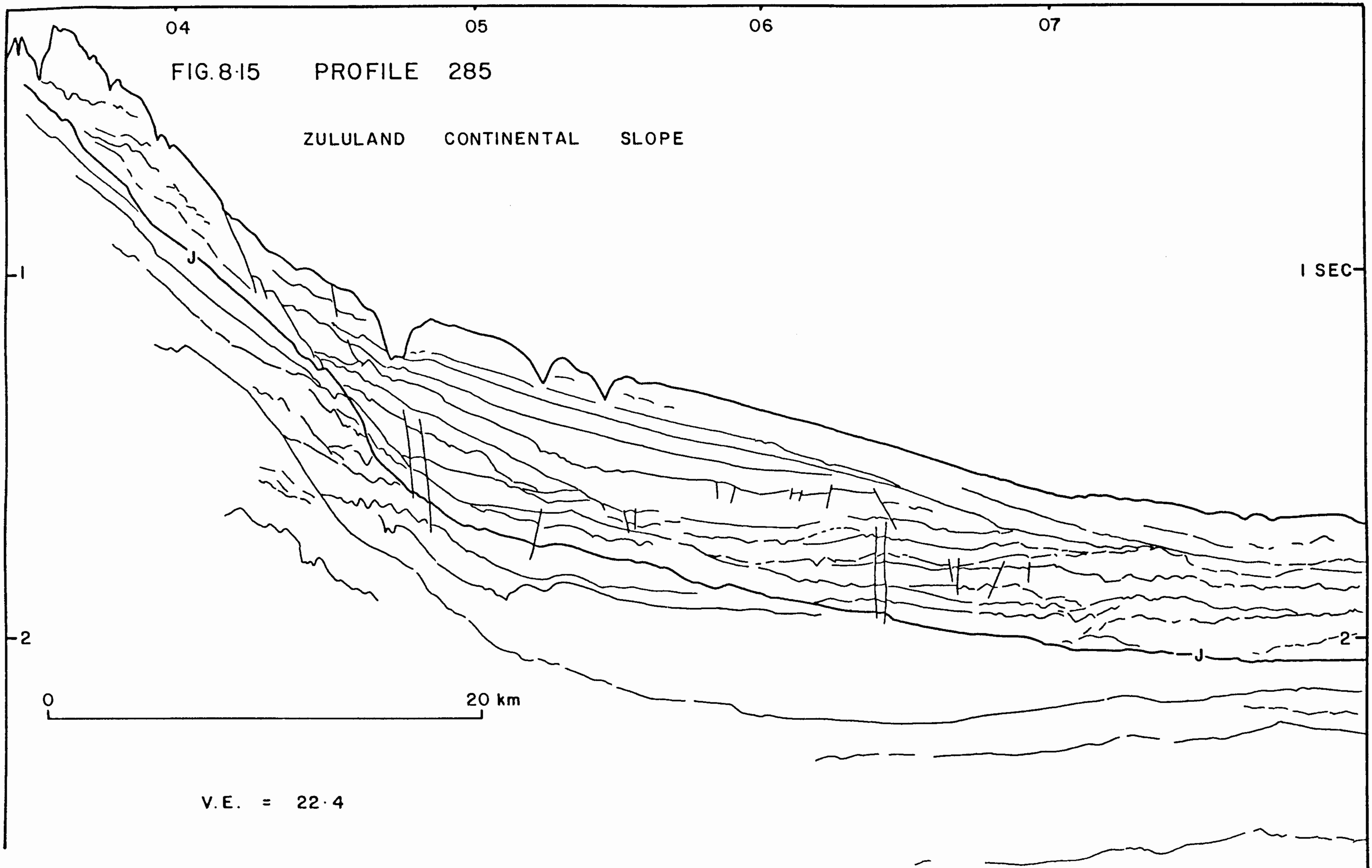
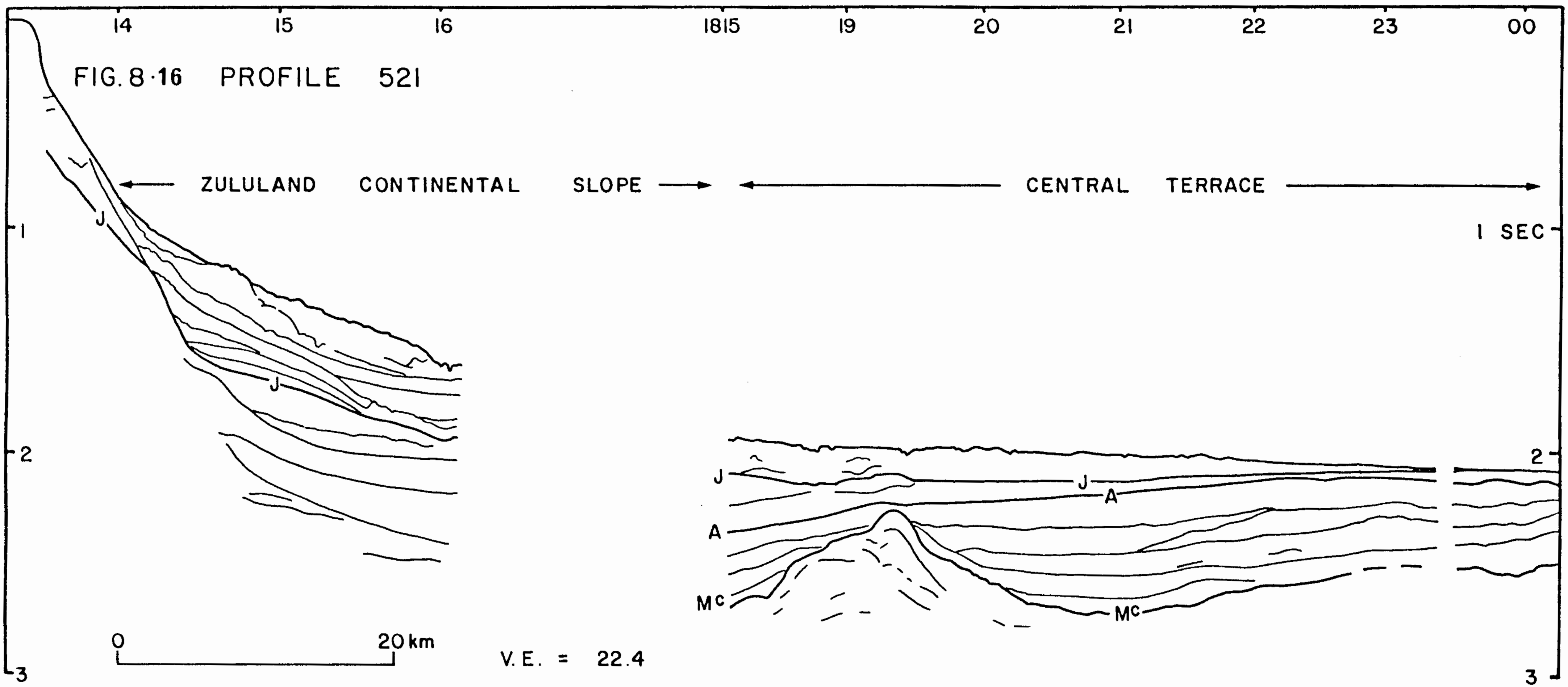


FIG. 8·15 PROFILE 285

FIG. 8·16 PROFILE 521



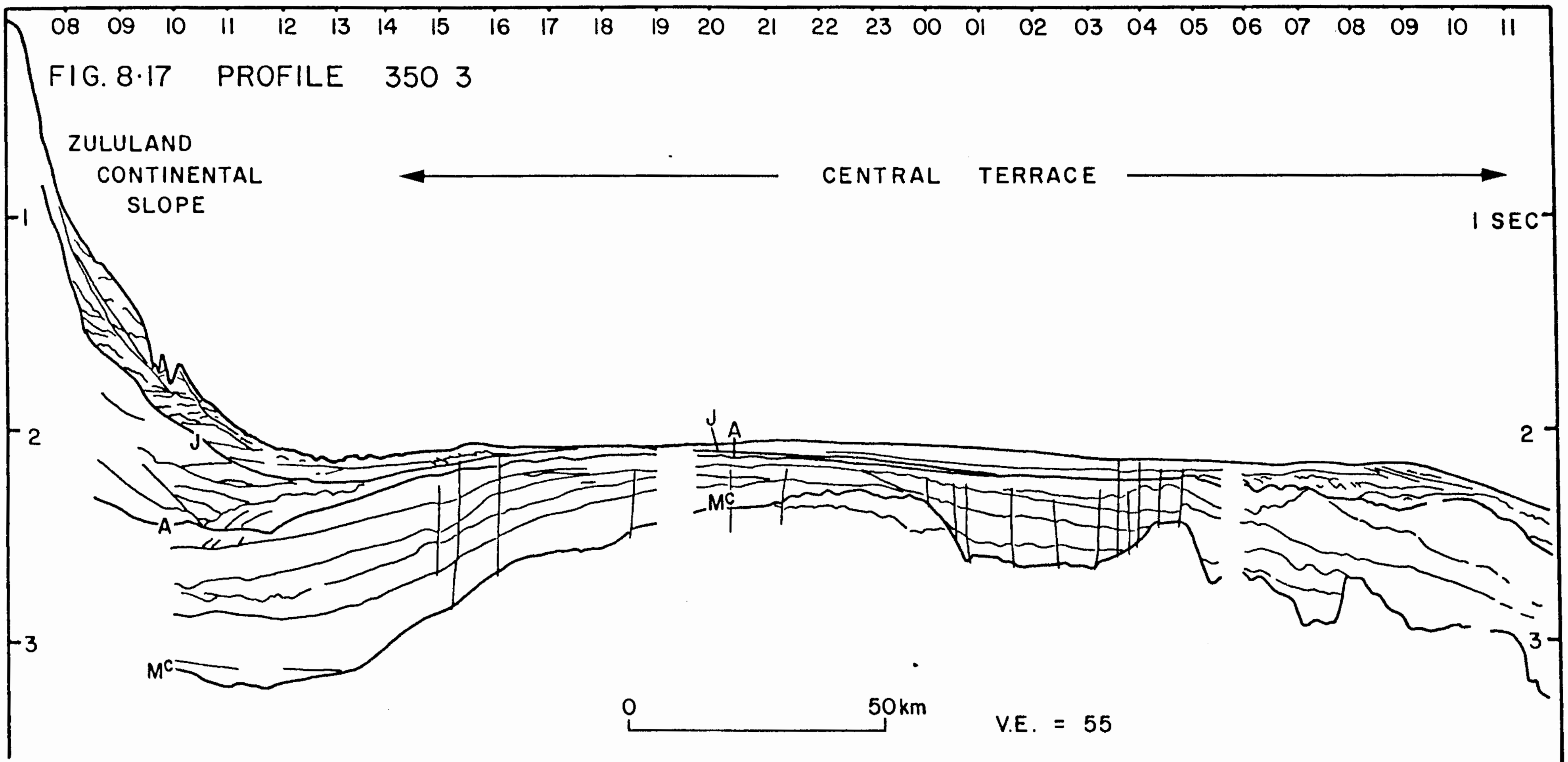


FIG. 8·17 PROFILE 350 3

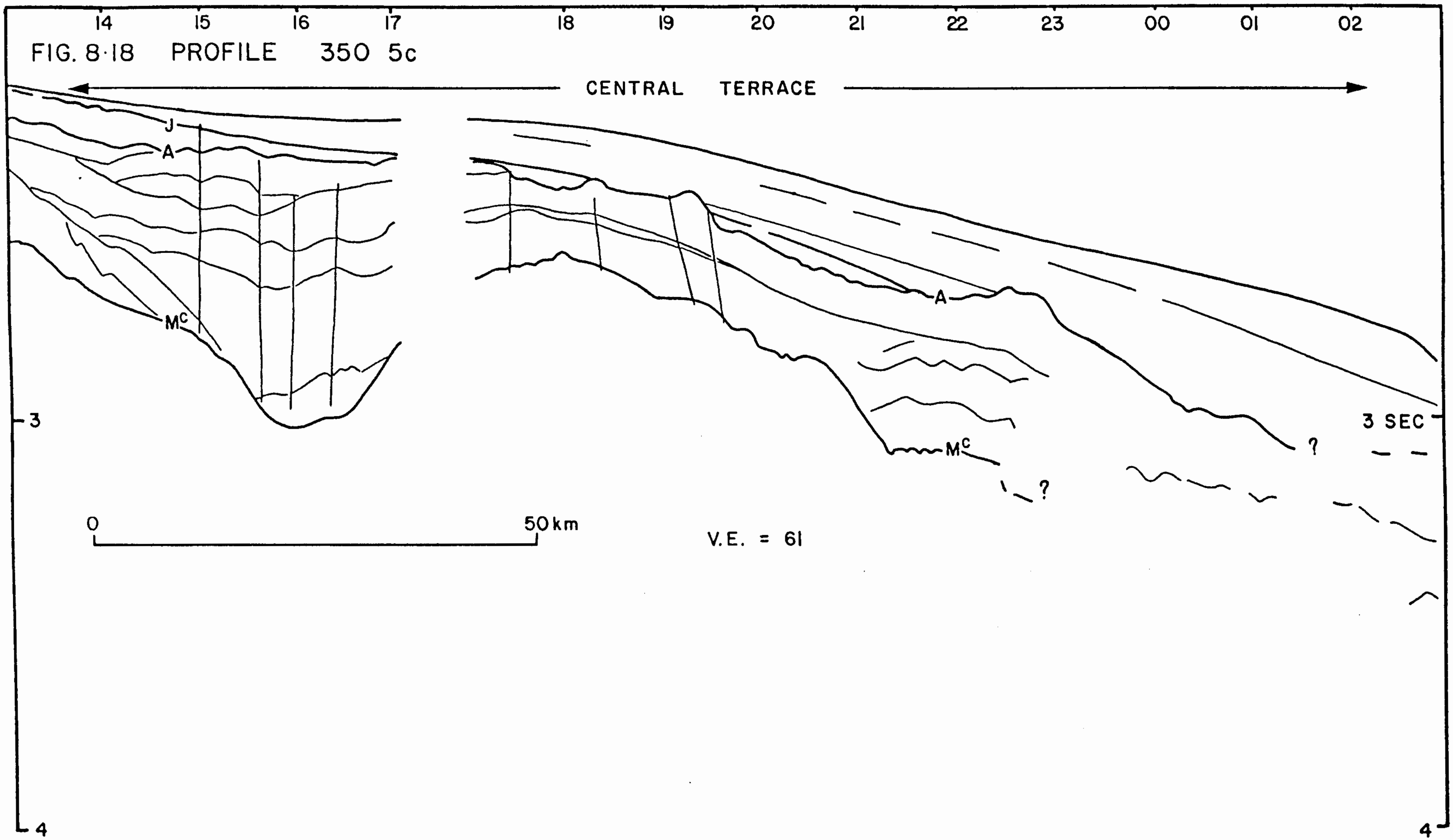


FIG. 8·18 PROFILE 350 5c

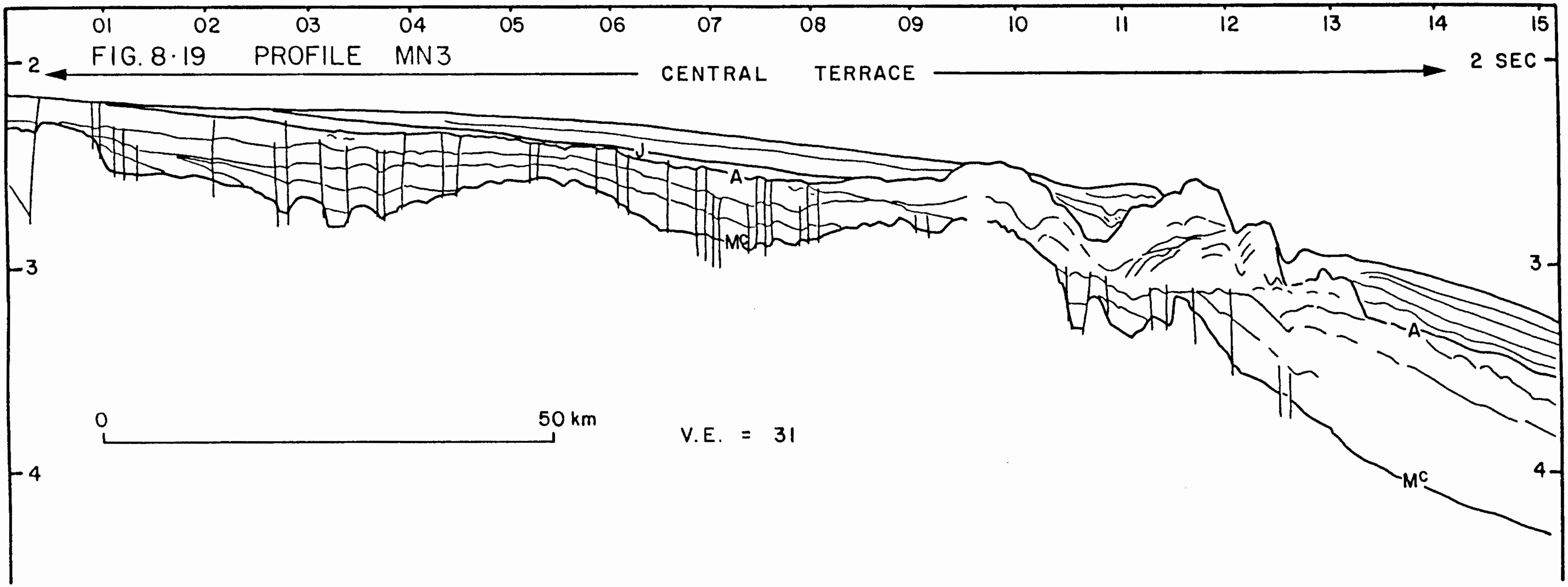


FIG. 8·19 PROFILE MN3

FIG. 8·20 PROFILE 371 7

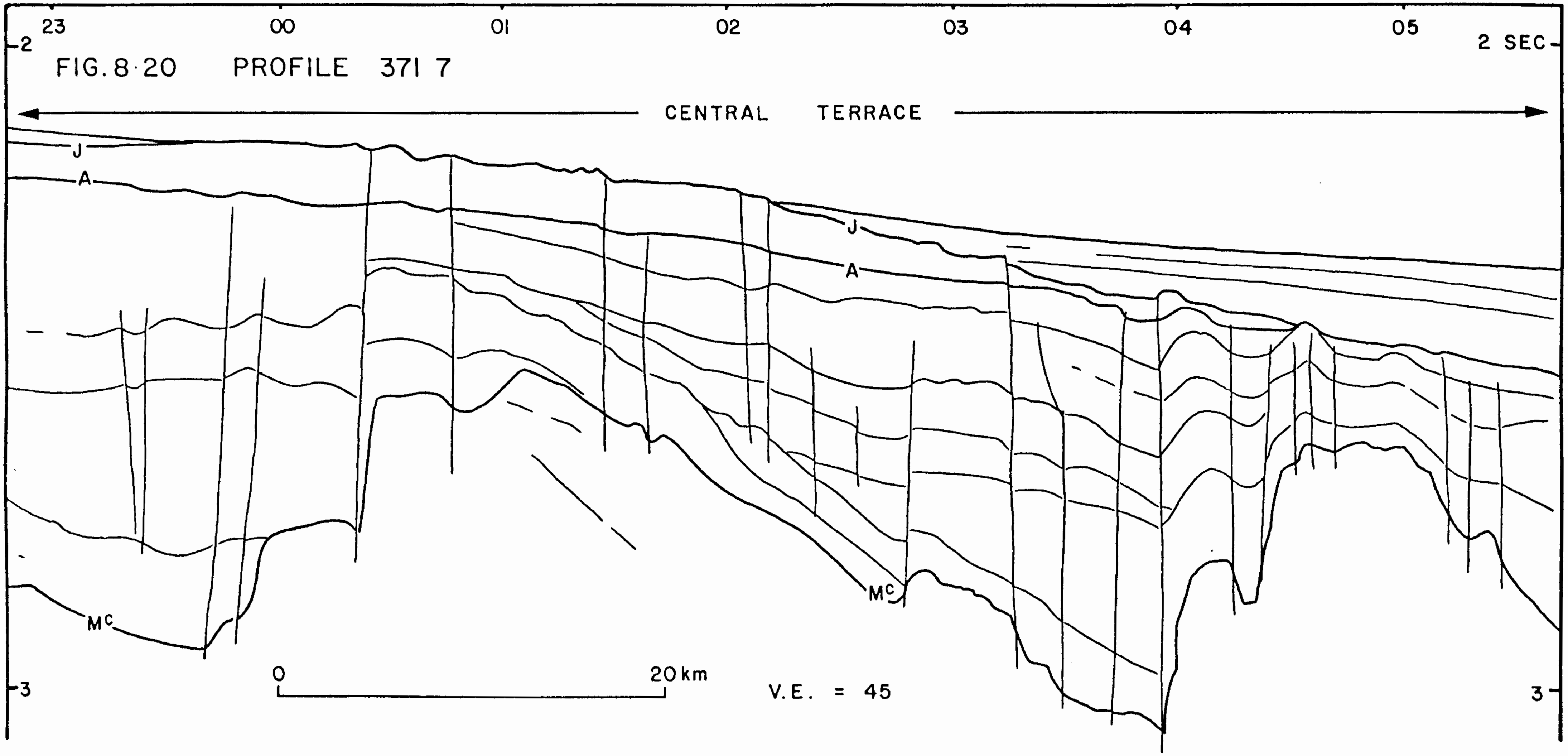


FIG. 8·21 PROFILE 284

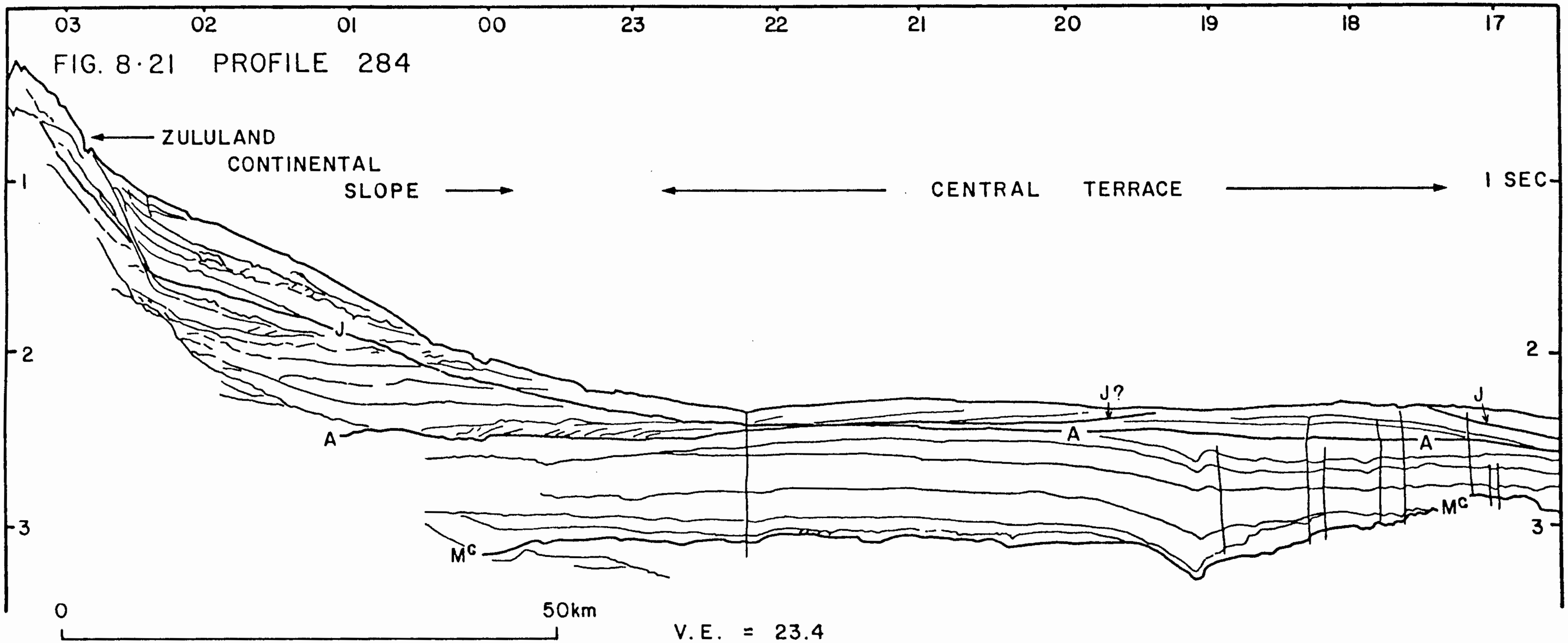
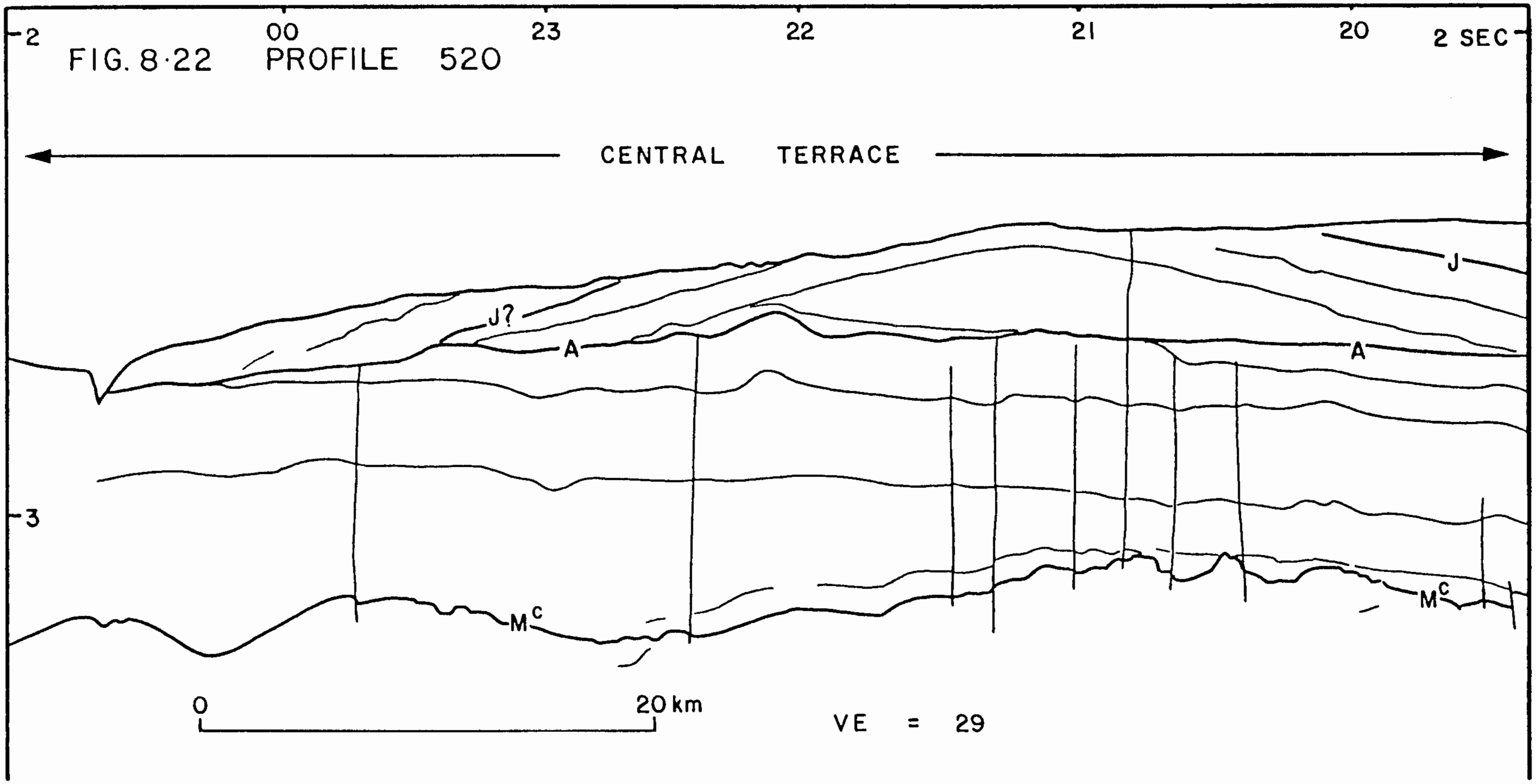


FIG. 8·22 PROFILE 520



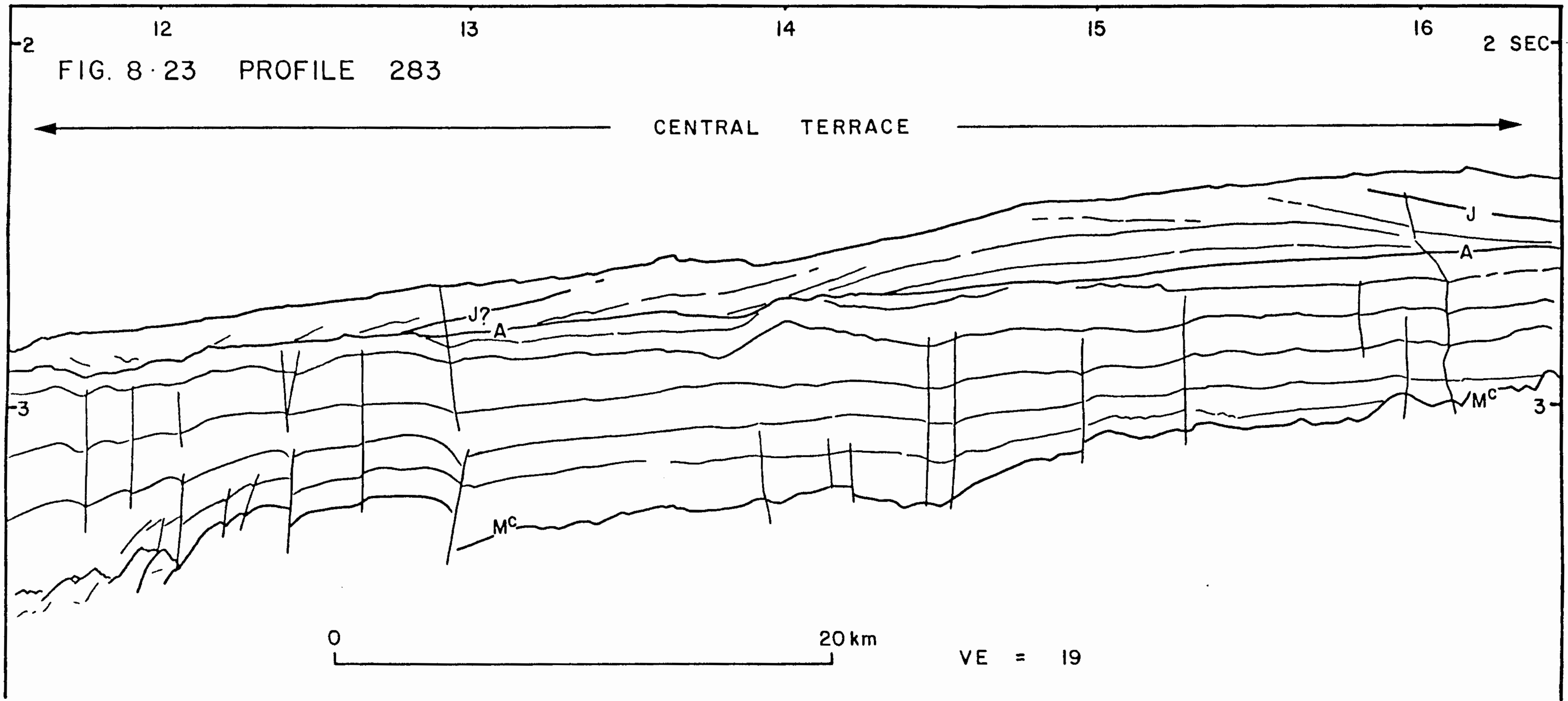


FIG. 8·23 PROFILE 283
CENTRAL TERRACE
12
13
14
15
16
2
2 SEC
3
3
J
A
J?
A
M^c
M^c
0
20 km
VE = 19

FIG. 8·24 PROFILE 287

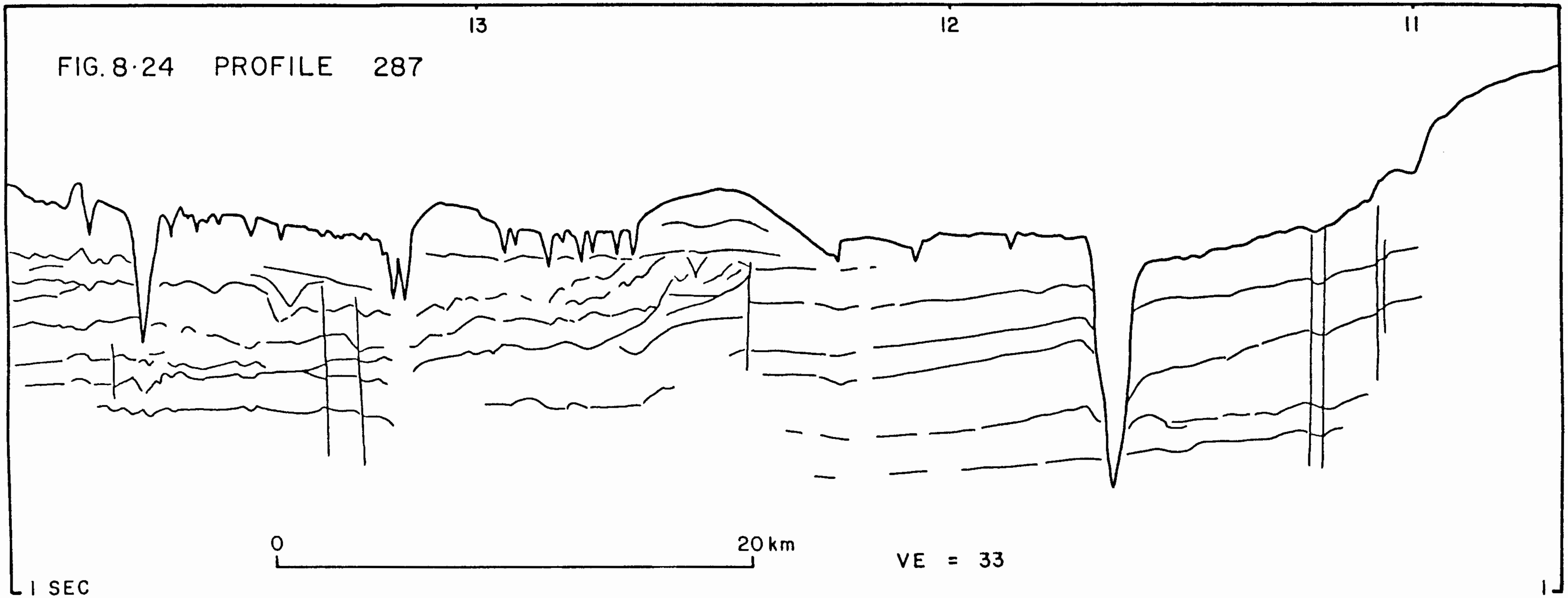


FIG. 8·25 PROFILE 371 6

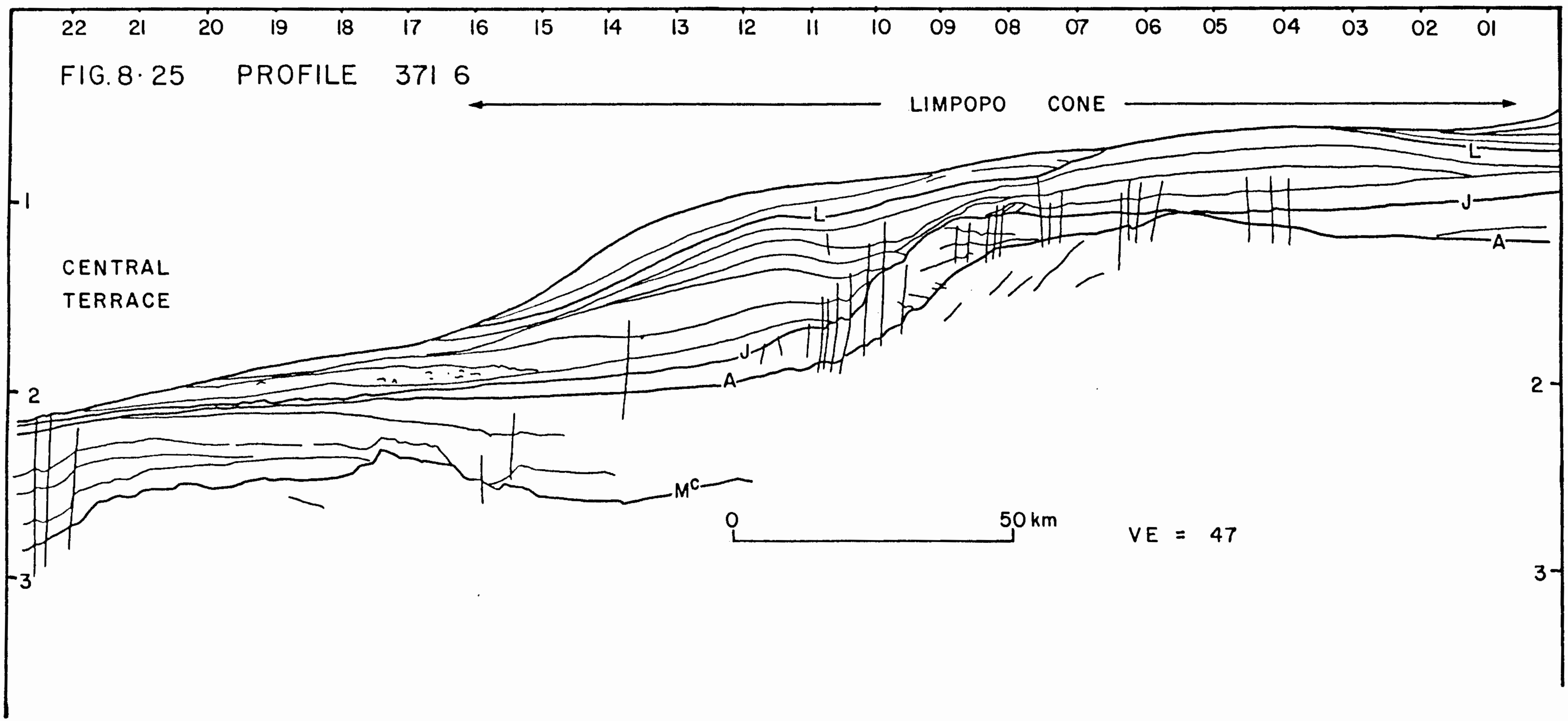
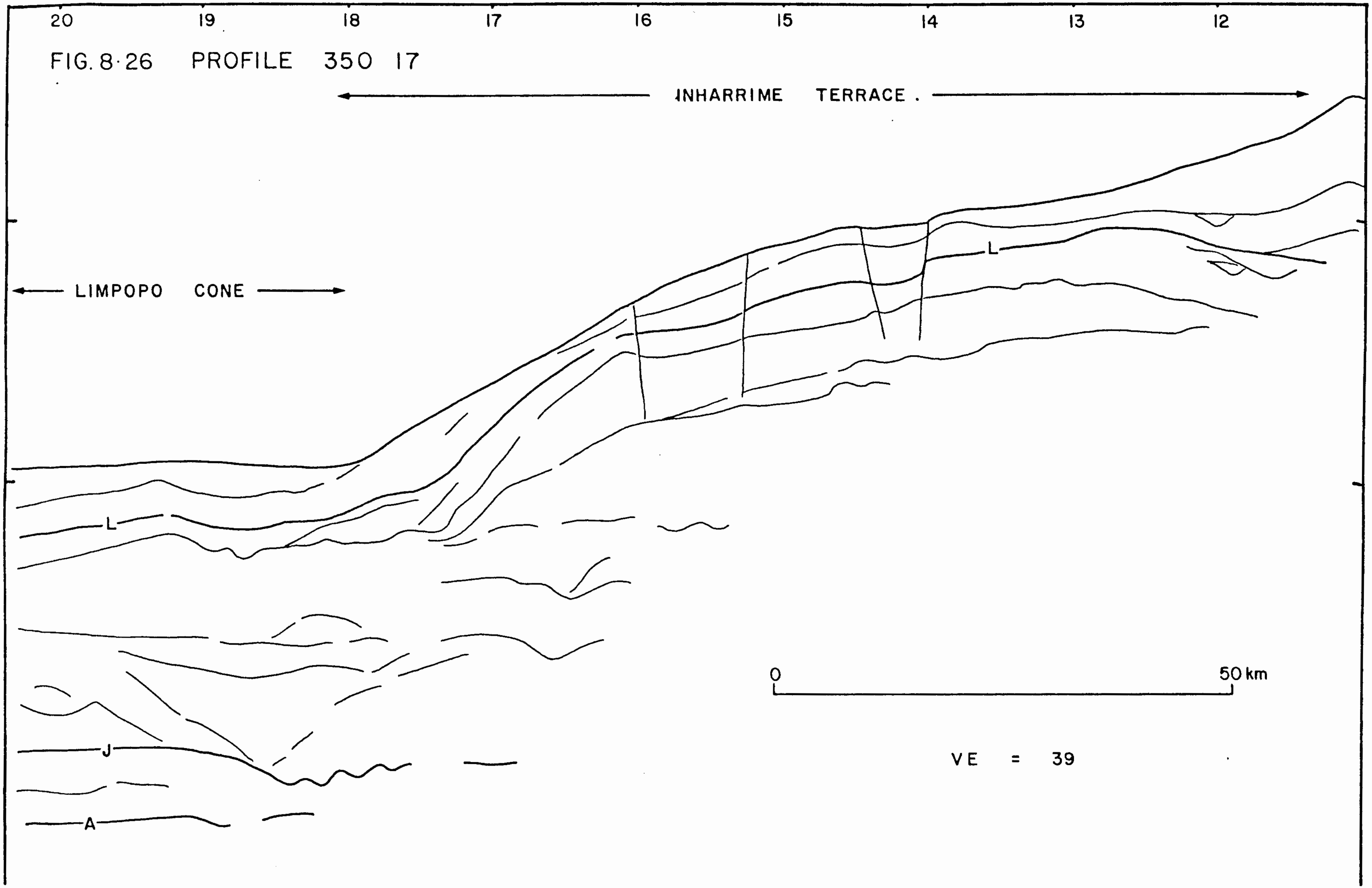


