GEOPHYSICAL SURVEYS AROUND MULL,

WESTERN SCOTLAND

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Thesis submitted for the degree of Ph.D.

Glasgow University

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November 1979

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ACKNOWLEDGEMENTS

I would particularly like to thank Dr. J. Hall for supervising this research.

R. McQuillin provided facilities at the Marine Geophysics Unit IGS, Edinburgh, ship time and equipment. P.J. Roberts D.K. Smythe and others at M.G.U. gave additional assistance.

Ialso extend my thanks to the Officers and crews of the R.R.S. Challenger and the R.R.S. John Murray and to the technicians of R.V.B. Barry , espécially R. Powell.

Many thanks are due to R.T. Cumberland for his considerable assistance on both cruises and throughout the project.

Mr. McLeod produced the dieline figures, Mr. McLean helped with the photographic work and Mrs. D Rae strained her eyes typing the manuscript .

R. Morrison and Prof. B.E. Leake provided facilities in Department of Geology, Glasgow University.

I would like to thank my parents for all the help and encouragement they have given me. Very special thanks are due to Dave for the many long hours spent in the darkroom on my behalf and for his support during this project.

This work was supported by a NERC research grant.

DECLARATION

The material presented in this thesis summarises the results of research carried out between October 1976 and October 1979 in the Department of Geology, University of Glasgow, under the supervision of Dr. J. Hall and in the Marine Geophysics Unit of the Institute of Geological Sciences, Edinburgh under the supervision of Mr. R. McQuillin. This dissertation is based on my own independant research and any published or unpublished material used by me has been given full acknowledgement in the text.

M. Wilson

November 1979

SUMMARY

Geophysical surveys have been carried out in the sea area around Mull on two cruises aboard NERC research vessels; RRS Challenger (May 1977) and RRS John Murray (February 1979).

The aim of the surveys was twofold:-

(a) to obtain seismic refraction data using an anchored sonobuoy with a seabed hydrophone, a method which proved to be an improvement on previous attempts using free-floating sonobuoys with neutrally-bouyant near-surface hydrophones. The results from 13 surveys are presented with geological interpretations.

(b) to locate the SW continuation of the Great Glen Fault by preparing a geological map of the area using geophysical data. This is clearly located on the depth to rockhead map where a deep trough in rockhead is observed trending 030°S from Loch Buie. The trend changes abruptly to 060° 14 km S of Loch Buie.

The course of the Moine Thrust through the area is also traced and its offset used to demonstrate a 55 km sinistral shift on the Great Glen Fault.

Other faults of NE-SW and NNE-SSW trend believed to be related to the Great Glen Fault are observed in the Firth of Lorne producing a faulted sequence of LOR lavas and sediments. Faults of similar trend are also observed between Colonsay and Jura, the most easterly is believed to be the northern extension of the Loch Gruinart Fault.

The relationship of the Great Glen Fault to other similarly trending faults in Scotland is discussed. It is believed that all these faults which are similarly orientated are the result of the same stress system produced by terminal events of the Caledonian orogeny.

Two relatively young active strike-slip fault systems are described and compared. The San Andreas Fault and The Alpine Fault. By comparison with these faults it is believed that the Great Glen Fault formed close to a plate margin; the segment north of the bend south of Loch Buie was oblique to the regional slip vector and the other segment was parallel to the regional slip vector.

CHAPTER ONE

INTRODUCTION

1.1 Aims

The aims of this thesis are two-fold:
(a) to obtain and interpret geophysical data from around Mull; in particular from across the S.W. extension of the Great Glen Fault in order to compare its tectonic style with those of other major transcurrent faults; and
(b) in the preparation of the data base for (a), to explore the use of seismic techniques, including processing of data recorded on magnetic tape, in determining the seismic velocity and structure of the sediments and/or rocks on the sea bed.

1.2 Regional Geology

1.2.1 Succession

Apart from rocks of Lower Palaeozoic age, representatives of the Geological Column from Precambrian to present day are found in the area. The succession is summarised in Table 1.1 and Fig. 1.1.

The Precambrian rocks can be divided into three major groups on the basis of age. The oldest group is the Lewisian gneisses, severely altered and metamorphosed sedimentary and igneous rocks. Those outcropping on Coll, Tiree and Iona probably belong to the Scourian Cycle (Drury, 1972; Westbrook, 1972). Unconformably overlying the Lewisian rocks are the Torridonian sandstones and their metamorphic equivalent, the Moine schists: they are separated by the Moine Thrust with the Torridonian rocks lying to the West. The now slightly-metamorphosed sandstones and grits of the Torridonian and the

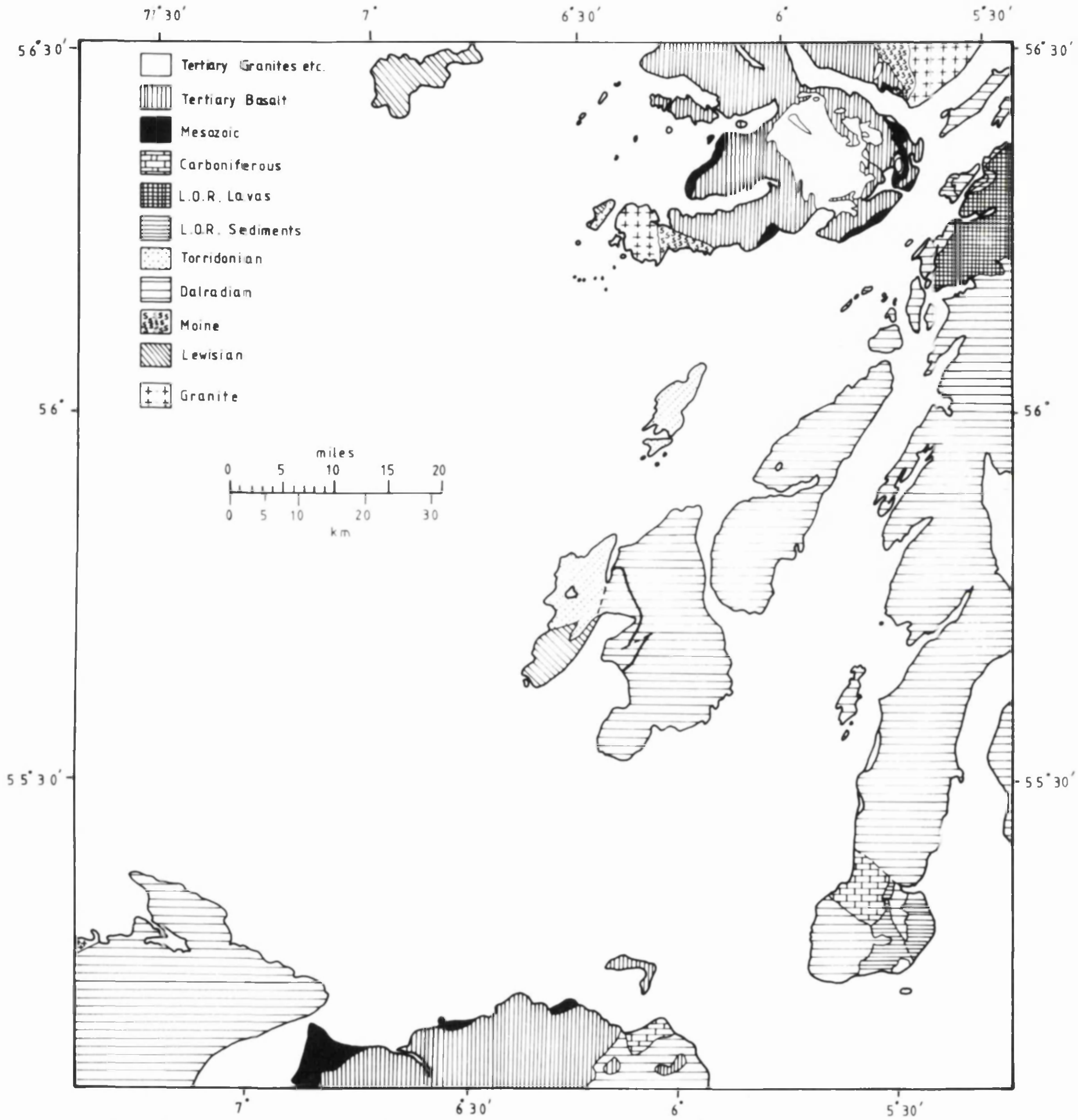


Fig 1·1 Regional Geology

Redrawn from Geological Survey "Ten-Mile" Map, sheet 1

ERA	System		Occurrence	Lithologies	Thickness (m)	Depositional Environment	Observed Field relations	Igneous Rocks
CAENOZOIC	Quaternary	Recent Pleistocene	Widespread	Fluvial sands & gravel boulder clay	variable variable	Fluvial, bead glacial		
	Tertiary	Eocene	Mull, Morvern Staffa & submarine outcrops S.of Mull					Basic plutons Acidic plutons Basaltic lavas (550 m)
MESOZOIC	Cretaceous	Upper	Gribun, Carsaig, Torosay, S. of Loch Don, Morvern	Chalk White sandstone Greensand	3 3 6	Shallow shelf Marine Marine	Unconformable on Jurassic	
		inf. Oolite	Loch Don	Sandy limestone, calcareous sandstone & fine grained sandstone	30			
		U. Lias	L. Don, Carsaig	Blue shale	9			
	Jurassic	M. Lias	Tobermory, Carsaig L. Spelve - L. Buie	Sandstones	60	Marine		
		L. Lias	Wilderness, Craignure Pt, L. Don	Black fossiliferous limestone & shales	145			
		Rhaetic	Gribun	Dark sandy limestones	12			
	Trias		Gribun, Inch Kenneth, Craignure L. Don, L. Aline	Conglomerates & sandstones	2000 L. Spelve 100 elsewhere	Continental	Unconformable on older irreg eroded surface	
	Permian		Islay (Port nan Gallan)	Breccia, blocks of qzite, lmst & schist in bright red matrix		Continental desert		Camptonite Monchequite dykes with E-W trend in Morvern
PALAEOZOIC	Carboniferous	Upper	Inninmore Bay Morvern	Pebbly sandstone occasional shale + thin coals.	791	Lagoonal/ deltail	Base not seen probably unconformable on Moine	With N-SE trend on Scarba and Colonsay
	Devonian	Lower Old Red Sandstone	L. Don, Kerrera, Oban	Local breccias with shales & sandstones	30-60	Continental	Lavas overlie sediments which thin to east	Lorne Plateau Lavas Hypersthene andesites Tuffs, felsitic flows Acid tuffs & basic flows Basic andesite
	Silurian		Ross of Mull					Ross of Mull Granite hornblende biotite diorite muscovite biotite granite - late Sil. early Dev. age contact metamorphosed local rocks
PRECAMBRIAN	Dalradian		Lismore, Scarba, Jura, Islay E. Mull (centre L. Don anticline	Metamorphosed psammitic, calcareous graphitic pelitic sediments Limestone (Lismore)	Variable 4000	Initially shallow water/ deltaic epicontinental deepwater	Apart from Lismore Limestone, belong to Iltay Nappe Group, many lateral variations in thickness and lithology	
	Torridonian		Oronsay, Colonsay, Iona Islay	Slightly metamorphosed sediments - mudstones, sandstones etc.	74000	Marine		
	Moine		Ross of Mull, Gribun	Pelitic schists, quartzite meta sediments	76000		Thought to be metamorphosed equivalent of Torridonian separated by Moine Thrust	
	Lewisian		Coll, Tiree, Iona, Colonsay	Pure gneiss coarse grained quartzofeldspathic banded gneiss				

Table 1.1 The Geological Succession

Sources: Bailey et al., 1924, 1925; Bennison & Wright, 1969; Cunningham-Craig et al., 1911; Johnstone, 1966; Knill, 1963; Lee & Bailey, 1925; Peach et al., 1909; Phemister, 1960; Rast, 1963; Richey, 1961.

semi-pelitic schists and gneisses of the Moine originate from separate depositional environments now brought into juxtaposition by movement along the Moine Thrust.