FIG. 8·26 PROFILE 350 17



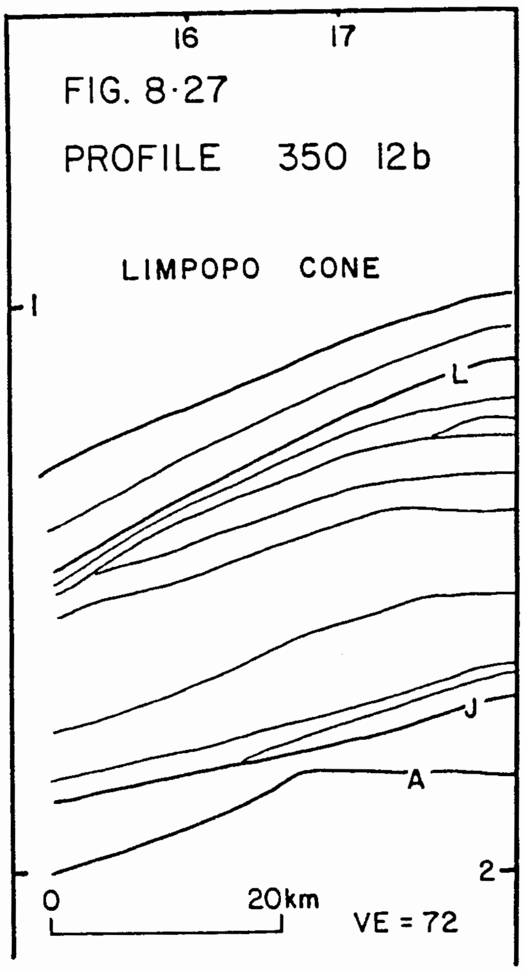


FIG. 8·27

PROFILE 350 12b

FIG. 8·29 PROFILE 371 8

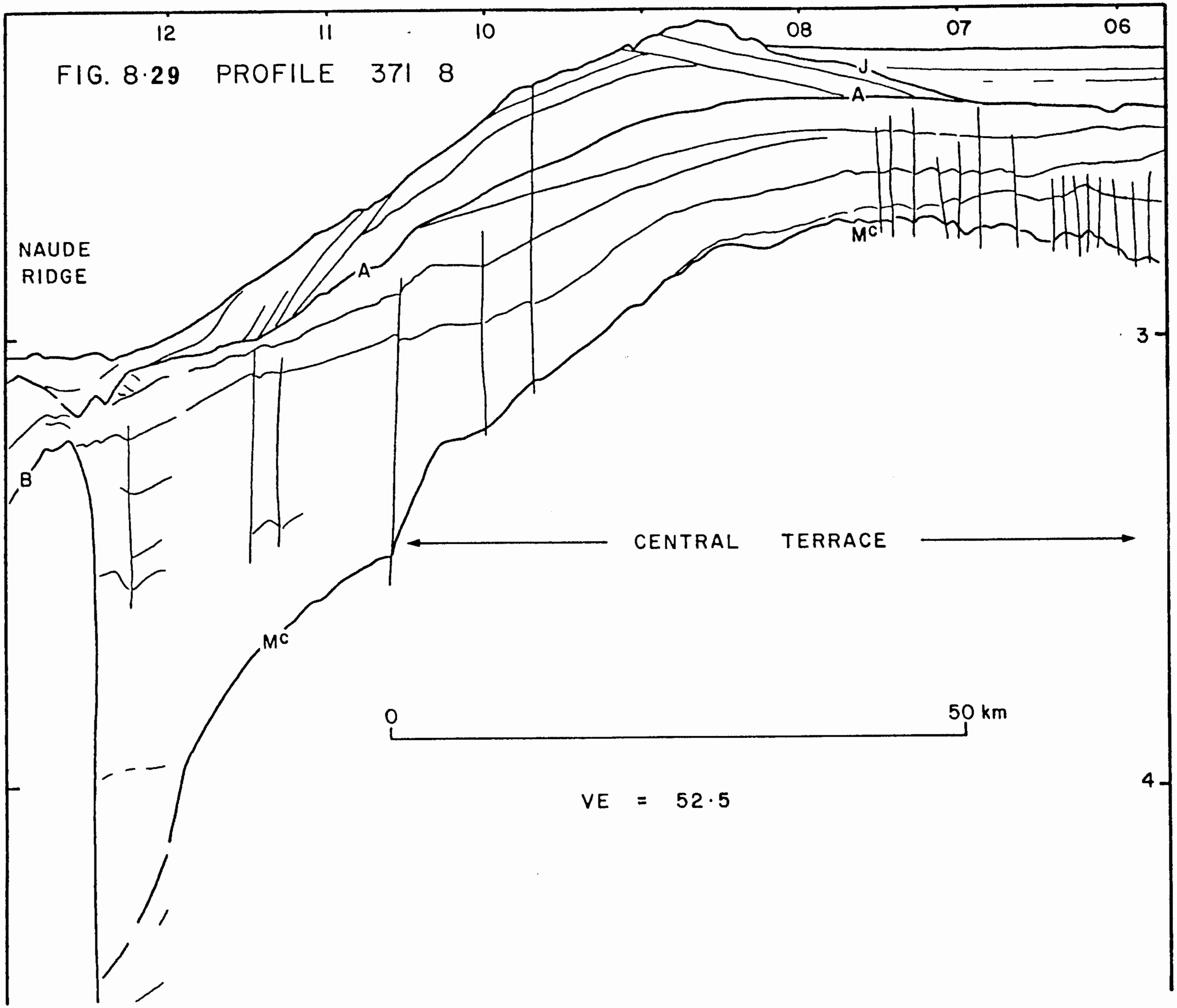
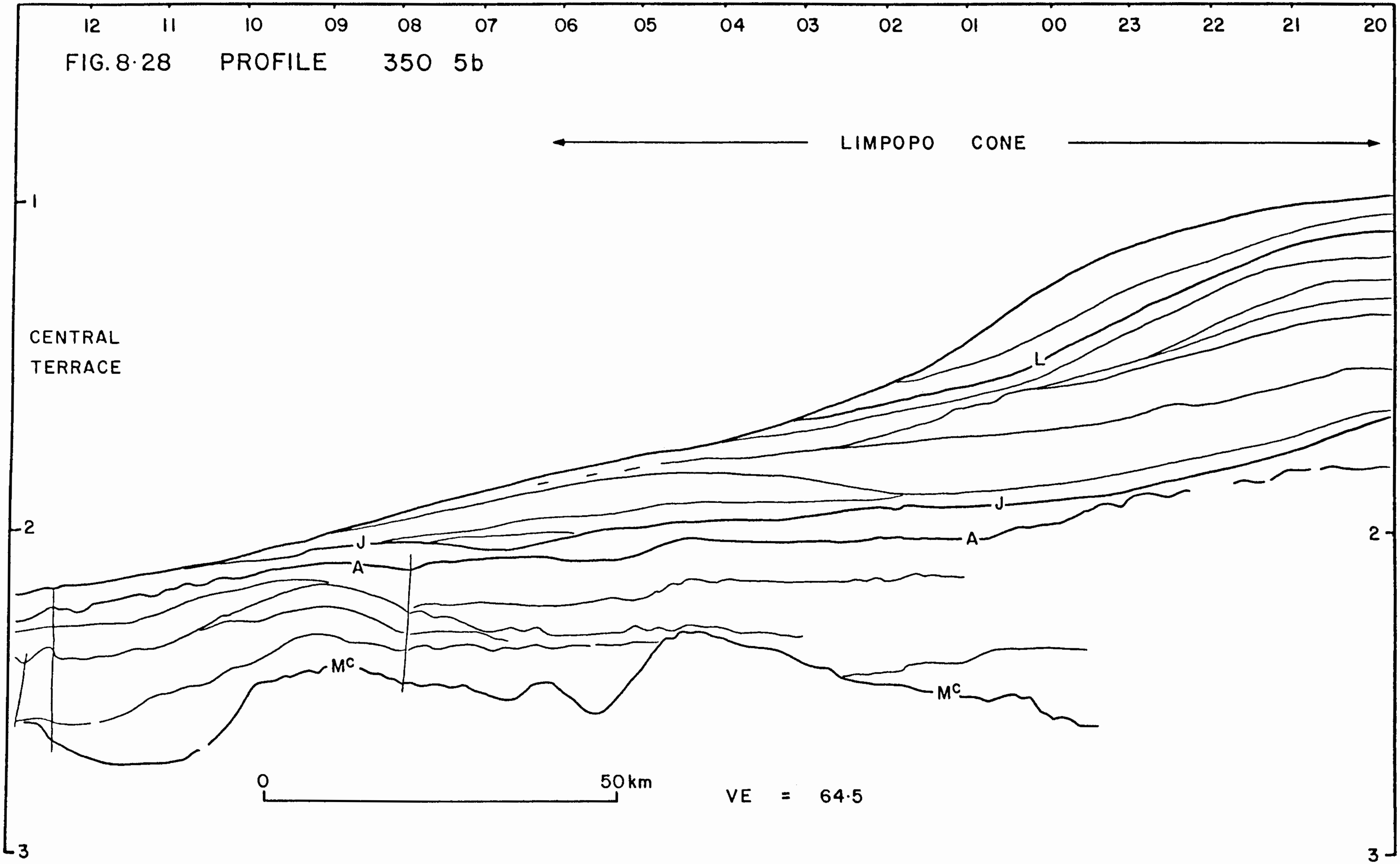


FIG. 8·28 PROFILE 350 5b



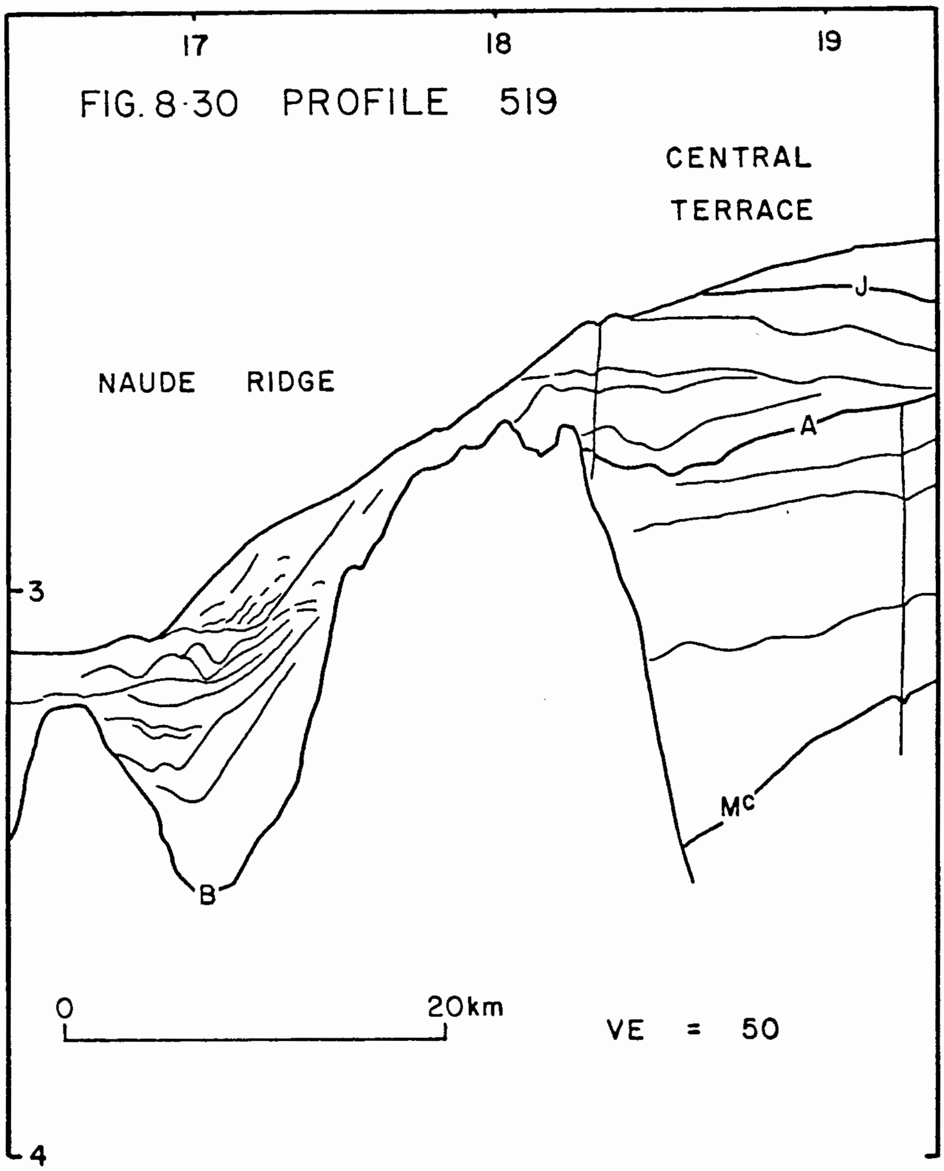
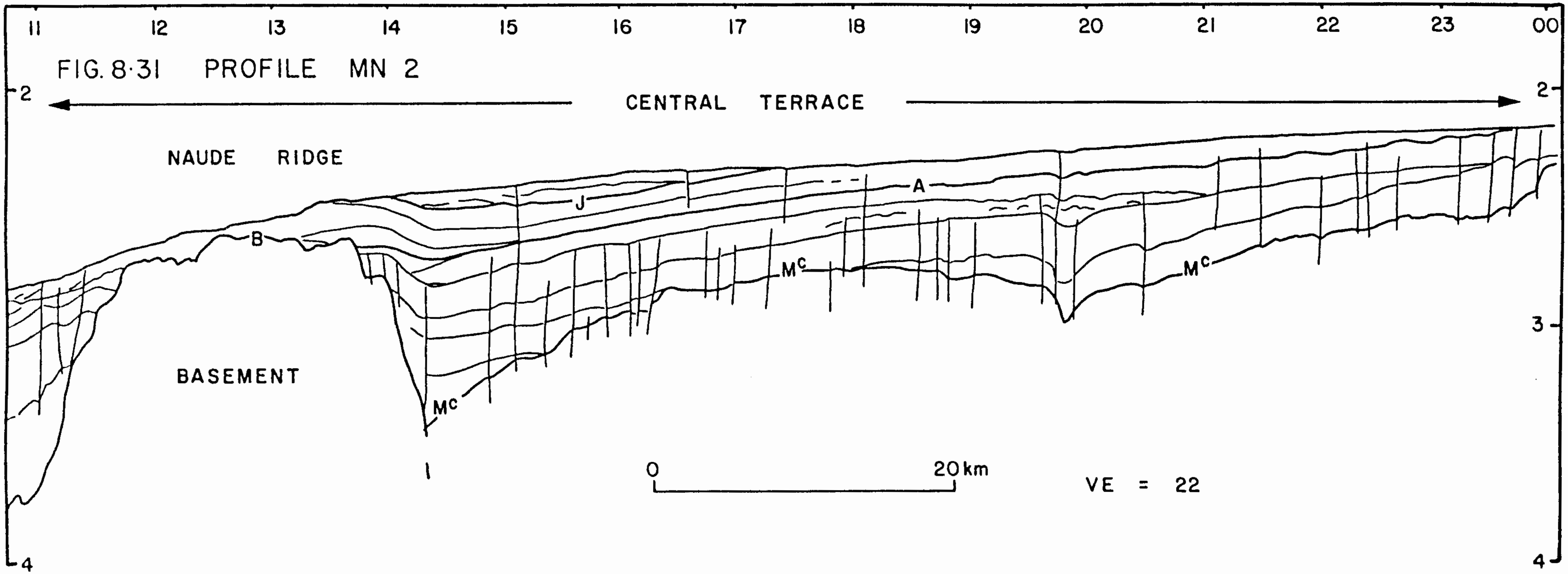

17
18
19
FIG. 8·30 PROFILE 519
CENTRAL
TERRACE
NAUDE RIDGE
J
A
3
B
Mc
0
20km
VE = 50
4

FIG. 8·31 PROFILE MN 2



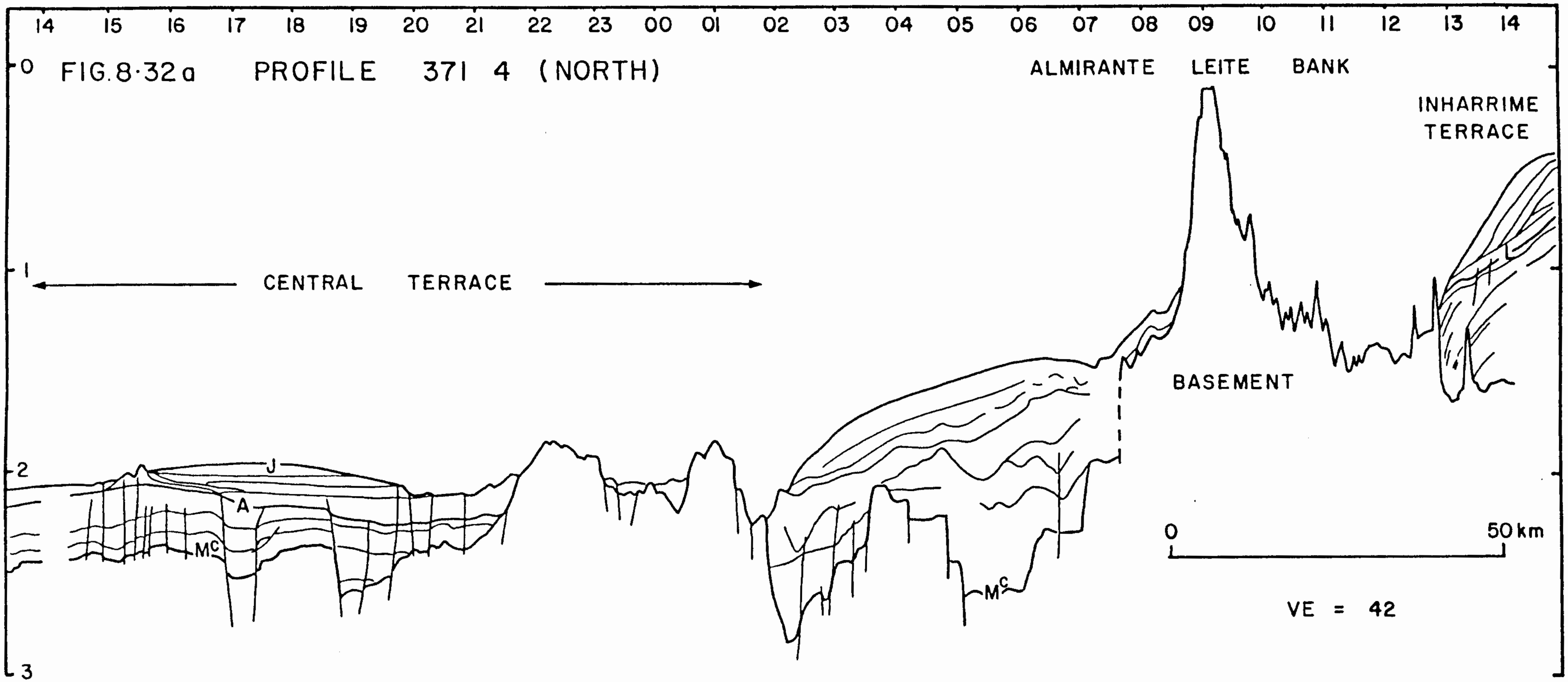
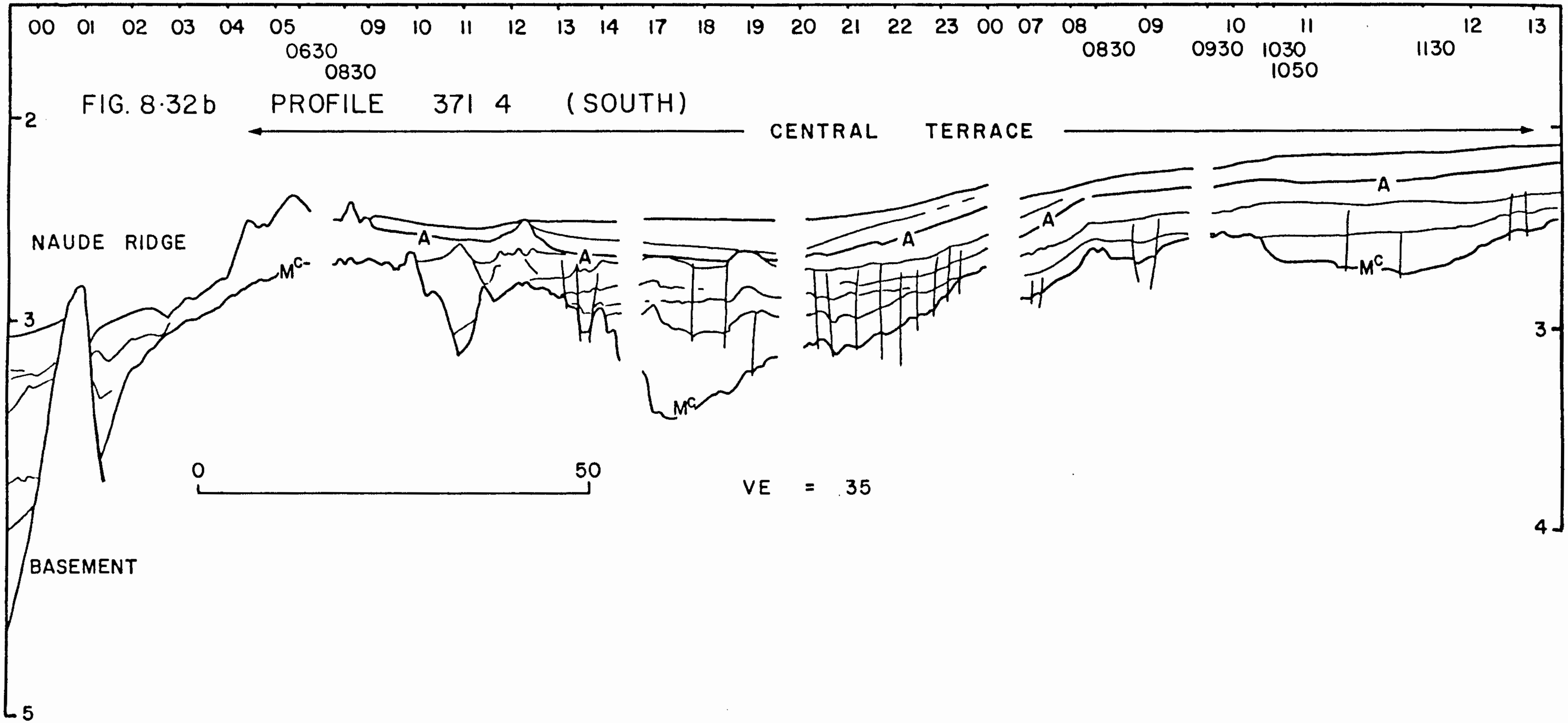


FIG.8·32a PROFILE 371 4 (NORTH)
14 15 16 17 18 19 20 21 22 23 00 01 02 03 04 05 06 07 08 09 10 11 12 13 14
0
1
2
3
ALMIRANTE LEITE BANK
INHARRIME
TERRACE
CENTRAL TERRACE
BASEMENT
J
A
M^c
0
50 km
VE = 42



00 01 02 03 04 05 09 10 11 12 13 14 17 18 19 20 21 22 23 00 07 08 09 10 11 12 13
0630
0830
0830
0930 1030
1050
1130
FIG. 8·32b PROFILE 371 4 (SOUTH)
CENTRAL TERRACE
NAUDE RIDGE
A
Mc
BASEMENT
0
50
VE = 35
2
3
4
5

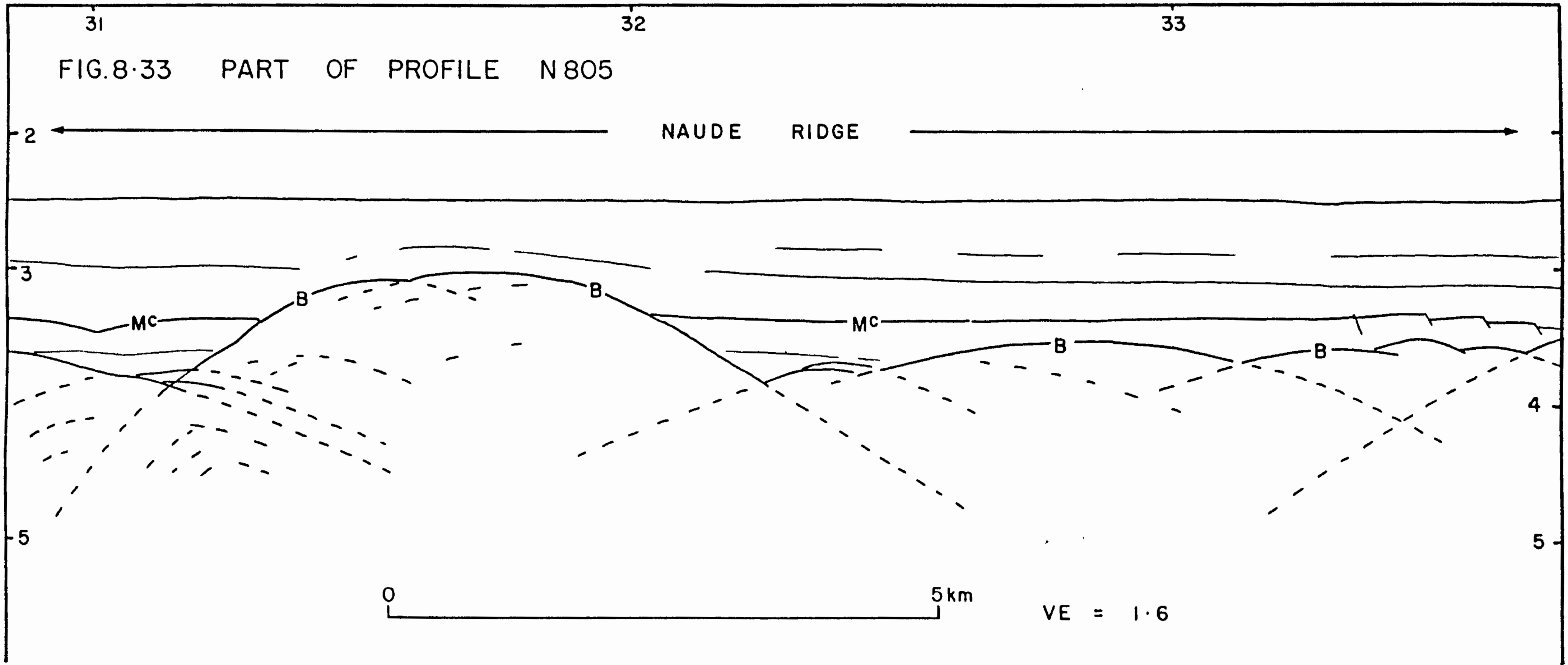


FIG. 8·33 PART OF PROFILE N805

area and lies, generally, around 1.25 sec depth (∿925 m). The region comprises over 80 seamounts, some of which are very shoal and steep-sided (section 2.3). The up-faulted region extends west-north-west and north under the Limpopo Cone and Inharrime Terrace, although here the shoalest known basement lies at -1.8 km (Fig.8.34; Beck & Lehner, 1974).

Two dredge-hauls were raised from this area. Sample 5739 comprised a thick manganese encrustation. Dredge-haul 5741 raised a variety of rock-types including vesicular lavas, porphyritic lavas, breccias and agglomerates of volcanic fragments and bombs. The mineralogy comprised feldspar (high albite/sanidine with some microcline) olivine (fayalite) hornblende, biotite, augite and nepheline, with minor accurrences of ore minerals, a sodalite group mineral Hauyne, and calcite in a replacement or in-filling situation. Flow structures of small phenocrysts in glass and undersaturation with respect to silica suggest classification as Nepheline trachytes.

On the northern part of the Central Terrace, basement lies generally between 2.5 and 3 sec depth. Relief, here, is more subdued. Although data are sparse, basement appears to descend to both west and east from a high at 34.4°E. The valley to the west of this trends north-south at 33.4°E. Southwards, basement descends to 3.25 - 3.75 sec before rising to ∿2.75 sec on the second major basement high - the Naudé Ridge. This is also a complex feature which forms the core to the southern Central Terrace. The Naudé Ridge is predominantly subsurface, with a single steep-sided (up to 27°) peak outcropping at 28° 51'S, 34° 42'E (profiles N805, MN4, 371 4 - Fig.8.32). Two more highs exist to the west of this (Fig.8.34) while the Naudé Ridge extends southwards to the Tugela Cone and the Deep Ocean Basin via three more separate peaks. The most westerly high (Fig.8.34) forms a narrow very steep-sided ridge (61° slope) trending WNW. The main part of the ridge forms a broad flat-topped ENE trending block whose flanks slope less steeply (8-16°). The general configuration of basement is partly reflected in the sediment

Table 8.1

Back-tracking of basement depths

Site	present depth (sec)	present depth (m)	water depth (sec)	water depth (m)	sediment thickness (sec)	sediment thickness (m)	loading	basement corrected depth	emplacement depth
(1) cont. marg. 28°S	3.40	4358	1.14	855	2.26	3508	2147	2211	+ 1289
(2) west valley 28°S 33.5°E	4.15	4553	2.35	1763	1.80	2790	1710	2843	+ 657
(3) C. Terr. 28°S 34.25°E	2.53	2194	2.16	1620	0.37	574	352	1842	+ 1658
(4) east valley 35.7°E	3.9	4221	2.28	1710	1.62	2511	1539	2682	+ 818

isopachyte map on basement (Fig.8.45). Use of Crough's (1983) empirical equation for correction of sediment loading shows that the configuration of high basement on the Central Terrace flanked by valleys is an original tectonic feature which has been enhanced by differential sediment loading (Table 8.1).

8.3.1.1 Regional Relationships

The data provided by Beck & Lehner (1974) indicate that the N.N.V. is floored by volcanic material which is an extension of the Lebombo volcanic suite. Boreholes on the broad coastal plain of Mozambique bottomed in volcanic material also considered to be an extension of the Lebombo basalts and rhyolites (Flores, 1973). In this region volcanic material lies at over 3000 m sub-surface, whereas in the Zandemala borehole (Fig.8.34) volcanic rock lies at 1768 m depth. It appears that a broad faulted high underlies the southernmost coastal plain of Mozambique and the northern Limpopo Cone, with the Almirante Leite Bank volcanic complex forming the shoalest part. Basement descends into valleys flanking the Central Terrace while the southernmost Central Terrace is cored by the Naudé Ridge. To the east, basement descends to over 5 sec before rising to between 3.5 and 2.5 sec on the rugged Mozambique Ridge. South of the Naudé Ridge, in the Deep Ocean Basin sub-province and its extensions, basement lies between 4.65 km (Ludwig et al., 1968; refraction site 160) and 7 km below sea-level (Chetty & Green, 1977). Available drilling, dredge-haul, seismic refraction and reflection data indicate that this entire area is underlain by volcanic material. In contrast, the inner Tugela Cone is underlain by Cape and Karoo rocks. Basement here descends to at least 3.3 sec (4.7 km using data from Ludwig et al., 1968; Du Toit & Leith, 1974) while under the southern face of the Tugela Cone, the South Tugela Ridge lies between 4.2 and 4.7 sec (Goodlad et al., 1982).

8.3.2 McDuff

Reflector McDuff lies at the top of a series of very prominent reflectors. It is best developed under the Central Terrace, being either masked by the first multiple or beyond penetration limits under the Limpopo Cone and Inharrime Terrace. McDuff has been extensively disrupted by faulting with throws ranging up to 0.7 sec (Figs.8.13, 8.17, 8.18, 8.19, 8.20, 8.21, 8.22, 8.23, 8.25, 8.28, 8.29, 8.30, 8.31, 8.32, 8.33). Most of the faults appear to be normal and the Central Terrace forms an irregular up-faulted block relative to flanking valleys to east and west (Figs.8.13, 8.17, 8.35). However locally at 28° 45'S, 33-33° 40'E (profile 283, 1000-1300 hrs, Fig.8.23) a series of imbricate thrust faults with throws up to 0.2 sec occur. Westerly blocks have ridden over easterly blocks. This compressive style of deformation has been demonstrated by Harding & Lowell (1979, Fig.7). Another unusual feature appears on profile 521 (Fig.8.16) where there is a smooth-sloped knoll of 0.5 sec relief, 11 km in extent, with slopes of 3.5°. The faint internal reflectors suggest a volcanic mound (see Mitchum et al., 1977, Fig.13 for a comparable feature) while the gentle slope suggests shield-type volcanic topography.

Regionally, the crest of the up-faulted block under the Central Terrace trends NNE at 34-34.5°E (Fig.8.35). From here, at a depth of 2.3 sec, it descends west, east, and steeply southwards, onlapping basement of the Naudé Ridge at >4.0 sec (profiles 371 8, 519, MN2 and 371 4, Figs.8.29-8.32). Horsts and graben up to 40 km long, with 0.5 sec relief, typify the Central Terrace. Fault-bounded blocks show relief of up to 1.0 sec on profile 371 4 where McDuff outcrops (Fig.8.32). Profiles 350 4 (Fig.8.13, 1755 hrs) and 371 4 (Fig. 8.32, 1835 hrs) intersect on a fault which is NW-oriented and dips NE. This trend is shown by the isochrons between 27° and 28°S and east of 35°E, although this may be a function of profile orientation and availability. A seismic bright spot occurs on reflector McDuff at 1745-1845 hrs on profile 371 6 (Fig.8.25).

8.3.2.1 Sediment unit 1: pre-McDuff

Because of the reflectivity of McDuff masking basement, the only data available on the basement - McDuff or Early Cretaceous sequence (section 8.2)

FIGURE CAPTIONS

Figs.8.34-8.45. Structure contour and isopachyte charts. Figs.8.34, 8.35, 8.37, 8.39 and 8.42 = structure contour charts on respectively, Basement, McDuff, Angus, Jimmy and 'L'. Figs.8.36, 8.38, 8.40, 8.41, 8.43, 8.44 and 8.45 = isopachyte charts on respectively, sedimentary unit 1 (Basement-McDuff), unit 2 (McDuff-Angus), unit 3 (Angus-Jimmy), units 4 & 5 (post-Jimmy), unit 4 (Jimmy-'L'), unit 5 (post-'L'), and total sedimentary thickness. All mearurements are in seconds two-way-time except for spot-values taken from Beck & Lehner (1974) which are in km and are shown on depth to basement (Fig.8.34) and total sedimentary thickness (Fig.8.45). These two Figures are contoured at 250 m/sec intervals while all others are contoured at 100 m/sec intervals. Values given on contours are m sec x 10. Ornament depicting sub-crops of Basement, McDuff, Angus, Jimmy and 'L' are the same as in Fig.8.46. Fine dotted lines in all diagrams indicate tracks where relevant data were available (c.f. Fig.8.1). On Figs.8.42 (depth to L) and 8.44 (post-L thickness) dot-dash line off Maputo indicates down-lap of 'L' on to a reflector which marks the top of the slump facies within unit 4.

Fig.8.46. = Geology of the area super-imposed on to a 100 m isobath chart. 1 = Basement; 2 = McDuff; 3 = Angus; 4 = Early Miocene; 5 = pre-Jimmy; 6 = undifferentiated post-Jimmy; 7 = pre-'L'; 8 = post-'L'.

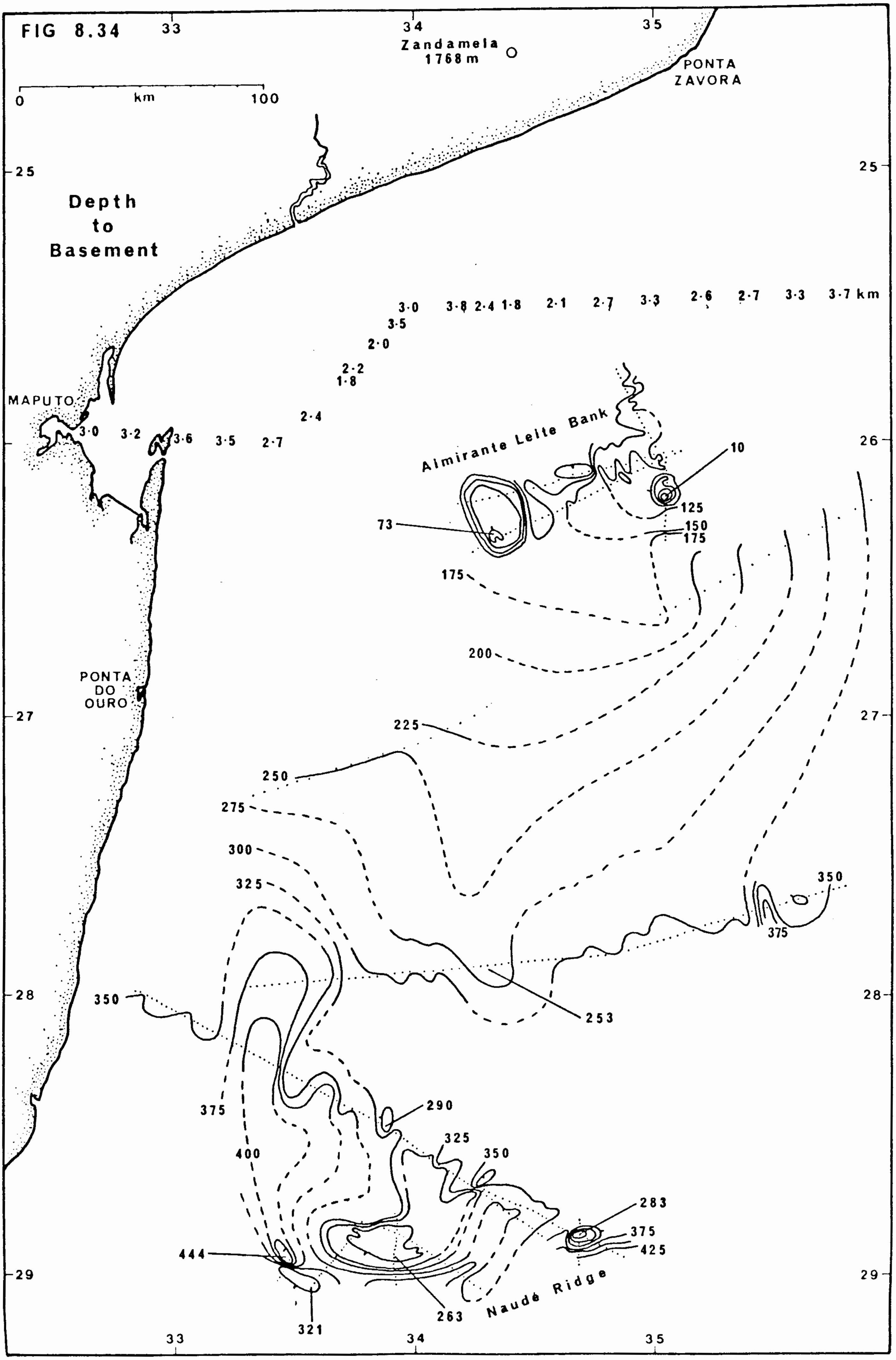
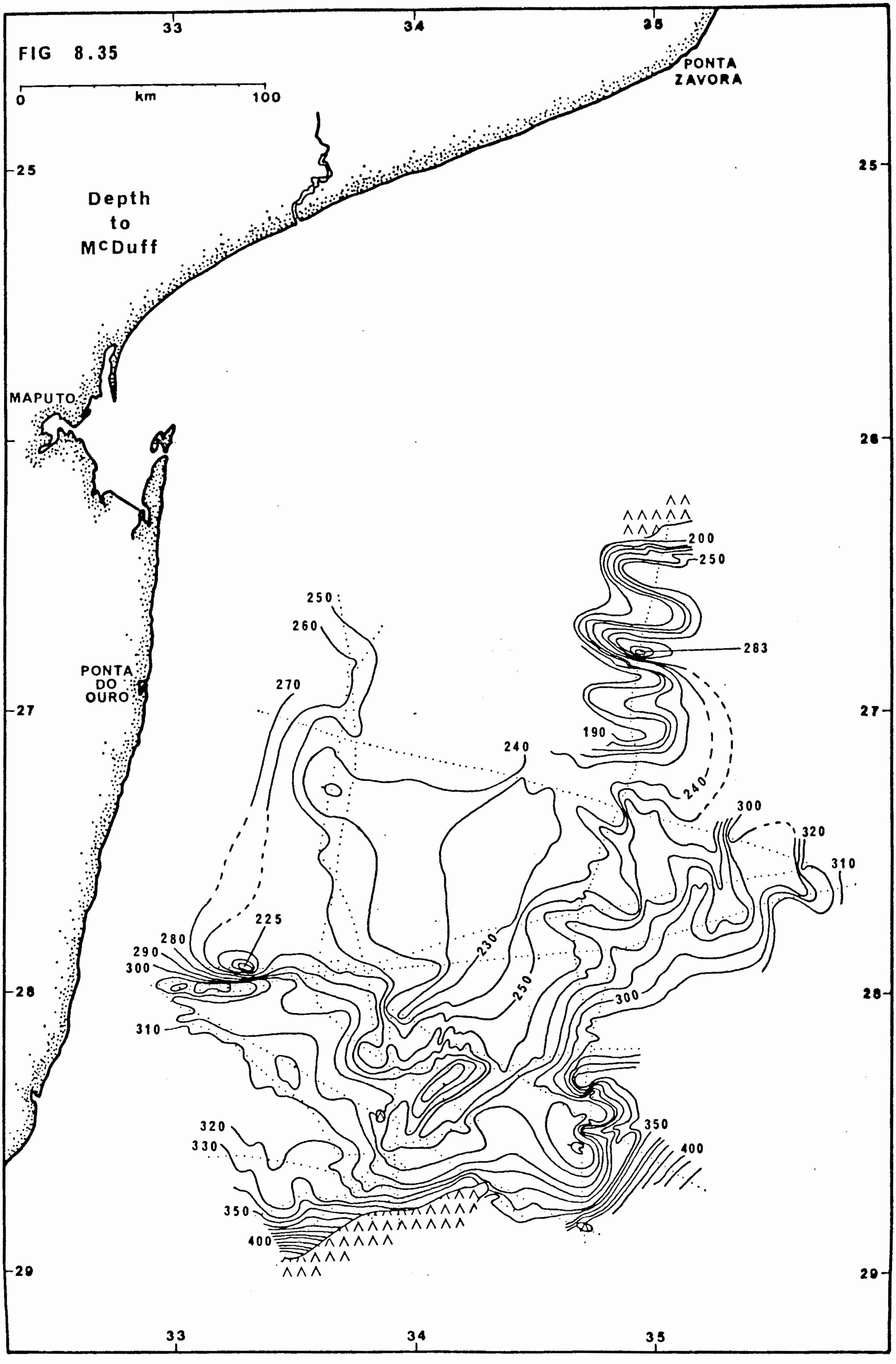


FIG 8.34
33
34
35
Zandamela
1768 m
PONTA
ZAVORA
0
km
100
25
Depth
to
Basement
3·0 3·8 2·4 1·8 2·1 2·7 3·3 2·6 2·7 3·3 3·7 km
3·5
2·0
2·2
1·8
2·4
MAPUTO
3·0 3·2 3·6 3·5 2·7
Almirante Leite Bank
10
26
73
125
150
175
175
200
PONTA
DO
OURO
27
225
250
275
300
325
350
375
253
28
350
290
375
325
350
400
283
375
425
444
29
321
263
Naudé Ridge

FIG 8.35



33
34
35
0
km
100
25
26
27
28
29
Depth
to
McDuff
MAPUTO
PONTA
DO
OURO
PONTA
ZAVORA
200
250
283
190
240
250
260
270
225
280
290
300
310
230
320
330
350
400

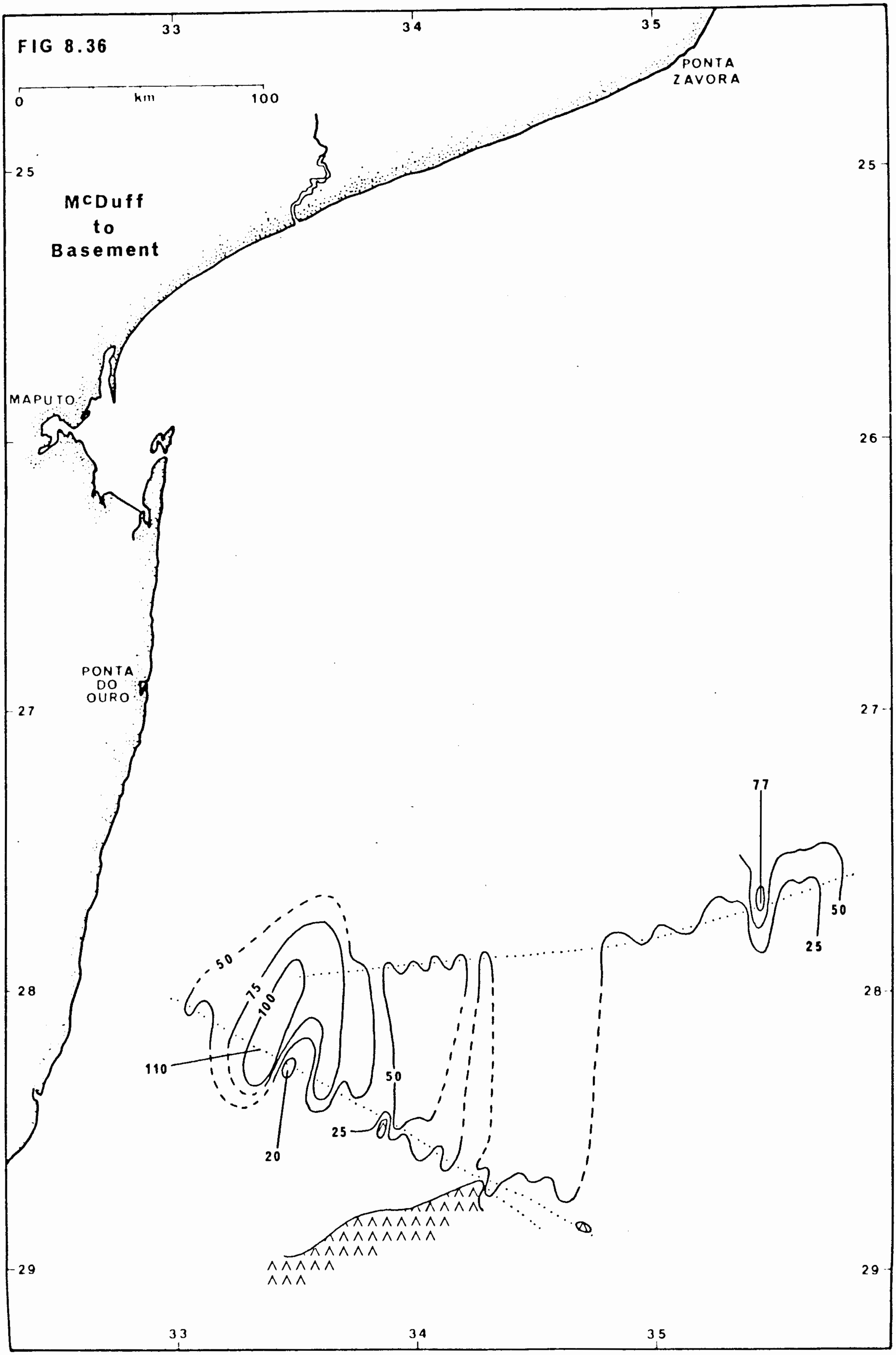


FIG 8.36
McDuff
to
Basement
0
km
100
33
34
35
25
26
27
28
29
PONTA
ZAVORA
MAPUTO
PONTA
DO
OURO
77
50
25
50
75
100
110
20
50
25

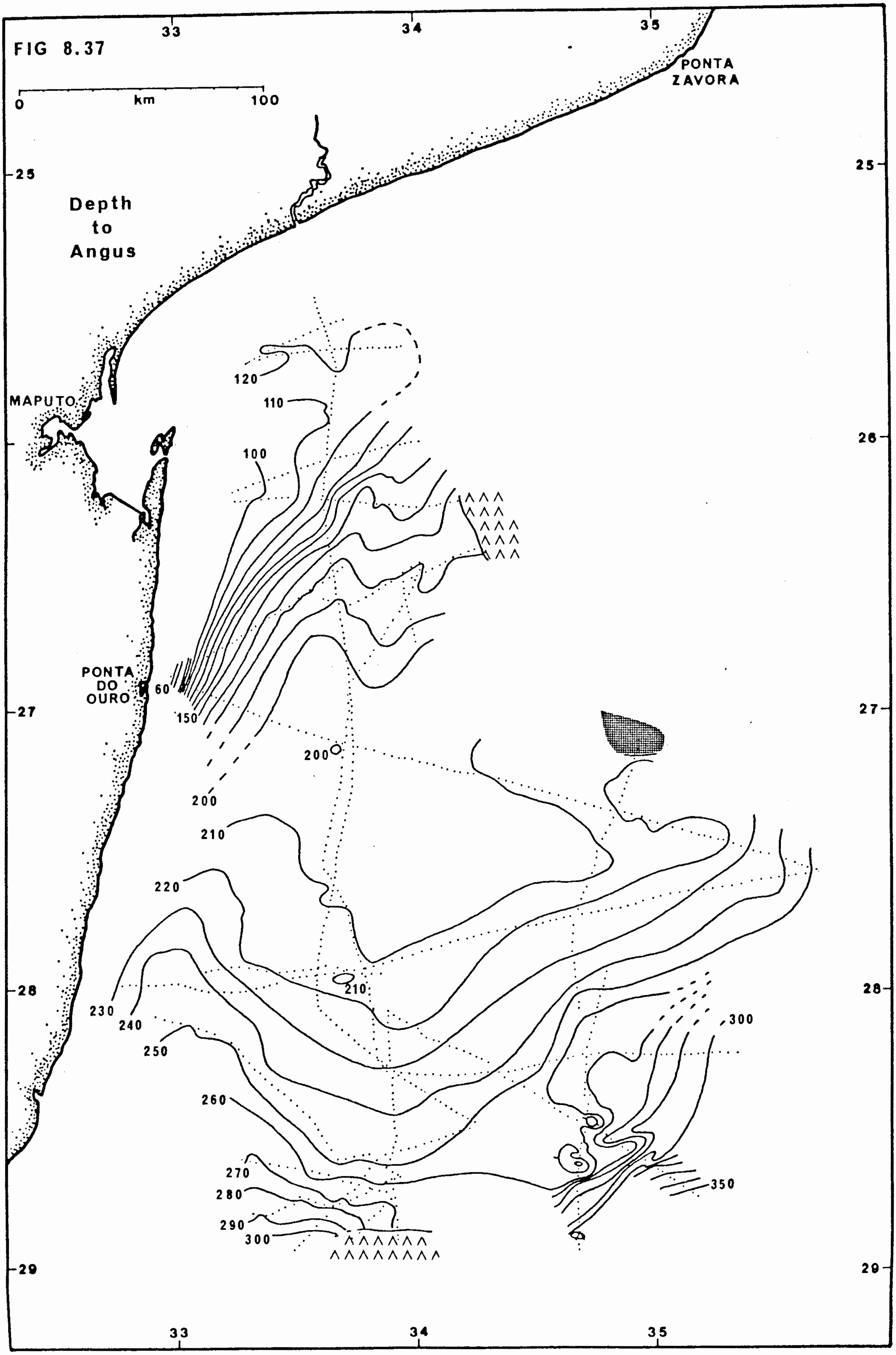


FIG 8.37
Depth
to
Angus
0
km
100
33
34
35
25
26
27
28
29
PONTA
ZAVORA
MAPUTO
PONTA
DO
OURO
60
100
110
120
150
200
200
210
210
220
230
240
250
260
270
280
290
300
300
350

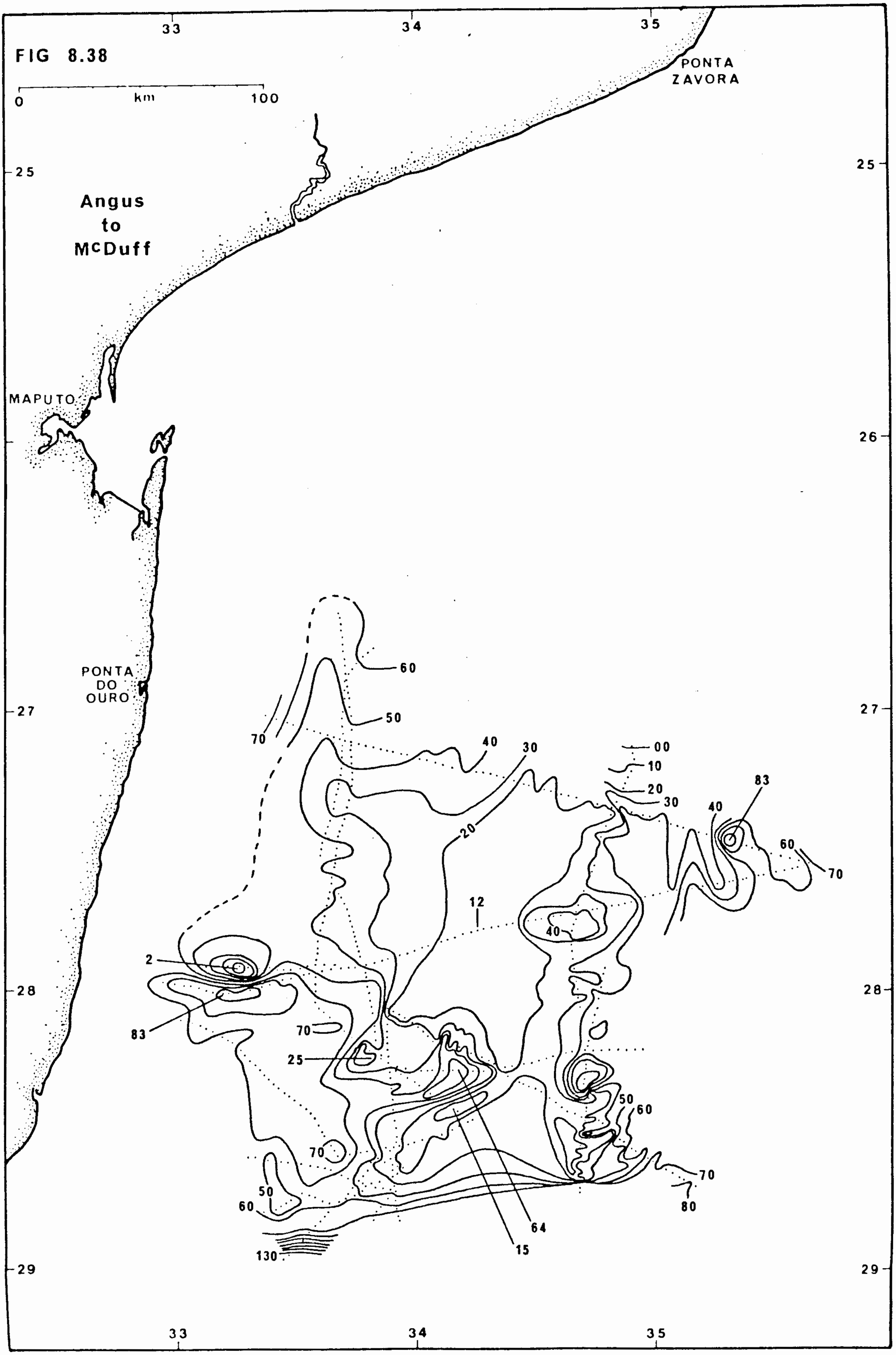

33
34
35
0
km
100
25
26
27
28
29
Angus
to
McDuff
PONTA
ZAVORA
MAPUTO
PONTA
DO
OURO
60
50
70
40
30
00
10
20
30
83
40
60
70
20
12
40
2
83
70
25
50
60
70
50
60
70
80
130
64
15

FIG 8.38

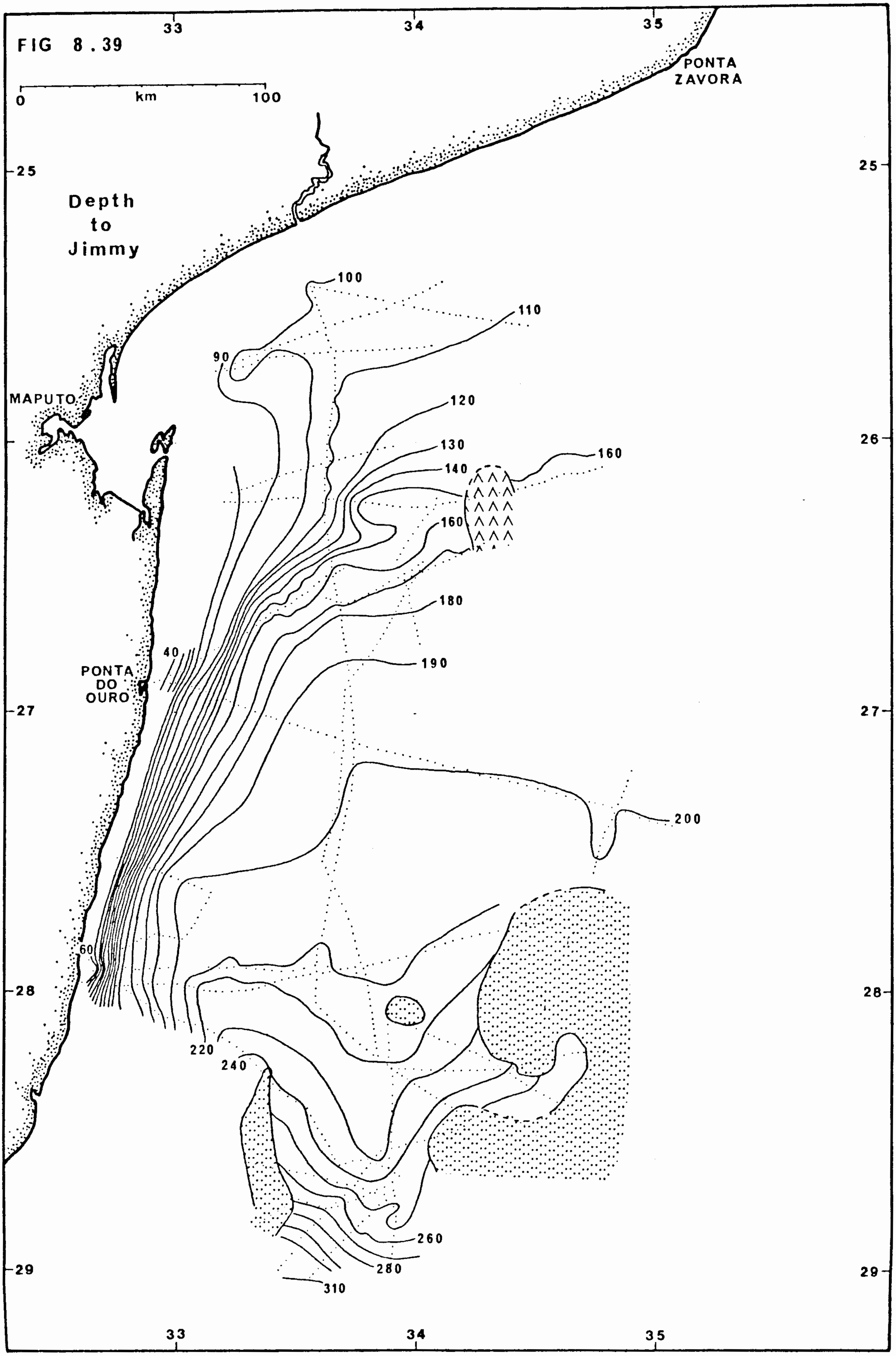
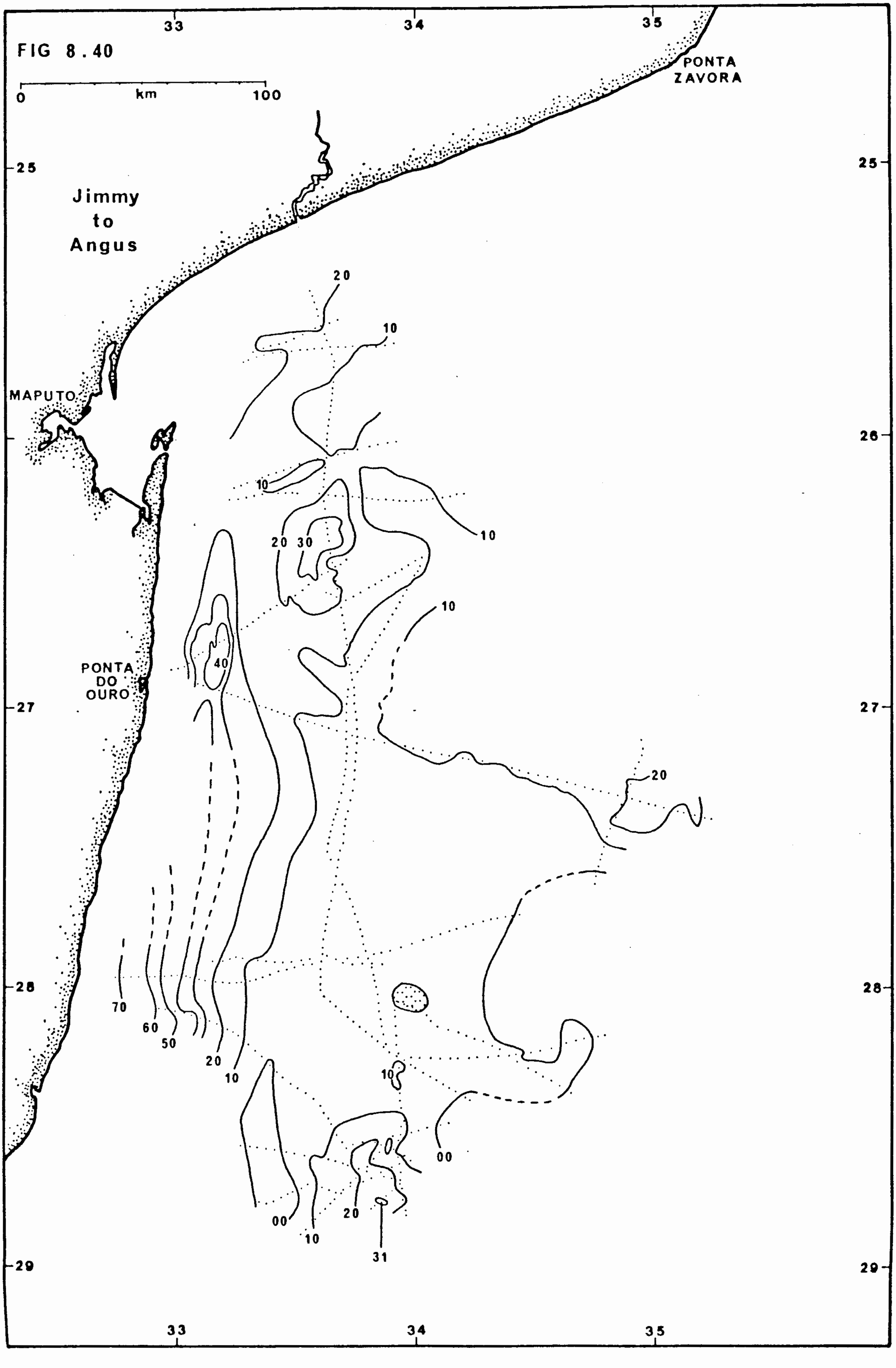


FIG 8.39
Depth
to
Jimmy
0
km
100
33
34
35
25
26
27
28
29
MAPUTO
PONTA
ZAVORA
PONTA
DO
OURO
40
60
90
100
110
120
130
140
160
160
180
190
200
220
240
260
280
310


Jimmy
to
Angus
0
km
100
33
34
35
25
26
27
28
29
MAPUTO
PONTA
DO
OURO
PONTA
ZAVORA
20
10
10
10
20 30
10
40
20
70
60
50
20
10
10
00
00
10
20
31
FIG 8.40

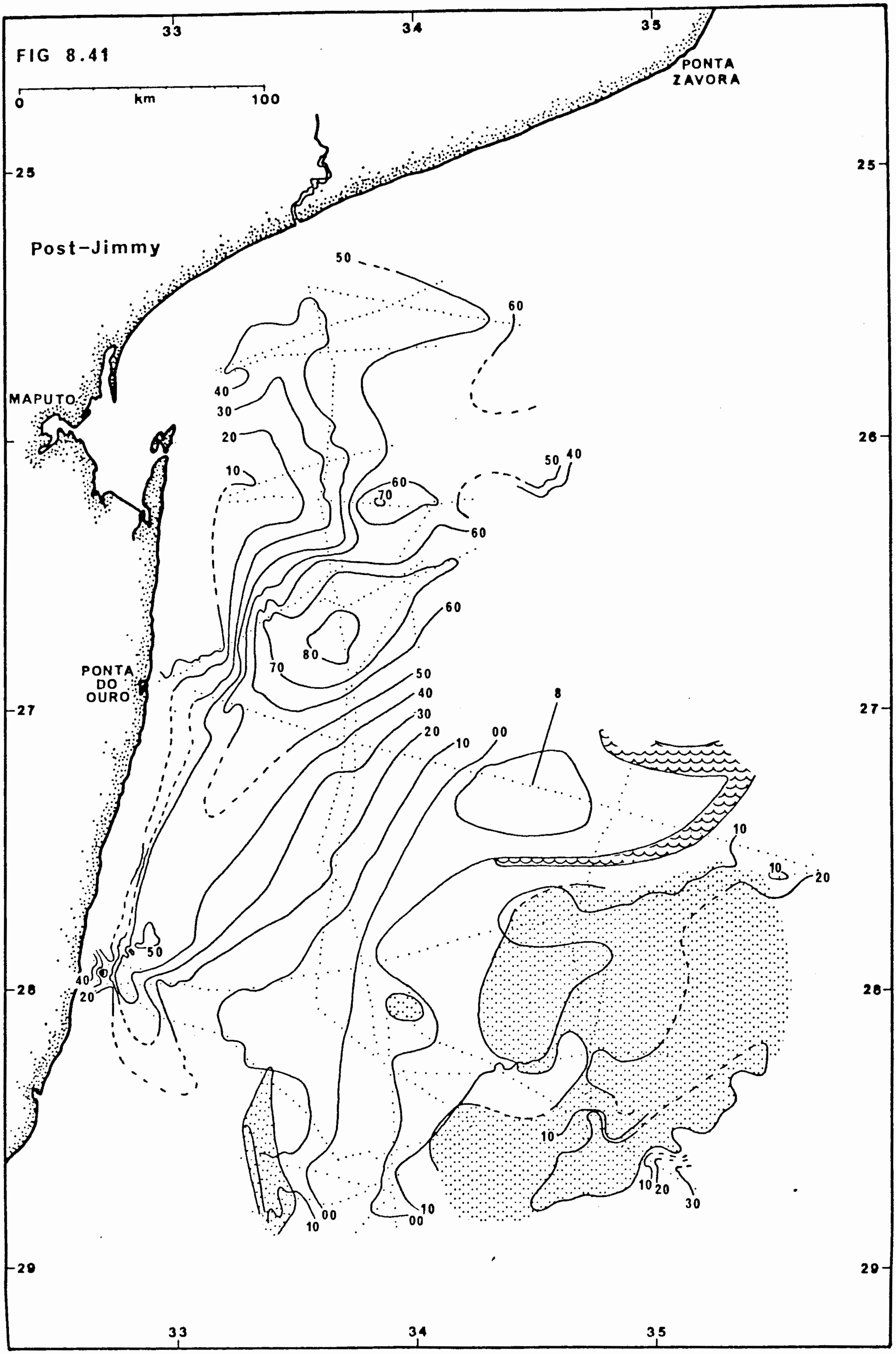


FIG 8.41
0
km
100
33
34
35
25
26
27
28
29
PONTA
ZAVORA
Post-Jimmy
MAPUTO
PONTA
DO
OURO
50
60
40
30
20
10
70
80
8
00

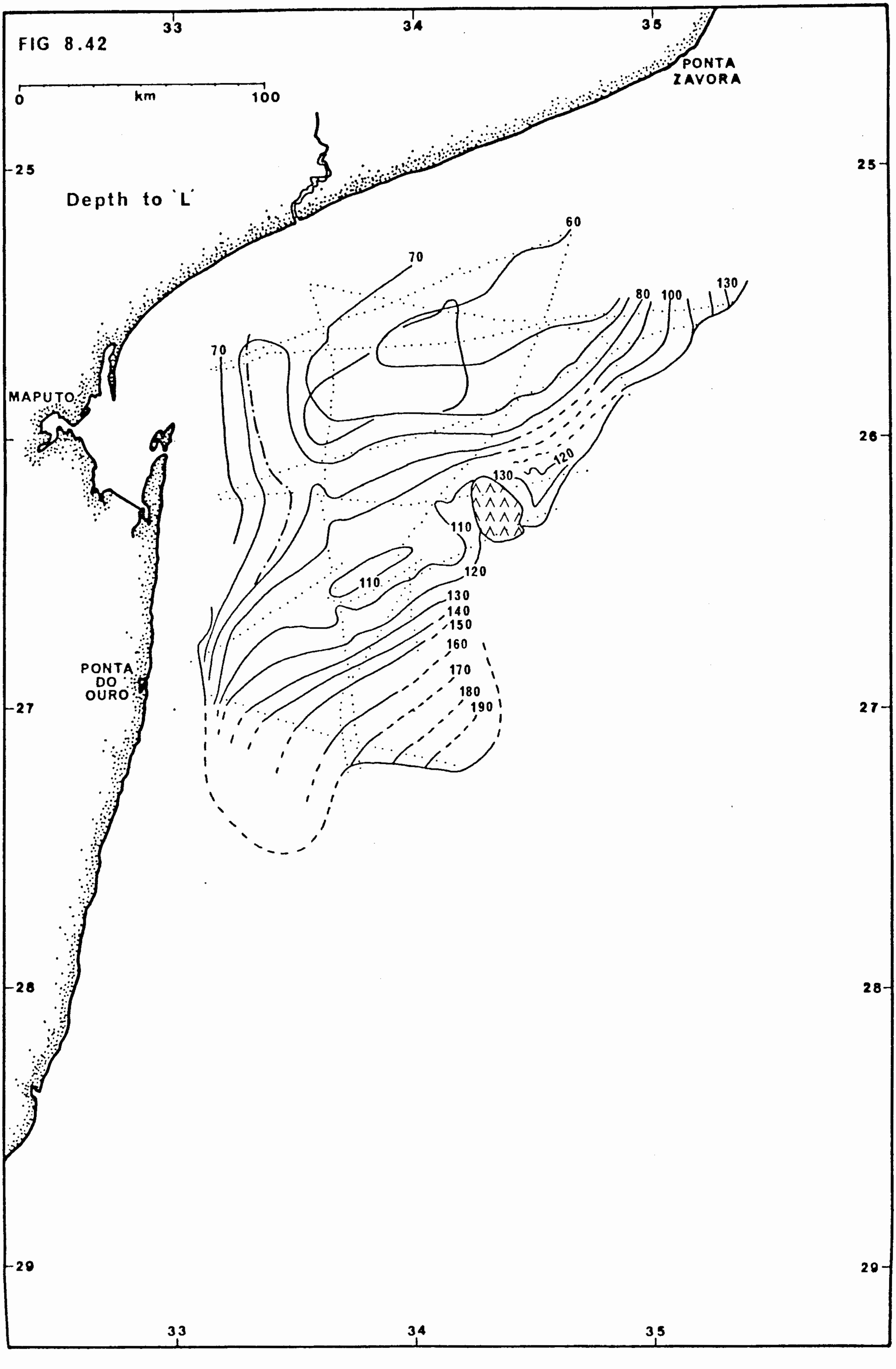


FIG 8.42
Depth to 'L'
0
km
100
33
34
35
25
26
27
28
29
PONTA ZAVORA
MAPUTO
PONTA DO OURO
60
70
70
80
100
130
120
130
110
110
120
130
140
150
160
170
180
190

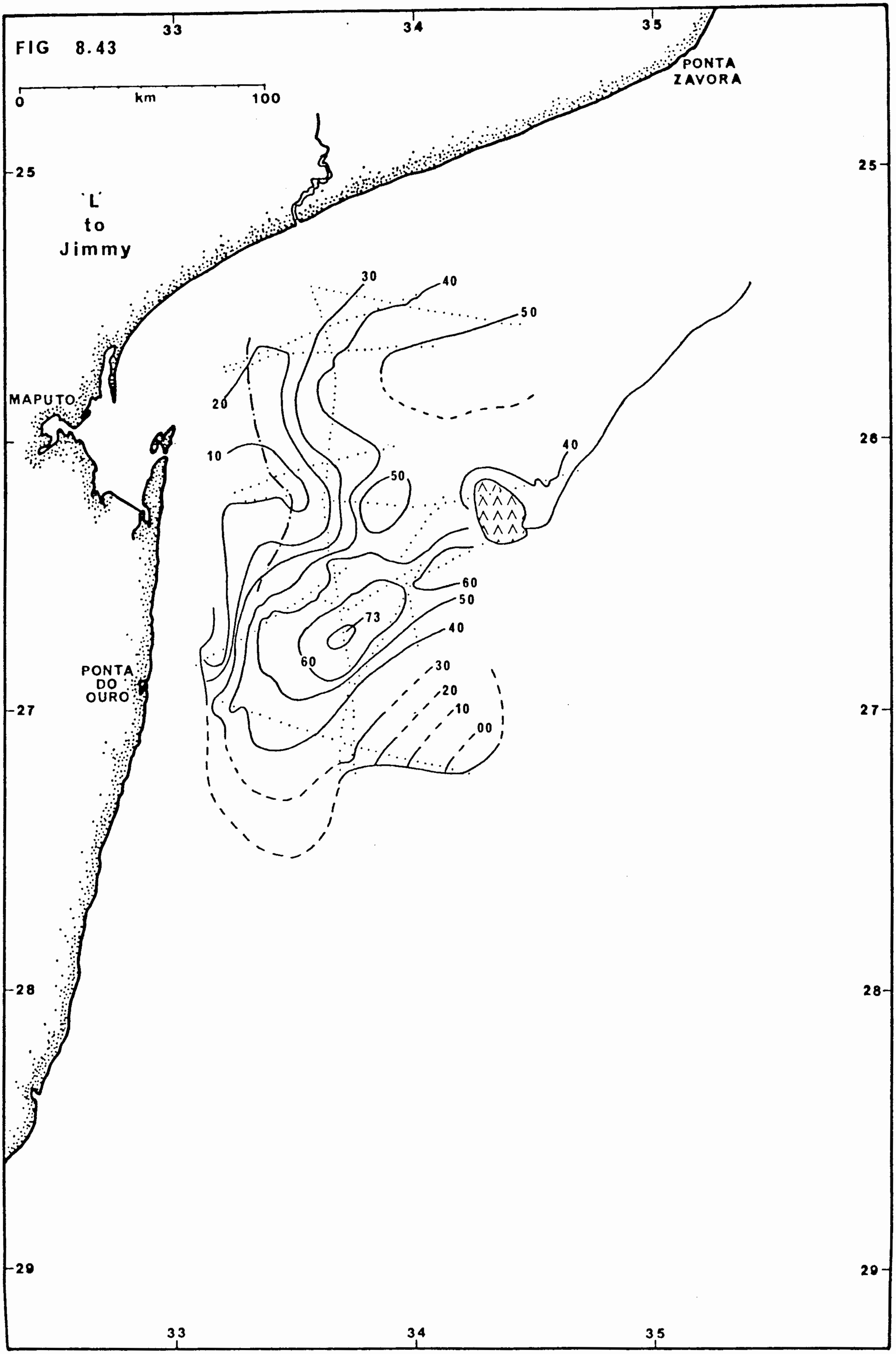


FIG 8.43
0 km 100
'L' to Jimmy
MAPUTO
PONTA ZAVORA
PONTA DO OURO
33
34
35
25
26
27
28
29
30
40
50
20
10
50
40
60
50
40
73
60
30
20
10
00