The fairly monotonous lithologies of the Moine and Torridonian Formations are succeeded by the lithologically diverse Dalradian Series spanning late Precambrian and Cambrian times. A detailed succession is given in Table 1.2. For convenience, the Dalradian has been divided by marker horizons into Upper, Middle and Lower. The base of the Upper Dalradian is marked by the Loch Tay limestone and the Portaskaig Boulder Bed marks the base of the Middle Dalradian. The depositional history, however, can be divided into two stages by the Jura Slate. Initially a shallow marine unstable shelf environment prevailed in which limestones, dolomites, shales (now slates) and cross bedded sandstones (now quartzites) were deposited. These strata can be traced discontinuously along strike from Islay to Ballachulish with little variation. Following this a sedimentary trough developed into which deep-water facies were deposited. The Scarba Conglomerate Group, deposited on a north-dipping slope by turbidity currents (Ander ton, 1977), fines upwards into the Easdale Slates in the north, indicating quieter deposition (Baldwin & Johnson, 1977). Both the Scarba Conglomerate Group and the Easdale Slates are overlain by the Port Ellen/Craignish Phyllites (Fig. 1.3).

The earliest Lower Palaeozoic rocks present in the area are those of Lower Old Red Sandstone age, occurring around Oban and Kerrera as an intercalated sedimentary and lava sequence deposited in a shallow syncline with a NE - SW axis (Waterston, 1965). They lie unconformably on Dalradian basement with the lower part of the sequence, principally sedimentary rocks ranging from limestones to breccias, thinning east-

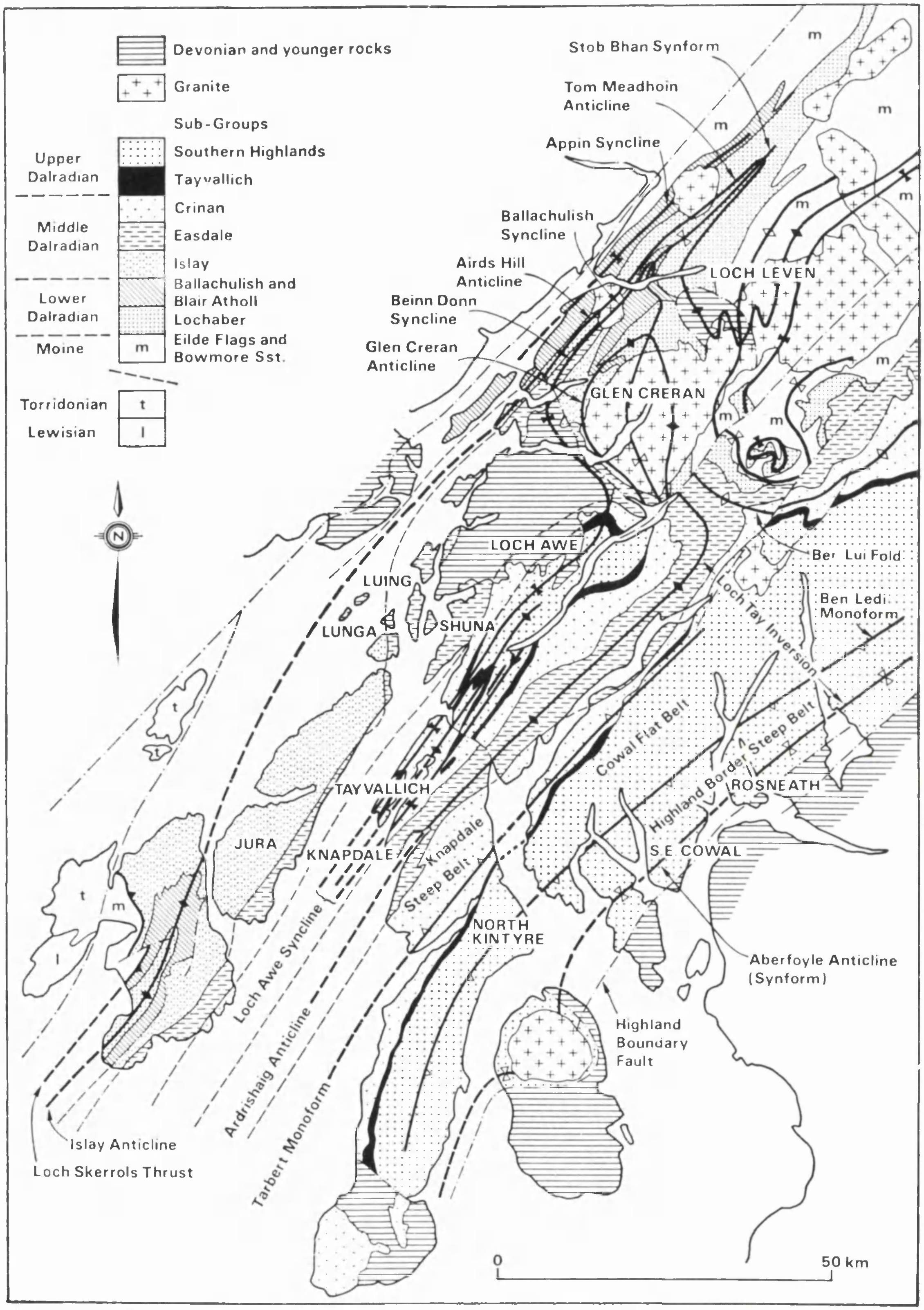


Fig 1.2 Simplified geological map of the Dalradian rocks in the South-west Highlands of Scotland, showing the axial traces of the major folds according to Roberts and Treagus (1977). (From Roberts and Treagus 1977)

		Islay & Tayvallich	Ballachulish & Loch Awe	Glen Creran	Knapdale & Cowal	Central Perthshire
Southern Highland (Upper Dalradian)		Kells Grit Tayvallich Volcanics	Loch Avich Lavas Loch Avich Grit Tayvallich Volcanics		Innellan Group with Loch Fad Conglomerate Bullrock Greywacke ——?——?——?—— Dunoon Phyllite Beinn Bheula Schist Green Beds Glen Sluan Schist	Highland Border Series Upper Leny Grit Leny Limestone and Shale Lower Leny Grit ? Aberfoyle Slate —— Correlation uncertain —— Ben Ledi Grit with Green Beds Pitlochry Schist
Argyll (Middle Dalradian)	Tayvallich	Tayvallich Limestone	Kilchrenan Conglomerate Tayvallich Slate and Limestone (thin ash bands)		Loch Tay Limestone	Loch Tay Limestone
	Crinan	Crinan Grit Ardmore Grit Conglomerate	Crinan Grit		Stonefield Schist Garnetiferous Mica Schist Erins Quartzite (Upper part)	Ben Lui Schist
	Easdale	Limestone and Laphroaig Quartzite Port Ellen Phyllite Scarba Conglomerate Jura Slate	Shira Limestone and Slate Ardrishaig Phyllite Degnish Limestone Easdale Slate	Ardrishaig Phyllite (Bonawe Succession) Pebbly Quartzite Group	Stronachullin Phyllite Erins Quartzite (Lower part) Ardrishaig Phyllite Easdale Slate	Farragon Beds Ben Lawers Schist Ben Eagach Schist Carn Mairg Quartzite
	Islay	Bonahaven Dolomite Port Askaig Tillite	Islay Quartzite Dolomitic Beds Tillite	Creagan Creran Bridge Quartzite Creran Flags	Perthshire Quartzite Series	Killiecrankie Schist Schichallion Quartzite Dolomitic Beds & Boulder Bed Schichallion Boulder Bed
Appin (Lower Dalradian)	Blair Atholl	Islay Lst {No correlation implied Mullach Dubh Phyllite Ballygrant Limestone Baharradail Phyllites	Lismore Limestone Cuil Bay Slate	Lismore Limestone	Pale Group Dark Group	Pale Limestone Banded Group Dark Limestone Dark Schist
	Ballachulish	Cnoc Donn Quartzite (Transition Group) Cnoc Donn Slate Kintra Limestone Mull of Oa Phyllite	Appin Phyllite and Lst. Appin Quartzite (Transition Group) Ballachulish Slate Ballachulish Limestone	Appin Phyllite and Lst. Appin Quartzite (Transition Group) Ballachulish Slate Ballachulish Limestone		Relationship Uncertain Kinlochlaggan Limestone Local Boulder Bed
	Lochaber (Transition)	Kintra Phyllite Maol an Fhithich Quartzite Bowmore Sandstone	Leven Schist Glencoe Quartzite Binnein Schist Binnein Quartzite Eilde Schist Eilde Quartzite Eilde Flags (Moine)	Leven Schist Glencoe Quartzite		Kinlochlaggan Quartzite Monadhliath Schist Eilde Quartzite Struan Flags (Moine)

TABLE I.2 Lithostratigraphy compiled for the Dalradian rocks in the South-west Highlands of Scotland after Harris and Pitcher (1976).
(from Roberts and Treagus 1977)

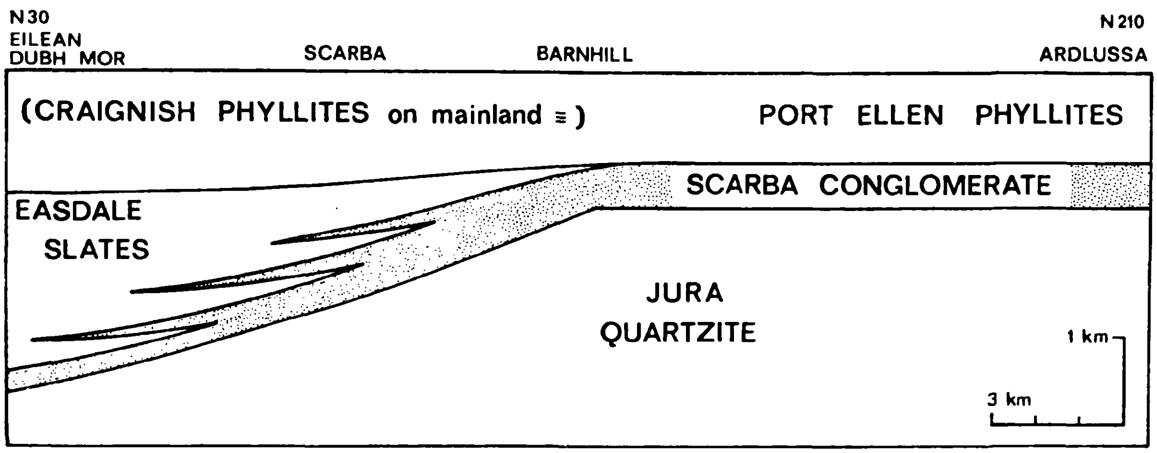


Fig 1·3 Stratigraphical relationships of the formations found on Jura and Scarba. (from Anderton 1977) ward where eventually the upper part of the sequence, mostly andesitic lavas, lies directly on the Dalradian.

Minor outcrops of Carboniferous and Permian sedimentary rocks occur in the area. Upper Carboniferous sandstones and shales with thin coalbands are seen at Inninmore Bay, Morvern, and are probably unconformable on Moinian rocks (MacGregor and Manson, 1933). A breccia, infilling a swallow hole on Islay, was described by Peach (1907) and initially assigned to lowermost Triassic. However, it is probably Carboniferous or Permian in age (Mykura, 1965; Pringle, 1952).