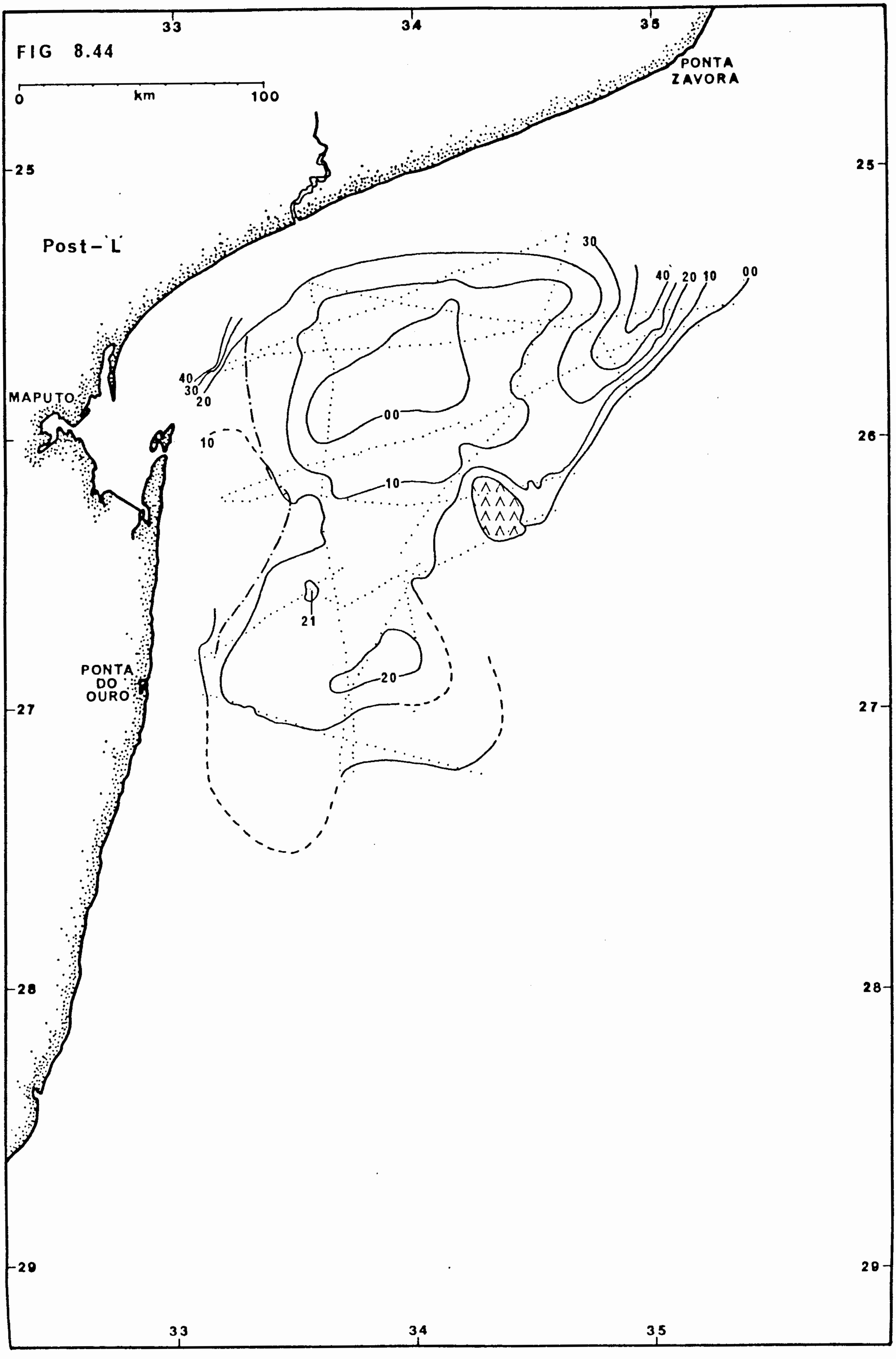
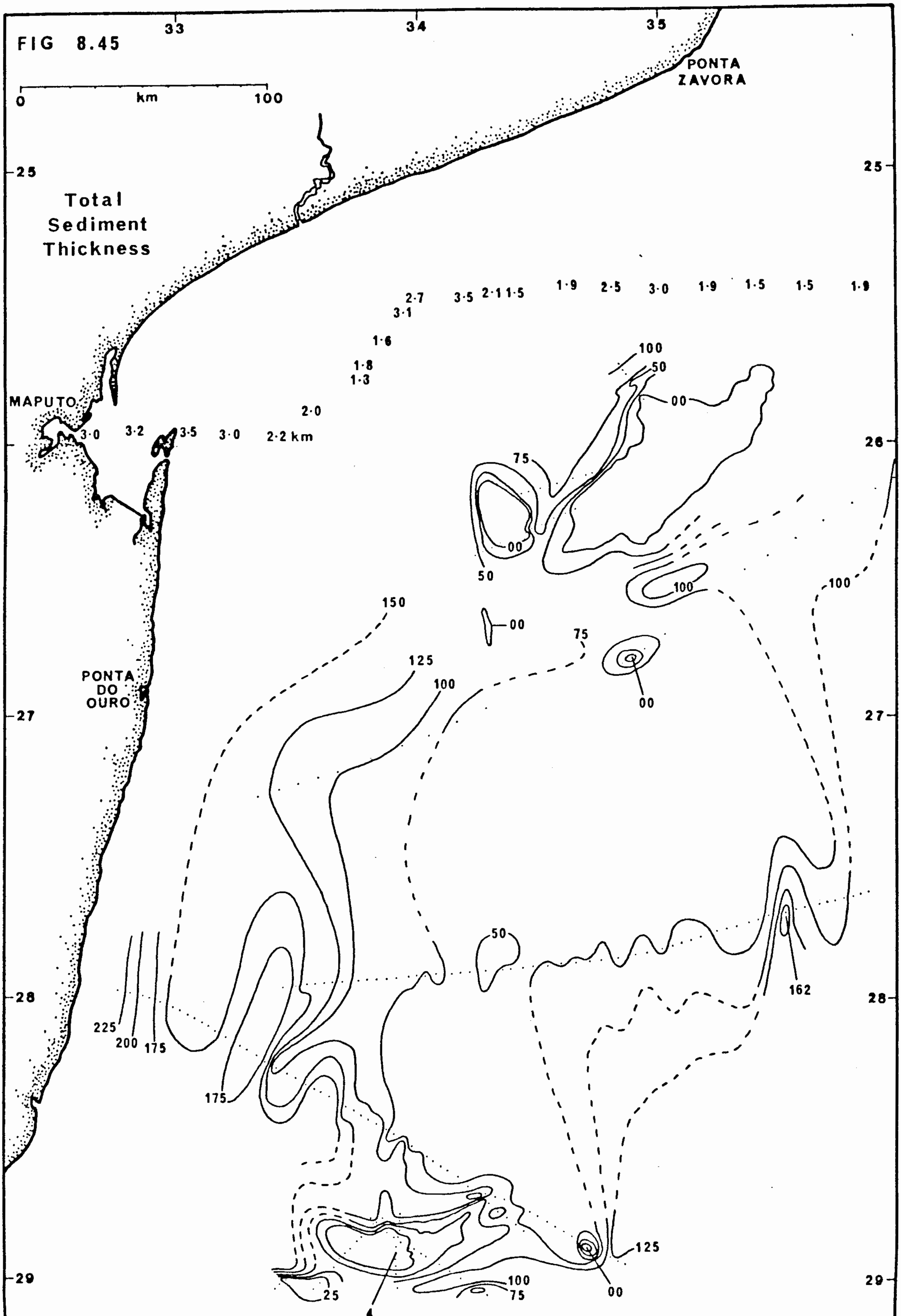


FIG 8.44
0
km
100
33
34
35
25
26
27
28
29
PONTA
ZAVORA
Post-'L'
MAPUTO
PONTA
DO
OURO
30
40
20
10
00
40
30
20
10
00
10
21
20

FIG 8.45



Total
Sediment
Thickness
0 km 100
33
34
35
25
26
27
28
29
PONTA
ZAVORA
MAPUTO
PONTA
DO
OURO
3·0 3·2 3·5 3·0 2·2 km
2·0
1·3
1·8
1·6
3·1
2·7
3·5 2·1 1·5
1·9 2·5 3·0 1·9 1·5 1·5 1·9
100
50
00
75
50
00
100
100
150
00
75
00
125
100
50
162
225 200 175
175
125
100
75
25
4
00

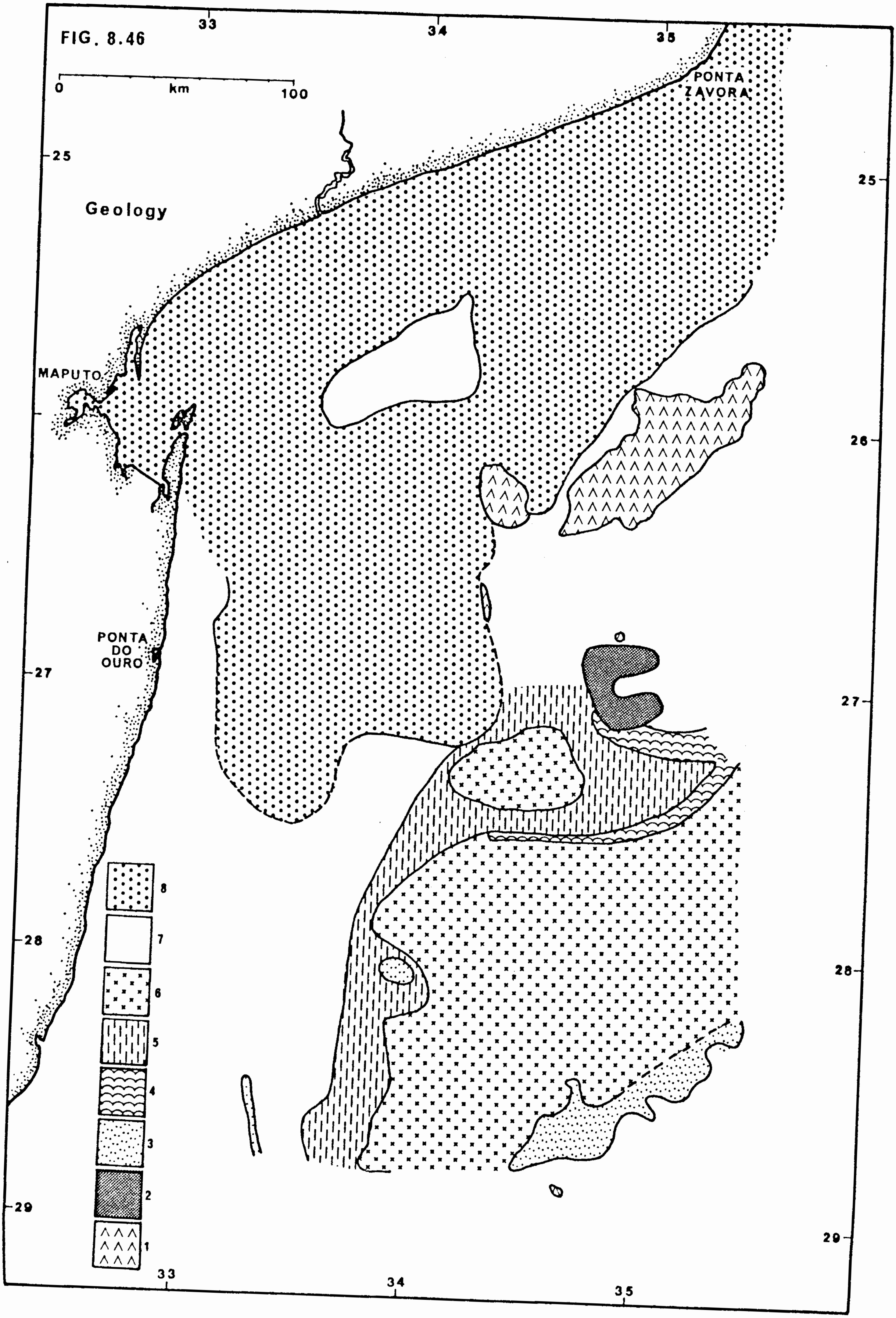


FIG. 8.46
Geology
0
km
100
33
34
35
25
26
27
28
29
MAPUTO
PONTA ZAVORA
PONTA DO OURO
8
7
6
5
4
3
2
1

is provided by multi-channel profiles N805, N805A and N806 (Figs.8.33 & 8.36). Unit 1 attains its greatest known thickness of 1.1 sec in the elongate depression on the northern side of the Naudé Ridge at 34° 25'E. West-dipping divergent reflectors on profile N806 and east-dipping divergent reflectors on profile N805 indicate increased rates of deposition and possibly contemporaneous subsidence in the depocentre axis. A second depocentre on the eastern Central Terrace is infilled by 0.77 sec of unit 1 sediments. On the central and southern Central Terrace infill is irregular and thinner. Unit 1 infills basement depressions through onlap and divergent fill (Fig.8.33). The closer association of McDuff and basement in the south-east of the region may be linked to the increased reflectivity of McDuff on profiles MN3 (0800-1100 hrs) and 371 4 (0300-1700 hrs)(Figs.8.19 & 8.32).

There is considerable variation of Early Cretaceous rock types drilled on the Mozambique Ridge (Simpson, Schlich et al., 1974). Volcanic ash-rich layers with up to 95% volcanic material alternate with silty claystones and limestones. The high contrast in transmitted sonic velocity (Christensen et al., 1974) probably accounts for the high reflectivity of this unit in seismic profiles. A similar lithologic scenario may account for similar high reflectivity of pre-McDuff sediment. This premise is qualitatively supported by a) a possible volcanic mound structure on profile 521 (Fig.8.16); b) the resemblance of McDuff to the volcanic-debris-strewn seafloor around seamounts of the Almirante Leite Bank. Faulting, syn-sedimentary subsidence suggested by depositional style, and volcanism evince an active tectonic regime.

8.3.3 Angus

Reflector Angus is a prominent unconformity which shows erosional truncation of well laminated beds below, and downlap of the overlying strata (particularly profiles 350 4, 350 3, 284, 283, MN2, Figs.8.13, 8.17, 8.21, 8.23 & 8.31). Although less reflective than McDuff, some areas of Angus exhibit increased reflectivity (bright spots, eg. profile 350 13 1730-1930 hrs and 2200-0000 hrs). In contrast to the relief shown by McDuff, reflector Angus is relatively smooth and gently dipping. By the time of its formation (Oligocene - section 8.2), the Limpopo Cone was already established in the north-western Natal Valley (Fig.8.37). The present-day valley on the western side of the Limpopo Cone (section 2.3) appears to have an Oligocene counterpart (profile 350 15, Fig.8.8 & Fig.8.37). From a depth of $\sim$1.0 sec Angus descends south-eastwards and southwards to 2.0 sec on the northern Central Terrace at a slope of 2°. On the Central Terrace, Angus is almost flat-lying except for some mounded relief to the east (profiles 350 4, 350 3, 350 5c, MN3 and 371 4, Figs.8.13, 8.17, 8.18, 8.19 & 8.32). These features form the main outcrops of Angus (Figs.8.19 & 8.32) whereas on the Central Terrace, a small area of pre-Angus sediment is exposed through erosion and/or non-deposition (profiles MN3 and MN2, Figs.8.19 & 8.31). A smaller Angus outcrop results from erosion in a V-shaped canyon at the base of the Zululand continental slope. To the south, Angus descends and on-laps basement (profiles 519 & MN2, Figs.8.30 & 8.31), whereas to the south-west Angus tops a subsidiary high of the Naudé Ridge (profile 371 8, Fig.8.29). To the south-east, the mounded pre-Angus sediment overlies very reflective areas of McDuff considered to closely overlie basement (profiles MN3 & 371 4, Figs.8.19 & 8.32).

8.3.3.1 Sediment Unit 2: McDuff-Angus

The basal part of unit 2 typically comprises well-laminated sediments having wavy parallel reflections. Faulting is very common on the Central Terrace where faults affecting McDuff have been repeatedly reactivated. Fault throws decrease significantly from the McDuff marker horizon to Angus (eg. throw decreases from 0.25 sec to 0.05 sec, 1830 hrs, profile 371 4, Fig.8.32). The variation in throw and increased bed thickness across a fault indicates faulting contemporaneous with sedimentation (growth faults - chapter 10). Faulting and disruption of reflectors is so common as to cause a "pseudo-megaripple" effect or regular wavy aspect to reflectors (eg. profile 371 8, Fig.8.29). A similar effect off the Mississippi has also been ascribed to frequent faulting by Stuart & Caughey (1977, Fig.13). The well-laminated sequence in-fills graben of McDuff age through on-lap, with some thickening of beds towards graben axes. This faulted on-lap configuration in graben has been called 'up-lap' by Brown & Fisher (1977, Fig.16; compare profile 371 7, Fig.8.20). More locally, at the base of the sequence (profiles 350 4, 284 and 350 5b, Figs.8.13, 8.21 & 8.28) some progradation to the east and south from the western margin of the basin has occurred.

Well-laminated on-lap or up-lap fill is superceded by a sequence of lenses and mounded fills (profiles 350 4, 521, 371 6, 350 5b, Figs.8.13, 8.16, 8.25 & 8.28). Inter-profile correlation shows that lenses of sediment in N-S oriented profiles correspond with the westerly sigmoidal progradation from the Central Terrace into a trough at 33.5°E (profile 350 3, 1000-2200 hrs, Fig.8.17). This configuration suggests a fan complex. Unit 2 sediments also prograde from the Central Terrace towards the valley to the east (profile MN3, Fig.8.19). On the Central Terrace, overlying the laminated on-lap fill and fan complex material, is a thin unit displaying small scale irregular or hummocky bedding (profiles MN3, 371 7, & MN2, Figs.8.19, 8.20 & 8.31). This completes unit 2 on the central area of the Central Terrace, while to the east it grades into and is overlain by thick (0.55 sec) irregular mounds. The asymmetry of these features suggest differential sedimentation under geostrophic current influence (contourite configuration of Mitchum et al., 1977 and Sangree & Widmier, 1977). These features unconformably overlie prograded laminated material (profile MN3, Fig.8.19) and outcrop in part on the seafloor (Fig.8.46).

The differential fill and various facies displayed by unit 2 is reflected in the isopachyte chart (Fig.8.38). Unit 2 thickness is very variable and in general mirrors the configuration of McDuff (Fig.8.35). Deposition is least on the upfaulted blocks of the Central Terrace and over the (volcanic?) mound on profile 521. The adjacent troughs have accumulated up to 0.83 sec of sediment mainly through on-lap fill associated with growth faults (up-lap). Maximum thicknesses reach 1.35 sec immediately north of the Naudé Ridge which has acted as a dam to southerly sediment transport. Sediment thickness increases both eastwards and westwards from the up-faulted Central Terrace. To the west, increased sedimentation is due to progradation from the west and north, and the north-east. To the east, greater sediment accumulation is related to progradation on to the eastward dipping faulted McDuff surface and to mounded contourite sedimentation.

8.3.4 Jimmy

Reflector Jimmy is a prominent unconformity which outcrops widely over the Central Terrace (Figs.8.40 & 8.46). To the south it lies at the top of a parallel laminated sequence which unconformably overlies the lower part of unit 3 (profiles 284, 520, 283, 371 8, 519, Figs.8.21, 8.22, 8.23, 8.29 & 8.30). In the north it marks a major change in construction of the Limpopo Cone (profiles 350 5a, 371 6, Figs.8.12 & 8.25). Off the Zululand coast around 28°S the unconformable nature of Jimmy is related to its exploitation as a slump glide plane (chapter 10).

The structure contour chart for Jimmy (Fig.8.39) shows similar features to the Angus chart (Fig.8.37). The Limpopo Cone which infills the north-western part of the basin, has expanded to the south-east. Along the Zululand coast, the continental margin has prograded eastwards. In this region Jimmy slopes eastwards at 5°, while in contrast, the Central Terrace is gently-sloping. The configuration of the Central Terrace is only slightly modified because unit 3 is thin or absent in this region: Jimmy slopes both to east and west from a central high. This southward facing convexity is sharper than in Angus times (Fig.8.39 c.f. Fig.8.37) because of the eroded outcrop of unit 3 sediment on the southern Central Terrace.

8.3.4.1 Sediment Unit 3: Angus-Jimmy

The most important depocentre for unit 3 sediment lies along the Zululand and Mozambique continental margin where up to 0.7 sec of sediments accumulated in a valley between the continental slope and the Central Terrace (Fig.8.40). In-fill of this valley was complex with oblique parallel prograded material transported from the east or north (profiles 350 3, & 284, Figs.8.17 & 8.21) as well as from the west. In other areas, sediment units appear to on-lap the continental slope, but this may be due to an older Oligocene/Miocene phase of slumping (chapter 10; profiles 285, 521 & 284, Figs.8.15, 8.16 & 8.24). On profile 350 4 (Fig.8.13) the toe of the slope-front fill is gently mounded. The general scenario appears to comprise low-energy slope-front fill with high-energy coalescing fan complexes probably emanating from canyons on the continental slope.

To the north, this depocentre extends towards the southern face of the Limpopo Cone, where two main areas of sedimentation occur. On profile 350 13 (Fig.8.10) irregularly bedded material progrades across the existing Limpopo Cone topography, spilling over into the trough between the Limpopo Cone and the Almirante Leite Bank. Both up-building and outbuilding has occurred on the outer slope of the Cone, with sedimentation rates increasing on the downthrown side of contemporaneous faults (profile 371 6, 0600-1100 hrs, Fig.8.25). On the upper, western surface of the Limpopo Cone a thick sediment pile has accumulated through complex sigmoid oblique progradation accompanied by faulting and slumping (profiles 350 13, & 350 5a (Figs.8.10 & 8.12)). To the east of this, unit 3 sediments thicken towards and onlap basement, with some current influence suggested by the shallow moat around the flanks of a basement high (Profile 350 13, Fig.8.10). This profile also displays a bright spot on reflector Jimmy (1630-1830 hrs). To the north, unit 3 thickens again off Maputo and the Limpopo River. Material appears to have come from both the east and west to infill a shallow valley at the base of the continental slope.

From the main depocentres on the northern and southern Limpopo Cone and the Zululand continental slope, unit 3 thins eastwards until it pinches out against Angus on the Central Terrace. To the east and south-west of the Central Terrace, unit 3 was either not deposited or has subsequently been eroded. Two depocentres of unit 3 do exist on the Central Terrace (Fig. 8.40). On the north-east Central Terrace, parallel-laminated sediments thicken to the north-east and have accumulated in fault bounded graben (profile 371 4, Fig.8.32). Early Miocene sediments are exposed on the southern up-faulted margin of this depocentre and in a scoured moat on the flank of an up-faulted outcrop of McDuff. A second depocentre exists on the southern Central Terrace where unit 3 thickens north of the Naudé Ridge. A parallel-laminated unit appears to drape the Naudé Ridge and its partial post-Angus cover (284, 520, 283, 371 8, 519 & MN2, Figs.8.21, 8.22, 8.23, 8.29, 8.30 & 8.31). Mounded sedimentation, and shallow moat-like depressions suggest current influence in the region of the Naudé Ridge (profile 519, Fig.8.30). Unit 3 is extensively eroded in this region, with various pre-Jimmy reflectors outcropping, particularly on the upper slopes of the valley to the west and south-west of the Central Terrace. This evinces potent current action, probably accompanied by slumping into the valley floor.

8.3.5 Post-Jimmy sediments; Units 4 & 5

Units 4 & 5 are separated by reflector 'L' which is recognised only on the Inharrime Terrace and the Limpopo Cone. No tie-line is presently available with which to check the correlation of reflector 'L' from the Limpopo Cone to the southern Zululand continental slope. On the basis of sediment recovery in cores (Fig.8.2) post-'L' material appears not to extend southwards of 28° 30'S (Fig.8.46). On the Central Terrace, it is not possible to differentiate units 4 and 5, since post-Jimmy sediments are everywhere separated from the Limpopo Cone sequence by outcrops of pre-Jimmy strata.

8.3.5.1 Undifferentiated Unit 4 & 5 sediments

On the east side of the N-S trending outcrop of Jimmy, on the southern Central Terrace, a parallel well-laminated sequence dips to the east very gently. This sequence fills the existing relief through on-lap fill (profiles 350 3, 350 5c, MN3, 371 7, 371 8 & MN2, Figs.8.17, 8.18, 8.19, 8.20, 8.29 & 8.31). The existing relief includes a fault scarp on reflector Jimmy of 0.02 sec (16 m) (profile 371 7, 0315 hrs). The post-Jimmy sequence is generally not affected by faults except immediately north of the Naudé Ridge (profile MN2, Fig.8.31). Because reflector Angus sub-crops on the Angus to Jimmy sequence, the post-Jimmy well-laminated material sometimes lies directly on Angus (profiles 350 5c, MN3 & 371 4, Figs.8.18, 8.19 & 8.32). In the eastern extremes of profiles 350 4, 350 3, 350 5c & MN3, it is uncertain whether reflector Jimmy remains at the stratigraphic level of Angus, or whether an Angus-Jimmy sequence is developed. However on profiles 350 5c and MN3, a well-laminated sequence, similar to that described above thickens into the valley to the east of the Central Terrace. Two other facies are displayed in this region: a) complex fill between the pre-Angus contourite mounds also displays current influence through migrating moats on the flanks of the mounds (profile MN3); b) south-east and east of the up-faulted block of Early Miocene sediment, lies a sequence of jumbled irregular sediments overlain by parallel-laminated east-dipping sediment (profiles 350 4 & 350 3, Figs.8.13 & 8.17). Note that the parallel-laminated material outcrops at the seafloor.

On the northern Central Terrace (profile 350 4, Fig.8.13) a thin (0.08 sec ∿64 m) though extensive (53 km wide) mound of sediment lies on a broad outcrop of reflector Jimmy. Despite its gentle relief (0.14°) the feature shows classic 'contourite' reflector configuration (sensu Mitchum et al., 1977; Sangree & Widmier, 1977). Its formation under the influence of geostrophic currents is substantiated by the re-worked nature of its surface sediments (Fig.8.5).

In contrast to the Central Terrace, up to 0.85 sec of post-Jimmy sediment has accumulated in a linear depocentre extending along the Zululand continental slope north-eastwards via the southern Limpopo Cone to the Inharrime Terrace (Fig.8.41). Thick piles of post-Jimmy material also accumulated off the Limpopo River, whereas east of Inhaca Island off Maputo, and along the upper slope to the south, units 4 and 5 are thin or absent. The general configuration of post-Jimmy sedimentation is similar to unit 3 deposition (Fig. 8.40) and represents a basin-ward progradation.

8.3.5.2 Reflector 'L'

'L' lies at the top of a sequence of correlatable reflectors underlying the Limpopo Cone and Inharrime Terrace. It marks the top of the major constructional phase of the Limpopo Cone. Unconformably-overlying post-'L' sediments are concentrated on the shallow continental slope off Maputo and on the Inharrime Terrace (profiles 371 5, 350 15 & 371 6, Figs.8.7, 8.8 & 8.25). Reflector 'L' outcrops centrally on the northern Limpopo Cone, while Jimmy-'L' reflectors outcrop on the eastern flank of the Inharrime Terrace and on the southern face of the Limpopo Cone. By 'L' times, the northern and north-western basin had been in-filled. Unit 4 sediments are dammed behind the northern Almirante Leite Bank, while elsewhere 'L' onlaps basement highs in a moated configuration (contrast profile 371 4, Fig.8.32, with profiles 350 13 & 350 12, Figs.8.10 & 8.11). On the lower continental slope off Inhaca Island, 'L' laps out against a lower sequence of Unit 4 (see below).

8.3.5.3 Sediment Unit 4: Jimmy-'L'

Unit 4 sedimentation can be divided into two main facies: (i) a lower sequence dominated by large-scale slumping and mass transport processes; (ii) a partly contemporaneous and partly overlying sequence comprising a large slope wedge (sensu Galloway & Brown, 1973).

8.3.5.3.1 Slump facies

Unit 4 sediments on the Zululand continental slope show many of the classic features of large submarine slumps (Lewis, 1971; Coleman & Garrison, 1977; Dingle, 1977; chapter 10). In the proximal region of the slump, these include main and subsidiary glide-planes, glide-plane scars, fissured areas, and tilted strata (Figs.10.1 & 10.2). Features in the distal slump include chaotic bedding, folds, low-angle thrusts and hummocky relief (profiles 286, 285, 521, 350 3, 284 and 287, Figs.8.14, 8.15, 8.16, 8.21 & 8.24). Within the slump, basal material appears more reflective (profiles 521 & 350 3, Figs. 8.16 & 8.17). The slump is detected by several profiles to the south of the region covered here, and extends from the Zululand continental slope to the northern Tugela Cone. It is either one very large feature of the order of magnitude described by Dingle (1980a), or comprises a series of individual adjacent slumps not resolved by the available track coverage. Because of the active valley at the base of the slope, it is difficult to discern the relationship between the toe of the slump and thin eroded post-Jimmy sediments dipping west on the south-western Central Terrace (profiles 284, 520 & 283, Figs.8.21 - 8.23). On the slopes of the lower reaches of this valley, south-west of the Naudé Ridge, relatively steep-dipping (1.5°) material may also be partly post-Jimmy in age (profiles 371 8 & 519, Figs.8.29 & 8.30). Some irregular bedding suggests material eroded or swept from up-slope was deposited in the valley via mass transport. In the valley, erosion and re-deposition is suspected via turbidity current action.

The toe of the slump is lens-shaped in strike-section (profiles 371 6 & 350 5b, Figs.8.25 & 8.28). The slumped nature of the material is evinced by irregular micro-relief and chaotic internal reflectors. (This was most clearly evident in a higher band-pass frequency recording (appendix 1)). Towards the north, slumped material is overlain by the upper part of unit 4 and unit 5 indicating that the slump is pre-'L' in age (profiles 350 4 & 371 6, Figs. 8.13 & 8.25). A thin sequence of reflective jumbled strata lies on the top-set surface of the pre-Jimmy Limpopo Cone just down-slope from what appears to be a glide-plane scar (profile 350 5a, Fig.8.12). Similar reflective, faulted, and jumbled strata indicative of down-slope mass transport exist all along the slope off Maputo (Fig.10.1, profiles 350 16, 350 15, 350 14 & 350 13, Figs.8.6, 8.8, 8.9 & 8.10). Sediment instability is also evinced by a broad (12 km) diapir of >80 m relief (profile 350 16, 2350-0045 hrs, Fig.8.6; compare Coleman & Garrison, 1977, Fig.3). As the only available tie-line lies just east of the slump (profile 371 6, Fig.8.25) it is at present impossible to determine whether the slump off Maputo is part of, or the same age as the large slump mapped on the Zululand margin.

8.3.5.3.2 Slope wedge facies.

On the southern face of the Limpopo Cone and on the Zululand continental margin, the slumped sediments are overlain by well-laminated sediments on-lapping the existing continental slope and Limpopo Cone (profiles 350 4, 371 6 & 350 5b, Figs.8.13, 8.25 & 8.28). In contrast, on the top-set surface of the Limpopo Cone, a thick lens of well-laminated unit 4 (Jimmy-'L') is partly contemporaneous with and partly overlies slumped sediments (profiles 350 16, 350 15, 350 14 & 350 13, Figs.8.6, 8.8, 8.9 & 8.10). Two major lobes are evident (Fig.8.43): the northern lobe is a WSW-oriented thick lens centred at 25° 45'S; the southern lobe lying on the south-eastern face of the pre-Jimmy Limpopo Cone is SW-oriented and is best developed between 26° and 27°S. Individual beds within the northern lobe are thickest in the central area, with reflectors converging towards the flanks. The resultant deposit is 'fan', 'wedge' or 'half-cone' shaped (Normark, 1970; Galloway & Brown, 1973; Rupke, 1978). The southern lobe progressively on-laps the existing Limpopo Cone, while down-slope, beds down-lap or pinch out towards the Central Terrace. The general configuration is a mounded type of slope-front fill.

Many reflectors become much more reflective close to the position of onlap (profiles 350 5a & 371 6, Figs.8.12 & 8.26). Apart from slumped beds which interfinger with the wedge (section 8.2.5.3.1), and some minor beds which thin towards the east and pinch out on the slope (profile 350 4, Fig.8.13) the main source of the submarine fan appears to be from the north-east around Ponta Zavora. A 3 km wide channel with associated levees up to 0.04 sec in relief (32 m) occurs in reflector 'L' (profile 350 13, 1345 hrs, Fig.8.10) while some channel migration and in-fill occurs under the Inharrime Terrace in upper unit 4 sediments (profile 350 17, Fig.8.26, 1200 hrs). This, in conjunction with the geographic position of the sequence, and by analogy with the post-'L' sediments of the Inharrime Terrace, suggests the 'slope submarine fan' or 'slope wedge' depositional model (Galloway & Brown, 1973; Rupke, 1978). Moats and mounds which are prominent around seamounts, are indicative of continued geostrophic current action. The sequence is characterised by various manifestations of sediment instability. Tensional graben of up to 45 km across, with fault throws of up to 0.025 sec (20 m) occur on the northern lobe. Increased throws with depth evince contemporaneous faulting and sedimentation. On the eastern flank of the Inharrime Terrace, pre-'L' sediments have slumped eastwards producing chaotic beds near the slump-toe. On the southern lobe, on the southern face of the Limpopo Cone the mounded nature of the beds is attributable to a) the mode of deposition, with beds thinning peripherally; b) down to basin growth faulting with concomitant collapse near the main fault resulting in a roll-over anticline (profile 371 6, Fig.8.25, 0930-1300 hrs).

8.3.5.4 Sediment Unit 5: post-'L'

Comparison of the structure contour chart of 'L' (Fig.8.42) with the present-day bathymetry (section 2.3), shows that in post-'L' time, the Inharrime Terrace has accreted and the continental margin off the Limpopo River has prograded. The Inharrime Terrace is a thick wedge or 'half-cone' of well-laminated reflective sediment, displaying a prominent set of leveed valleys (profiles 350 15, 0530 hrs, 350 14, 1530 hrs, 350 17, 1200 hrs, Figs.8.8, 8.9 & 8.26). Migration and in-fill of the channels results in a nested sequence of progressively younger channels and associated levees ranging from reflector 'L' upwards to the sea-bed (0.7-0.35 sec = 560-260 m). The channels are up to 4 km wide from levee crest to crest, with a relief of up to 0.07 sec (56 m). The youngest channel is in-filled and no active channel has been detected. Post-'L' material pinches out both on the western and eastern flanks of the Inharrime Terrace. A second major depocentre occurs at the base of and on the shallow continental slope off Maputo and the Limpopo River. Initially, the valley separating the continental slope from the pre-'L' Limpopo Cone was in-filled from the east, with well-laminated reflective strata dipping gently west-north-west (profiles 350 16, & 371 6, Figs.8.6 & 8.25). Later material is derived predominantly from the north and west. It is uncertain whether some irregular channel-like bedding (west end profile 371 5, Fig.8.7) is due to leveed valleys. Farther south, the base-of-slope valley is complexly in-filled (profile 350 14, Fig.8.9). The modern valley on the west side of the Limpopo Cone is filled with densely laminated ponded material similar, though on a smaller scale, to scoured channel fill at the base of the Mozambique Ridge flank south of the area covered here (c.f. Dingle et al., 1978). A third depocentre, on the southern Limpopo Cone, also comprises reflective, finely-laminated sediments. In this region prominent moats attest to continued erosion and re-deposition around seamount flanks. Pre-'L' sediments outcrop on the southern face of the Limpopo Cone and on the upper continental slope (profiles 350 5a, 350 4, 371 6 & 350 5b, Figs.8.12, 8.13, 8.25 & 8.28).

8.3.6 Total sediment thickness

Data on the total sediment thickness are sparse because of the lack of penetration to bedrock and thus the total sediment isopach chart partly mirrors the basement structure contour chart (compare figures 8.34 and 8.45). The main depocentre is a north-oriented trough lying sub-parallel to the Zululand margin. Here, at 28°S, sediment thickness reaches 2.26 secs (3.4 km assuming 0.5 sec of unit 1) and 3.5 km off Inhaca Island (Beck & Lehner, 1974). This depocentre broadens under the Limpopo Cone and extends northwards to the Mozambique coastal plain where over 3 km of sediments overlie volcanic rocks (Flores, 1973). The Zululand continental margin and Limpopo Cone appear to have been depocentres from the inception of the basin to the present day (Figs.8.36, 8.38, 8.40, 8.41, 8.43 & 8.44; Flores, 1973; Beck & Lehner, 1974). Both basement highs, the Almirante Leite Bank complex and the Naudé Ridge have acted as sediment dams: thick sediment piles have accumulated on their northern flanks while sediment is thin or absent on their apices. On the Central Terrace, sediment thickness is very variable, and generally thinner than under the continental slope, reaching a minimum of 0.37 secs on the up-faulted central area (compare Figs.8.38 & 8.45). This region is predominantly in-filled by sedimentary units 1 and 2. By comparison the valley to the east of the Central Terrace has accumulated up to 1.9 secs of sediment, whereas on the irregular Mozambique Ridge sediment thickness ranges from 0 - 1.5 sec. On the inner Tugela Cone, 4.6 km of sediment is proven by the data of Du Toit & Leith (1974), while the eastern Tugela Cone is underlain by 2.9 km of sediment (Ludwig et al., 1968). In the Deep Ocean Basin, sediment thickness decreases southwards from 3.4 km at 32°S to 1.2 km at 35°S, where the Natal Valley opens out into the Transkei Basin (Chetty & Green, 1977). Thus the Natal Valley sedimentary basin has not accumulated as thick a sedimentary pile as the west or south coast margins, where deposits over 7, 8 and 9 km thick respectively occur in the Walvis, Orange and Outeniqua Basins (Leith & Rowsell, 1979; Dingle, 1979; 1982).

8.4 DEPOSITIONAL RATES AND STYLES

8.4.1 Depositional rates

Maximum thicknesses of sediment, and sedimentation rates for each stratigraphic unit are given in Table 8.2. Each unit has a minimum thickness of zero as a result of either non-deposition on basement highs which acted as sediment dams or through non-deposition and erosion on the Central Terrace caused by geostrophic currents (chapter 9).

The maximum sedimentation rate for stratigraphic unit 1 is relatively high, declines for units 2 and 3, and then greatly increases for units 4 and 5. Note that an arbitrary age mid-way in the post-Jimmy era has been used to calculate separate sedimentation rates for unit 4 and unit 5.

The relatively high sedimentation rate in unit 1 may be related to several effects. Initial break-up is generally considered to provide some topographic relief adjacent to the incipient ocean (eg. Hay et al., 1981). This would be expected to help generate a vigorous drainage system especially during warm humid Late Jurassic and Cretaceous times (Barron et al., 1981). This effect may have been tempered partially because the oceanic crust was initially emplaced sub-aerially. In Late Jurassic and Cretaceous times, prior to any possible Late Tertiary uplift of Africa which may have led to new watersheds in Botswana (section 9.3.1) it is possible that the drainage system of the inland Okavango Basin in Botswana linked up with the Zambezi and Limpopo Rivers, as it occasionally does today in a limited way during floods. If this were the case, the proto-Limpopo drained a far larger catchment with a resultant enormous sediment supply. An alternative postulate is that subsidence of margins increased due to a world-wide taphrogenic/epeirogenic episode stretching from at least the Late Jurassic (and sometimes from the Permo-Triassic) to the Aptian/Albian (Kent, 1977; Dingle, 1982). This allowed higher sedimentation rates within proximal areas of marginal basins as opposed to progradation on to more distal parts.

Table 8.2

Maximum sedimentation rates

Sediment unit	Duration m.y.	Sonic velocity km/sec	Maximum thickness location	(sec)	(m)	sedi rate m/m.y.
1	35	4.0	west valley	1.1	2200	63
2	64	2.7	north flank NaudéRidge	1.35	1823	28
3	29	2.2	Zululand cont. slope	0.70	770	27
4&5	5	1.6	Limpopo Cone	0.85	680	136
4	2.5?	1.6	Limpopo Cone	0.73	584	234
5	2.5?	1.6	Inharrime Terrace & Maputo Slope	0.44	352	141

Age of basement, for the calculation of sediment unit 1, is taken as 133 m.y. This is half-way between the initiation of spreading of Antarctica and the initiation of a triple junction at 122 m.y. (chapter 6). The area of maximum accumulation os sedimentary unit 1 is close to the proposed position of a fracture zone east of the Lebombo Mountains, while M10 (122 m.y.) crust may have been emplaced near this locality in this spreading compartment. McDuff is taken as basal Cenomanian (98 m.y.). Angus is taken as basal Oligocene (34 m.y.). Jimmy is taken as Miocene/Pliocene boundary (5 m.y.). 'L' is arbitrarily taken as 2.5 m.y.

Reduced sedimentation rates in units 2 and 3 may be due to declining rates of subsidence (Table 8.5). The general increase in Late Tertiary sedimentation rates relative to Palaeogene and Cretaceous rates is mirrored by local and global DSDP results (Girdley et al., 1974, p.738; Davies et al., 1977; Fig.9.6). This has been related to a variety of effects (section 9.3.4) but in general, increasing contents of aluminium oxide suggest increased terrestrial weathering associated with cooler climate (Donnelly, 1982). This effect is noticeable in other east coast basins of Africa, while basins of the west coast show an opposite effect (Dingle, 1982, Tables 2 & 3). The reduced sedimentation rates on the west coast are related to reduced subsidence of the continental margins with age and increased aridity of the Saharan and Namibian hinterlands (section 9.4).

Although sedimentation rates have increased in the Late Tertiary (Table 8.2), sediment accumulation rates do not tally with modern river inputs and terrestrial sediment supply. Modern sediment supply figures (Table 8.3) are based on calculation of catchment areas and estimates based on average carbonate contents (Flemming, 1981; Flemming & Hay, 1983). An estimate of the volume of post-Jimmy (sedimentary units 4 & 5) material in the N.N.V. basin was made by summing areas of the isopachyte chart (Fig.8.41) between the various isopachs. The value is probably an under-estimate since a seismic velocity of 1.6 km/sec was used and so no correction for compaction was made. Incorporation of coastal plain and dune barrier deposits (Frankel, 1972; Flores, 1973; Hobday, 1979) does not alter the conclusion that present-day sediment supply is 25 times as high as Pliocene-Recent accumulation rates. Similarly for the total sediment pile in the N.N.V., present-day supply is 30 times greater than the average accumulation rate. Similar conclusions are reached when comparing the entire Mozambique coastal plain/Natal Valley/Transkei Basin depocentre (Tables 8.3 & 8.4).

There are three possible explanations: 1) present-day supply has been over-estimated and the deposits under-estimated; 2) sediment is transported

Table 8.3

Modern supply of sediment

(10^6 m^3)

	Terrigenous input	Biogenic input	Total
Northernmost Natal Valley	78.251	34.43	112.681
East coast margin	133.105	58.566	191.671

Figures based on Flemming & Hay (1983) except that the biogenic input is taken as 44% of the calculated terrigenous input because the average carbonate content of samples in the N.N.V. (Fig.8.2) = 44%.

Table 8.4

Comparison of modern sediment input with calculated accumulation rates

NORTHERNMOST NATAL VALLEY

$m^3 x 10^{12}$

	Offshore	coastal plain	dune ridges	Total	sedi rate ($m^3 x 10^{12}$/m.y.)	% present-day supply
Post-Jimmy sediments	22.85	0.43	0.18	23.46	4.49	4.0
Total sediments	234.123	252.5	-	486.623	3.6	3.2

NATAL VALLEY/TRANSKEI BASIN

$m^3 x 10^{12}$

	N.N.V.	Tugela Cone	S.Natal Valley	Port St. Johns Basin		$m^3 x 10^{12}$/m.y.	
Total sediments	486.623	131.25	695.8	3.6	1317.27	10.8	5.6

out of the area - this does not happen through dissolution as the entire Natal Valley appears to be above the carbonate compensation depth (although it may not always have been); rather, it may happen, particularly in the case of the fine fraction, through the action of geostrophic currents; 3) present-day terrigenous supply is much higher than in the past either because of natural causes or through agricultural malpractices. From a consideration of the erosion of the Tugela Valley, Murgatroyd (1979) concluded that present-day erosion is 28 times higher than Cretaceous-Recent rates. His figures based on terrestrial criteria agree well with Tables 8.3 and 8.4. In the eastern United States agricultural malpractices are considered responsible for a ten-fold increase in erosion, and appear to be under control now (Meade, 1982). The scale of increase in southern Africa, by comparison, is alarming.

8.4.2 Depositional styles

The tectonic style of the N.N.V. basin varies through time. Back-tracking of basement emplacement depths and palaeo-bathymetry (Tables 8.1 & 8.5) indicate that initially, volcanic material was emplaced sub-aerially. This is in agreement with on-land sedimentological observations: the initial sediments over central and southern Mozambique comprise continental sandstones and conglomerates (Flores, 1973; Forster, 1975); on the Zululand continental plain, shallow marine sediments and interfingering fluviatile sandstones and conglomerates rest on lavas showing evidence of sub-aerial weathering (Kennedy & Klinger, 1975). As subsidence proceeded, sedimentary unit 1 (pre-McDuff) was deposited in shallow marine sub-basins bounded by irregular basement highs (Fig.8.33), some of which were probably fault-controlled (Beck & Lehner, 1974, Fig.4). Sedimentary unit 2 (McDuff-Angus) is dominated by faulting and extends over the Central Terrace. This 'uplap' style is succeeded by offlap and onlap of sedimentary units 3, 4 and 5 which are predominantly restricted to the Zululand continental margin, Limpopo Cone and Inharrime Terrace. Available data do not clearly elucidate the upper slope and shelf,

Table 8.5

Back-tracking of palaeo-depths

Site	basement corrected depth	basement emplacement depth	sediment accumulation	subsidence cooling	subsidence loading	palaeo-depth	depth
a) McDuff Times							
(1) cont.marg. 28°S	-2211	+1289	1000	2075	700	-486	
(2) west valley	-2843	+657	2200	2075	1540	-758	
(3) C.Terrace	-1842	+1658	420	2075	295	-287	
(4) east valley	-2682	+818	1540	2075	1078	-795	
b) Cret/Tertiary Boundary							
(1)			405	2800	225	-1031	
(2)			405	2800	225	-1303	
(3)			81	2800	45	-976	
(4)			472	2800	252	-1300	
c) Present-day							depth
(1)			1495	3500	655	-891	855
(2)			595	3500	295	-1703	1763
(3)			145	3500	61	-1592	1620
(4)			632	3500	302	1670	1710

The calculated palaeo-depth for four sites is given for McDuff times, the Cretaceous-Tertiary boundary and the present-day. Actual present-day bathymetry is given for comparison, and agreement is good.

and thus relationships with sea-level variations are unclear (c.f. Vail et al., 1977). This succession of tectono-depositional styles (from uplap to offlap-onlap) is similar to that observed in Brazilian rift and pull-apart basins and is considered typical of rift and post-rift basins (Brown & Fisher, 1977, Fig.14).

Three basic sedimentary models have been proposed for deep-water basins (Gorsline, 1978; 1980; Nardin et al., 1979): a) continuous processes of the 'oceanic' facies where particle settling and resuspension by currents dominate; these processes are evinced by relatively high carbonate contents and current-moulded features (chapter 9); discontinuous processes may either encompass b) the slope and base-of-slope model where mass-gravity processes dominate (chapter 10); or c) the submarine canyon-fan model. This latter suite of processes is best displayed at two localities - the Inharrime Terrace and the Zululand margin south of 27°S. In the former locale, leveed valleys occur at depths ranging from ∿250-1000 m, suggesting that they are predominantly submarine features, rather than having formed sub-aerially during sea-level low-stands (Fig.9.3). Nested valleys indicate a predominantly depositional regime. These fan-valleys (sections 8.3.5.3.2 and 8.3.5.4) show similar characteristics to those described from submarine fans of various scales. On the Navy fan (Normark, 1970; Normark et al., 1979) valleys are 15-50 m deep and ∿400 m wide, while on the enormous Bengal fan, valleys are up to 100 m deep and up to 27 km wide (Curray & Moore, 1971). Dammed coastal lakes near Ponta Zavora (Fig.2.5; Frankel, 1972) and the drainage pattern suggest that a river (possibly the Limpopo) formerly entered the sea at Inharrime. In this respect, the Limpopo Cone and Inharrime Terrace are similar to 'abyssal cones' which lie off major deltas (Rupke, 1978) except that in this case depths are intermediate. On the Zululand margin many valleys cut the shelf and slope. Their upper reaches can sometimes be related to former river channels now choked by the coastal barrier-lagoon system (Hill, 1975). On the slope, 'V'-shaped valleys cut across a slumped mass (chapter 10) and appear to coalesce into a steep canyon running south-eastwards along the broad valley on the west flank of the Central Terrace. In the area covered here, the valleys appear erosional. The bathymetry (section 2.3) suggests that material transported via these canyons is deposited in the lobe shaped area in the 'Deep Ocean Basin' region south of the Central Terrace (Fig.2.3).

8.5 COMPARISON WITH THE PLATE TECTONIC MODEL

Seismic evidence in the N.N.V. is not in conflict with the plate tectonic model described in part 2. The Almirante Leite Bank and crust underlying the Inharrime Terrace, Limpopo Cone and Central Terrace is volcanic. In the model given in chapter 6, this was erupted at a spreading ridge between Africa and Antarctica in M21-M10 times. Back-tracking of palaeo-bathymetry assuming a normal oceanic crustal cooling curve (Parsons & Sclater, 1977) explains the delay between initial spreading and marine incursion: volcanic material was initially emplaced sub-aerially. Crust in the Deep Ocean Basin sub-province (Fig.2.3) lies at more normal levels (Chetty & Green, 1977). Perhaps excess volcanism ended and more normal rates of crustal emplacement followed the initiation of a stable triple-plate spreading regime after M10 times.

8.6 IMPLICATIONS FOR OIL POTENTIAL

The hydrocarbon potential of the continental margins of southern Africa has been assessed by Dingle (1980b). Some additional points arise from the present study.

Coastal Mozambique and the Natal Valley are underlain by oceanic crust, and sediments in the area have built upwards and outwards since continental rupture. Coastal Mozambique is therefore analogous to the Niger Delta - another area of the African continent extending on to oceanic crust, and now a giant oil field. Although the sedimentary pile is over 12 km thick in the Niger Delta (Lehner & De Ruiter, 1977; Dingle, 1982) compared with 3.5 km in Mozambique and the N.N.V., a depth of only 2 km is sufficient at normal thermal gradients to reach temperatures capable of maturing hydrocarbons (Leith & Rowsell, 1979; Tissot et al., 1980). Potential is enhanced through the extensive role of strike-slip tectonics. Such areas (Fig.7.4) typically produce abundant and extensive trap structures such as en échelon faulted plunging anticlines (Harding, 1973; 1974; Harding & Lowell, 1979; Reading, 1980). Potential source rocks were deposited in wide areas of the evolving margins and basins around southern Africa in pre-McDuff times (section 8.2).

A variety of trap situations are displayed by seismic profiles in the N.N.V., while several seismic 'bright spots' exist. These are indicative of large contrasts in acoustic impedance and have often been ascribed to gas-clathrate interfaces (Tucholke et al., 1977; White, 1977). To date hydrocarbon discoveries around South Africa are sub-economic (Light et al., 1982) but are derived from the stratigraphic equivalents of Argentinian and Chilean oil fields of the Magellan Basin (Olea & Davis, 1977). No exploration drilling has been undertaken in the Falkland Plateau Basin (the continuation of the Outeniqua Basin - chapter 3) but a high degree of maturity exists in source rocks (Ludwig, Krasheninnikov et al., 1980). In the adjacent Malvinas Basin, an oil-flow of 5000 barrels/day was tested (Deal, 1982). These facts point to hydrocarbon potential around southern Africa in the Outeniqua Basin, coastal Mozambique and the Natal Valley.

8.7 SUMMARY

Five regional acoustic horizons - basement, McDuff, Angus, Jimmy and 'L' - are respectively dated as pre-Barremian, ~Cenomanian, Oligocene, Miocene/Pliocene boundary and Pliocene-Recent. These divide the sedimentary succession into five units.

Seismic and dredging data indicate volcanic basement is in continuity with Lebombo Group rocks. Almirante Leite Bank and the Naudé Ridge form two ENE-oriented complex basement highs.

By McDuff times, the Central Terrace forms a clearly-defined spur which is irregularly up-faulted relative to flanking valleys. Sedimentary unit 1 is up to 1.1 sec thick and infills basement depressions via on-lap and divergent fill.

By Angus times, the Limpopo Cone had developed, relief was smoother, and sediment had topped part of the Naudé Ridge. Unit 2 is very variable in thickness (max. 1.35 sec north of Naudé Ridge) because of uneven McDuff topography. Growth faults are almost ubiquitous, while upwards within the unit fan complexes and contourites occur to west and east of the Central Terrace respectively.

Unit 3 (Angus-Jimmy) forms a progradation of the Limpopo Cone and Zululand continental slope (max. thickness = 0.7 sec), while over the eastern Central Terrace this unit is absent.

Units 4 and 5 cannot be distinguished on the Central Terrace where post-Jimmy sediment is thin re-worked or absent. Over the Limpopo Cone, reflector 'L' marks major changes in deposition. Unit 4 (Jimmy-'L') is subdivided: the lower part is dominated by massive slumping; this is partly overlain by a large submarine fan forming the major expression of the Limpopo Cone (max. thickness = 0.73 sec). Unit 5 (post-'L') sediments of the Inharrime Terrace also form a submarine fan (0.44 sec thick) displaying a nested sequence of leveed channels. Post-'L' material also occurs on the Maputo continental slope and the southern Limpopo Cone.

Total sediment thickness is 3.5 km under the Limpopo and Maputo continental slopes. This is less than other marginal basins around Africa. Sedimentation rates are relatively high for unit 1, low for units 2 and 3 and high for units 4 and 5. Present-day sediment supply rates are 20-30 times higher than accumulation rates calculated from total and Pliocene-Recent sediment volume. Either sediment is exported from the basin, or modern supply rates have increased dramatically.

Depositional style evolves from uplap-dominated to an alternation of off-lap on-lap. Oceanic, base-of-slope and canyon-fan deep-sea sedimentary facies are exhibited.

CHAPTER 9

THE INFLUENCE OF GEOSTROPHIC CURRENTS ON SEDIMENTATION

9.1 INTRODUCTION

Large areas of the N.N.V. aseismic marginal plateau are shallower than 2000 m (section 2.3). The area is swept by part of the South Indian Ocean Western Boundary Current, the Agulhas Current, where flow extends to depths up to 2000 m (section 2.5). As a result, the current has influenced sedimentation over a wide area: current-controlled features are abundant, and sediment has been concentrated in various depocentres while other areas have been scoured (section 8.3). Moreover, the current mobilises large amounts of bed-load sediment on the narrow continental shelf where large submarine dunes migrate over an unconformable surface (Flemming, 1978; 1980; 1981). The Agulhas Current is thus a major control on hiatus development both on and off the shelf.

The influence of deep geostrophic currents on sediment dispersal, particularly in the western North Atlantic, has been well documented over the last two decades (Heezen et al., 1966; Schneider et al., 1967; Hollister & Heezen, 1972; Heezen, 1977; Stow & Lovell, 1979). Usually, cold bottom waters spreading from the Arctic or Antarctic via Western Boundary Undercurrents are invoked as the controlling agents. Similarly, in deeper areas of the South West Indian Ocean, cold undercurrents of Antarctic Bottom Water move northwards into the various basins (section 2.5) and influence abyssal sedimentation (Kolla et al., 1976; 1980; Dingle & Camden-Smith, 1979; Johnson & Damuth, 1970; Camden-Smith et al., 1981). These currents do not penetrate into the relatively shallow (<2000 m) N.N.V. (Warren, 1974; Kolla et al., 1976). Intensification of the Indian Ocean circulation system in the geological past has been invoked to explain widespread hiatuses found in Deep Sea Drilling cores (Vincent, 1974; Leclaire, 1974; Davies et al., 1975; Moore et al., 1978).

The purpose of this chapter is three-fold:

1) to demonstrate the influence of the Agulhas Current on sedimentation by comparing hydrographic and geological features;

2) to review and critically evaluate proposed mechanisms for the development of regional hiatuses; consideration is given to sea-level fluctuations, African hinterland configuration and drainage, palaeo-circulation and deep-sea sedimentation rates;

3) to assess the significance of N.N.V. hiatuses and sedimentation patterns in relation to the palaeo-circulation of both the Agulhas Current and the Indian Ocean as a whole.

9.2 CURRENT CONTROLLED SEDIMENTATION

9.2.1 The Agulhas Current

Physical oceanography in the region is described in section 2.5. Recapping, the main points of interest are: a) both the continental margin and the Mozambique Ridge act as western boundaries (Harris & Van Foreest, 1977; Harris, 1978; Lutjeharms et al., 1981); b) two main types of water are supplied: i) warm, relatively saline 'tropical' water is supplied to the Agulhas Current by the Mozambique and East Madagascar currents (Harris, 1972; Lutjeharms et al., 1981); ii) cooler, less saline water, more 'subtropical' in character is recycled from the Mozambique Basin supplied partly by the Return Agulhas Current (Grundlingh, 1977; Bang & Pearce, 1978); c) four features of the Agulhas Current (Figs.2.18, 9.1) dominate the physical oceanography of the N.N.V.: 1) Route A flows south-westerly diagonally across the area impinging on the shelf near 28°S; 2) Route B flows south above the eastern Central Terrace and the Mozambique Ridge; 3) a counter-current flows north-north-east up the coast from about 27.5°S; 4) a cyclonic eddy, partly defined by the counter-current and Route A, is most clearly marked off Ponta Zavora, and is associated with dynamic upwelling.

9.2.2 Geology

The bathymetry of the area is discussed in detail in section 2.3. Physiographic features are distributed in a simple asymmetric pattern comprising two main terraces: one lying in the north western basin between 400 and 800 m is formed by the Inharrime Terrace and the Limpopo Cone, while the second, the Central Terrace lies between 1300 and 2000 m.

The basin configuration and sedimentary in-fill are discussed in section 8.3. Sediment distribution is asymmetric with build-up of the Limpopo Cone contrasting with low sedimentation rates on the Central Terrace. This asymmetry is more pronounced in Late Palaeogene and Neogene times (units 3,4 & 5). Although the asymmetry is primarily related to adjacent terrestrial sediment supply, current-influence is considered a control because a moderate pelagic sedimentation rate is expected throughout the area, given the high biogenic supply.

9.2.3 The Match of Physiographic and Oceanographic Features

Figure 9.1 shows that several semi-permanent features of the Agulhas Current are closely related to the asymmetry of the sediment distribution pattern. Areas of fast accumulation are associated with sediment sources and with either sluggish currents, up-welling, or shear between opposing flows. Areas of slow accumulation or non-deposition are related to more powerful currents.

9.2.3.1 Areas of non-deposition and/or erosion

Route A, the inner southerly flow-path of the Agulhas Current, coincides with the eastern edge of the Inharrime Terrace and the southern edge of the Limpopo Cone, which together form the limits of thick Pliocene-Recent sediments. That is, the outer slope of this sedimentary pile forms part of a western boundary. This flow-path appears to keep the northern part of the Central Terrace and the volcanic sea-mounts of the Almirante Leite Bank almost sediment free, and restricts fast sedimentation to the north-western corner of the Natal Valley.

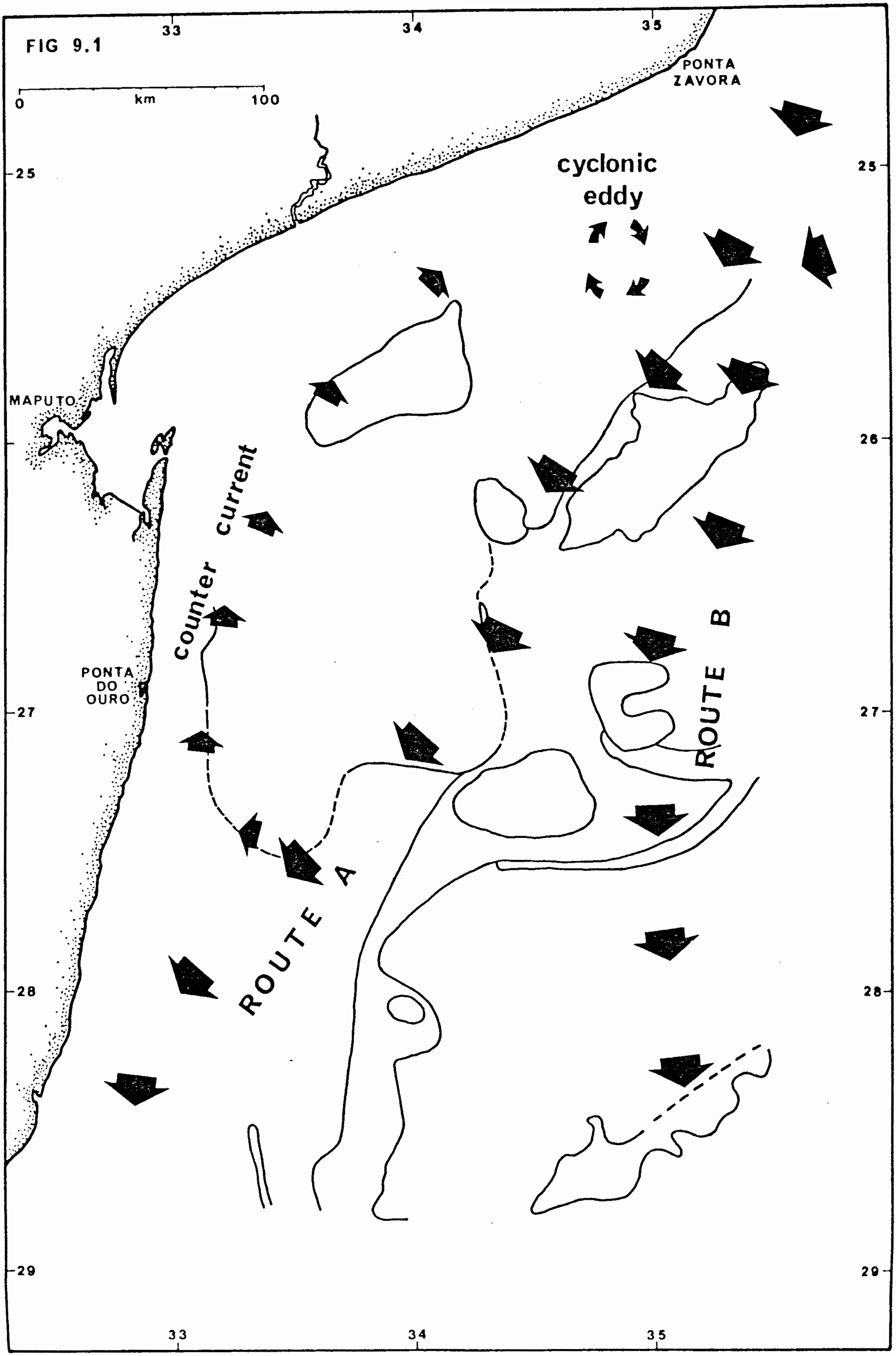
FIGURE CAPTIONS

Fig.9.1. Main water flow-paths in the area super-imposed onto a trace of the geology map (Fig.8.46). Route A and Route B are two main flow-paths of the Agulhas Current. Route A and the counter current define a cyclonic eddy which is best defined off Ponta Zavora.

Fig.9.2. Simplified traces of profiles 350 15 (Fig.8.8) and 350 4 (Fig.8.13) on to which the data of hydrographic profiles 2 and 4 (Fig.2.18) have been projected. Hydrographic data adapted from Lutjeharms (1971) and Harris (1972).

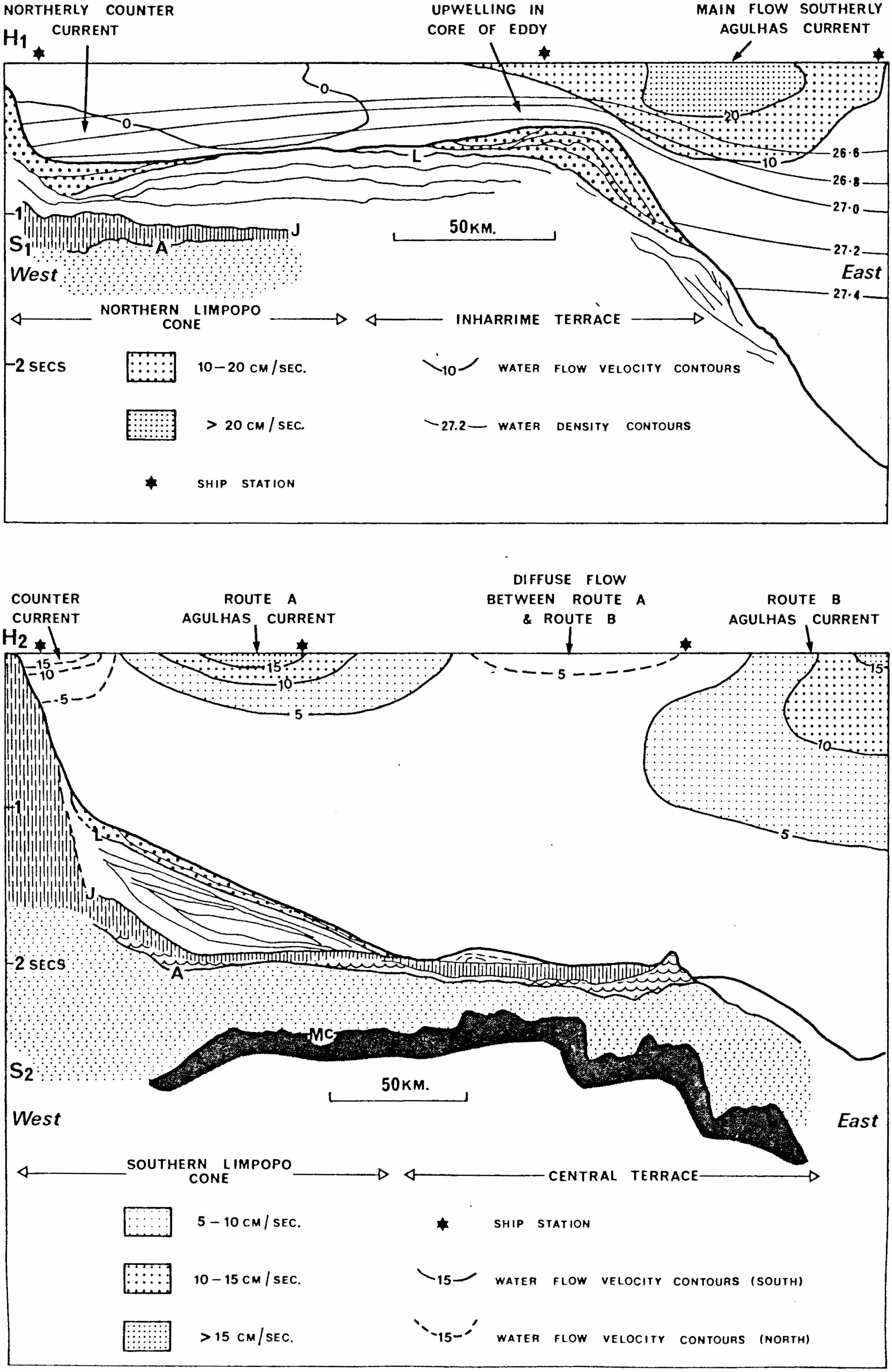
Fig.9.3. a) Comparison of global sea-level fluctuations (Vail et al., 1977); b) Southern African transgression/regression cycles (Siesser & Dingle, 1981); c) N.N.V. hiatuses; d) western Indian Ocean hiatus frequencey based on DSDP data (Moore et al., 1978). Dates of N.N.V. hiatuses from Fig.8.4. In d) Cretaceous penetrated by four holes only - data from Simpson, Schlich et al. (1974). Black line - hiatus; open rectangles - dated sequences; open cubes - undated sequence.

FIG 9.1



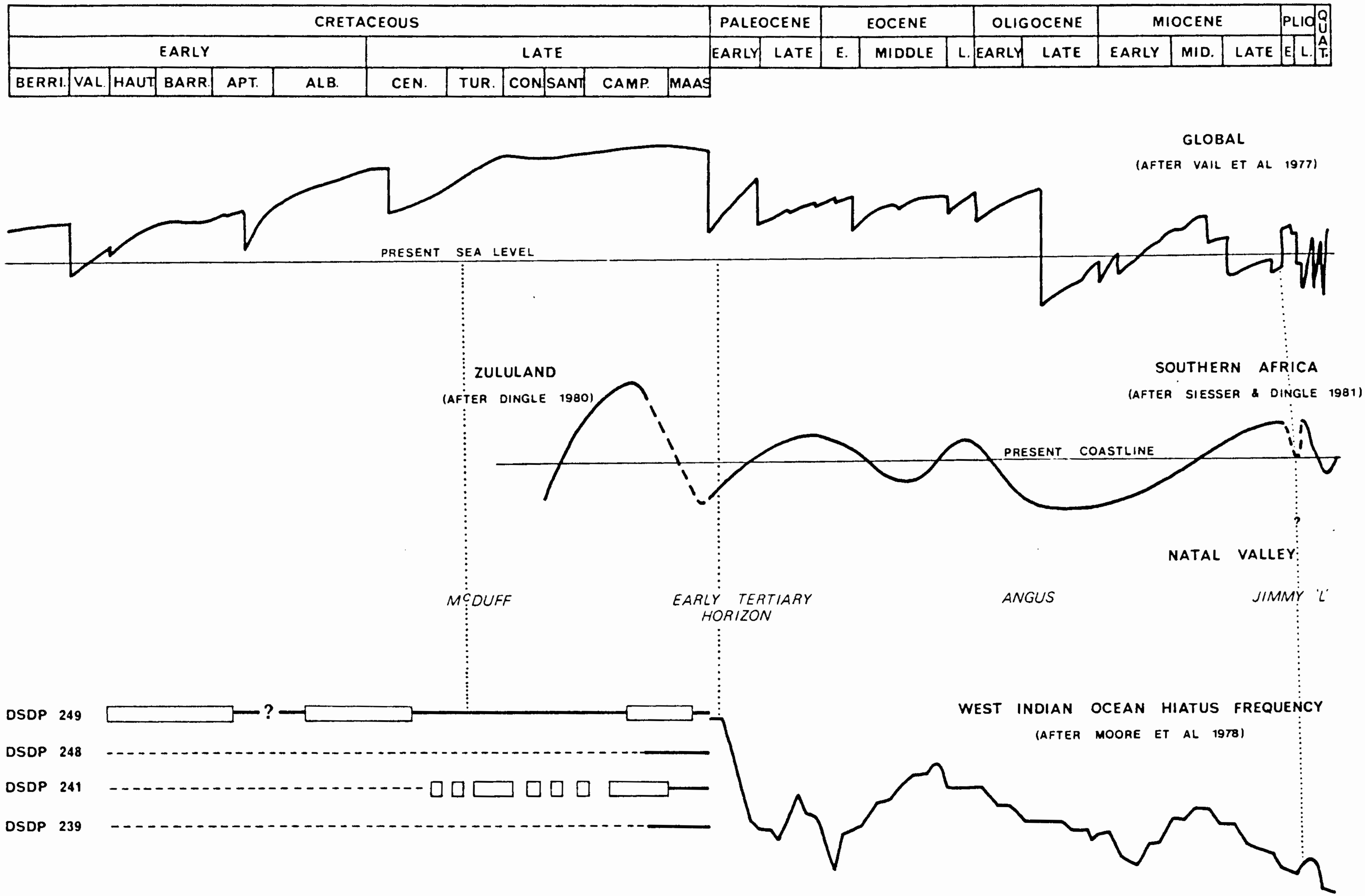
33
34
35
0
km
100
PONTA
ZAVORA
-25
25
cyclonic
eddy
MAPUTO
26
counter
current
ROUTE B
PONTA
DO
OURO
-27
27
ROUTE A
-28
28
-29
29
33
34
35

FIG 9.2



NORTHERLY COUNTER CURRENT
UPWELLING IN CORE OF EDDY
MAIN FLOW SOUTHERLY AGULHAS CURRENT
H1
0
20
L
10
26·6
26·8
27·0
27·2
27·4
1
J
S1
A
West
50KM.
East
NORTHERN LIMPOPO CONE
INHARRIME TERRACE
2 SECS
10–20 CM/SEC.
> 20 CM/SEC.
10 WATER FLOW VELOCITY CONTOURS
27.2 WATER DENSITY CONTOURS
SHIP STATION
COUNTER CURRENT
ROUTE A AGULHAS CURRENT
DIFFUSE FLOW BETWEEN ROUTE A & ROUTE B
ROUTE B AGULHAS CURRENT
H2
15
10
5
1
L
J
2 SECS
A
MC
S2
50KM.
West
East
SOUTHERN LIMPOPO CONE
CENTRAL TERRACE
5 – 10 CM/SEC.
10 – 15 CM/SEC.
> 15 CM/SEC.
SHIP STATION
15 WATER FLOW VELOCITY CONTOURS (SOUTH)
15 WATER FLOW VELOCITY CONTOURS (NORTH)

FIG 9.3



CRETACEOUS
EARLY
LATE
BERRI.
VAL.
HAUT.
BARR.
APT.
ALB.
CEN.
TUR.
CON.
SANT.
CAMP.
MAAS.
PALEOCENE
EARLY
LATE
EOCENE
E.
MIDDLE
L.
OLIGOCENE
EARLY
LATE
MIOCENE
EARLY
MID.
LATE
PLIO
E.
L.
QUAT.
GLOBAL
(AFTER VAIL ET AL 1977)
PRESENT SEA LEVEL
ZULULAND
(AFTER DINGLE 1980)
SOUTHERN AFRICA
(AFTER SIESSER & DINGLE 1981)
PRESENT COASTLINE
NATAL VALLEY
McDUFF
EARLY TERTIARY HORIZON
ANGUS
JIMMY 'L'
DSDP 249
DSDP 248
DSDP 241
DSDP 239
WEST INDIAN OCEAN HIATUS FREQUENCY
(AFTER MOORE ET AL 1978)

The eastern Central Terrace lies under Route B of the Agulhas Current. In this case, the Mozambique Ridge is probably providing a western boundary. Here, McDuff, Angus, Early Miocene and Pliocene (Jimmy) strata outcrop over wide areas. The up-faulted blocks of pre-McDuff material have been scoured free of sediment and are flanked by moats. The linear part of the Early Miocene outcrop is also related to a combination of faulting and subsequent non-deposition through current action. Elsewhere on the Central Terrace, outcrops of older strata appear to be related solely to erosion and/or non-deposition.

That current action is capable of erosion, or of preventing sedimentation, is demonstrated by the scour moats surrounding many of the sea-mounts of the Almirante Leite Bank. Profile 371 4 (Fig.8.32) shows current-scoured channels on either side of McDuff and volcanic outcrops while profiles 350 13 & 350 12 (Figs.8.10 & 8.11) show a sea-mount flanked by impressive scour-moats which have sub-bottom reflector outcrops on the moat-walls. Similar moated sea-mounts are well-known from other parts of the world (Rogalla, 1960; Dietrich & Ulrich, 1961; Heezen & Johnson, 1963; Lowrie & Heezen, 1967; Johnson & Lonsdale, 1976; Normark & Spiess, 1976; Von Stackelberg et al., 1979) and it is generally agreed that moats are the products of current erosion or, at least of differential sedimentation. The deepest moat is on the upstream (NE) flank of the sea-mount, and appears to deepen towards the north (Fig.2.5). Roberts et al. (1974) have suggested that flow around a sea-mount produces a deeper channel on the right-hand side of the sea-mount (looking downstream in the southern hemisphere), and this is consistent with currents flowing from the east or north-east past the sea-mount in question here. On the south-eastern Central Terrace, moats occur around the pre-Angus outcrops which are themselves contourite mounds (profile MN3, Fig.8.19). Similar features are abundant on the Mozambique Ridge.

On the northern Central Terrace, profile 350 4 (Fig.8.13) shows a thin contourite mound partly capping reflector Jimmy. Sediment samples from this feature comprise Holocene foraminiferal sand with pteropods with up to 5%

reworked Miocene and Pliocene foraminifera (Fig.8.5). (Incidentally, the occurrence of pteropods extends their known Indian Ocean occurrence as mapped by Berger, 1978). Hydrographic data (Figs.9.1 & 9.2) suggest that current scour caused by Route A and Route B on either side of this feature, has isolated it from other post-Jimmy depocentres. Hydrographic measurements taken in a different month (Van Foreest, 1977) showed Route A and the counter-current farther east with a broad region of current-shear and up-welling between these two opposing flows situated above the sedimentary outlier. These fluctuating hydrographic conditions may be responsible for the slow build-up of the lens ($\sim$12 m/my) and the reworked nature of its sediments, as well as the shallow surrounding scour-moats.

Farther south, on the central area of the Southern Central Terrace, the more extensive deposit of thin post-Jimmy sediments may be accumulating in response to a similar hydrographic regime, with current scour caused by Routes A and B exposing the outcrops of older strata on the upper slopes of the flanking valleys.