Mesozoic rocks from the Triassic upwards can be found around the Mull coastline and at Loch Don, though a complete succession is not seen anywhere. About 2000 m of Triassic rocks are observed at Loch Spelve (Rast *et al.*, 1968) but elsewhere the thickness of the Triassic is of the order of 100 m. In general the thicknesses of the Mesozoic on Mull are less than those observed further north on Skye and Raasay.

The Tertiary period was dominated by igneous activity; although some sedimentary rocks of that age are found interbedded with the Basaltic Plateau Lavas e.g. Ardtun Leaf Beds of Mull (Bailey *et al.*, 1925).

The region suffered successive glacial advances during the Pleistocene giving rise to raised beaches, 'U'-shaped, over-deepened valleys and substantial deposits of boulder clay.

1.2.2 Igneous Activity

Three major episodes of igneous activity can be discerned; (a) Caledonian, during this orogeny a magmatic cycle similar to that described by De Sitter (1956) is recorded (Johnstone, 1966) in (1) pre-tectonic basic rocks, (2) syntectonic migmatites, and (3) post-tectonic intrusive and volcanic rocks. The Ross of Mull Granite belongs to the third stage as do the Foyers and Strontian granites (Phemister, 1960) which followed a migmatitic phase (Phemister, 1960) when pegmatites and pegmatitized pelitic gneisses were injected into the Moine Schists. North-South trending Caledonian dykes are seen on Colonsay, whilst NE - SW trending dykes are seen on Jura and Argyllshire. These are typically porphyritic microdolerite in composition (Mercy, 1965). At the same time as the Lower Old Red Lavas were being extruded, passive granites (Read, 1961) were being emplaced.

(b) <u>Permo-Carboniferous</u>. A regional linear WNW-trending dyke swarm of doleritic composition was intruded into the Mull, Lismore and Morvern region (Speight & Mitchell, 1979).

(c) <u>Tertiary</u>. Intensive prolonged igneous activity occurred during this time throughout the Inner Hebrides and NW across the North Atlantic Igneous Province. The earliest volcanic activity was the outpouring of huge quantities of basalt as lava flows. Some 1800 m of this basalt still remains on Mull in spite of subsequent intense erosion. Following this, large explosive vents associated with rhyolitic magma were formed. These were then followed by the intrusion of plutonic masses as a succession of intrusions around the igneous centres. Accompanying this was the formation of NW - SE-trending linear dyke swarms focussing on the igneous centres following lines of weakness in the crust. The dykes cut both the lavas and intrusive masses and extend for great distances (e.g. the Cleveland Dyke extends for 400 km (Anderson, 1951)). Descriptions of these volcanic centres can be found in Richey (1961) and Stewart (1965).

1.2.3 Structure

The area is dominated by the NE - SW Caledonian grain. Those structures not formed as a direct result of the Caledonian stresses have followed weaknesses initiated at that time.

Folding. Prior to the deposition of Palaeozoic and Mesozoic rocks the Dalradian rocks underwent a complex multi-phase structural modification during the Caledonian Orogeny (Table 1.3, Fig. 1.2). Apart from the Lismore Limestone, all the Dalradian rocks in the area belong to the Iltay Nappe Complex which encompasses all the metamorphic rocks between the Iltay Boundary Slide and the Highland Boundary Fault. The complex is regarded by Johnstone (1966) as consisting of two main sub units:

(a) Northern Nappe Zone (Islay, Jura and Scarba) a transported sheet (comprising several folds) contiguous to and overlying the Iltay Boundary Slide. In this sheet all early formed recumbent anticlines recorded close to the NW.

(b) The Tay Nappe (lying to SE) forms the remainder and is an early formed recumbent anticline closing to the SE.

The rocks of the North Grampians Complex lying between the Iltay Boundary Slide and the Great Glen Fault are mainly Moinian and lack distinctive marker horizons: as a result they are less well-defined structurally.

A series of concentric folds in the country rocks around the Early Mull Caldera were probably the result of early intrusion of pyroxene granophyres along the caldera margin (Stewart, 1965).

Faulting. NE - SW dip-slip faults are observed on Kerrera and around Oban intersected by a NW - SE-trending set (Lee & Bailey, 1925), cutting both the Dalradian schists and the Lower Old Red sedimentary and lava sequence.

The Moine Thrust is the uppermost thrust of a group of near horizontal fracture planes known collectively as The Moine Thrust Zone. The zone runs NNE - SSW and a combined horizontal movement of at least 30 km (Clough, 1907) to the west has brought the Moine Schists into contact with the Torridonian Sandstones.

A set of NE-and NNE-trending strike-slip faults cross the Central Highlands, of these the Great Glen Fault is probably the best known. Fig. 1.4 shows the main faults of this system. They occupy straight depressions containing crushed and shattered rock. Sinistral movement has been demonstrated for some of these faults:

(a) Strathconon Fault (Clough, 1910; Hinxman & Crampton, 1914) with a displacement of $1\frac{1}{2}$ - 5 km;

(b) Ericht-Laidon Fault (Read, 1923) 6 - 8 km displacement; and

(c) Loch Tay Fault (Wilson, 1905) displacement 6 km.

Evidence from LANDSAT imagery indicates that a great number of lineations occur parallel to this set indicating the possible presence of faults as yet undetected by land geological work (Johnson & Frost, 1978).

1.3 The Geology of the Great Glen Fault

1.3.1 Land-based Studies

The Great Glen Fault is the best known of the set of dominantly sinistral strike-slip faults described in the previous section. The fault runs from the Moray Firth where it trends at 035^{o} through Loch Ness and Loch Lochy and out to sea via Loch Linnhe in the south. There it trends at 040^{o} making its outcrop slightly concave to the north-west. Its physical expression on land is a striking topographic trough --

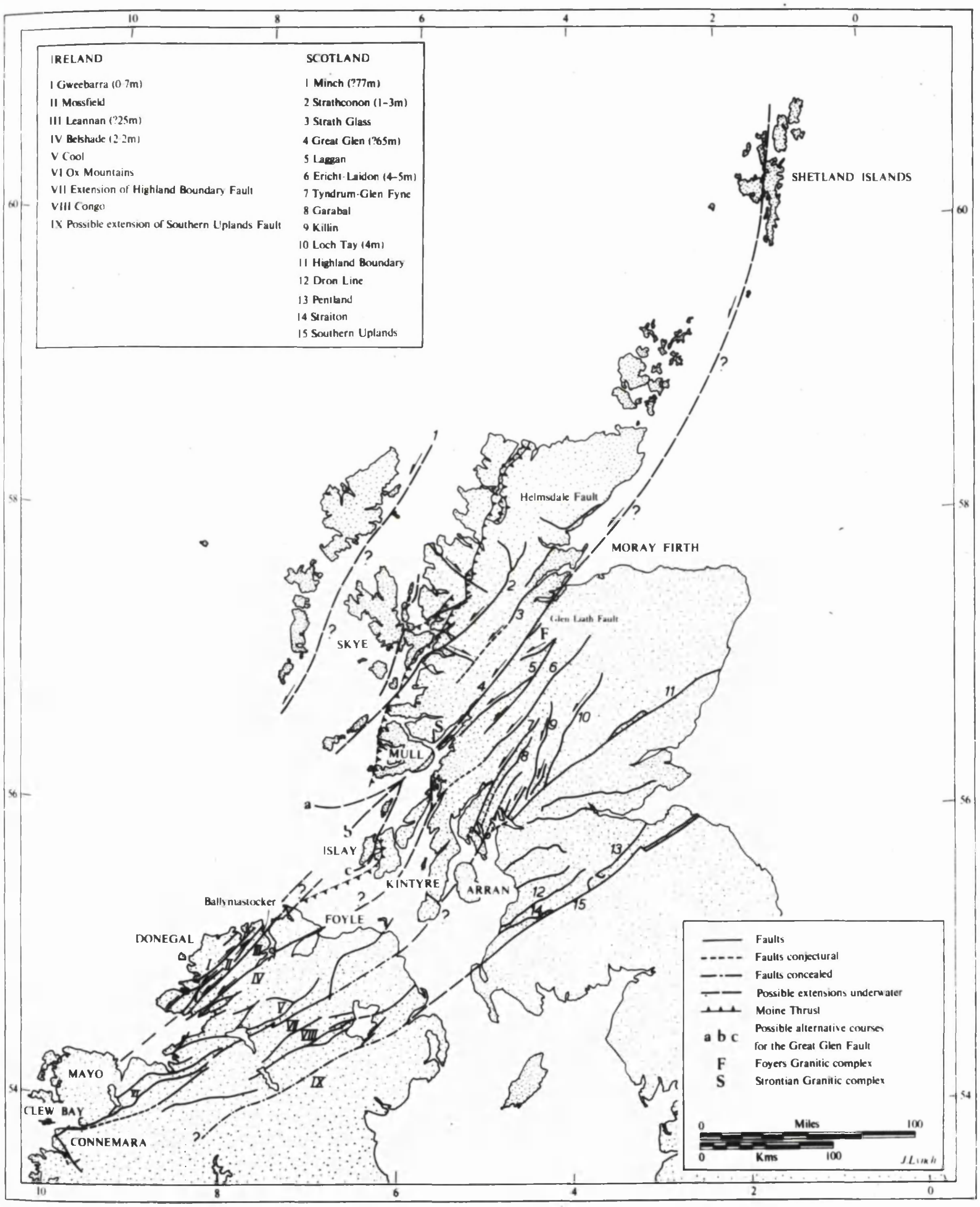


Fig 1.4 NE-SW faults in Scotland (from Pitcher 1969)

an over-deepened, 'U'-shaped valley formed by recent glacial activity. This 'trough' is partially filled by glacial sediments, in excess of 100 m at Inverness (Horne & Hinxman, 1914). The fault is represented by a zone of intensely sheared and crushed rocks 1 - 2 km wide (Geike, 1865; Kennedy, 1946; Eyles & MacGregor, 1952). Records of sporadic seismic activity associated with the fault began in 1768 (Davison, 1924) with epicentres lying in three main groups on the south side of the fault:

(a) around Inverness (most active up to 1934);

(b) around Fort William; and

(c) Oban district (could be related to Great Glen Fault or one of the parallel faults).

These data may be a feature of an observational bias caused by population distribution.

An historical summary on the development of ideas concerning the Great Glen Fault is given in Table 1.3. Early workers (Miller, 1841; Murchison & Geike, 1861) could reach no conclusions regarding the nature of displacement on this fault. Geike (1865) working in the Moray Firth area proposed a downthrow to the SE on the basis of Jurassic rocks to the SE being adjacent to Middle Old Red Sandstone rocks to the NW. This was reinforced by the work of the Geological Survey (Horne & Hinxman, 1914) working SW of Inverness, who suggested a minimum downthrow of 1800 m to the SE. In the same memoir Cunningham-Craig gives the first indication that lateral displacement may have taken place on the Great Glen Fault when he described horizontal slickensides on a fault, belonging to the Great Glen Fault System, at Allt Mor, suggesting that the lateral movement proven may also be applicable to most if not all of the NE - SW faults.