On the narrow continental shelf south of Maputo, Flemming (1978) has discovered large submarine dunes moving in response to the Agulhas Current,while on the outer shelf and rocky upper slope, the seafloor is scoured sediment-free (Figs.8.2, 8.13 & 8.46). A bed-load parting near 27° 30'S indicates northerly sediment transport near Ponta do Ouro and southerly sediment movement south of 28°S (Flemming, 1981). This proves the competency of the Counter Current and Route A as sediment-transport agents. On the shelf, in the bed-load parting area, fluctuating currents have caused southward and northward moving bedforms to be super-imposed. This has important implications for interpretation of ancient environments: evidence for alternating currents has traditionally been ascribed to tidal currents whereas in this case it is due to oscillation of southerly flow of the Agulhas Current and northward flow of counter-currents (Flemming, 1981). It is likely that bedforms also exist in off-shelf regions influenced by the current (c.f. Kolla et al., 1976). Evidence for alternating flow may thus be found in the Southern Limpopo Cone, too, at intermediate depths.

9.2.3.2 Areas of fast accumulation

The Inharrime Terrace and Limpopo Cone form one of the western boundaries for the Agulhas Current. The formation of the Inharrime Terrace in post-'L' times altered the western boundary and probably resulted in a shift in the main current régime. The close correspondence between the cyclonic eddy described by buoy 1116 (section 2.5) and the shape of the Inharrime Terrace suggests a dynamic relationship. On the eastern flank of the Inharrime Terrace, horizon 'L' is exposed under the main southerly flow, while the main post-'L' depocentre lies under the axis of upwelling. Note that the isopycnals (Fig.9.2) have been drawn from the data of only 3 ship's stations, and the zone of upwelling is sometimes narrower (Orren, 1963; Fig.16). Core 5113, taken on the Inharrime Terrace contains up to 40% glauconitic and phosphoritic particles. Glauconite pellets are very fragile and lobate in character, forming casts of foraminiferal and pteropod tests. Similar deposits have been found under the up-welling region of the Benguela Current on the west coast of South Africa. There, the authigenic material is considered to be in situ because of the fragile nature (Birch, 1971) and its "immature" chemistry (Birch et al., 1976). However, recent uranium series dating on the Inharrime Terrace samples (J. Thomson, personal communication, 1982) indicates that the pellets are in excess of 800 000 years old, although the biogenic component includes Late Pleistocene species. The glauconitic material has probably been reworked by turbidity currents in the fan-valley system off Ponta Zavora (section 8.4). Although the exact mode of formation is unknown it is possible that up-welling in conjunction with river in-put provided high nutrient and iron values in the water column leading to authigenic production. This contributed to the build-up of the Inharrime Terrace. This situation, within an up-welling cell in a western boundary current provides a possible mechanism for the origin of east coast phosphogenic provinces.

A second thick deposit of post-'L' sediment lies under the shallow continental slope off the Limpopo River mouth (Fig.8.44). High terrigenous input is indicated by low carbonate content which reaches a minimum in core 5114 (average carbonate content for core 5114 = 23%, minimum value = 8%). The in-shore counter current probably sweeps sediment northwards and eastwards, restricting sedimentation to the base of the continental slope and accounting for the exposure of reflector 'L' on the north-central Limpopo Cone.

The third post-'L' depocentre caps the southern Limpopo Cone. Sediments are probably the distal equivalents of the Inharrime Terrace. In addition, this depocentre lies between the in-shore counter current and the inner southerly flow-path (Route A). Here, maximum current shear between the two opposing flows results in settling out of particles from sediment-laden currents.

9.3 A REVIEW OF HIATUS DEVELOPMENT MECHANISMS

Hiatus development is related to the balance between sediment supply and re-distribution or dissolution. Major controlling factors are sea-level fluctuations, hinterland configuration and drainage, and changes in palaeo-oceanographic circulation and surface water productivity.

9.3.1 Sea-level fluctuations

Because regional transgression/regression cycles in Southern Africa correspond with independent Australian and global data, eustatic sea-level change is considered the main controlling factor (Siesser & Dingle, 1981). Cretaceous data adapted from Dingle (1980c) likewise fit the global curve (Figs.9.3a & b).

During high sea-level phases, rivers debouch on to the continental shelf (Gibbs, 1981) whereas during low-stands, rivers dump sediments directly into the deep-sea with river-mouths near the shelf break. Thus, a sea-level low-stand tends to generate a hiatus on the shelf and upper slope (Rona, 1973; Vail et al., 1977) while increased off-shelf sedimentation reduces the probability of a deep-sea hiatus. When a greater width of the shelf is flooded, the influence of shelf circulation may be affected.

9.3.2 Hinterland Configuration

The history of tectonism and epeirogenesis in Africa is controversial. King (1972) considered that uplifts of the African hinterland caused all but one of the regressions on the Natal coastal plain (emerged shelf). Peneplanation and resulting sediment supply reduction is invoked for the long Early Tertiary hiatus (Fig.9.3b). This hypothesis has been criticised because of (i) lack of evidence for tectonism on the coastal plain (Frankel, 1972); (ii) repeated uplifts and the present-day configuration of Jurassic Karoo strata indicate an untenable Karoo landscape (De Swardt & Bennett, 1974); (iii) eustatic sea-level changes account for the gross distribution of Mesozoic and Tertiary deposits (Siesser & Dingle, 1981). It is partly supported by i) hypsographic curves which suggest uplift of Africa in the Late Tertiary (Bond, 1978; Harrison et al., 1981) although the main changes have occurred in North and West Africa (Bond, 1978; Fig.9); ii) differential uplift is indicated by the variety of elevations of Late Miocene/Pliocene littoral deposits which rise from 45-90 m in the western Cape Province and in Zululand to over 300 m in the eastern Cape near Port Elizabeth (King, 1972; Siesser & Dingle, 1981; Tankard et al., 1982, Fig.14.2); iii) Smith (1982) has proposed that ∿70 m.y. ago 'High Africa' moved across an asthenospheric anomaly related to Indian Ocean break-up; the resultant heating led to a phase-change in the upper mantle causing uplift 50-60 m.y. later (ie. ∿10 m.y. ago).

The complicated break-up pattern in this region (part 2, chapters 3 - 6) and particularly the preceding Karoo volcanism and attendant high heat values, caused uplifts associated with rifting followed by thermal contraction and subsidence. Since then, sea-level changes and isotatic adjustments to water and sediment loading have been the main control on hinterland configuration and drainage. A single pulse of epeirogenesis may have occurred in the Miocene/ Pliocene, explaining antecedent drainage in Natal.

9.3.3 Oceanic Palaeo-circulation

The evolution of global oceanic circulation is related to climatic events and the creation or destruction of sea-ways through seafloor spreading (Berggren & Hollister, 1974; Kennett, 1978; Schnitker, 1980). Analysis of Deep-Sea Drilling data shows that phases of hiatus development occurred in the Cenomanian/Turonian (Davies et al., 1975) at the Cretaceous/Tertiary boundary, in the Late Eocene/Early Oligocene, in the Early-Middle Miocene, and in the Pliocene (Moore et al., 1978)(Fig.7d). These were ascribed to current-induced erosion or non-deposition because of (i) lack of evidence for tectonic control (Vincent, 1974; Davies & Kidd, 1977), although tectonically induced examples exist (Leclaire, 1974; Luyendyk & Davies, 1974; Whitmarsh et al., 1974); (ii) enhanced build-up of sediment drifts in local areas and chemical evidence (Moore et al., 1978); (iii) hiatuses being more clearly defined in areas influenced by Western Boundary Currents (Davies et al., 1975).

9.3.3.1 Cenomanian/Turonian

Examination of the evolution of the Indian Ocean through seafloor spreading indicates that (i) Madagascar had moved to its present position just after 108 m.y. ago; (ii) the Astrid Ridge rifted away from the Mozambique Ridge around 113 m.y. but the two only cleared each other through strike-slip movement ∿100 m.y. (intermediate between Figs.6.4 & 6.5); (iii) the Falkland Plateau cleared the tip of Africa ∿98 m.y. ago. Deep-water connections between the South Atlantic, Antarctic, Indian and Tethys Oceans were thus established by Albian/Cenomanian times (Fig.9.4).

A change from euxinic to open-marine sedimentation in the western and eastern Indian Ocean occurred after the Cenomanian hiatus and on the Falkland Plateau and South Atlantic in the Albian (Girdley et al., 1974; Veevers and Johnstone, 1974; Barker, Dalziel et al., 1977; Natland, 1978). Aptian/Albian euxinic sediments have been considered the result of a) excessive volcanism (Vallier, 1974), b) stagnation either in local fault-controlled basins (Flores, 1973) or on an ocean-wide scale (Andrews, 1977; Natland, 1978;),

FIGURE CAPTIONS

Fig.9.4. Cenomanian circulation patterns. Reconstruction based on chapter 6 and references in the introduction to Part 3 of the thesis. Currents after references in text but particularly Gordon (1973) and Berggren & Hollister (1974).

Fig.9.5. Eocene circulation patterns in the Indian Ocean after Davies & Kidd (1977), Moore et al. (1978) and Frakes (1979). Some opening between Australia and Antarctica is shown (Cande & Mutter, 1982).

Fig.9.6. a) Indian Ocean and Pacific Ocean total sediment and carbonate sedimentation rates (Davies et al., 1977). Accumulation rates in metres/m.y. A recent global compilation shows that the Santonian and the Cenomanian/Turonian were also times of low sedimentation rates (Hay et al., 1981). Re-assessment of data (Worsley & Davies, 1979a, b), with accumulation rates in grams/cm^2/1000 years calculated using porosity and composition data. This gives an allowance for the compaction of older sediments and is probably a better estimation of accumulation rates. b) Shows the sea-level curve of Vail et al. (1977) as in Fig.9.3 and a smoothed version.

FIG 9.4

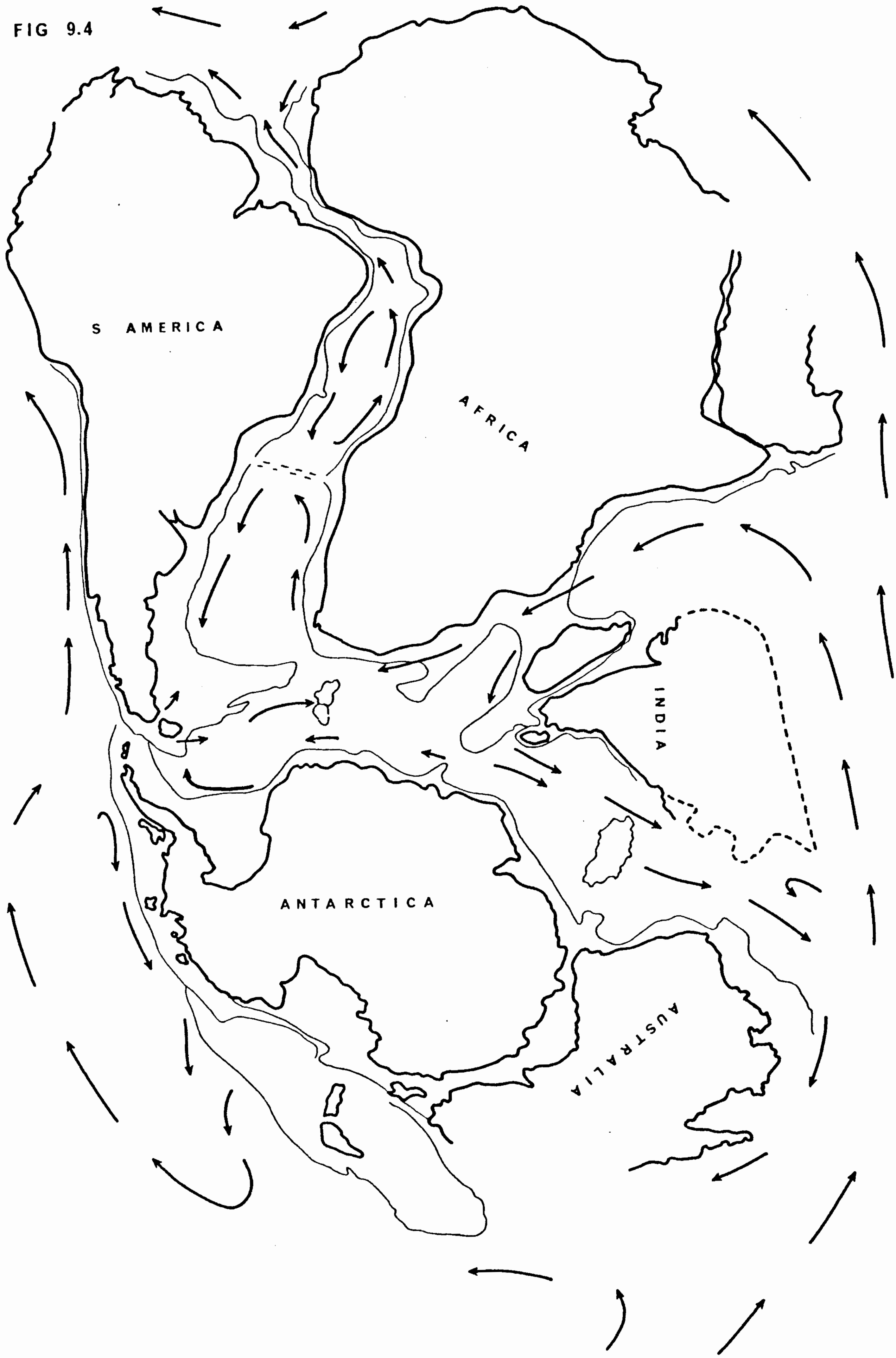


FIG 9.5

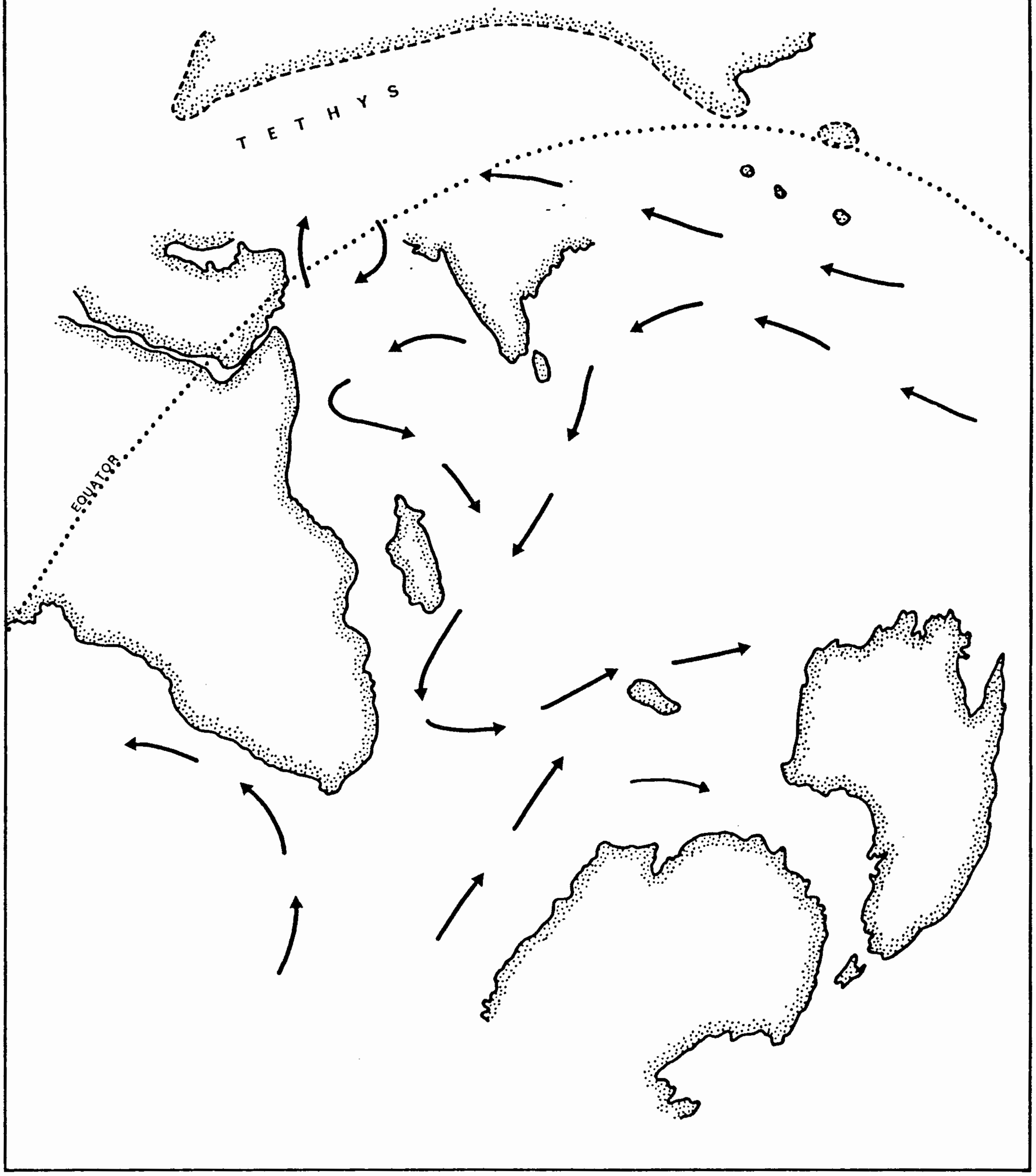
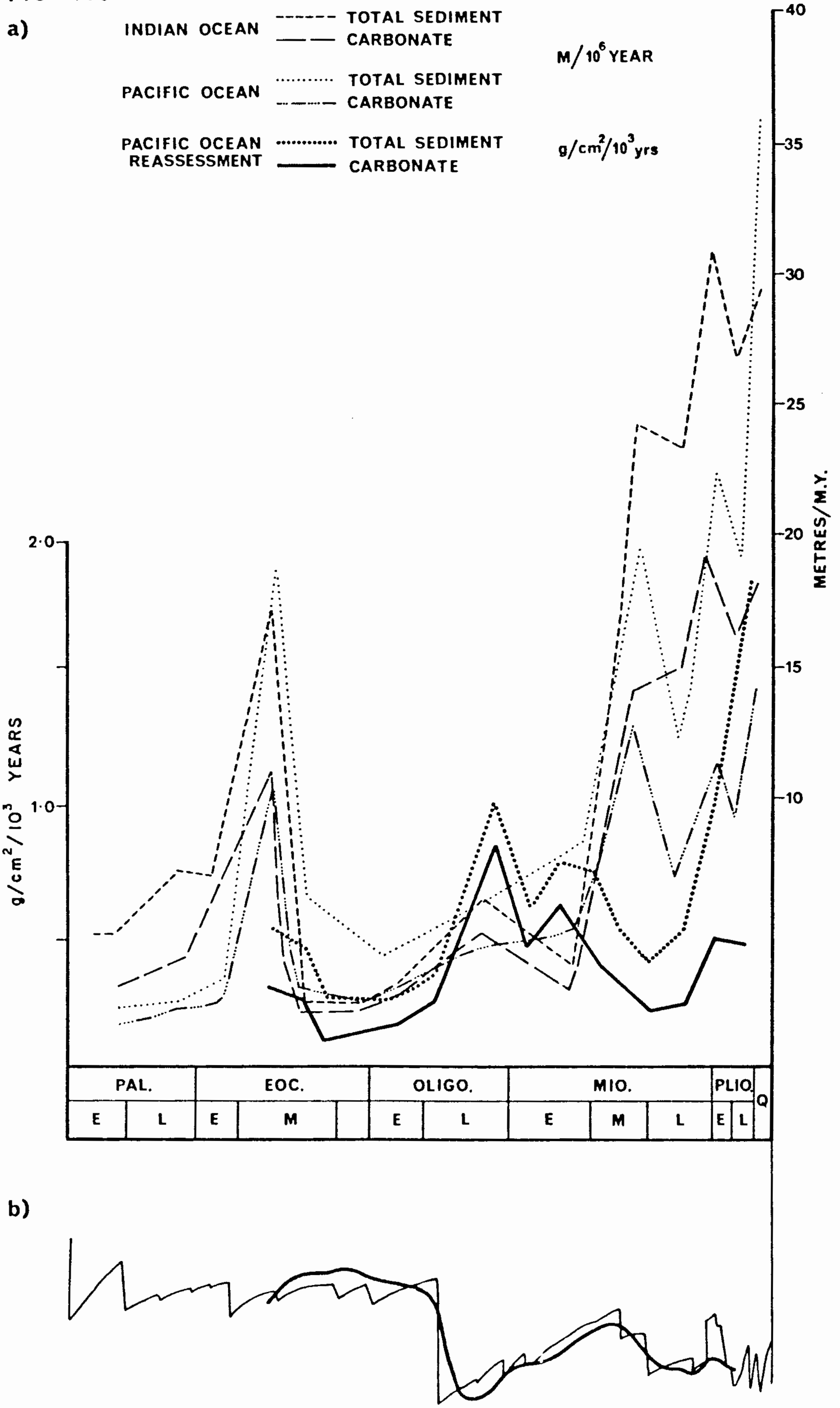


FIG 9.6


a)
INDIAN OCEAN
TOTAL SEDIMENT
CARBONATE
PACIFIC OCEAN
TOTAL SEDIMENT
CARBONATE
PACIFIC OCEAN REASSESSMENT
TOTAL SEDIMENT
CARBONATE
M/10⁶ YEAR
g/cm²/10³ yrs
METRES/M.Y.
40
35
30
25
20
15
10
2·0
1·0
g/cm²/10³ YEARS
PAL.
EOC.
OLIGO.
MIO.
PLIO.
Q
E
L
E
M
E
L
E
M
L
E
L
b)

c) impingement of oxygen-minimum layers on the sediment water interface (Schlanger & Jenkyns, 1976; Barker, Dalziel et al., 1977); d) reasonably vigorous oceanic circulation driven by warm saline bottom water production which is low in oxygen (Brass et al., 1982; Saltzman & Barron, 1982; Southam et al., 1982). This last approach differs from the assumption that low temperature differentials between high and low latitudes indicates sluggish thermohaline circulation (Douglas & Savin, 1975; Margolis et al., 1977).

Variations in temperature (Coplen & Schlanger, 1973) and in the carbonate compensation depth (Sclater et al., 1977) may indicate a phase of enhanced flow through the new sea-ways leading to the facies change and hiatus (c.f. Natland, 1978; Cool, 1982). Alternatively, euxinic deposition may be terminated by a change in organic carbon production caused by a regression (Jenkyns, 1980).

9.3.3.2 Cretaceous/Tertiary

Splitting of Greater New Zealand from Antarctica between 80 and 56 m.y. ago (Coleman, 1980) formed new sea-ways in the Ross Sea region where bottom waters are produced. Greater contrast between high and low latitude temperatures (Douglas & Savin, 1975) may point to thermo-haline circulation. Chemical evidence (Moore et al., 1978) indicate dissolution of biogenous material by bottom currents originating in Antarctic source regions (Schnitker, 1980). Amelioration of global climate at the Palaeocene/Eocene boundary and Middle-Late Eocene (Haq et al., 1977) has been linked to reduced current vigour and hiatus termination (Davies et al., 1975).

9.3.3.3 Eocene/Oligocene

The psychrosphere (cold bottom layer in the ocean) was initiated near the Eocene/Oligocene boundary in association with a sharp drop in Sub-Antarctic temperatures (Kennett & Shackleton, 1976) and marked changes in benthic fauna (Benson, 1975; Corliss, 1979, Cavelier et al., 1981)(Fig.9.5). Simultaneously, the carbonate compensation level dropped due to invigorated circulation of "young" bottom waters less corrosive to calcareous organisms (Van Andel et al., 1975, 1977). Erosion caused by this phase of current vigour was brought to an end by (i) closure of Tethys 18 m.y. ago (Berggren & Hollister, 1974) and the formation of one North Indian Ocean gyre (Leclaire, 1974); (ii) restriction of bottom water flow by the uplift of the South-West Indian Ocean Ridge (Sclater & Harrison, 1972); (iii) re-direction of bottom water flow to the circum-Antarctic current as the Tasman Rise cleared Antarctica 30 m.y. ago and the Drake Passage opened 22 m.y. ago (Kennett et al., 1974; Kennett, 1978; Barker & Burrell, 1977; Barker, Dalziel et al., 1977).

9.3.3.4 Middle Miocene

Enhanced circulation in the Middle Miocene is linked to the final lowering of the Greenland-Iceland ridge and the formation of North Atlantic Deep Water (Schnitker, 1980). (Today NADW flows round Africa into the Indian Ocean - Kolla et al., 1976). Increased siliceous content of Late Miocene sediments show that the high productivity belt of the Somali Basin was already established by that time (Leclaire, 1974; Vincent, 1974). Lessening of current activity has been linked to the closure of the gap between Australia and Indonesia 12-14 m.y. ago (Moore et al., 1978).

9.3.3.5 Pliocene

Intermittent phases of invigorated circulation in the Gulf Stream from the Miocene/Pliocene boundary to the present have been related to a Miocene/Pliocene pulse in Antarctic ice growth and Pliocene-Recent Arctic glaciation (Kaneps, 1979). Moore et al. (1978) suggest world-wide circulation stimulation and deep-sea hiatus development is linked to Arctic glacial build-up initiated 4.5 m.y. ago (Herman & Hopkins, 1980; Margolis & Herman, 1980) with large-scale development 3.2 m.y. ago (Shackleton & Opdyke, 1977). New bottom water flow patterns in the Indian Ocean are suggested by a Miocene/Pliocene drop in the carbonate compensation level (Sclater et al., 1977). This phase of world-wide current activity is still on-going.

9.3.4 Deep-Sea Sedimentation Rates

Large variations in sedimentation rates occurred through the Cenozoic and these are synchronous in all the major oceans (Davies et al., 1977). Global reduction in precipitation and supply of river loads is postulated as the cause of reduced sedimentation rates (eg. Late Eocene/Early Oligocene, Fig.9.6). Floral evidence and reduced lignite occurrences support the reduced Oligocene rainfall model but do not explain all facets of the sedimentation rate variation (Frakes, 1979. p.189-215). A later assessment (Worsley & Davies, 1979a, b) which allowed for sediment compaction, significantly changes the pattern, and fast deep-sea sedimentation rates and low sea-levels were correlated (Fig.9.6b). A similar relationship on a broader scale was noted in the North Atlantic by methods discounting compaction (Thiede et al., 1980). This relationship suggests that sediments are trapped on continental shelves during high sea-level stands and then flushed to the deep-sea during low sea-levels. Sediment budget calculations, however, show that the shelf cannot accommodate all the terrigenous detritus and some sediment must always reach the deep-sea during high sea-level. Certainly in the Natal Valley, large quantities of sediment by-pass the shelf at the present day (Flemming, 1980, 1981). Worsley & Davies (1979a) therefore postulated that during high sea-levels, biogenous sediments are laid down on shelves, decreasing the supply of dissolved nutrients to the deep-sea. This in turn reduces biogenous productivity and hence sedimentation rates. Conversely, during low sea-level, nutrients are flushed to the deep-sea and biogenous sedimentation rates increase. Comparison of Figures 9.3b and d indicates that Eocene-Recent Indian Ocean data partly support this hypothesis, while Cretaceous and Palaeocene data do not.

9.3.5 Conclusions on hiatus development models

By concentrating sediment into localised depocentres (Flemming, 1981; Martin, 1981) enhanced current action may create widespread hiatuses. Sedimentation rates at drill-sites in conjunction with seismic profile grids are required to measure volumes of sediment deposited per unit time. This would help to determine whether reduced sediment supply (Davies et al., 1977) or enhanced circulation (Moore et al., 1978) was the main cause of global hiatuses. The possibility of phases of reduced rain-fall controlling detritus supply is partly supported by climatic evidence (Frakes, 1979).

If high off-shelf sedimentation rates are associated with low sea-level (Rona, 1973; Vail et al., 1977; Worsley & Davies, 1979a) then shelf hiatuses should alternate in time with deep-sea hiatuses. Rona (1973) however, found a correlation between deep-sea hiatuses and regressions, and suggested that oceanic circulation is linked to sea-level. Off-shelf hiatuses in the N.N.V. appear to correlate with coastal regressions and proposed phases of invigorated currents (Figs. 8.4 & 9.3) and both mechanisms may have been active simultaneously. Similarly in Australia, an Oligocene shelf non-sequence is attributed to low sea-level while off-shelf hiatuses are associated with current action (Carter, 1978). When the sea submerges only a narrow strip of the outer shelf (as it does today in the N.N.V.) current action is a contributory factor in hiatus formation both in on-shelf and off-shelf areas.

New sea-ways created by continental drift may cause eustatic sea-level lowering by flooding new basins (e.g. Hsu and Winterer, 1980) and simultaneously alter oceanic circulation patterns. In the case of glacio-eustatic sea-level lows, climatic cooling may be linked to current stimulation (e.g. Kaneps, 1979). However, the relationship between ice-build up and invigorated currents is not clear-cut (Ledbetter et al., 1978; Ledbetter, 1979; Hutson, 1980).

9.4 SIGNIFICANCE OF NATAL VALLEY HIATUSES

9.4.1 Cenomanian/Turonian hiatus - McDuff

Applying oceanic crustal subsidence curves (Parsons & Sclater, 1977) to Late Jurassic/Early Cretaceous oceanic crust emplaced sub-aerially (Table 8.1; c.f. Sclater et al., 1977) suggests that the N.N.V. had subsided to relatively shallow depths by the Cenomanian (Table 8.5), with structural highs remaining sub-aerially exposed. Faulting of up to 0.7 sec throw and relief of 1.0 sec on reflector McDuff (section 8.3.2) and a slight angular unconformity (1°-2°) on the Zululand coastal plain provide evidence for tectonism up to and immediately after McDuff times. Plate tectonic re-organisations and hot-spot activity provide mechanisms. A combination of sea-level lowering (possibly up to 500 m - Dingle, 1980c) and tectonism may thus explain the McDuff hiatus and possibly the change from euxinic to open marine sedimentation (Hypothesis of Jenkyns, 1980). In contrast, hiatuses and facies changes at deep sites in the South Atlantic and Indian Oceans are better explained by invigorated circulation (Natland, 1978).

9.4.2 Cretaceous/Tertiary boundary hiatus

High Palaeogene sedimentation rates in Leg 25 DSDP sites have been related to increased Zambezi River loads resulting from uplift in Kenya and Mozambique (Girdley et al., 1974). However, Southern Mozambique was tectonically quiescent at this time (Flores, 1973). The N.N.V. seafloor lay at intermediate depths at this stage (Table 8.5) and thus the off-shelf hiatus developed at considerable depth beyond the zone directly affected by coastal regression. Depocentre shifts due to sea-level changes and current action were likely controls.

9.4.3 Oligocene hiatus - Angus

Current-moulded mounds began accumulating late in Unit 2 time (profiles MN3 & 3714, Figs.8.19 & 8.32). Similarly, pre-Angus densely laminated sediments formed in scour moats at the base of the Mozambique Ridge. These observations are consistent with current influence detected around southern Africa from Eocene times (Tucholke & Embley, in press). Reflector Angus is clearly associated with current-moulded features suggesting that Agulhas Current action strengthened in the Oligocene at a time when the psychrosphere developed worldwide. This endorses the premise that tropical/sub-tropical gyres are stimulated by increased bottom water production.

Late Eocene/Miocene uplift in northern Mozambique (Flores, 1973) and Miocene uplift in Kenya (Kent, 1972) have been linked with high Miocene sedimentation rates in the Western Indian Ocean (Girdley et al., 1974). Alternatively, the sea-level low-stand may have controlled the pulse in sedimentation (Fig.9.6). Southern Mozambique was tectonically quiescent (Flores, 1973) and Angus-Jimmy construction of the Limpopo Cone proceeded at moderate rates (Table 8.2). Continued current action throughout this phase is evinced by: (1) reworked faunas in Early and Late Miocene horizons of the Central Terrace (Salmon, in press); (2) reworking and winnowing of Mozambique Ridge Miocene sediments (Leclaire, 1974); (3) exposures of Middle Miocene sediments on the Mozambique Channel seafloor (Saito et al., 1974); (4) the absence of unit 3 sediments on parts of the Central Terrace. Current-action and sediment supply are both important factors in the development of Oligocene hiatuses in the south-west Indian Ocean (compare Kent, 1974b; Girdley et al., 1974).

9.4.4 Early/Middle Miocene hiatus

Scrutiny of DSDP columns indicates this hiatus occurred in the north-western Indian Ocean as a discreet event, while in south-western areas the Oligocene hiatus extended into the Miocene. Alteration in the North Indian Ocean circulation system affected the Mozambique Current (a source of the Agulhas Current) from the Early Miocene (Leclaire, 1974). The formation of a local Early Miocene horizon sampled on the Central Terrace (Fig.8.5) may be associated with these changes.

9.4.5 Miocene/Pliocene boundary hiatus - Jimmy

New current flow-paths in the N.N.V. similar to those of today were established after the deposition of horizon Jimmy, $\sim$5 m.y. ago. The contourite mound (sensu Vail et al., 1977) of reworked sediment on the Central Terrace (profile 350 4, Fig.8.13; Fig.9.1) began to accumulate at this time indicating that the dual flow-paths Route A and Route B were established. Asymmetric sedimentation was exaggerated, with scour on the eastern and western Central Terrace contrasting with high accumulation rates on the Limpopo Cone (Table 8.2). These changes are linked to new global patterns of Bottom Water production related to a pulse in Antarctic ice growth 5 m.y. ago (Kaneps, 1979) and Pliocene Arctic ice build-up from 5 m.y. ago to 3.2 m.y. ago (Shackleton & Opdyke, 1977; Herman & Hopkins, 1980). General stability of Agulhas Current flow-paths in the Pliocene-Recent is in accord with lack of change in global circulation patterns (Schnitker, 1980), with current fluctuations probably linked to climatic cycles.

Fast post-Jimmy sedimentation rates on the Limpopo Cone compare well with other east coast basins (Dingle, 1982) and with high global sedimentation rates (Fig.9.6). This may be controlled either by the Miocene/Pliocene boundary regression and the Late Pliocene regression or by possible hinterland uplifts (Smith, 1982) as suggested by antecedent drainage in Natal (King, 1972; Bristow, 1982). Continued current activity in post-Jimmy times is shown by the correspondence of current regimes and sedimentation patterns (Fig.9.1) and reworking on the Central Terrace, Mozambique Ridge and Mozambique Channel (Fig.8.5; Saito & Frey, 1964; Simpson, Schlich et al., 1974; Vincent, 1976).

9.4.6 Horizon 'L'

Although the main patterns of sedimentation have remained fairly constant in post-Jimmy times, depocentres shifted in the northern Limpopo Cone after the formation of the undated horizon 'L' (eg. profile 350 15, Fig.8.8). The pre-'L' depocentre, situated centrally on the Limpopo cone, has been replaced by the post-'L' Inharrime Terrace while the depocentre under the continental slope off Maputo has encroached eastwards. The pre-'L' depocentre is an older broader equivalent of the Inharrime Terrace. Currents adjusted to these new physiographic features, suggesting that the up-welling cell between the main southerly flow-path of the Agulhas Current and the counter current occupied a more westerly position in pre-'L' times.

9.5 PALAEO-ECOLOGICAL IMPLICATIONS OF THE EVOLUTION OF THE AGULHAS CURRENT

9.5.1 Direct current influence

Changes in the current regime must have directly influenced the marine biota off the east coast of South Africa. There are three main types of specialised faunal response to the Agulhas Current (Heydorn, 1978).

1) Temperature: warm tropical water brought south along the south-eastern coast of Africa extends the geographic realm of tropical Indo-Pacific groups such as marine fish (some may be found far south of their normal coral reef habitat), warm water planktonic foraminifera (Bé & Hutson, 1977) and a variety of intertidal flora and fauna (Stephenson & Stephenson, 1972); in contrast temperate forms such as the pilchard Sardinops ocellata and one of its predators Pomatomus saltatrix (Bluefish) move northwards along the coast with cooler inshore water in winter.

2) Current transport mechanism: pelagic larvae of spiny lobsters, bluefish and eel elvers (Jubb, 1964) are dispersed passively by the current; some free-swimming juvenile forms such as young loggerhead turtles and some species of sharks use the current as an aid to dispersal; the re-cycling nature of the current (section 2.5) is important in maintaining populations in various groups: bluefish which move north-eastward in cool in-shore water release larvae further off-shore where they are moved southwards by the main current; the movement of spiny lobster larvae in the sub-tropical gyre results in a closer affinity between populations found off East Madagascar and Natal than either area shows with the Mozambique Channel; this is partly explicable in terms of the East Madagascar Current/Agulhas Current/Return Agulhas Current on the one hand and the Mozambique Current/Counter Current circulation on the other hand (section 2.5; Figs.2.17, 2.18, 9.1).

3) Larval retention via marine/estuarine interaction: some species of estuarine fish spend only their juvenile phase in estuaries and pass their

adult life in the open ocean. Spawning occurs in-shore of the main southerly current where counter currents and eddies prevent southerly transport of the larvae to potentially lethal cooler waters to the south. Juveniles retained in a geographic area by this mechanism then migrate into nearby estuaries.

These specialised adaptations must have evolved through geologic time as the current system itself evolved. Close links between evolution of marine biota and development of current systems exist in the Southern Ocean (Kennett, 1978; 1980) and a similar inter-relation probably exists in the Agulhas System too.

9.5.2 Palaeo-climatic influence

Solar insolation on earth is greater at low latitudes while energy loss through radiation from earth is greatest at the poles. Ocean currents are responsible for 30-40% of the heat transfer from the equator to the poles needed to maintain the energy balance on earth (Frakes, 1979). The interactions between global circulation, heat transfer and climate are not clearly understood (eg. Ledbetter, 1979; Crowley, 1983). However large-scale changes in major currents such as the Agulhas, (marked by horizons Angus and Jimmy) and Antarctic climatic events with which they are linked, may have influenced the climate of mainland Africa.

Cold temperatures in Antarctica needed for the production of Antarctic Bottom Water at the Eocene/Oligocene boundary must have led to increased hemispheric temperature gradients (equator to pole temperature difference). Theoretically, and by analogy with the last inter-glacial (Nicholson & Flohn, 1980) this should have strengthened the mid-latitude westerlies and the Hadley Cell and associated sub-tropical high pressure belts, causing more powerful south-east trade winds. This resulted in enhanced aridifying effect of the down-welling air of the sub-tropical highs. The inter-tropical convergence zone is the 'meteorological equator'. Increased thermal contrast between the two hemispheres such as may have been caused by Antarctic cooling would tend to push the meteorological equator north of the geographic equator. Thus the rainy areas associated with the inter-tropical convergence zone moved northwards.

Both surface and bottom waters at high latitudes were cooled at the Eocene/Oligocene boundary by ∿5°C (Kennett & Shackleton, 1976; Keigwin, 1980). Northward movement of cold bottom water and enhanced wind circulation stimulated up-welling of cooler water in low latitudes. Estimates of the drop in tropical Pacific sea-surface temperatures vary from 1°C to 5°C (Savin, 1977; Keigwin, 1980). If a similar situation prevailed in the Indian Ocean, cooler water would have been spread southwards by the palaeo-Agulhas Current resulting in a general drop in Oligocene Ocean temperatures relative to the Eocene. A change in surface temperature of 5°C can cause a 20-30% reduction in evaporation (Frakes, 1979). If evaporation was decreased, air circulating in around the South Indian Ocean High Pressure Cell towards Southern Africa would have become less moisture-laden. Today, over wide areas of sub-tropical Africa, even in Botswana and Namibia, rainfall is derived from air-flows bringing moisture in from the Indian Ocean (Tyson, 1969; Heydorn & Tinley, 1980; section 2.3). Thus both the atmospheric and oceanographic changes had aridifying effects and may have been widespread. Certainly, fossil plant and sedimentological evidence suggests that in various parts of the globe less rain fell in the Oligocene than in the Eocene (Frakes, 1979).