Table 1.3 Historical summary of Great Glen Fault discussions
Table 1.3 Historical summary of Great Glen Fault discussions

WHO	TYPE OF MOVEMENT	EVIDENCE	EXTENSIONS	COMMENTS
MURCHISON & GEIKIE 1861	Normal, not much displacement in spite of the appearance of a mega shear.	Deep crack as movements due to Lisbon earthquake - seen on Loch Lomond surface		
HORNE In Horne and Hinxman 1910	Normal 204.0m SE	Jurassic against schists and Devonian sediments at Eathie		
CUNNINGHAM-CRAIG 1914	Inferred horizontal	Horizontal sinistral fault in Allt Mor		
KENNEDY 1946	Sinistral transcurrent 104 km(65 m) post MORS pre U Carboniferous.	1)Foyers and Strontian granites are part of the same stock. 2)Line up of Dalradian metamorphic zones. 3)Horizontal slickenslides 4)Mylonite present 5)Offset of Moine injection complexes.	Extended West by correlating the L.Skerrols Thrust with the Moine Thrust.	Shand(1951) could not find any slickenslides or mylonite. Several authors question whether Foyers and Strontian granites are parts of the same stock.
LEEDAL 1951	Sinistral movement minor amount movements post Permian.	Displacement of 3 complexes - monchiquitic dykes of Permian age plus horizontal slickenslides.		
FLINN 1961, 1969	Sinistral then dextral	Bathymetric and aeromagnetic evidence.		
MYKURA 1965		Devonian palaeogeography.	Walls Boundary Fault dextral.	
PITCHER et al. 1964	Sinistral 40 km	Proved sinistral displacement Leannan Fault.	Leannan Fault	

STORETVELDT 1974	A minimum sinistral displacement of 200 - 300 km during late Devonian-Permian.	Position of ORS geomagnetic poles in Norway and Caithness are not coincident.	Walls Boundary Fault.	Method criticised by Turner *et al.* as being inexact. Walls Boundary Fault thought to be dextral(Flin,Mykura & Young). Also Mykura says reconstruction puts ORS sediments of different origins adjacent
DONOVAN ET AL. 1974/5	No major sinistral movement post Devonian Minor dextral movement after. NB.Does not rule out possibility of sinistral movement before Devonian.	Old Red Sandstone sediments are very similar either side of fault.		
CHESHER & BASON 1975	Normal downthrownto south active throughout Mesozoic. No transcurrent movement.	Moray Firth seismic evidence down/cuts Jurassic against ORS.		
MUNRO 1973		Foyers - Strontian		
PIDGEON & AFTALION 1978		Ages of Foyers and Strontian different		Only analysed few samples.

Table 1.3 contd.

Kennedy (1946) presented a comprehensive argument for a 100 km sinistral shift along the Great Glen Fault centred on the former unity of the Foyers and Strontian granites. Several lines of evidence were presented to support this thesis, divided into three main categories they are:

(1) Indirect evidence

(a) Physical features of the fault. It is almost straight with a wide shatter belt of intense deformation.

(b) Tectonic association. It is one of a set of similarly trending fractures, some of which show clear sinistral displacement, e.g. Strathconnon and Enicht-Laidon faults.

(2) Direct evidence

(a) The north - south belts of the Moine injection complexes are offset sinistrally by 100 km.

(b) The sillimanite and kyanite metamorphic zones are similarly displaced.

(c) The Foyers and Strontian Granites are the displaced fragments of the same intrusion as they consist of identical rock types and are structurally homologous. Their situations relative to the sillimanite isograd and the injection complex are also identical.

(3) Contributory evidence

The Moine Thrust may reappear to the SE of the Great Glen Fault as the Loch Skerrols Thrust on Islay, this view had also been proposed by Bailey (1917) although he assumed a normal throw on the Great Glen Fault.

Kennedy suggests that the main displacement took place after Middle Old Red Sandstone times but before the Upper Carboniferous. Minor movements occurred in the Jurassic and Tertiary (Bailey, 1925; Kennedy, 1946).

Kennedy's arguments have subsequently been both contested and supported. Shand (1951) doubted the existence of a major strike-slip fault since he could not find the diagnostic evidence he considered to be essential, namely horizontal slickensides and mylonitisation. Eyles and MacGregor (1952), in defence of Kennedy, described evidence of intense cataclasis (locally developed into mylonite) from five sites along the course of the Great Glen Fault.

In her study of the Foyers Granite Mould (1946) noted some differences from the Strontian Granite as described by MacGregor and Kennedy (1931). These differences are:

(a) hornblende is present only in the Foyers Granite; and

(b) the pelitic schists around the Foyers granite have suffered a high degree of contact metamorphism not evident around the Strontian granite.

Ahmod (1967) has shown that the two granites are geophysically dissimilar. The Foyers granite is more radioactive and less strongly magnetised than the Strontian granite. The estimated shapes and orientation of the stocks are different as deduced from the magnetic anomaly contours, the Foyers Granite being 3 x 6 km in plan view with the strike of the major axis N 50° E, whereas the Strontian Granite is 6 x 13 km in plan view with a north - south striking major axis. Ahmed concludes that the seismic (Davison, 1924) and geological evidence is consistent with dip-slip movement on the fault (presumably to the SE (Horne & Hinxman, 1914)) and suggests that a stress pattern with a dominant vertical component was introduced at the time of or as a result of the Newer Granite intrusions.

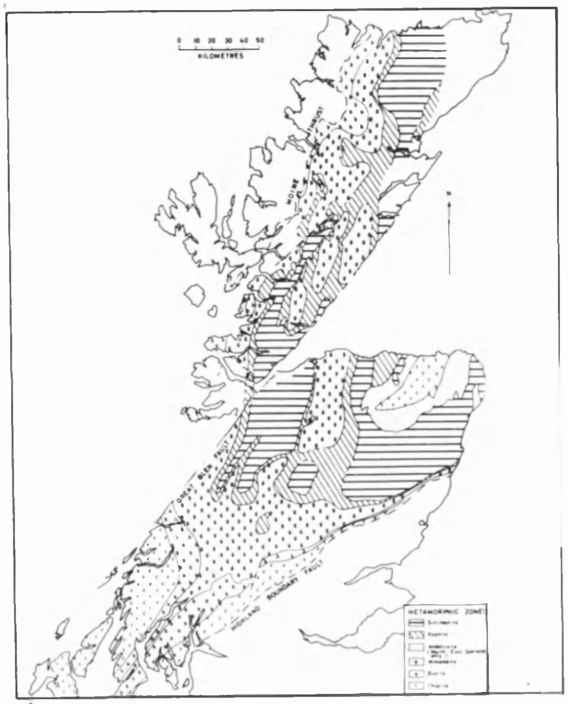
On the basis of structural and petrological comparisons, Munro (1965, 1973) and Marston (1967) proposed that although the granites show broad similarities, they differ in detail to an extent that throws doubt on Kennedy's assertion of their equivalence. Marston (1970) suggests that these differences, together with those of magnetic field, could be accounted for by the different levels to which the granites are presently exposed, the Strontian granite being exposed at a much lower level compared to the Foyers granite.

K/Ar dates for the granites (Miller & Brown, 1965; Brown et al., 1968) are similar, 400 ± 18 my for the Foyers granite and 410 ± 18 my for the Strontian granite. Using Lead-Uranium isotope studies, Pidgeon and Aftalion (1978) showed that the zircons of the two granites are chemically distinct with old inherited zircons occurring only in the Foyers granite. However, these results cannot be regarded as conclusive because of limited sampling.

The weight of present evidence seems to suggest that the Foyers granite and Strontian granite are two separate intrusions of the same suite of granites intruded at the close of the Caledonian Orogeny.

There is still support for the essence of Kennedy's hypothesis of a major sinistral displacement along the Great Glen Fault. However, a variety of modifications and alternative hypotheses have been proposed (Table 1.3); the most variant of these suggests a dextral movement of 200 km in Lower Old Red Sandstone times (Garson & Plant, 1972). Garson and Plant's palinspastic reconstruction (Fig. 1.5) aligns the North coast of Scotland with the South coast of the Moray Firth and also the metamorphic zones. However, Winchester (1974) has shown that a closer matching of the metamorphic zones is achieved by assuming a sinistral displacement of approx. 130 km.

Chinner (1978a) has argued that the metamorphic isotherms



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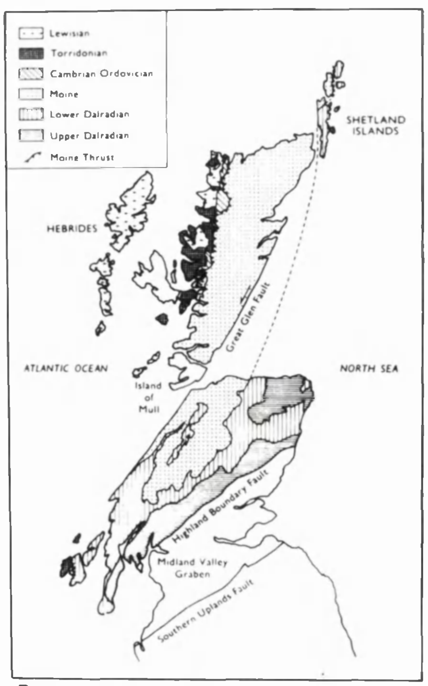

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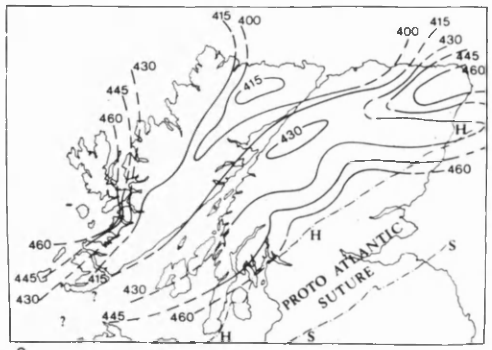

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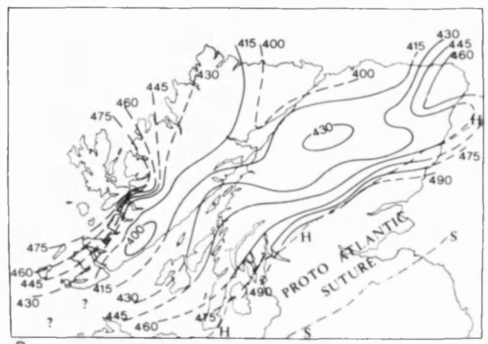

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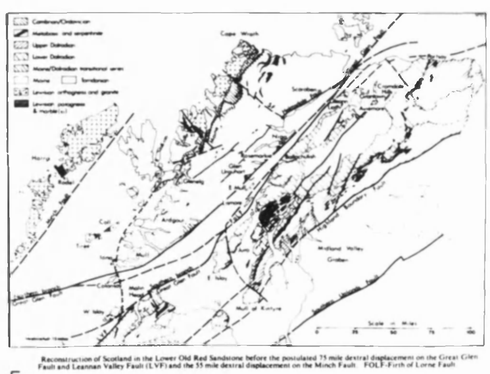


E

A-after Winchester (1974)
B-Storetvedt (1974)
C-Brown & Hughes (1973) biotite dates
D- " " K-Ar isochrons
E-Garson & Plant (1972)

Fig 1.5 Palinspastic reconstructions on the Great Glen Fault can be used to support either dextral or sinistral movement and that Winchester's (1974) reconstruction requires the thermal surfaces to be vertical to avoid offset by the post Devonian downthrow. Winchester (1978) replied that his (1974) reconstruction was based on the correlation of metamorphic zonal structures and not on matching isograds. However, Chinner (1978b) warns that the use of metamorphic zones is inspecific (unlike correlating stratigraphic horizons). Winchester (1978b) accepts this but reiterates his claim that the metamorphic zonal structures are the least unreliable means of assessing movement on the Great Glen Fault, in the absence of structural or lithological correlations.

Further support for Garson and Plant's hypothesis comes from a study of the K/Ar dates of Caledonian metasediments and related intrusions. Contouring these ages on a dextral palinspastic reconstruction, Brown and Hughes (1973) obtained what they consider to be a more satisfactory arrangement than that obtained for the sinistral reconstruction (Dewey & Pankhurst, 1970). However, Brown and Hughes's interpretation is dependent on the following assumptions: that the Proto-Atlantic suture lies in the Midland Valley (Gunn, 1973); that Benioff zones dipped to the north-west and south-east; and that K/Ar ages of plutonic and associated metamorphic rocks decrease inland from the plate margin, as has been observed in modern examples (e.g. Brown, 1973).

Holgate (1969), whilst supporting an overall sinistral movement of 100 km along the Great Glen Fault, proposes an initial sinistral displacement of 130 km (during Lower Old Red Sandstone time) followed by a dextral shift of 30 km during the Lower Eocene; the latter offsetting Mesozoic sediments, Tertiary dykes and drainage systems and camptonite-monchiquite dykes. Holgate's view is supported by Donovan et al. (1976) on the basis of the palaeogeographic reconstructions for the Middle Old Red Sandstone of the Inverness area.

Using palaeomagnetic results from the Devonian of Norway and Scotland, Storetvedt (1974) has proposed a 200 - 300 km sinistral shift along the Great Glen Fault, seen as the Melby and Walls Boundary Faults on Shetland. The reliability of this evidence was questionned by Turner et al.,(1976) who also opined that if movement between Norway and Scotland had taken place, it need not have been along the Great Glen Fault. Mykura (1975) also disagreed with Storetvedt, firstly because Storetvedt's (1974) reconstruction would juxtapose Old Red sediments of different origins, as suggested by different contained faunas; and secondly because the movement on the Walls Boundary and Melby Faults can be shown to be dextral.

In their proposed sequence of movements for the Great Glen Fault, Garson and Plant (1972) suggested a 125 km movement along the Leannan Fault, through Islay between Lismore Island and Strontian to Fort William and the Moray Firth ignoring the demonstrable 40 km sinistral shift on the Leannan Fault (Pitcher et al., 1964). Durrance (1976) was able to show that the Loch Gruinart Fault on Islay downthrows to the south-east and suggested a dextral offset. However, the distribution of the 194 gravity stations on which his work is based allows considerable latitude in drawing the contours. A magnetic study (Westbrook & Borradaile, 1978) confirms the S.E. downthrow proposed by Durrance (1976) but prefers a 9 km sinistral shift. Both interpretations suggest that it is unlikely that the Loch Gruinart Fault is the main extension of the Great Glen Fault; however, it may be a major splay.

In spite of the voluble debate following Kennedy's hypothesis of sinistral movement along the Great Glen Fault no firm conclusions can be made since there is a lack of unequivocal evidence to support or disprove 'leftist' or 'rightist' views. It is, however, generally believed that the Great Glen marks the line of a major transcurrent fault with a component of downthrow to the south-east and that the major displacement took place in Lower Old Red Sandstone times with possible minor movements during late Mesozoic or Tertiary.

1.3.2 Offshore Extensions of the Great Glen Fault and seabed geology to S.W.

In this section extensions to the Great Glen Fault to the S.W. will be discussed in detail as part of a review of the seabed geology in and around the Firth of Lorne. Some reference will also be made to the N.E. extension.

Bathymetrically the north-west continental margin can be divided into three zones:

(i) The inner shelf - the bathymetry describes a ragged topography formed during the last ice age (ending around 8000 years ago). The Firth of Lorne contains glacially over-deepened valleys, some filled in places with thick Quaternary deposits (Hall & Rashid, 1977). The orientation of these valleys follows two main trends: (a) the NE - SW strike of the Dalradian metamorphic basement, and (b) NNE - SSW parallel to faults in the Lower Old Red sediments. Hall and Rashid (1977) also suggest the presence of a -30 m erosion surface of glacial origin. However, this is considered unlikely in the light of sparker data, obtained during this study, which shows irregular rockhead beneath a smoothing wrap of sediments in areas (e.g. south of Kerrera) where this surface was believed to exist.

In general the troughs may mark the lines of faults or

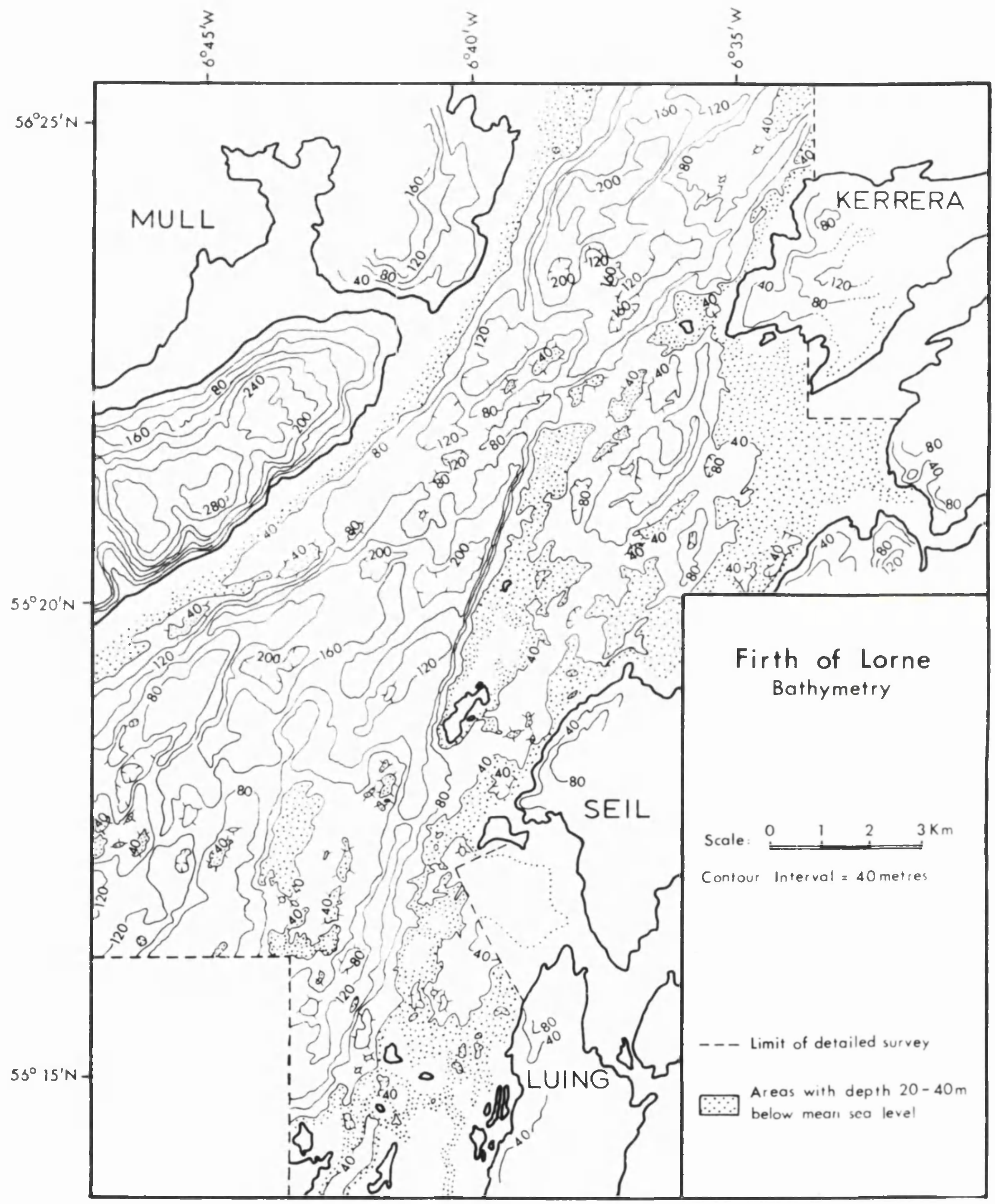


FIG. 1.6 Bathymetry of the Firth of Lorne and topography of surrounding land, both contoured at 40 m interval. The area of sea bed lying at depths between 20 m and 40 m below sea level is shown shaded and corresponds to the submerged platform discussed in the text and shaded in histogram of Figure 2.

(from Hall & Rashid 1977)

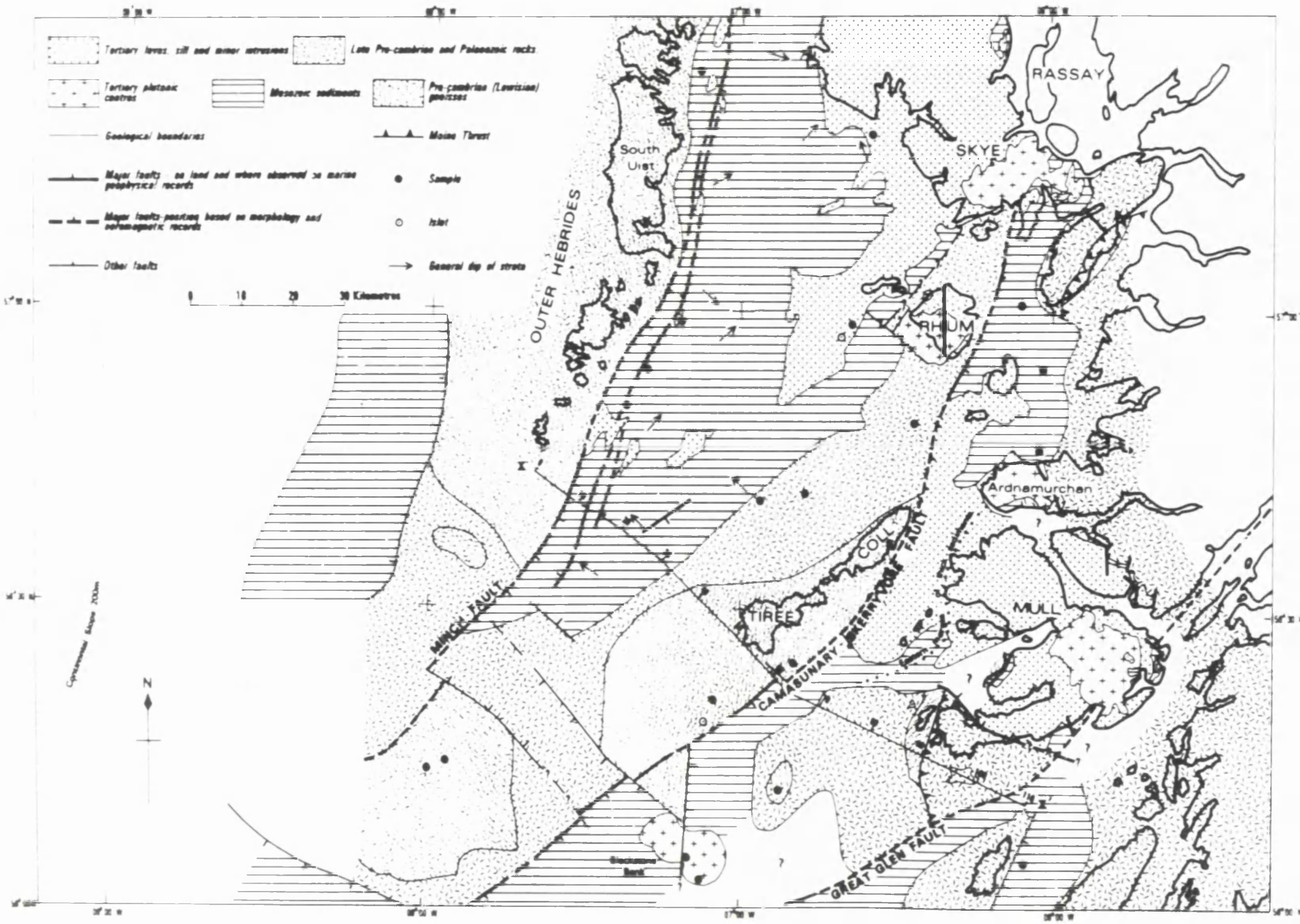


Fig. 1·7 Pre-Quaternary geology of the Sea of Hebrides. (from Binns et. al. 1973) zones of softer more-easily weathered rock, whilst the highs are likely to be formed of more resistant rocks. This philosophy has been pursued by Rashid (1978) who has proposed a complex arrangement of Dalradian, Lower Old Red lavas and sediments and Mesozoic rocks on the evidence of shallow seismic and magnetic data for the Firth of Lorne.