By the Early Miocene Africa had moved northwards towards Europe, closing the Tethys ocean and reducing equatorial latitudinal oceanic circulation (Berggren & Hollister, 1974). Near the Oligocene/Miocene boundary, Antarctica moved clear of Patagonia, opening the Drake Passage and allowing circum-Antarctic Current flow (Barker & Burrell, 1977; Kennett, 1978). The Antarctic may have been cooled sufficiently by Oligocene events for ice build-up but low precipitation inhibited substantial glaciation. Greater precipitation caused by North Atlantic Deep Water reaching Sub-Antarctic regions and altering up-welling and circulation may have caused ice build-up in Antarctica in the middle Miocene (Schnitker, 1980).

At the Miocene/Pliocene boundary the pulse in Antarctic ice build-up again increased hemispheric temperature gradients and increased the thermal contrast between the hemispheres. This enhanced the aridifying effect of sub-tropical highs. Intensified trade-winds would tend to accentuate up-welling of the north-north-westerly moving Benguela Current or South Atlantic eastern boundary current. Upwelling in the Benguela Current, which appears to date from the Eocene (Shackleton & Boersma, 1981; Fig.2), intensified from the early Late Miocene through the Miocene/Pliocene boundary (Fig.2, Siesser, 1978a; 1978b; 1980). The sub-antarctic convergence was pushed north some 300 km at this time (Kemp et al., 1975). It is uncertain whether the sub-tropical convergence was also pushed north to near its present position but the development of the Cape fynbos vegetation ∿5 m.y. ago (Coetzee, 1978; Hendey, 1981) suggests that it was. During the last glacial period, ∿18 000 years ago, the sub-tropical convergence moved <2° north around southern Africa (Hays et al., 1976; Prell et al., 1979), but increased re-cycling by the Agulhas Return Current reduced water temperatures in the south-west Indian Ocean (Hutson, 1980; Prell et al., 1980). Route A and Route B of the Agulhas Current are associated with in-flow from the Mozambique Channel and the Mozambique Basin respectively (Harris & Van Foreest, 1977). By analogy with Late Pleistocene glacial times, increased re-cycling of sub-tropical water from the Mozambique Basin about 5 m.y. ago may have been a factor in the clear establishment of the dual flow-paths Route A/Route B in post-Jimmy times (post-5 m.y.). A more northerly position of the sub-tropical convergence may have reduced the geographic extent of the Indian Ocean gyres without altering temperatures within the tropical areas (Thunell, 1981). However cooler sea temperatures in sub-tropical areas resulting from enhanced re-cycling and the more northerly position of the sub-tropical convergence again led to reduced evaporation and less rainfall over sub-tropical southern Africa. At the same time the increasing effect of the up-welling Benguela

Current caused aridity of the Namib desert and Namaqualand (Siesser, 1978a). In contrast, if the sub-tropical convergence migrated north, this may have marked the on-set of winter rains in the southern Cape.

9.5.3 Effect on terrestrial eco-systems

Recently, it has been emphasised that oceanic temperatures influence the rainfall characteristics of adjacent land-masses and hence the vegetation in these areas (Milewski, 1981). The vegetation of Africa has evolved through the Cretaceous and Tertiary as a result of changing climate and environment (Axelrod & Raven, 1978) and terrestrial faunas have adapted to different habitats.

Major faunal turnovers in African mammals occurred at the Eocene/Oligocene boundary, the earliest Miocene and at the Miocene/Pliocene boundary (Maglio & Cooke, 1978). Closing of the Tethys ocean in the Early Miocene as Africa collided with Eurasia must have reduced maritime influence in the region, leading to enhanced continentality and new habitats. New African faunas in the Early Miocene also resulted from migration from Europe as the two land-masses converged. In contrast, the Eocene/Oligocene and Miocene/Pliocene saw local evolution involving mainly autochthonous and endemic groups (Maglio, 1978). In Africa, although the fossil record for this period is restricted to the north, a variety of groups such as Rodentia (Lavocat, 1978), Suidae (Cooke & Wilkinson, 1978), apes (Simons et al., 1978) and elephants (Coppens et al., 1978) were involved in the Oligocene evolutionary turnover indicating ecological changes were pervasive through a variety of habitats. Thus the low latitude terrestrial climatic effects attendant upon Antarctic Bottom Water production and invigorated Agulhas Current action may have put ecological pressure on land animals resulting in large changes seen in the fossil record.

At the Miocene/Pliocene boundary increasing numbers of grazing antelopes indicate that grassland on the High veld became prevalent at the expense of bushveld type scenery (Vrba, 1980; Brain, 1981). At the same time the Cape fynbos flora flourished at the expense of Palmae (tropical forest types) (Coetzee, 1978; Hendey, 1981). These changes are linked to drier interior climate and wetter southern Cape conditions associated with northward movement of the climatic belts and cooler sea temperatures (Brain, 1981; Hendey, 1981). This marked a huge increase in the success of animals adapted to savannah grassland such as alcelaphine antelopes (Vrba, 1980), long-necked giraffes (Churcher, 1978), and carnivorous cats (Savage, 1978). Hippopotamidae (Coryndon, 1978) and elephants (Coppens et al., 1978) were also involved in faunal radiations. The increase in grassland at the expense of wooded country may also have constituted the ecological pressure which pushed ancestral apes (noted in 8 m.y. old deposits) out on the grassland, resulting in speciations producing hominid types which characterised the Pliocene (Brain, 1981). Again therefore Antarctic events and new current regimes around Africa may have affected main-land climate and ecology.

9.6 SUMMARY

1) The Northernmost Natal Valley (N.N.V.) represents a marginal plateau with wide areas at intermediate depths. The area is swept by the Western Boundary System of the South Indian Ocean Sub-tropical Gyre. The combination of these factors has resulted in wide-spread current influence leading to asymmetric depositional patterns. Fast accumulation of sediment since the Early Pliocene is associated with either sluggish currents, shear between opposing flows or up-welling. The Limpopo Cone is restricted to the north-west corner of the basin, while the glauconites of the Inharrime Terrace are a response to a nutrient-rich up-welling cell. In contrast, powerful currents have scoured parts of the shelf and upper continental slope, exposed Cretaceous and Tertiary strata on parts of the Central Terrace and prevented deposition on the Almirante Leite volcanic seamounts (chapter 8; Figs.9.1 & 9.2). Sub-tropical

surface/near-surface current action (as opposed to the well-documented influence of abyssal currents) is therefore instrumental in forming hiatuses in both on-shelf and off-shelf areas.

2) Post Gondwana break-up sedimentary basin in-fill has been examined to evaluate the significance of depocentre shifts and the development of regional hiatuses. N.N.V. hiatuses correlate with regressions on the African coastal plain, with proposed phases of invigorated Indian Ocean Currents and with phases of slow deep-sea sedimentation (Figs.8.4, 9.3, 9.6). An assessment of mechanisms for hiatus development suggests that the influence of sea-level on sediment supply cannot fully explain the apparent time correspondence between shelf and deep-sea hiatuses. Current action, in conjunction with variations in sediment supply and sea-level changes, appears to have been a major control.

3) Current activity was a possible contributory cause of a Cenomanian/Turonian (McDuff) hiatus and facies change, and also of a Cretaceous/Tertiary boundary hiatus. An Early Oligocene hiatus (Angus) is clearly enhanced by invigorated Agulhas Current action. Increased Antarctic Bottom Water production and establishment of a cold bottom layer in the ocean (the psychrosphere) stimulated tropical/sub-tropical surface currents of the Indian Ocean at this time. Continued current action from Angus times onward is demonstrated by asymmetric sediment distributions and reworking. A change in the Agulhas Current regime occurred at or just after the Miocene/Pliocene boundary hiatus (Jimmy). This is linked to new patterns of bottom water production related to changes in the extent of the Antarctic ice-sheet and Pliocene build-up of Arctic ice. General stability of the main flow-paths since this time is in accord with lack of change in the global circulation system since the Pliocene (Schnitker, 1980). Pliocene-Recent fluctuations are probably associated with climatic cycles (Hutson, 1980).

The counter current and up-welling cell off Ponta Zavora moved eastwards in response to depocentre shifts in the northern Limpopo Cone in post-'L' times.

4) Horizons McDuff, Angus and Jimmy correlate with major geological/ palaeontological events in the last 100 m.y. (Thierstein & Berger, 1978). An abundance of hypotheses varying from sudden temperature changes, sea-level fluctuations, chemical poisoning and extra-terrestrial phenomena have been advanced to explain episodic events having drastic biological effects (eg. Herman, 1981). Without questioning these hypotheses and discussing their inter-relationships, the influence of changing current regimes is emphasised here.

The Agulhas current controls the sea-surface temperatures off the east coast of Southern Africa. Sea-water temperature affects sea-surface evaporation rates and thus the moisture content of the air passing over the Indian Ocean. This in turn influences the climate over much of sub-tropical Africa which receives its rainfall from in-flowing Indian Ocean air. A probable drop in Indian Ocean sea-temperatures at the Eocene/Oligocene boundary, associated with freezing conditions around Antarctica and invigorated oceanic and atmospheric circulation led to reduced rainfall in the Oligocene relative to the Eocene. Similarly at the Miocene/Pliocene boundary a northward movement of the sub-tropical convergence and increased re-cycling of the Agulhas Return Current reduced sea-surface temperatures east of Southern Africa. Again a terrestrial aridifying effect resulted from reduced oceanic evaporation. These climatic effects were contributary causes of faunal turnovers occurring in the Early Oligocene and at the Miocene/Pliocene boundary.

CHAPTER 10

SEDIMENT INSTABILITY

10.1 INTRODUCTION

Submarine slumping has been recognised as an important process on continental slopes of active, passive rifted, and passive sheared margins (Uchupi, 1967; Moore et al., 1970; Normark, 1974; Moore et al., 1976; Dingle, 1977 ; Hampton & Bouma, 1977; McGregor & Bennett, 1979). Mass-movement processes have affected large proportions of the continental slope (Embley, 1980). Basin-ward movement of sediments on continental slopes is dominated by gravity processes, and this has been identified as one of three major sedimentological models for deep-water basins (Gorsline, 1978; 1980; Nardin et al., 1979). Similarly, slumping has been a major factor in the formation of the continental margin off the west and south coasts of southern Africa (Dingle, 1977; 1980a; Summerhayes et al., 1979; Embley & Morley, 1980). Seismic data reported here (chapter 8) indicate that slumping is an important process in the evolution of the east coast margin too.

Mass-transport processes are important in other environments such as deltas (Walker & Massingill, 1970; Prior & Coleman, 1981) fjords (Prior et al., 1981) and aseismic plateau flanks (Roberts, 1972). A variety of instability-related phenomena have been described (Coleman & Garrison, 1977; Prior et al., 1979). In addition to large slumps, the northernmost Natal Valley exhibits a range of features such as growth faults, tensional graben and diapirs (Fig.10.1). The purpose of this chapter is to discuss the characteristics, distribution, age and causes of the various mass-movement-related features.

10.2 LARGE SLUMPS

10.2.1 Diagnostic features

Consequent upon downward and lateral movement of sediment, the 'head region' or 'proximal slump' displays tensional features, while the 'toe' or 'distal slump' shows compressional features (Fig.10.2; Lewis, 1971; Dingle, 1977; 1980a). Characteristics readily recognised on seismic profiles include:

a) fissured zone - upslope from the main glide plane scar lies an area of v-shaped depressions related to tensional faults or subsidiary glide planes;

b) glide plane scar - a steep scarp of truncated bedding is exposed by the downward movement of the slumped mass;

c) tensional depression - a valley is situated between the glide plane scar and the body of the slump;

d) slumped mass - the main body of material is bounded on its lower side by a curved glide plane which may be parallel to undisturbed beds below; in the proximal area, coherent strata are tilted and sometimes faulted antithetically; in the distal part compression is manifested in folding and thrusting; the sea-floor (or the top of the slump in older buried features) is irregular and hummocky in compressional areas becoming smooth in the most distal region.

10.2.2 Distribution

Five areas within the N.N.V. are affected by slumping - i) the Zululand continental slope south of the Limpopo Cone, ii) the continental slope off Maputo adjacent to the Limpopo Cone; iii) the eastern slope of the Inharrime Terrace; iv) and v) the valleys flanking the Central Terrace to east and west (Fig.10.2, Table 10.1).

10.2.2.1 Zululand slump

This is the largest mass-transport feature in the region, and has been mapped over an area of 20 800 km^2 extending from the southern Limpopo Cone along the Zululand continental margin. The total area of this feature appears to be $\sim$34 000 km^2, as it broadens to the south, encompassing much of the north-eastern Tugela Cone. In the region considered in detail here, it reaches a thickness of 0.52 msec (416 m). In magnitude it rivals the largest slumps recorded elsewhere on the southern African margin which are the largest described to date in the literature (Dingle, 1980a).

FIGURE CAPTIONS

10.1. Distribution of sediment-instability features. 1 = glide planes; 2 = fissures; 3 = chaotic bedding, folds & thrusts; 4 = faults defining tensional graben with downthrow marked. Thickness of Zululand and Maputo slumps contoured at 100 m sec intervals two-way-time, values m sec x 10.

Fig.10.2. Characteristic features of a slump after Lewis (1971); Dingle (1977, 1980a).

Fig.10.3. Characteristic features of a) growth fault with attendant roll-over anticline after Bruce (1973). b) Tensional graben after Prior et al. (1979). c) Diapir after Coleman & Garrison (1977).

Fig.10.4. Distribution of earthquakes. Black dot = >5 on the Richter Scale; open circle = <5 on the Richter Scale. After Fernandez (1983) and Fernandez & Guzman (1973) et seq. Slumps outlined by dashed lines (as in Fig.10.1). M.S.= Maputo Slump; I.S. = Inharrime Slump; E.C.T.S. = East Central Terrace Slump; W.C.T.S. = West Central Terrace Slump; Z.S. = Zululand Slump.

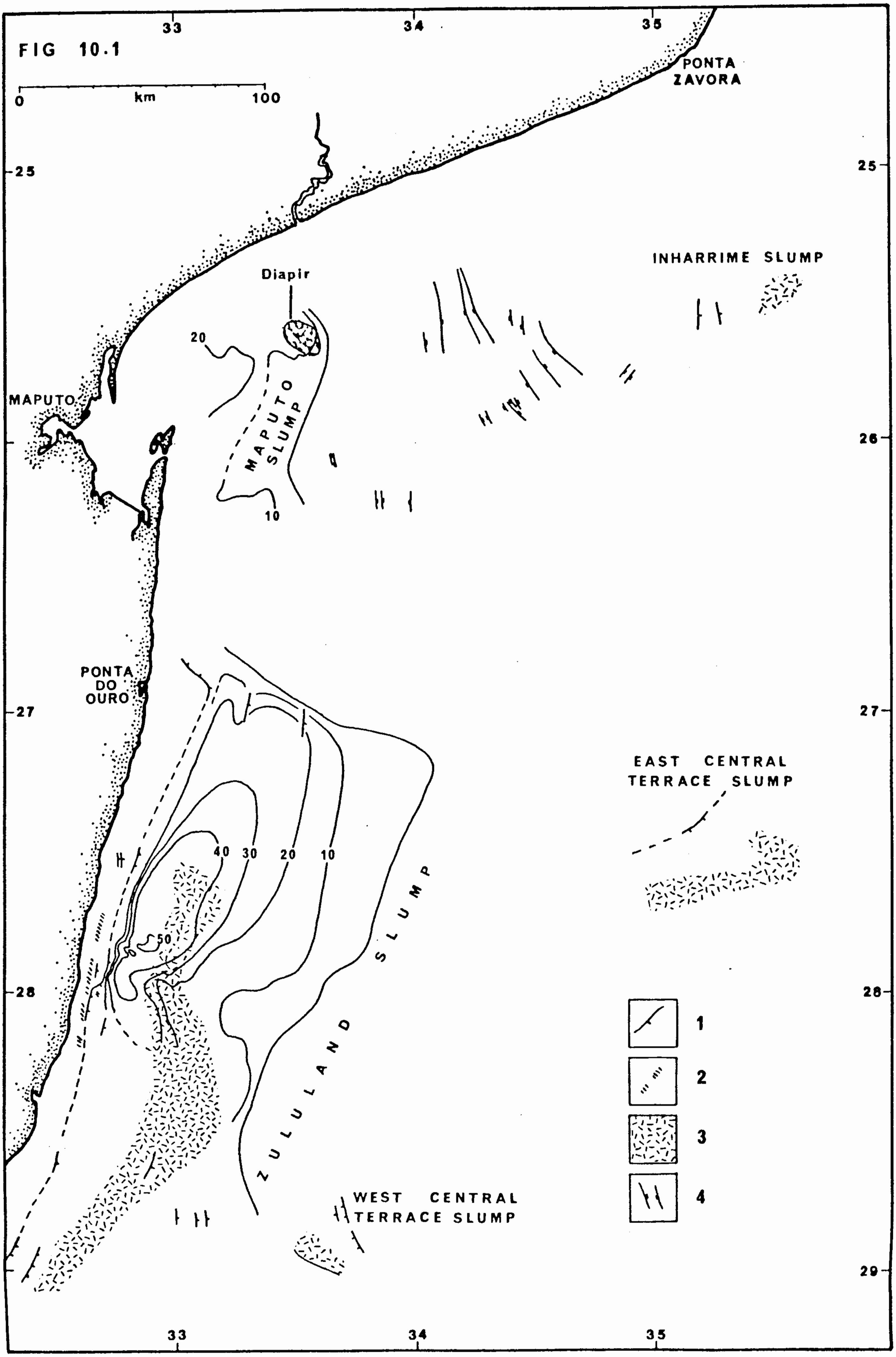


FIG 10.1
0
km
100
33
34
35
25
26
27
28
29
PONTA ZAVORA
INHARRIME SLUMP
Diapir
20
MAPUTO
MAPUTO SLUMP
10
PONTA DO OURO
EAST CENTRAL TERRACE SLUMP
40
30
20
10
50
ZULULAND SLUMP
WEST CENTRAL TERRACE SLUMP
1
2
3
4

FIG. 10.2

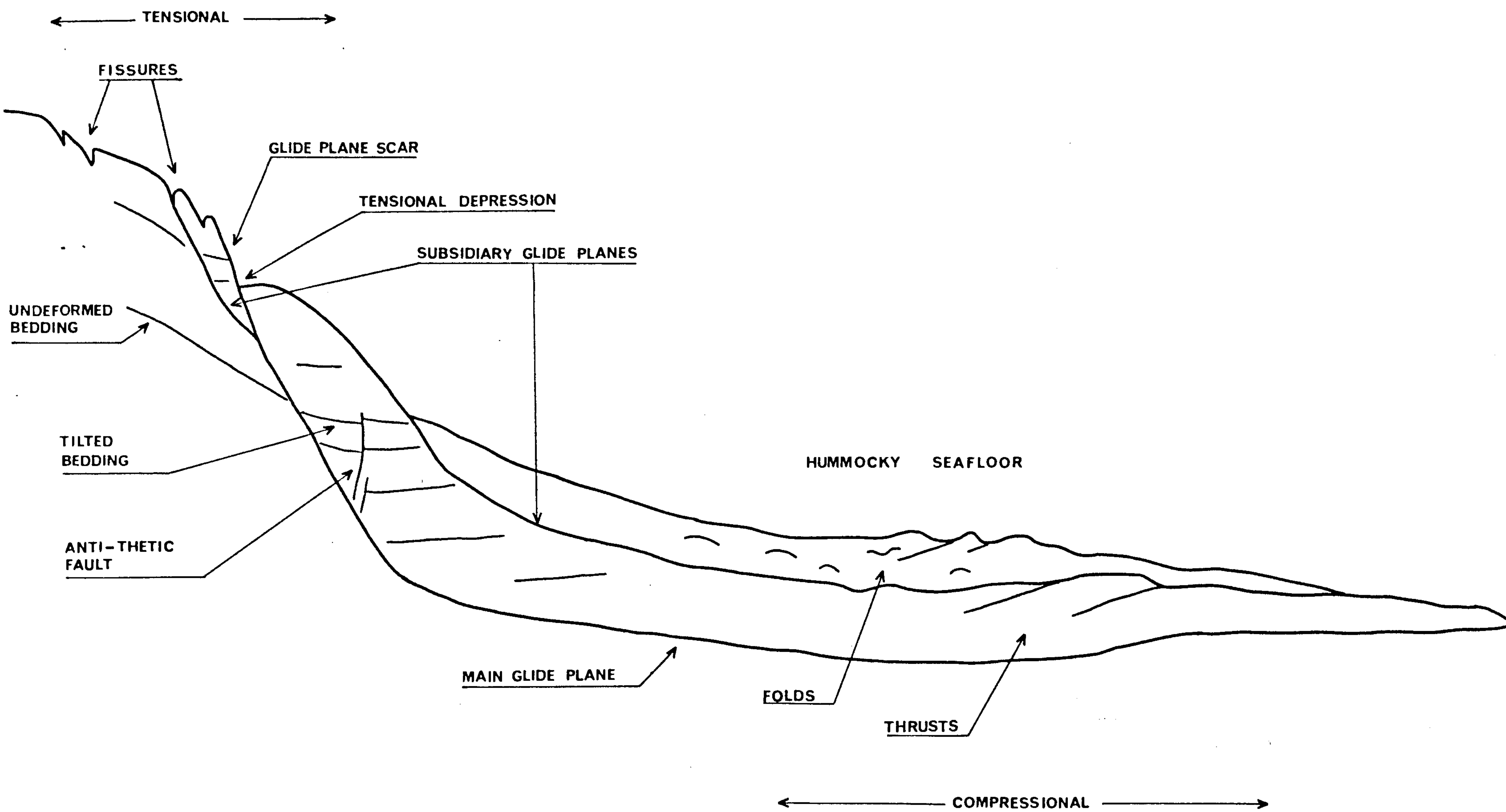
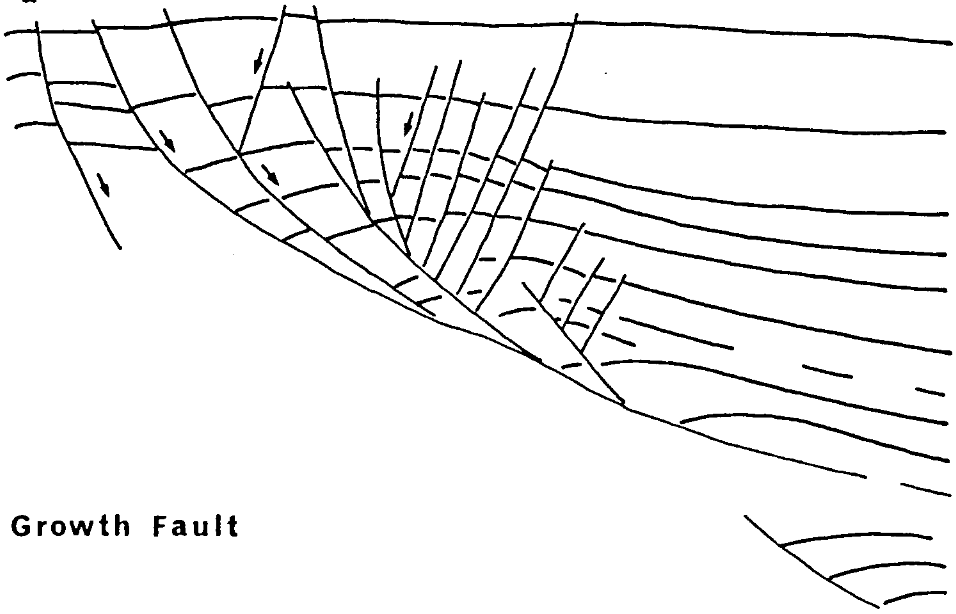

TENSIONAL
FISSURES
GLIDE PLANE SCAR
TENSIONAL DEPRESSION
SUBSIDIARY GLIDE PLANES
UNDEFORMED BEDDING
TILTED BEDDING
ANTI-THETIC FAULT
HUMMOCKY SEAFLOOR
MAIN GLIDE PLANE
FOLDS
THRUSTS
COMPRESSIONAL

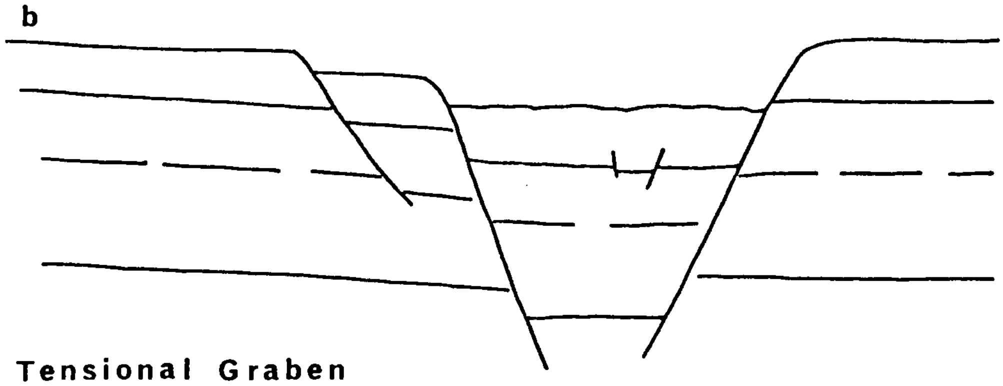
FIG 10.3

a



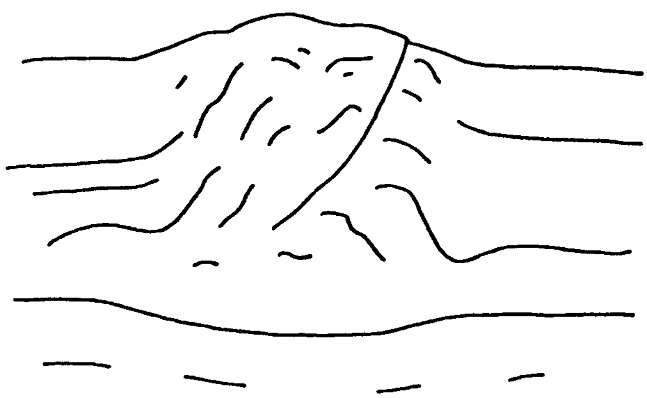
Growth Fault

b



Tensional Graben

c



Diapir

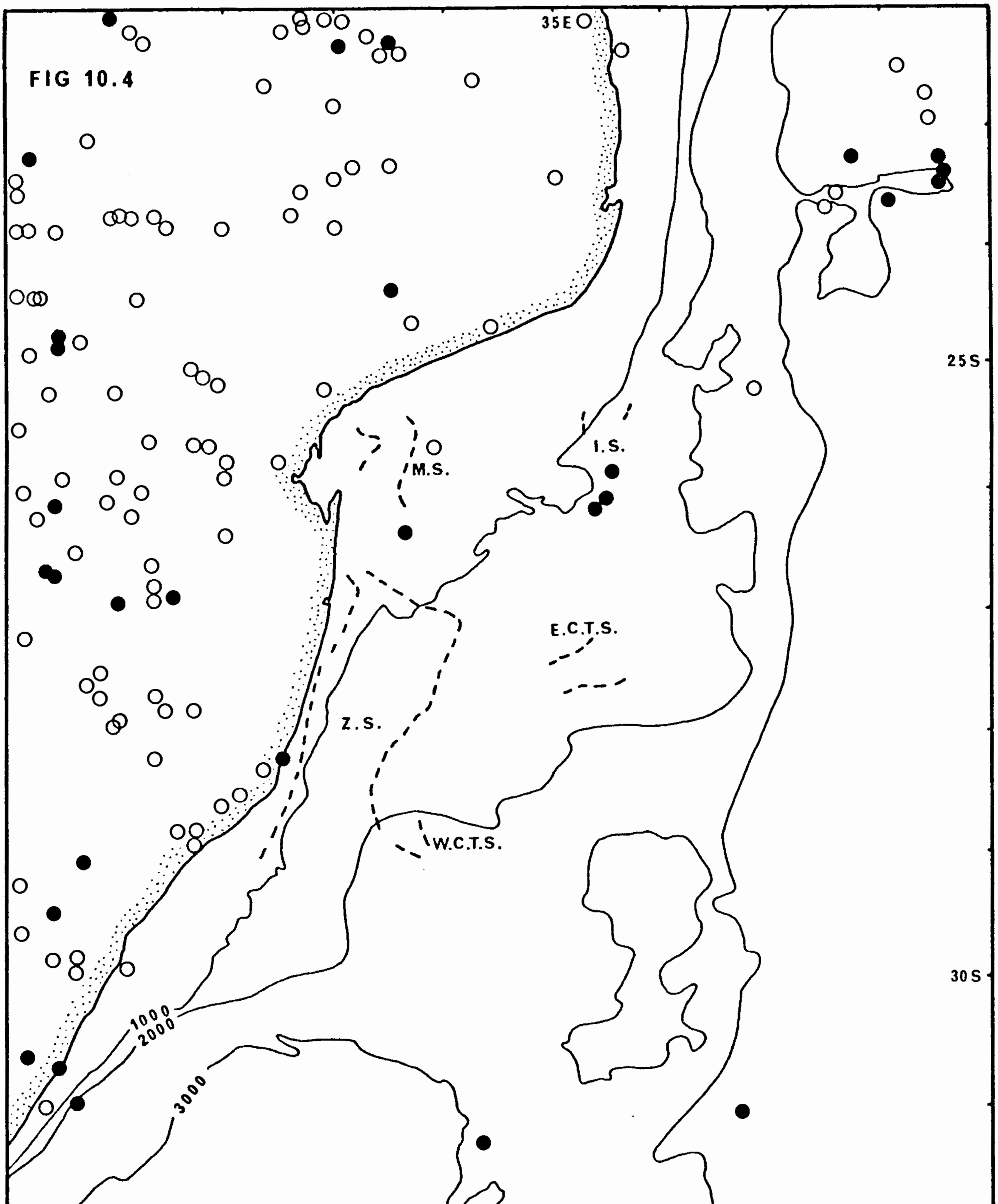


FIG 10.4

The main glide plane scar forms a 250 km long scarp which is the steepest part of the continental slope (section 2.3). Subsidiary glide planes which occur near the main glide plane, and sometimes down-slope from compressional features are also marked by elongate steep scarps (compare Fig.10.1 with section 2.3 Figs.2.6-2.14 and Goodlad, 1978). Upslope of the main glide plane lies a broad fissured zone associated with subsidiary glide planes (profiles 286, 285 & 287, Figs.8.14, 8.15 & 8.24). Some fissures are up to 112 m deep (c.f. Herzer, 1979; Dingle, 1980a). Even deeper v-shaped notches (315 m) occur on profile 287 (Fig.8.24) but these may either represent a) points where this coast-parallel line briefly intersects a major glide-plane scar; or b) deeply incised canyons superimposed on the slump-mass. Certainly, elsewhere on the slump, particularly between 27° 20'S and 28° 30'S, a system of canyons cuts the slump surface, contributing to the varied microtopography of the area (sections 2.3 and 8.4). Tensional depressions are not particularly well-developed reaching a maximum relief of ∿22 m (profile 284, Fig.8.21). To the north, under the southern Limpopo Cone, the slump-mass is buried by the slope-wedge facies of upper unit 4 age (profile 350 4, Fig.8.13), while to the south, it is largely exposed. The main glide plane coincides with reflector Jimmy over much of its extent. The exception is the northern toe of the slump which is bounded on its lower side by a reflector just above Jimmy (profiles 371 6 & 350 5b, Figs.8.25 & 8.28). Bedding in the proximal area is fairly coherent, although it is tilted and faulted - sometimes antithetically (profile 286, Fig.8.14). Downslope, bedding becomes more chaotic,and complicated compressional features including folds, thrusts and chaotic bedding are abundant. In cross-section the toe of the slump is lensoid, resembling the geometry of a distal turbidite deposit or 'fan-complex' (Sangree & Widmier, 1977). To the south, canyons and gullies interfere with the toe of the slump, masking relationships.

10.2.2.2 Maputo slump

This slump is situated on the shallow continental slope off Maputo (Fig.10.1). Data are sparse (profiles 350 16, 350 15, 350 14 & 350 13, Figs.8.6, 8.8, 8.9 & 8.10) and the slump has been mapped over only 2800 km². It is not known whether it is connected to the Zululand slump 50 km to the south, and present data do not cover the head region. It is thus suspected that the slump is areally more extensive than indicated here. The slump reaches a maximum thickness of 0.28 sec (224 m) on profiles 350 16 and 350 15 (Figs.8.6 & 8.8). Here, material appears to have flowed and cascaded eastwards, terminating in a 12 km wide diapir. The toe of the slump merges with slope-wedge facies material of the Limpopo Cone. To the south (profiles 350 14 & 350 13, Figs.8.9 & 8.10) small-scale movement is evinced by faulting and chaotic bedding. The base of the slump again coincides with reflector Jimmy, while the slump-mass is buried by sediments of upper unit 4 age.

10.2.2.3 Inharrime Slump

This slump is situated on the eastern flank of the Inharrime Terrace, with the head region in water depths of 800 m. Its areal extent is unknown as only two profiles are available in the area, but it is at least 45 km wide (profiles 350 15 & 350 14, Figs.8.8 & 8.9). The main glide plane outcrops near the distal extent of the post-'L' Inharrime Terrace (unit 5), and is associated with a group of normal down-to-the-east faults (c.f. Lewis, 1871, Fig.2). A buried tensional depression occurs on profile 350 15 (Fig. 8.8, 0240 hrs). The slump-mass is up to 0.4 sec (320 m) thick and has moved along two major glide planes. The proximal part of the slump has affected the flank of the pre-'L' slope-wedge facies which forms a terrace between 800 and 1000 m on the eastern flank of the Inharrime Terrace (section 2.3). Downslope, compression is evinced through chaotic bedding and a hummocky seafloor. The toe of the slump appears to be east of the area covered by available profiles.

Table 10.1

Quantitative data on slumps

	Zululand	Maputo	Inharrime	West Central Terrace	East Central Terrace
Length of head (km)	250	-	-	-	-
head-toe (width) (km)	90	50	45	27	>50
area (km^2)	20,800	>2,800	-	-	-
max. thickness (m)	416	244	320	136	168
volume (km^3)	3370	>248			

10.2.2.4 Slumps on the flanks of the Central Terrace

The Central Terrace is bordered by valleys both to east and west. Both of these areas appear to be influenced by slumping. To the east of the Central Terrace lies an upfaulted block of Early Miocene foraminiferal sand (Fig.8.46). The eastern face of this feature may well be a glide plane scar with a relief of 0.18 sec (135 m) (profile 350 4, Fig.8.13, 1500-1530 hrs). Comparison with the bathymetric map (section 2.3) indicates that this feature is gently curved and oriented east-north-east. To the south-east lies a tensional depression and the proximal part of a slump-mass. The eastern part of profile 350 3 (Fig.8.17) lies sub-parallel to the fault (glide-plane-scar?) and shows irregular bedding in post-Angus material. More data are required to unequivocably determine whether these features are slump-related.

A small slump may exist at the base of the slope of the valley to the west of the Central Terrace, near the western part of the Naudé Ridge (profiles 371 8 & 519, Figs.8.29 & 8.30). The exposed unit 3 horizon (profile 371 8, 0945-1030 hrs) may represent a glide plane scar. A subsidiary glide plane outcrops at 1045 hrs on profile 371 8. Steeply dipping sediments downslope from this are disturbed. The slump-mass is up to 0.17 sec thick (136 m) and 27 km wide. The toe of the feature is cut by a gully which is a lower extension of the canyon system on the Zululand continental margin.

If, as suspected from limited data, these two features are slumps then their situations are more akin to those of aseismic-plateau-flank slumps than continental slope slumps.

10.3 OTHER INSTABILITY-RELATED FEATURES

10.3.1 Growth Faults

These faults (fig.10.3a) have been variously termed 'regional faults', 'contemporaneous faults', 'down-to-basin faults', 'detached normal faults' and 'growth faults' (Harding & Lowell, 1979). There is a close relationship between sedimentation and fault displacement: sediment units increase in thickness on the downthrown side of faults, and there is a progressive increase in throw with depth (Bishop, 1973; Bruce, 1973; Weimar & Davis, 1977; Rider, 1978).

In the N.N.V., these structures are most common on the Central Terrace and are best displayed in the McDuff-Angus sequence (unit 2). Sediments thicken dramatically into down-faulted areas. Basins developed through this process are oriented east-north-east and north-south (Fig.8.38). Fault throws decrease from 0.25 sec to 0.02 sec from the stratigraphic level of McDuff to Angus. Some faults have been active in post-Jimmy times too. In many areas, faults are closely spaced and roll-over anticlines form on down-faulted sections (eg. profiles 350 5c and 371 7, Figs.8.18 & 8.20).

The largest growth fault structure occurs on the southern face of the Limpopo Cone (profile 371 6, Fig.8.25). Here, an anticlinal structure is partly resultant from the lensoid shape of the deposit and partly due to downthrow adjacent to faults. This reverses the basinward dip leading to a roll-over anticline. The crest of the anticline tends to migrate basin-ward with depth (c.f. profile 371 6, Fig.8.25, 1000-1130 hrs, with fig.10.3a). As displacement occurs along the down-to-basin or 'synthetic' faults material collapses towards the upfaulted side, resulting in antithetic faults (Bruce, 1973; profile 371 6, 1045 hrs). Up-faulted areas may be associated with buoyant material such as overpressurised shales. In the Limpopo Cone, pre-Angus (unit 2) material may be buoyant relative to post-Angus and post-Jimmy sediments.

10.3.2 Tensional Graben

Coleman & Garrison (1977) report flat-bottomed to U-shaped, faulted depressions termed 'radial graben' or 'tensional graben'. Off the Mississippi delta in water depths between 5 & 90 m these features are typically 400-700 m wide and 8 to 10 km long with throws of 10-15 m. Over 1950 km^2

of the northern Limpopo Cone and the Inharrime Terrace are characterised by broad tensional graben up to 45 km across with throws up to 0.025 sec (20 m) (Fig.10.1). The faults here affect sediment of units 4 and 5 in water depths between 350 and 500 m. There is some evidence for re-activation of faulting (growth faulting) as throws increase slightly with depth (0.015 - 0.025). The tensional graben are complex features with several faults, sometimes down-thrown towards a central graben (profile 350 15, Fig.8.8, 0950-1340 hrs), and sometimes with alternating throws (350 14, Fig.8.9, 1025-1350 hrs). Individual faulted blocks and graben within the system range down in size to 1.5 km in width, which is approaching the scale of feature described by Coleman & Garrison (1977). Most of the faults are exposed on the sea-bed, although on the western flank of the Inharrime Terrace, some are buried (profile 350 14, Fig.8.19). Thus there are many comparisons between Missi-ssippi delta front and Limpopo Cone features, with the major contrasts being the size of the graben and water-depth.