Fig 1.8 is part of the Hebrides and Minches Bouguer anomaly map of the I.G.S. (Binns et.al.1975) and includes both onshore and offshore data. Onshore the relationship between geology and Bouguer anomalies is evident, particularly over the Mull centre where a concentric anomaly reaching a maximum of 65 mgal is observed. Another gravity high is centred over the west coast of Iona coinciding with the outcrop of Lewisian gneiss (density 2.7 g cm^{-3}) which is denser than any of the neighbouring rocks (quartzite density 2.6-2.7 g cm^{-3}; granite density 2.5-2.7 g cm^{-3}). These density values would also account for the small gravity low observed over the Ross of Mull.

An elongate gravity low is located over Jura, which is composed of Dalradian quarzite of fairly uniform composition. To the north of Jura the thickness of the Jura quartzite diminishes whilst that of the Easdale slates increases (see Fig 1.3). The density of slate is greater than that of quartzite (slate density approximately 2.7 g cm^{-3}). Taking the difference between maximum slate and quartzite thicknesses of 2 km and a density contrast of 0.1 g cm^{-3} a gravity difference of 8.4 m gals is calculated (using the simple slab formula $g = 2\rho Gh$) which fits reasonably well with the values observed at Jura (0.0 mgal) and at Luing (approximately 10 mgal) especially as the Mull centre anomaly may be adding slightly to the Luing value.

Offshore an elongate low of 5-10 mgal is observed between Colonsay and the Ross of Mull. Much of this anomaly may be attributed to the presence of low density sediments of Recent and Mesozoic age (e.g. sandstones; density 1.8-2.7 g cm^{-3}) infilling the trough in rockhead (Rashid 1978; this study p.49 and Fig 3.3.) which underlies this anomaly. It is clear that this anomaly is not solely due to the trough as the anomaly is broader (5-10 km) than would be expected from a near surface source of 2-3 km width indicating a contribution from deeper seated intra-crystalline-basement contrasts. The effects of such contrasts have already been demonstrated, above, for Jura and Luing.

The above discussion of Fig 1.8 illustrates the problems in using the Bouguer anomaly map to define the course of the Great Glen Fault zone in an area where strong features such as the Mull centre dominate the local gravity field and where intra-basement contrasts are also present.

Another probable fault zone juxtaposing Torridonian basement against Dalradian basement lies to the east of Colonsay where a bathymetric trough (Rashid, 1978) and a linear magnetic anomaly concur (Faruquee 1972; see Fig. 1.9). This has been correlated with the Loch Gruinart Fault on Islay (Dobson et al., 1975; Westbrook and Borradaile, 1978). Durrance (1977) and Westbrook and Borradaile (1978) on the basis of gravity data and an upward continuation model of the I.G.S. aeromagnetic map, respectively, have proposed that the Loch Gruinart Fault has a downthrow to the south-east of about 1 km with respect to the displacement of the Permo-Triassic rocks(also proposed by Dobson et.al.1975)and probably a greater throw affecting the surface of the Lewisian basement.

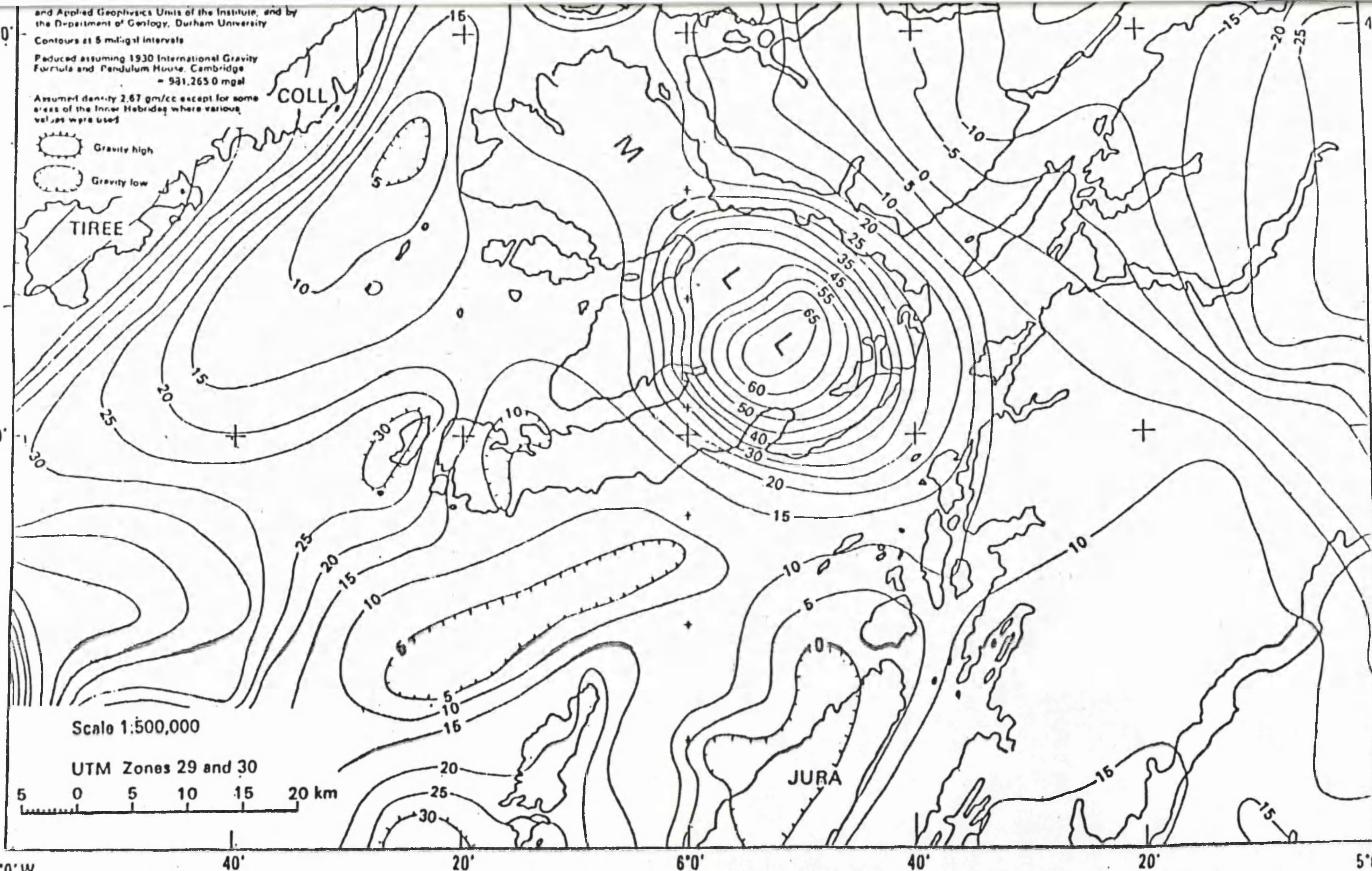


Fig 1.8 Part of the I.G.S. Hebrides and Minches Bouger Anomaly Map (Binns <u>et.al.</u> 1975)

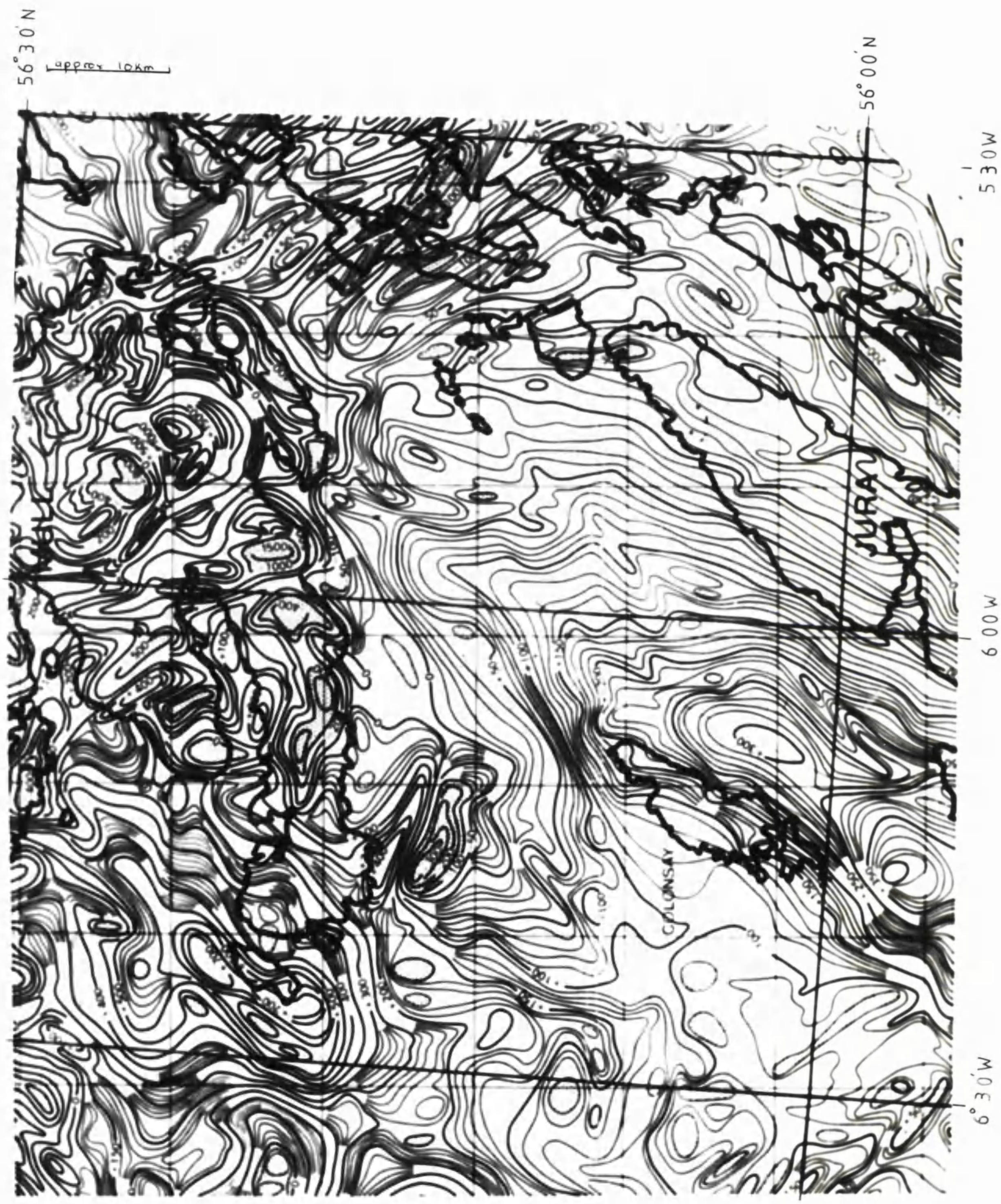


Fig 1.9 Part of IGS sh 10 Aeromagnetic map

and an upward continuation model of the I.G.S. aeromagnetic map, respectively, have proposed that the Loch Gruinart Fault has a downthrow to the south-east with a throw of about 1 km with respect to the displacement of the Permo-Triassic rocks (also proposed by Dobson <u>et al</u>., 1975) and probably a greater throw affecting the surface of the Lewisian basement. In addition to the vertical movement on the Loch Gruinart Fault, Durrance (1976) proposes that a 25 km dextral shift has also taken place whilst Westbrook and Borradaile (1978) prefer a 9 km sinistral shift. These minor movements of 25 km or 9 km indicate that it is unlikely that the Loch Gruinart Fault is the extension of the Great Glen Fault, however, it may be a major splay.

On the basis of gravity, magnetic and shallow seismic profiling (Riddihough, 1965, 1968; Riddihough & Young, 1971; Dobson <u>et al</u>., 1974, 1975), Dobson <u>et al</u>., (1975) have suggested a southward extension of the Loch Gruinart Fault across the Malin Sea. Although offset by several minor sinistral faults, trending NW - SE, it is considered to link up with either the Lag or Leannan Faults in Ireland.

(ii) <u>The outer shelf</u> - in contrast to the inner shelf the bathymetry is generally smooth due to the blanketing effect of recent sediments. The shoals of Stanton and Blackstone Banks are due to the outcropping of resistant igneous rocks; they are also the sites of magnetic and strong gravity anomalies (McQuillin & Binns, 1975).

Several linear troughs cross the area and have been attributed to assymetric, fault bounded sedimentary basins or half grabens (McQuillin & Binns, 1973; Binns <u>et al</u>., 1973; McQuillin & Binns, 1975; Binns <u>et al</u>., 1975). The principal faults involved are the Minch Fault, the Camasunary -

Skerryvore Fault and the Great Glen Fault with the course of the latter being the least well defined, Fig. 1.7 These troughs are floored by Precambrian or Palaeozoic rocks and contain a Mesozoic fill. A Faults presence is also often indicated by the concurrence of gravity and/or magnetic anomalies with troughs.

Smythe and Kenolty (1975) identified the 'Great Glen Fault' on a deep seismic reflection profile. They recorded it as bisecting a large NE - SW basin of Tertiary sediments (probably Oligocene in age and lying south and east of the Blackstones bank) which is supposedly downthrown to the south-east by 0.5 km. These sediments were considered to be underlain by lavas and Mesozoic sediments. This suggestion for the Great Glen Fault line follows that proposed by McQuillin and Binns (1973) on the same evidence. However, the author has examined the data on which these interpretations are based and found no evidence of a major fault in the region claimed by the above authors.

Magnetic data has been used (Riddihough, 1964, 1965, 1968; Riddihough & Young, 1971) to indicate a number of features:
(a) areas of Tertiary lavas (north of Donegal);
(b) several faults including three possible courses for the extension of the Great Glen Fault; none of which link with any on Ireland or Islay (Fig.1.10). These three proposed fault lines converge at a point above 30 km west of the north tip of Colonsay.

Riddihough (1964) suggested two possible marine extensions for the Leannan Fault, one passing to the west of Islay as a possible splay of the Great Glen Fault, the other linking with the Loch Gruinart Fault.

It should be noted that other faults in the area probably exist but are not detectable as there is no magnetic suscepti-

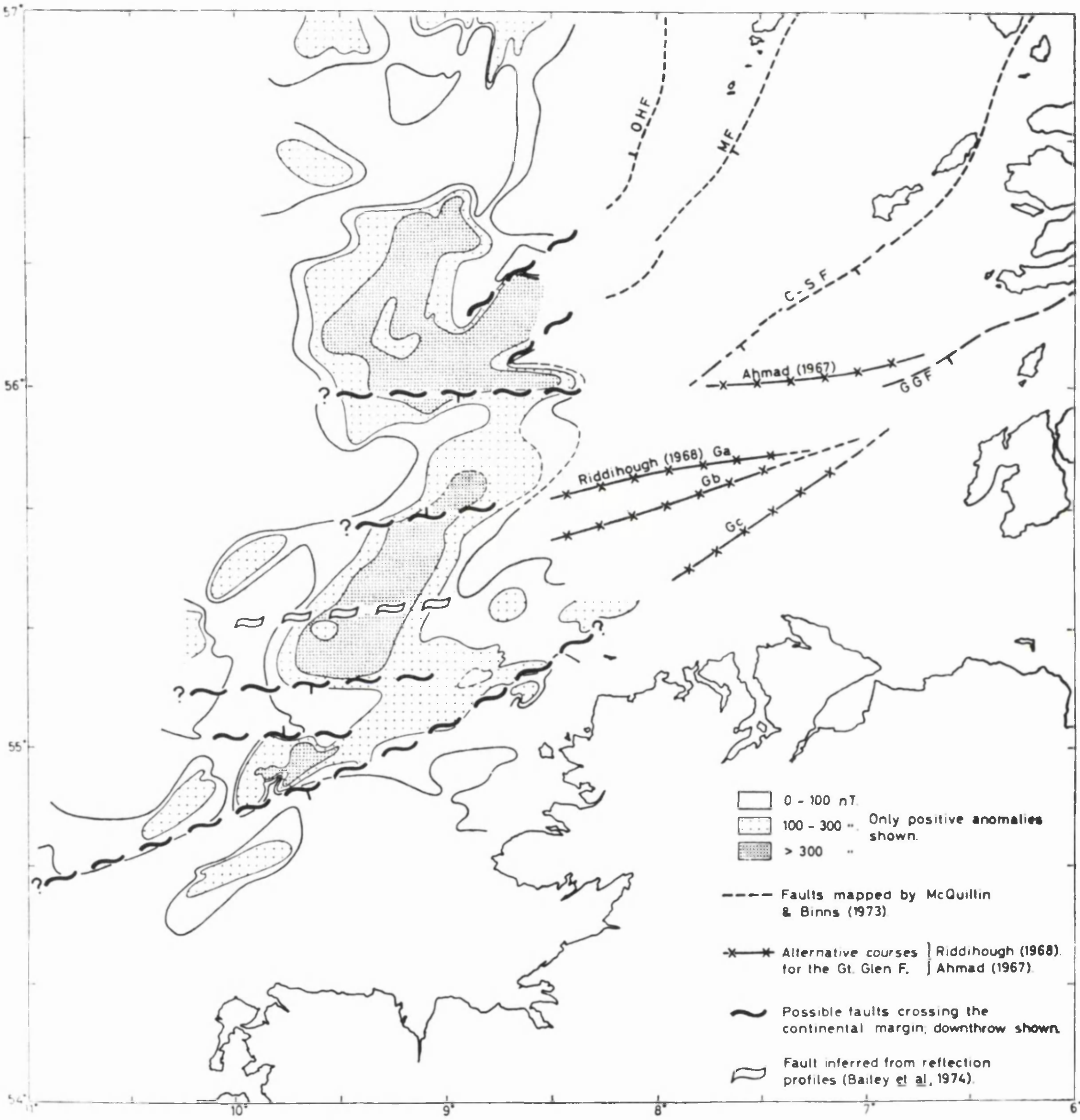


Fig 1·10 Simplified total field anomaly contour map showing possible faults inferred from the magnetic data, and those previously postulated to traverse the inner shelf. OHF = Outer Hebrides Fault; MF = Minch Fault; C–SF = Camasunary–Skerryvore Fault; GGF = Great Glen Fault (McQuillin & Binns 1973)

from Bailey et al. (1974)

bility or density contrast across them, e.g. the Iltay Boundary slide within the Dalradian which is seen on Islay (Bailey, 1925).

(iii) The Continental Slope - at the eastern edge of Rockall Trough the top occurs at about 140 - 180 m below sea level and has a uniform slope of 6° down to 1200 m below sea level (Roberts, *pers. comm.* in Binns *et al*., 1973).

Bailey *et al*., (1974) have demonstrated a 3-fold succession based on shallow seismic reflection profiles;

(a) acoustic basement with some structure visible and an irregular surface - not recognised seaward of the shelf break;

(b) prograding sedimentary sequence, dipping towards the shelf edge;

(c) horizontal sedimentary series unconformable on prograding sediments.

On the supposition that the Rockall Trough was opening 180 my - 80 my ago (Bailey *et al*., 1971; Laughton, 1971), Bailey *et al*. (1974) suggested that probable ages for these units are as follows: (a) acoustic basement - Triassic - Mid Jurassic, (b) prograding sediments - Late Cretaceous, and (c) horizontal sediments - Eocene to Recent.

There is evidence of a more complex history on some profiles showing older sedimentary wedges between the acoustic basement and prograding sediments. A thick apron of sediments is developed on the slope, with slumping on the upper slope. These sediments are not correlatable with the landward sediments due to the slumping and faulting.

At latitude $55^{\circ}30'$ N the margin shows a pronounced E - W dextral offset of 30 km. This offset is also displayed by the contours of the acoustic basement (Bailey *et al*., 1974, Fig. 5B) indicates movement took place post Jurassic. This structure lines up with possible courses for the Great Glen Fault, named by Riddihough (1968) as Gb and Gc and is also consistent with the proposed Tertiary dextral shift (Holgate, 1969) or the proposed post Devonian shift (Donovan et al 1976). The amount of movement indicated is less than that proposed by Garson and Plant (1972). For this fault a downthrow to the north, following the dextral shift, is indicated from the N - S lines of Bailey et al. (1974).

Wilson (1962), in the most extreme view, has suggested that the Great Glen Fault might link up with the Cabot Fault of Newfoundland via the Gibbs Fracture Zone. This was questionned by Holgate (1963) since movements on the Great Glen Fault are earlier than those observed on the Cabot Fault. Wilson (1963) replied that earlier movements on the Cabot Fault were possible, but as yet had not been recognised.

1.3.3 NE marine extensions

A marine deep seismic reflection survey in 1972 to investigate the Mesozoic sedimentary basin indicated by gravity survey in the Moray Firth (Chesher & Bacon, 1975) has revealed the presence of a NNE-trending fault zone, of variable width, following the northern coastline. Several stages of only normal movement appear to have taken place along this zone during the Mesozoic, as indicated by the increase in displacement of reflectors with age, seeFig. 1.11.

Flinn(1961) extended the Great Glen Fault northwards through Shetland along the Walls Boundary Fault and across the northern continental slope following an offset in the bathymetric contours which he attributed to a post slope formation sinistral movement of 80 km.