10.3.3 Diapirs

These generally result from differential weighting of salt or clay. In the Mississippi delta, sands prograding over prodelta muds cause flowage of the underlying mud (Coleman & Garrison, 1977). Off Maputo in the toe region of the Maputo slump, eastward flowage has resulted in a diapir up to 80 m in relief and 12 km across (c.f. profiles 350 16 & 350 15, Figs.8.6 & 8.8, with Fig.10.3c). If this feature is a result of differential loading, the overlying material has prograded from the east initially (upper unit 4 material) and then from the north and north-east (unit 5). In contrast, Coleman & Garrison (1977) suggest that muds flow in the same general direct-ion as overlying sands prograde.

A less clear-cut example of diapirism occurs in unit 3 material (Angus-Jimmy) at the base of the pre-Jimmy Limpopo Cone where it is overlain by thick sediments of unit 4 (profile 350 5a, Fig.8.12, 1830 and 2000 hrs).

10.4 AGE

The various instability features are dated in relation to the stratigraphy erected in section 8.2 (Table 10.2). Several phases of instability are evident. The main mass of the Zululand slump forms the lower part of sediment unit 4, and is overlain towards the north by pre-'L' sediments of the upper part of unit 4. Over most of the slump, the main glide plane coincides with reflector Jimmy. The main mass of the Maputo slump shows the same stratigraphic relationships and the two may even be connected (Fig.10.1). Here considerable thicknesses of pre-'L' (upper unit 4) sediment overlie the slump. Reflector 'L' remains undated but has been arbitrarily placed midway in the Pliocene-Recent sequence (post-Jimmy) because of the thickness of post-'L' sediments (section 8.2). The large Zululand and Maputo slumps are younger than 5 m.y. old and approximately 3 m.y. old.

These slumps are underlain by an older phase of slumping and diapirism. This phase is pre-Jimmy and post-Angus (34-5 m.y. old).

In the case of the slumps on the flanks of the Central Terrace, the glide plane of the western example exposes an horizon within the Angus-Jimmy sequence. This slump too is post Angus but no younger limit can be set on its age. The slump to the east (Fig.10.1) exposes Early Miocene sediments. Again no younger limit to its age can be set.

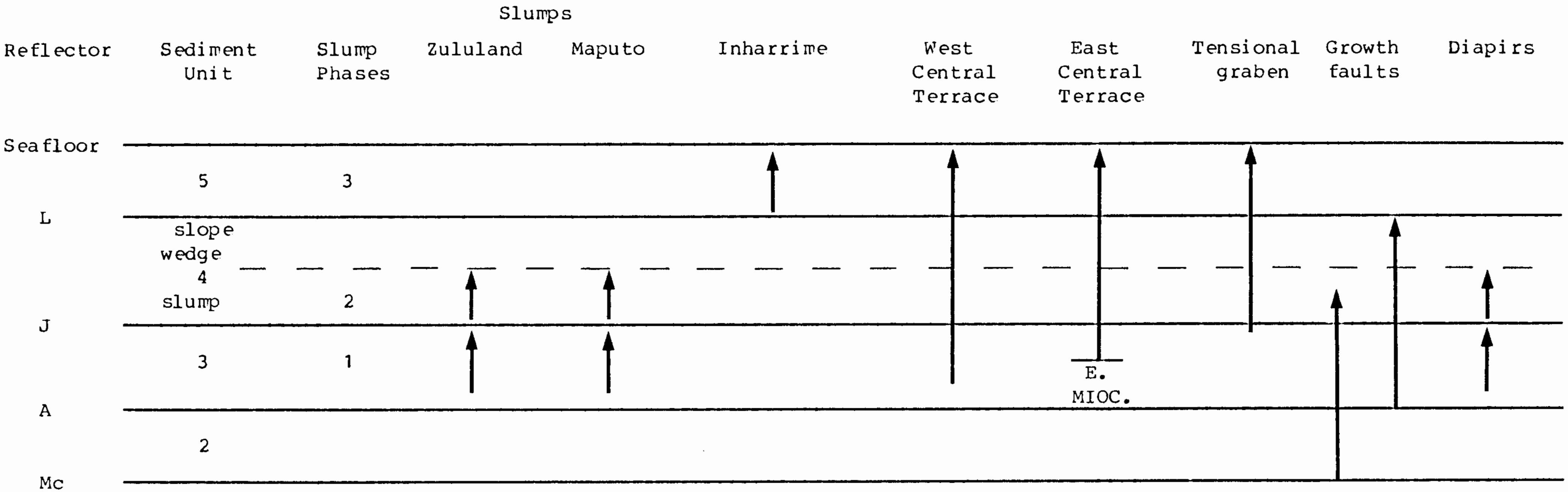
The Inharrime slump is post-'L' in age. Over most of its length it is exposed on the seafloor, but the upper head region is overlain by the youngest prograding material of the post-'L' Inharrime Terrace.

Growth faulting on the Central Terrace is best developed in the McDuff-Angus (unit 2) sequence but some faults have offset reflector Jimmy and remained active in post-Jimmy time. The large roll-over anticline and growth fault structure of the Limpopo Cone is best developed in the Angus-'L' (units 3 & 4) sequence.

Tensional graben are widespread in the Limpopo Cone and affect sedimentary units 4 and 5. Many faults are exposed on the seafloor and may still be

Table 10.2

Age of Sediment Instability



The arrow lengths indicate the best constraint available on dating the various features. The two arrows under growth faults refer to (on the left) the Central Terrace and (on the right) the southern Limpopo Cone.

active.

Diapirism is associated with both the pre-Jimmy and post-Jimmy phases of the Maputo slump, and the northern Zululand slump.

Elsewhere on the southern African margin, four phases of large-scale slumping have been recognised (Dingle, 1980a). These are dated as Lower Albian, Palaeocene, Eocene-Pliocene and Pliocene-Recent. To date, on the east coast margin Palaeocene and Cretaceous slumps have not been recognised. The older phase of sediment failure underlying the Zululand and Maputo slumps (Oligocene-Pliocene) may be equivalent to Eocene-Pliocene slumps on the west coast. The major Pliocene-Recent slumps of the south and west coast are suspected to be as young as Late Pleistocene (Summerhayes et al., 1979; Dingle, 1980a). In contrast the main post-Jimmy Zululand and Maputo slumps are pre-'L' in age and may be ∿3 m.y. old. Slumps on the Central Terrace flanks are post-Angus and post-Early Miocene and may be relatively modern. The Inharrime slump is post-'L' in age and these latter three <u>may</u> be equivalents of Pliocene-Recent, possibly Late Pleistocene slumps on the west and south coast margins.

10.5 TRIGGER MECHANISMS

Trigger mechanisms for sediment mass-gravity-transport have been reviewed by Middleton & Hampton (1976) and Rupke (1978). Failure along a slip plane will result when the shear stress exceeds the shear strength. Therefore either an increase in shear stress or a decrease in shear strength may trigger movement, and this may be achieved by a) thickening of the sediment pile by gradual deposition beyond a limiting factor or by sudden re-sedimentation; b) oversteepening of the slope due to undercutting by currents or waves or sediment failure downslope; c) increases in pore pressure due to compaction, gas/clathrate phase changes, or gas production (Carpenter, 1981); d) fluidisation or thixotropy induced by shocks or impacts such as earthquakes, tsunamis or hurricane waves. Certain localities where sedimentation rates are high, sediment type is variable, and gas is biogenically produced may be more prone to sediment failure (Coleman & Garrison, 1977). Dingle (1980a) has pointed out that lower sea-level may be an important modifier to several of these processes in that a) it may cause isostatically-induced instability and increased seismicity; b) it results in modified sedimentation patterns (c.f. section 9.3.2); c) it brings the shelf edge and upper slope within the effects of storm-wave accelerations.

In the case of growth faults, tensional graben, and diapirs, proposed mechanisms for movement are related to variability of sediments which build out into unconfined depocentres such as deltas or continental slopes. Growth faulting which is most common in the McDuff-Angus sequence is not associated with high sedimentation rates (Table 8.2) but may be related to the suspected euxinic nature of pre-McDuff material (over-pressurisation of black shale?). Tensional graben are most likely due to seaward flowage of material resulting in tensional collapse. The site of the most clearly defined diapir suggests that differential loading by overlying on-lapping sediment caused instability.

Considering the large slumps in the area, all four mechanisms listed above may have been causal. The early phase of slumping on the Zululand margin and the slump on the west flank of the Central Terrace affected sediments which accumulated relatively slowly, whereas post-Jimmy slumps occurred in regions of high accumulation rates (Table 8.2). This, and variations in sediment type may have led to instability. Down-slope under-cutting by current action is also a possibility since the Agulhas Current has been an active agent in sediment dispersal at least since the Oligocene (chapter 9). Numerous bright spots are evident on the seismic profiles. Although this is indicative only of a large impedance contrast, it is possible that the contrast may be due to gas/clathrate interfaces (White, 1977; 1979; Tucholke et al., 1977; Shipley et al., 1979). Gas-related or pore pressure changes may have been important triggers. Lastly, shocks may result from earthquakes or from the passage of severe storms. Many tropical cyclone tracks cross Madagascar and reach the Mozambique coast (Davies, 1972, Fig.21). These would not be expected to exert influence in the water depths considered here, unless a similar climatic scenario existed during low sea-level stands. On the other hand, earthquakes are quite frequent in the area, even though it is a mid-plate region. At least 30 events of magnitude greater than 5 on the Richter scale have been recorded or estimated in the last 350 years (Fig.10.4)(Fernandez, 1983; Fernandez & Guzman, 1973, et seq.) Offshore, the Almirante Leite Bank (26°S, 35°E) and a bathymetric high centred at 22° 30'S, 38°E are seismically active. The seismicity in Mozambique is considered to be a southerly continuation of tectonism associated with the east African rift system (Fairhead & Girdler, 1971; Fairhead & Henderson, 1977). The rift valleys have been active at least throughout Neogene times and may have a tectonic history extending back to the Jurassic (Burke & Whiteman, 1973), and beyond (McConnell, 1980). It is likely therefore that similar seismicity has prevailed during the known history of slump activity (Table 10.2). Both the Zululand and Inharrime slumps are buried to a greater or lesser degree by later sediments and cannot have been triggered by the nearby historical seismicity (Fig.10.4) but similar or larger shocks may have initiated them in the geologic past.

10.6 SUMMARY

A variety of sediment instability and mass-gravity features occur within the N.N.V. These include a) large slumps, b) growth faults and associated roll-over anticlines, c) tensional graben and d) diapirs. Five large slumps have been mapped, three of which occur on the continental slope while two flank the Central Terrace. The largest is in excess of 20 000 km^2 in area, possibly up to 34 000 km^2, and is amongst the largest cited in the literature to date. Glide plane scars up to 250 km in length form the most prominent scarps on the continental slope. Proximal areas comprise large rotated sediment masses up to 416 m thick, while downslope, chaotic bedding and thrusts result from compression. Trigger mechanisms may include earthquakes, sediment undercutting through current action, or relatively fast deposition of variable sediment types. At least three phases of slumping can be recognised: Angus-Jimmy, Lower Jimmy-'L', and post-'L' (34-5 m.y., ∿5-3 m.y., and post ∿2.5 m.y.).

Growth faulting is almost ubiquitous under the Central Terrace in sedimentary unit 2 (McDuff-Angus ∿ 98-34 m.y. ago), with some reactivation of faults in post-Jimmy times (<5 m.y.). A large growth-fault complex with associated roll-over anticline occurs on the southern Limpopo Cone, predominantly in post-Jimmy sediments. complex tensional graben extend over an area of 1950 km^2 on the Inharrime Terrace and upper Limpopo Cone. Several faults are exposed on the sea floor and affect both Jimmy-'L' and post-'L' sediments. Diapirs occur in pre-Jimmy and Lower Jimmy-'L' sediments of the Limpopo Cone. Differential loading and flowage of sediments deposited in an unrestricted depocentre are the most likely causes of the latter three types of instability.

PART 4

CHAPTER 11

SUMMARY

11.1 REGIONAL SETTING

The Natal Valley is a N-S trending sedimentary basin lying between the south-eastern coast of Africa and the Mozambique Ridge. It is a southerly extension of the Mozambique coastal plain basin, while to the south it merges with the abyssal Transkei Basin. Bathymetrically, it is atypical of passive Atlantic-type continental margins - broad regions lie at intermediate rather than abyssal depths. The climate is governed by a sub-tropical high pressure cell. Hinterland run-off is moderate. Physical oceanography is dominated by the Agulhas Current which is part of the south Indian Ocean western boundary system.

11.2 PLATE TECTONIC RECONSTRUCTIONS

11.2.1 West Gondwanaland refit

A new reconstruction for South America and south central Africa is derived: euler pole 46.75°N, 32.65°W, 56.4° counter clockwise rotation of South America relative to Africa. This is based on fitting the north-eastern apex of the Falkland Plateau into the re-entrant formed by the sheared Agulhas margin and the southern face of the Tugela Cone. Simultaneously, known Precambrian outcrops in northeastern Brazil and in the Gulf of Benin are juxtaposed rather than overlapped. The fit is also constrained by alignment of at least three tectonic features crossing from one continent to another: (1) the geophysically defined eastern and western boundaries of the submarine Jurassic Outeniqua Basin of South Africa and the Falkland Plateau Basin; (2) the Late Precambrian transcurrent fault and mylonite belts of Pernambuco (Brazil) and Foumban (West Africa); (3) the Triassic northern tectonic front of the Cape Fold Belt and the major morphological feature on the Falkland Plateau with which it is closely lined up. Isotopic ages of Falkland Plateau gneisses correspond to Cape Pluton and Cape fold Belt ages, suggesting their palaeoposition lies within the realm of the Cape Fold Belt. The fit implies that ~200 km of relative movement has occurred between north-west Africa and south-central Africa. The reconstruction is considered pre-drift rather than pre-rift.

11.2.2 Mesozoic magnetic seafloor spreading anomalies in the Natal Valley

The above revised reconstruction is corroborated by the identification of anomalies M10-M0 in the Natal Valley. Poorly formed magnetic anomalies close to the southern face of the Tugela Cone may represent anomalies M10N-M12 deformed by the rifting process. Half-spreading rates are 1.3 - 1.4 cm yr^{-1} for M0-M4 and 1.6 cm yr^{-1} for M4-M10. These anomalies are offset ~1300 km from their equivalents in the southern Cape Basin. This is an improved estimate of the offset along the Falkland Agulhas Fracture Zone (FAFZ).

11.2.3 Falkland Plateau palaeopositions

Using Natal Valley and Georgia Basin anomalies, I have computed a sequence of Cretaceous palaeopositions of the Falkland Plateau relative to southern Africa. In all cases conjugate sets of Cape and Argentine Basin anomalies were simultaneously juxtaposed. The offset along the FAFZ remained ~1300 km long throughout M10-M0 time, and by anomaly 34 time, the offset was ~1270 km. Therefore no major ridge-jumps had occurred by then.

Dating M0 as 108 m.y. (Larson & Hilde, 1975) and anomaly 34 as 80 m.y. (LaBrecque et al., 1977) the average half-spreading rate immediately south of the FAFZ for the Cretaceous normal polarity epoch is 4.2 cm yr^{-1}. Using this, I date: a) the change in early pole of rotation at 105 m.y.; b) a reconstruction which juxtaposes salt boundaries in the Brazil and Angolan Basins at 103.7 m.y.; c) final separation of the Falkland Plateau from southern Africa at 98.3 m.y.; d) formation of the oceanic northern part of the Agulhas Plateau at 97.3-90.7 m.y.

An alternative position for the continent-ocean-boundary (COB) in the southernmost Cape Basin which is implied by the model summarised in 11.2.1 - 11.2.3, has recently been substantiated by multi-channel seismic profiling

(Austin & Uchupi, 1982). This refutes the reconstruction of Rabinowitz & LaBrecque (1979). The pre-rift reconstruction of Vink (1982) is invalidated because: a) the South Atlantic rift did not necessarily propagate from south to north along its entire length; b) crustal extension is not required to increase with progressive propagation of a rift. Comparison of magnetic data with the COB positions in the southernmost Cape and Argentine Basins and the Georgia Basin suggests continental separation began 122-127 m.y. ago, with undeformed magnetic anomalies dating from 122 m.y. (M10). These dates are consistent with micropalaeontological and sedimentological data around southern Africa.

11.2.4 Evolution of the Agulhas Plateau and its relationship to other aseismic plateaux

This feature is not the site of a ridge-jump. Islas Orcadas Rise in the Georgia Basin, Meteor Rise in the southern Agulhas Basin and Agulhas Plateau off South Africa are all ~93 m.y. old and all originated together as a result of excess volcanism on the spreading ridge. Meteor Rise and Islas Orcadas Rise, forming a composite feature, separated from the Agulhas Plateau between 93 and 80 m.y. ago. Between 65 and 58 m.y. Islas Orcadas and Meteor Rise were separated by a well-documented ridge-jump (Du Plessis, 1977; LaBrecque & Hayes, 1979; Barker, 1979). It is suggested that the southern continental part of the Agulhas Plateau moved to its present position relative to Africa between 122 m.y. and 108-105 m.y. while it was part of the South American plate.

11.2.5 A revised fit of east and west Gondwanaland - plate tectonic status of the northernmost Natal Valley

A revised sequence of successively older Cretaceous and Jurassic reconstructions has been prepared for East Antarctic relative to Africa.

Reconstruction of a triple junction configuration south of Madagascar for anomaly 29 implies that identified anomaly 34 crust of the Antarctic plate is matched by anomaly 34 crust east of the Prince Edward Fracture Zone on the African plate. The resultant anomaly 34 reconstruction implies that identified Mesozoic crust off Dronning Maud Land lies west of the Prince Edward Fracture Zone and is equivalent to the easternmost identified Mesozoic crust in the Mozambique Basin. Application of the M2-M21 rotation of Madagascar (Segoufin & Patriat, 1980) to the new M2 fit for Antarctica results in a revised reconstruction for east and west Gondwanaland: euler pole 1.67°S 35.99°E, rotation angle 53.43°. Although the refit was achieved from interpretation of seafloor spreading and fracture zone anomalies it places East Antarctica adjacent to the Lebombo mountains, and south of and sub-parallel to the Sabi acid volcanic suite, bearing out the premise that these rocks mark a Mesozoic plate boundary. Moreover the fit is supported by the alignment of two pre-drift tectonic features: a) the northern front of the Cape Fold Belt and its Antarctic equivalent marked by the outcrop limit of the Whiteout Conglomerate of the Ellsworth Mountains in their relocated position and the Gale Mudstone in the Pensacola Maountains; b) the western front of Pan-African reactivation of the Mozambique Mobile Belt and the Dronning Maud Land equivalent marked by the western limit of the Sverdrupfjella metamorphic suite. As in the West Gondwanaland Case (11.2.1), the continents have not broken along pre-existing lineaments - rather they have cut across them. This runs contrary to the assumption within the 'Wilson Cycle' hypothesis that new rifts exploit pre-existing zones of weakness.

Initially oceanic crust was emplaced in coastal Mozambique and the northernmost Natal Valley, with East Antarctica moving along a transform fault zone immediately east of and sub-parallel to the Lebombo mountains and the eastern Tugela Cone. At M10 times a triple junction formed near the tip of the Falkland Plateau implying that the oldest magnetic anomalies in the eastern Weddell Sea cannot pre-date M10. At M2 times the Mozambique and Astrid Ridges separated in a ridge-jump episode. If the section of spreading ridge involved in the jump extended east of the Mozambique Ridge, this could explain the existence of high amplitude magnetic anomalies in the southern Mozambique Basin. Gunnerus Ridge, Conrad Rise and Del Cano Rise are all linked tectonically with the Madagascar Ridge.

11.2.6 Structural style in coastal Natal

The above revised Gondwanaland reconstruction implies that most of the continental margins of eastern Africa are sheared rather than rifted. Strike-slip controlled margins include the Agulhas margin, the Lebombo Mountains and the Davie Fracture Zone. Areas adjacent to these margins have traditionally been considered to have rifted apart under tensional forces. Correspondence of the dating of the faults of Coastal Natal with magnetic dating of continental rifting suggests that the faults are associated with Gondwanaland break-up. The nature and orientation of faulting in Natal seems to be consistent with a dextral strike-slip model. The following features are diagnostic: en échelon fault configuration; reverse 's' plan of fault traces (sensu Wilcox et al., 1973); reversal of fault throw along strike; faults which are curved, 'dog-legged', intersecting, splayed or anastomosing; pull-apart graben resulting from side-stepping faults; upraised tilted blocks possibly caused by convergence of curved faults; down-dropped basins due to diverging faults; conglomeratic lacustrine facies in fault bounded basins. The fault-controlled nature of the Natal coast south of Durban is substantiated. A major problem in this interpretation is the lack of reported thrust faults in Natal. Offshore seismic data has however revealed a series of thrusts affecting a mid-Cretaceous horizon (chapter 8).

11.3 THE NORTHERNMOST NATAL VALLEY SEDIMENTARY BASIN

11.3.1 Basin configuration and in-fill

The northernmost Natal Valley is a continuation of the Mozambique coastal plain sedimentary basin. Five physiographic sub-provinces are recognised within the region: continental shelf, Limpopo Cone, Inharrime Terrace, Central Terrace and Almirante Leite Bank. Acoustic stratigraphy is established in relation to five regionally developed acoustic horizons: basement is pre-

Barremian, reflector McDuff is ∿Cenomanian, reflector Angus is Oligocene reflector Jimmy is Late Miocene/Early Pliocene, and reflector 'L' lies within the Pliocene-Recent. Structure contour and isopachyte charts have been constructed for these horizons and the sedimentary units delineated by them.

Seismic and dredging data indicate that basement is volcanic and is in continuity with the Lebombo volcanic suite. Basement lies generally between 2000 and 4500 m below sea-level except in the Almirante Leite Bank where volcanic seamounts shoal to 71 m depth. This volcanic complex and the Naudé Ridge form two basement highs. Back-tracking of bathymetry suggests basement was emplaced sub-aerially, explaining the 27 m.y. delay between initiation of seafloor spreading and marine incursion.

Sedimentary unit 1 (basement - McDuff) infills basement depressions through on-lap and divergent fill, reaching a maximum known thickness of 1.1 secs in a valley on the west flank of the Central Terrace just north of the Naudé Ridge. By McDuff times the Central Terrace is more clearly delineated as an irregularly up-faulted spur relative to flanking valleys, with a relief of over 1.0 sec.

The configuration of reflector Angus is smoother. The Limpopo Cone was already developed and sediment had topped part but not all of the Naudé Ridge. Thickness of sedimentary unit 2 (McDuff - Angus) is very variable as a result of the uneven McDuff surface, reaching a maximum of 1.35 sec just north of the Naudé Ridge. Sedimentation is dominated by growth faulting in irregular horsts and graben. A fan complex infills the valley to the west of the Central Terrace at the base of the continental slope. To the east in the uppermost part of unit 2, sediments are deposited in contourite mounds (sensu Vail et al., 1977).

Reflector Jimmy is also generally smooth. Sedimentary unit 3 (Angus - Jimmy) represents a progradation of the Zululand continental margin and the Limpopo Cone, with maximum thickness in the former area reaching 0.7 sec.

On the Central Terrace unit 3 drapes the Naudé Ridge to the southwest, but is absent to the east.

Sedimentary units 4 and 5 cannot be differentiated over the Central Terrace, because reflector 'L' is not developed there. Post-Jimmy sediment is thin re-worked or absent in this area.

Reflector 'L' is a smooth acoustic horizon marking depocentre shifts in the Limpopo Cone. Unit 4 (Jimmy - 'L') is sub-divided: the lower sequence is dominated by massive slumps on the Zululand and Maputo continental slopes; partly contemporaneous and partly over-lying this is a large 'slope-front wedge' (sensu Galloway & Brown, 1973) emanating from Ponta Zavora in the north-east. Maximum thicknesses (0.73 secs) developed on the Limpopo Cone.

Unit 5 sediments of the Inharrime Terrace form a thick 'slope-front wedge' (0.44 sec) with a nested sequence of leveed channels. Post-'L' material also occurs off the Limpopo River mouth and on the southern Limpopo Cone.

Total sediment thickness reaches a maximum of 3.5 km under the continental slope off Maputo and Zululand. The sediment pile is not as thick as in other marginal basins around Africa (c.f. Dingle, 1982). Sedimentation rates are relatively high for sediment unit 1, low for units 2 and 3, increasing over five-fold in units 4 and 5. Accumulation rates calculated from total sedimentary basin volume and Pliocene-Recent volume are between 30 and 25 times less than estimated present-day sediment supply rates. Either sediment was exported from the basin (via current action), or present-day rates of supply have increased drastically relative to the geologic past.

Depositional style evolves from uplap-dominated in units 1 and 2, to an alternation of off-lap on-lap. This is typical of rift and post-rift basins (Brown & Fisher, 1977). All three deep basin facies models are displayed: 1) continuous or 'oceanic' processes - biogenic sedimentation and geostrophic current dispersal; 2) slope and base-of slope processes dominated by mass-gravity events; 3) canyon-fan processes.

11.3.2 The influence of geostrophic currents

Broad areas of the N.N.V. which are shallower than 2000 m deep are influenced by the Agulhas Current - part of the western boundary system of the South Indian Ocean. Current action has: a) prevented sedimentation on Almirante Leite Bank seamount flanks; b) exposed Cretaceous and Tertiary outcrops and re-worked thin Neogene deposits on the Central Terrace; c) scoured the outer shelf and upper continental slope; d) helped restrict fast Neogene sedimentation to the north-west of the basin; e) led to glauconite and phosphorite authigenesis under an upwelling cell. This demonstrates the influence of sub-tropical gyres as opposed to the well-documented effect of cold abyssal currents.

Current action was possibly a contributing factor in the development of mid-Cretaceous (McDuff) and Cretaceous-Tertiary boundary hiatuses. Clear current influence dates from late unit 2 times (pre-Angus) and was a definite factor in the formation of an Oligocene hiatus (Angus). This corroborates the premise that establishment of the psychrosphere near the Eocene-Oligocene boundary invigorated surface currents. Changes in the current regime occurring near the Miocene/Pliocene boundary are associated with Antarctic ice-sheet growth, Arctic ice build-up and new bottom-water circulation patterns. The counter-current and up-welling cell moved east in post-'L' times in response to depocentre shifts.

The Agulhas Current controls sea-surface temperatures and hence evaporation rates off the east coast of southern Africa. This in turn influences the climate over much of the southern African sub-continent which receives its rainfall from in-coming air from the South Indian Ocean. Changes in the Agulhas Current system associated with cooler sea-temperatures and Antarctic climatic events led to reduced rainfall in the Oligocene and at the Miocene-Pliocene boundary. This constituted a forcing factor in Early Oligocene and Miocene-Pliocene boundary terrestrial faunal turn-overs.

11.3.3 Mass-gravity processes

Sediment instability in the N.N.V. is manifested in large slumps, growth faults and associated roll-over anticlines, tensional graben and diapirs. Five large slumps have been delineated. Two occur on the Zululand and Maputo continental slope, one occurs on the east flank of the Inharrime Terrace, while two less well documented features occur on the flanks of the Central Terrace. The biggest slump rivals the largest known slumps in the world being at least 20 800 km^2 in area, possibly extending over 34 000 km^2. Proximal areas, comprising large rotated sediment masses are up to 416 m thick, while distal areas display compressional features such as thrusts and folds. Three phases of slumping are recognised: Angus-Jimmy, Lower Jimmy-'L', and post-'L'. Trigger mechanisms may include earthquakes, sediment undercutting through current action, or fast deposition of variable sediments.

Growth faulting is almost ubiquitous under the Central Terrace in sediment unit 2. A large growth fault complex occurs on the southern Limpopo Cone, predominantly in post-Jimmy material. Complex tensional graben affect over 1950 km^2 of the Inharrime Terrace and upper Limpopo Cone. Faults affect unit 4 and 5 sediments and sometimes are exposed on the seafloor. Diapirs occur in pre-Jimmy and Lower Jimmy-'L' sediments of the Limpopo Cone. Deposition of variable lithologies in an unrestricted basin results in differential loading, flowage, and tension leading to the latter three types of instability.

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APPENDIX 1

METHODS

A 1 PLATE TECTONIC ROTATIONS

This work is based on the theory of seafloor spreading (Hess, 1962) and the production of marine magnetic anomalies through magnetic field reversals (Vine & Mathews, 1963). Plate tectonics on a sphere (McKenzie & Parker, 1967; Morgan, 1968) relies on Euler's theorem which states: "if a rigid body is turned about one of its points taken fixed, the displacement of this body from one given position to another is equivalent to a rotation about some fixed axis going through the fixed point". Flaws in the rigid plate hypothesis are discussed in section 4.4. Reviews of how the theory of continental drift gained respectability and evolved into the plate tectonic paradigm are given by Marvin (1973) and Cox (1973). Rotations were performed using programmes written by C.J.H. Hartnady based on a method set out by Phillips & Forsyth (1972). Finite rotations on a sphere are performed by co-ordinate transformation. The co-ordinates of any point after rotation about a pole are found by multiplying the position vector of the point (in Cartesian co-ordinates) by a tensor representing the rotation. The rotation tensor, T can be regarded as the product of three tensors: the first, A,transforms the point into the co-ordinate system of the rotation pole; the second, R performs the rotation about this pole; the third transforms back to the original co-ordinate system and is the inverse of the first tensor, A^{-1}. As these tensors are orthogonal $A^t = A^{-1}$ where A^t is the transpose of A and the rotation tensor is $T = A^tRA$ Further calculations are given by Phillips & Forsyth (1972, p.1593-1594).

Co-ordinates of features were grouped into files and then a file was created for the co-ordinates of the rotated points of each trial rotation. Files were either viewed on the computer graphics package for selection, rejection or adjustment or were sent to the printer. Computer printouts were evaluated and selected diagrams draughted.

A 2 CONTINUOUS SEISMIC PROFILING

This technique dates from the second world war when sub-bottom features were recorded by echo-sounders (Hersey& Ewing, 1949; Hersey, 1963). Since then a variety of sound sources have been developed such as air-guns sparkers, boomers and pingers.

A2.1 Principle of operation

An energy supply activates a transducer producing a sound source. Sound waves are partly transmitted and partly reflected by interfaces in acoustic impedance of the transmitting media. A receiver converts reflected pressure waves to electrical signals. These are amplified, filtered and recorded, usually graphically and magnetically.

Four parameters of the sound source affect performance of a system.

1. Frequency spectrum.

   Lower frequency, longer wavelength sound penetrates farther into sand and rock (e.g. Peters, 1967; McQuillan & Ardus, 1977).

2. Pulse length

   The pulse length equals pulse duration (time) multiplied by the sound velocity. The minimum layer thickness resolved equals half the pulse length. This situation is complicated by the tow-configuration of the transducer, which may lengthen the effective pulse length. Because the sound produced is either omni-directional or only partly focussed, various ray-paths occur between sound source and hydrophone array (e.g. D'Olier, 1979).

3. Repetition rate

   When a sound pulse is transmitted, various reflected signals are received from various depths and recorded at the appropriate "two-way-travel time" position. Each pulse produces a series of returns. These are successively recorded graphically on recorder paper as it passes a stylus, and a seismic profile is built up. The density of coverage on the profile is therefore a function of repetition rate of the sound source, paper feed rate and ship's speed over ground. If the energy source is capable of activating the sound source very quickly then more pulses per unit time are transmitted and more data received.

4. Energy introduced

A related factor is the energy transmitted to the transducer each pulse. This may be reduced to allow more frequent repretition or vice versa. The more power transmitted the greater the penetration. High-frequencey transducers cannot handle the high energy used to activate low-frequency devices. In general, therefore penetration is lost and resolution improves as higher sound frequencies are used.

A.2.2 Energy Sources

In sparkers, boomers and pingers, electrical power is used. During the Natal Valley work, single-channel seismic work was carried out using Bolt air-guns. Energy is supplied by a compressor, and air is released explosively into the water by an electrically triggered chamber. Firing rate and head of pressure are controlled by compressor capacity. A drawback in the system relates to oscillation and collapse of air-bubbles which increases pulse length, reducing resolution (eg. McQuillin et al., 1979). In our profiles the 'bubble-pulse' was generally 100 msec behind the initial pulse, allowing distinction of the two.

A.2.3 Acoustic receivers

Arrays of hydrophones were towed in oil-filled flexible buoyant tubes. Hydrophones convert pressure pulses to electrical signals.

A.2.4 Recording and processing

Incoming signals were amplified and passed via pass-band filters to recorders. Generally the frequency used was in the range 25 - 300 Hz although Mr. J. Engelbrecht fitted two recorders and separately recorded 25 - 300 and 185 - 700 Hz sections of the noise spectrum. EPC graphic recorders with electro-sensitive paper were used, and unfortunately no magnetic records were taken. Data from an older cruise was recorded on a Mufax wet-paper system.

A.2.5 Seismic profiles - distortions and interpretation

Seismic profiles are usually vertically exaggerated, scale distorted and marred by acoustic artefacts. Horizontal scale depends on recorder settings and survey speed while vertical scale depends on two-way-travel time of sound in various media (eg. Palmer, 1967; McQuillin & Ardus, 1977; E.G.&G., 1977; McLelland Engineers, 1982).

Various acoustic artefacts arise from the geometry of the system. The water-air interface has a high acoustic impedance and reflected signals are again reflected downwards towards the sea-floor, resulting in multiple travel paths. Along sloping reflecting horizons, the slope of the most common type of the multiple is twice that of the first return, and is thus easily recognised.

Some sound sources are omni-directional and even focussed-beam systems transmit a cone-shaped beam of sound towards the sea-floor. Thus reflections can be recorded from features not directly below the transducer. As a transducer is towed over a protusion on the sea-floor, signals are returned both from the sea-floor directly below the transducer and from the protusion reflecting signals from the outer area of the transmitted cone of sound. This produces a characteristic hyperbolic return. Where several prominences are closely spaced, hyperbolic returns are caused by each, giving a characteristic profile fro a rough sea-bed or reflecting horizon (eg. Krause & Kanaev, 1970).

When profiling over a sea-floor depression, the cone of transmitted sound results in three returns from the same pulse. A characteristic "bow-tie" profile is the result. If the curvature of the hollow or syncline is greater than its depth below the transducer, a bow-tie is not produced, but reflections are received from various parts of the hollow, resulting in the feature appearing shallower than it is in reality.

A.2.6 Cruise data

Sesimic data from cruises 246, 267, 277, 291, 350 and 371 of the Thomas B. Davie, cruise AII 93 of the Atlantis II and cruise 76/28 of the Meiring Naudé were collated. Cruises TBD 350, 371 and MN 76/28 were run by Professor R.V. Dingle, Mr S.W. Goodlad and me, specifically to gather data in the Natal Valley. For cruises 267 and 277 only photographs of profiles were made available (courtesy the late E.S.W. Simpson) and detailed reference to these have been omitted. In addition, the Southern Oil Exploration Corporation (SOEKOR) provided multi-channel profiles N801-N806. However the quality of reproduction is poor and the scale of presentation (2.5 cm = 1 second two-way-time) inadequate for detailed interpretation. As a result only reflectors McDuff and acoustic basement could be reliably identified on multi-channel profiles. I recommend that SOEKOR study good quality copies of profiles N802 and N803 with a view to delineating the continent-ocean boundary on the Tugela Cone.

APPENDIX 2

PUBLICATIONS RELATED TO THIS WORK

Various parts of this thesis have been published in international journals and in the annual reports of the Marine Geoscience Unit. Because some of these articles are multi-authored, this appendix provides notes on the contributions of various authors.

Part 1

The bathymetry of the area was described in the annual report (Martin, 1978) while a fuller review of the physical oceanography appeared the following year (Martin, 1979).

Part 2

A paper containing the main points of chapter 3 was published in 1981 (Martin, Hartnady, Goodlad, 1981). Dr Hartnady contributed the Univac computer package which he and I used to produce the revised reconstruction by trial and error. He also provided input on West Africa (section 3.2.2). Mr Goodlad provided the bathymetry of the Tugela Cone area. I plotted all pertinent features, produced the final reconstructions, wrote the paper and dealt with all the required revisions..

When I had re-interpreted magnetic anomalies in the Natal Valley as the Mesozoic sequence M0-M10, I brought this fact to the attention of Mr Goodlad. He informed me that he had recently and independently arrived at a similar conclusion. The fact that our independent interpretations matched very closely gave us confidence in the anomaly identifications, and it was agreed that he write a paper reporting the discovery (Goodlad, Martin, Hartnady, 1982). I am unsure why Mr Goodlad included Dr Hartnady as a co-author. I then produced a paper on the palaeopositions of the Falkland Plateau and the implications thereof (Martin, Goodlad, Hartnady, Du Plessis, 1982). In this case I produced all the reconstructions, diagrams and script, dealing also with all revisions. I included Dr Hartnady as a co-author because I had used his programmes. Mr Goodlad I included because he, too, had arrived at a similar re-interpretation of the anomalies. Mr Du Plessis, who was head of the Geological Survey group within the Marine Geoscience Unit at the time, provided original magnetic profiles having originally led the survey cruise.

Chapter 5 appeared in the annual report (Martin, Hartnady, Murray, 1982). Dr Hartnady again was included because of the use of his programmes. I approached Mr Murray with the prediction that the present-day offset of the Falkland Agulhas Fracture Zone (FAFZ) is 165 km. On checking Mr Murray's recent magnetic profiling in the South Atlantic we discovered that his new data, closer to the FAFZ than existing data, did indeed indicate an offset of 170 km. This constituted his contribution.

Chapter 6 stems from work carried out in conjunction with Dr Hartnady, which was originally presented in January 1983 at the South African National Committee for Oceanographic Research (SANCOR) symposium (Hartnady & Martin, 1983) and later published in an annual report (Martin & Hartnady, 1983). Dr Hartnady performed the majority of the rotations while I plotted the material, presented the work at the conference in poster and lecture form, and wrote both the abstract and the paper quoted above.

An earlier version of chapter 7 was submitted for publication in the Transactions of the Geological Society of South Africa. A revised version was solicited for publication by the journal editor. In the meantime the work has appeared in an annual report (Martin, 1983).

Part 3

A preliminary report on the Natal Valley project was published in the early stages of the programme (Dingle et al., 1978). The bathymetry was compiled by Mr Goodlad and myself. I produced the geological map on the basis of seismic profiling interpretations carried out in conjunction with Mr Goodlad, while he prepared figures 3 and 4 of that paper, again based on our combined seismic work. I then wrote a preliminary version of the script. Professor Dingle produced a more polished manuscript which was edited by all three authors.

A paper describing the influence of the Agulhas Current was presented at the 1979 SANCOR symposium (Martin, 1979) and was published later (Martin, 1981a). Reviews by Professor Dingle substantially improved this paper.

The palaeo-climatic implications of Agulhas Current palaeo-oceanography were published in a local journal (Martin, 1981b). An accompanying diagram which I designed was used as the cover illustration of that issue of the journal.

A fuller treatment of basin in-fill and palaeo-oceanography appeared the following year (Martin, Goodlad, Salmon, 1982). Mr Salmon dated key samples micropalaeontologically on the basis of which I designed figure 8.5. Mr Salmon also worked on samples selected by Mr Goodlad from more southerly areas of the Natal Valley. I wrote and re-wrote the paper but reviews by Messrs Goodlad and Salmon improved the manuscript.

These papers incorporated the division of the area into sub-provinces, the acoustic stratigraphy, and chapter 9. The bulk of chapter 8 and chapter 10 have not appeared elsewhere.

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