After proposing a dextral offset for the Devonian rocks

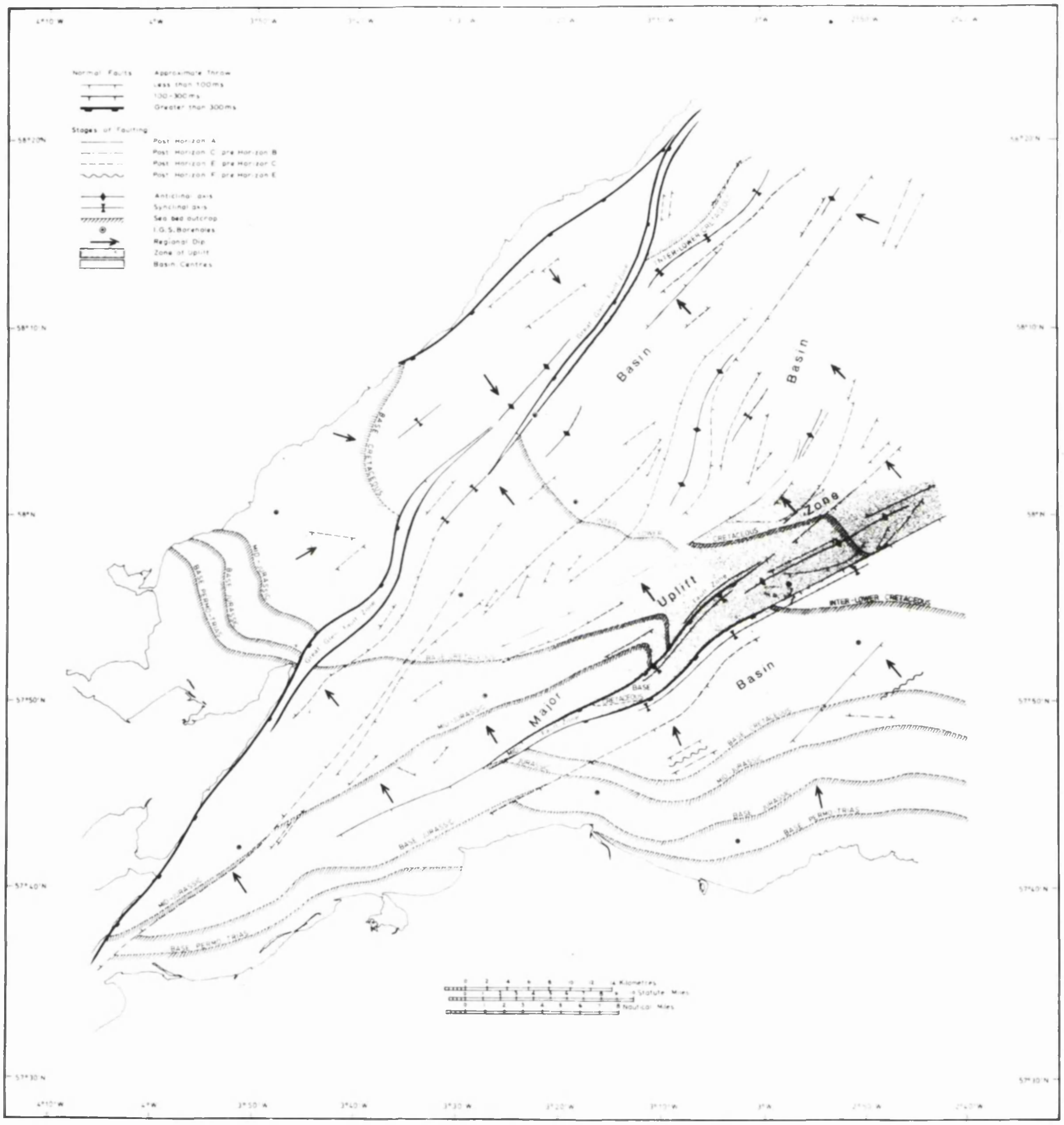


Fig 1·11 Summary structural map of the Moray Firth
(from Chesher & Bacon 1975)

of Shetland from aeromagnetic evidence (I.G.S. 1972) agreeing with Mykura's (1975) proposals for the Melby and Walls Boundary faults, Flinn (1969) modified his earlier view in proposing an initial pre-Devonian 165 km sinistral shift followed by a later Devonian dextral shift.

1.4 Tectonic Evolution

There have been two tectonic events of major significance in determining the distribution of rocks in this area. The first was the Caledonian Orogeny in Late Precambrian and Lower Palaeozoic times involving intense deformation, major faulting and igneous activity. The second event was the opening of the North Atlantic, also associated with faulting and igneous activity but with considerably less overall deformation.

The culmination of the Caledonian Orogeny was the suturing of continents across the Iapetus Ocean resulting in deformation and metamorphism of geosynclinal deposits (Harland & Gayer, 1972). Evidence for such an ocean is demonstated by comparing the faunal distribution in Cambrian and Ordovician rocks of the Lake District and Girvan (Williams, 1972; Skevington, 1974). The history of the Iapetus Ocean is extremely complex involving several openings and closings. It is only for the most recent events associated with the final closure and suture that evidence allowing reasonable models to be formulated exists. Many models for this final event have been proposed (Dewey, 1969; Fitton & Hughes, 1970; Phillips et al., 1976; Wright, 1976; Lambert & McKerrow, 1976; McKerrow et al., 1977; Moseley, 1977). All these models involve at least two subduction zones, running north-east - south-west beneath the Southern Uplands or Midland Valley and the Lake District and often a third zone beneath the Lake District together with a fault to the north of the Southern Uplands. Two sites have been proposed for the suture zone: over all the Southern Uplands (Dewey, 1969; Mitchell & McKerrow, 1975; McKerrow et al., 1977) and beneath the Midland Valley (Gunn, 1973; Garson & Plant, 1974).

One model (Phillips et al., 1976) is used to explain the cause of sinistral movement on the Great Glen Fault (Fig. 1.12). In this model suturing was initiated in the NE and progressed south-westwards. At the same time subduction was taking place further north resulting in the collision of the subduction zones ($sd_1$and sd_2 on Fig. 1.12) at depth. On this model, therefore, collision would have occurred first in SW and progressed NE. As the subduction of oceanic crust (SD2) would be stronger than the subduction of lighter continental material sinistral strike slip movement would be more likely. The lighter continental crust to the north would show a relative upthrow.

In the area of study, red beds were deposited during the Devonian. During the Carboniferous and Permian the region was also a land area. At this time the Variscan Orogeny was taking place with intense deformation further south in Britain and in Europe. Little deformation occurred in north-west Britain though it is possible that the Caledonian lines of weakness were reactivated and minor movements took place along them, e.g. post Devonian movements along the Great Glen Fault (Donovan et al., 1976).

During the Mesozoic the region of study was once again under the sea and deposition in fault bounded basins took place (e.g. McQuillin & Binns, 1973). Deltaic and estuarine conditions prevailed during the Jurassic, reverting to fully marine conditions in the Cretaceous.

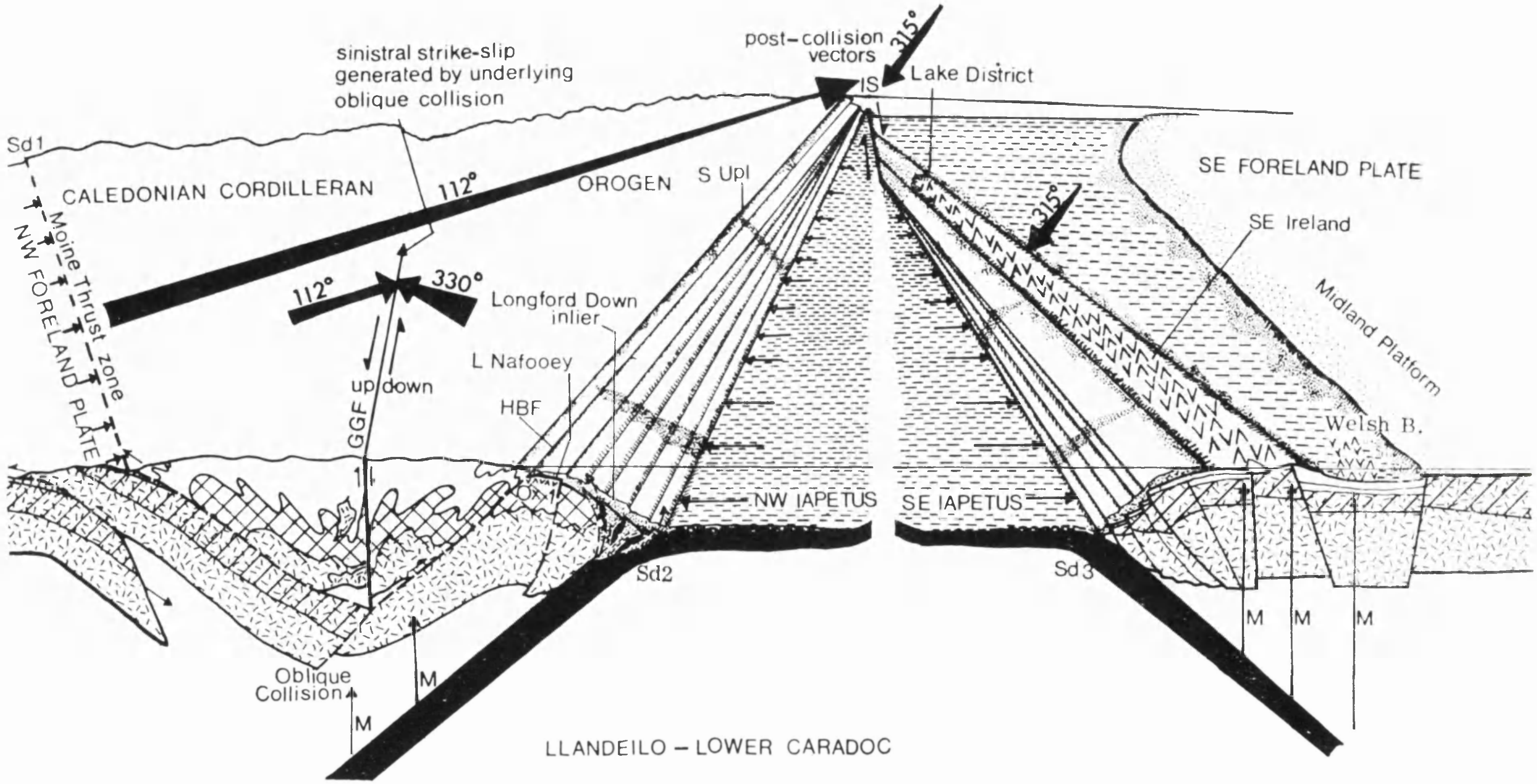


Fig 1·12 Palaeogeographical and structural reconstruction for the Llandeilo–lower Caradoc. Large arrows indicate plate movement directions inferred from tectonic trends. Small arrows indicate stratigraphic younging direction. Arrows with M indicate rising basic magma. (from Phillips et.al. 1976)

Sometime during the Mesozoic the North Atlantic began to open, initially along Rockall Trough. There is some uncertainty as to when the formation of Rockall Trough began, the earliest age proposed is early Permian (Russell, 1976). However, Bott and Watts (1971) suggest a Permo-Trias age although the concensus favours a late Jurassic or Cretaceous opening (Hallam, 1971; Bailey et al., 1971; Laughton, 1971; Roberts, 1971, 1975). The history of the North Atlantic is more complex than that of the South Atlantic and probably involved several stages of spreading at different locations; Rockall Trough, Labrador Sea, Reykjanes Ridge. Reconstructions for the opening of the North Atlantic are dependent on identification and dating of magnetic anomalies and on the identification of oceanic and continental crust.

Bott (1978) proposed the following history for North Atlantic evolution:

(1) About 230 - 180 my ago (Triassic) - early rift stage predates split between N. America and N. Africa.

(2) About 180 - 115 my ago (Jurassic and early Cretaceous) - separation of N. America and N. Africa, with left lateral shear movement between Europe and Africa.

(3) About 115 my ago to present - separation of Grand Banks and Iberia.

(4) About 115 - 75 my ago - opening of the Bay of Biscay and possibly the Rockall Trough - Faeroe - Shetland channel during this interval (details still obscure).

(5) About 80 - 45 my ago - opening of Labrador Trough.

(6) About 60 my ago to present - separation of the Rockall - Faeroe micro continent from Greenland, with contemporaneous opening of the Rockall Trough for the first 15 my of this interval.

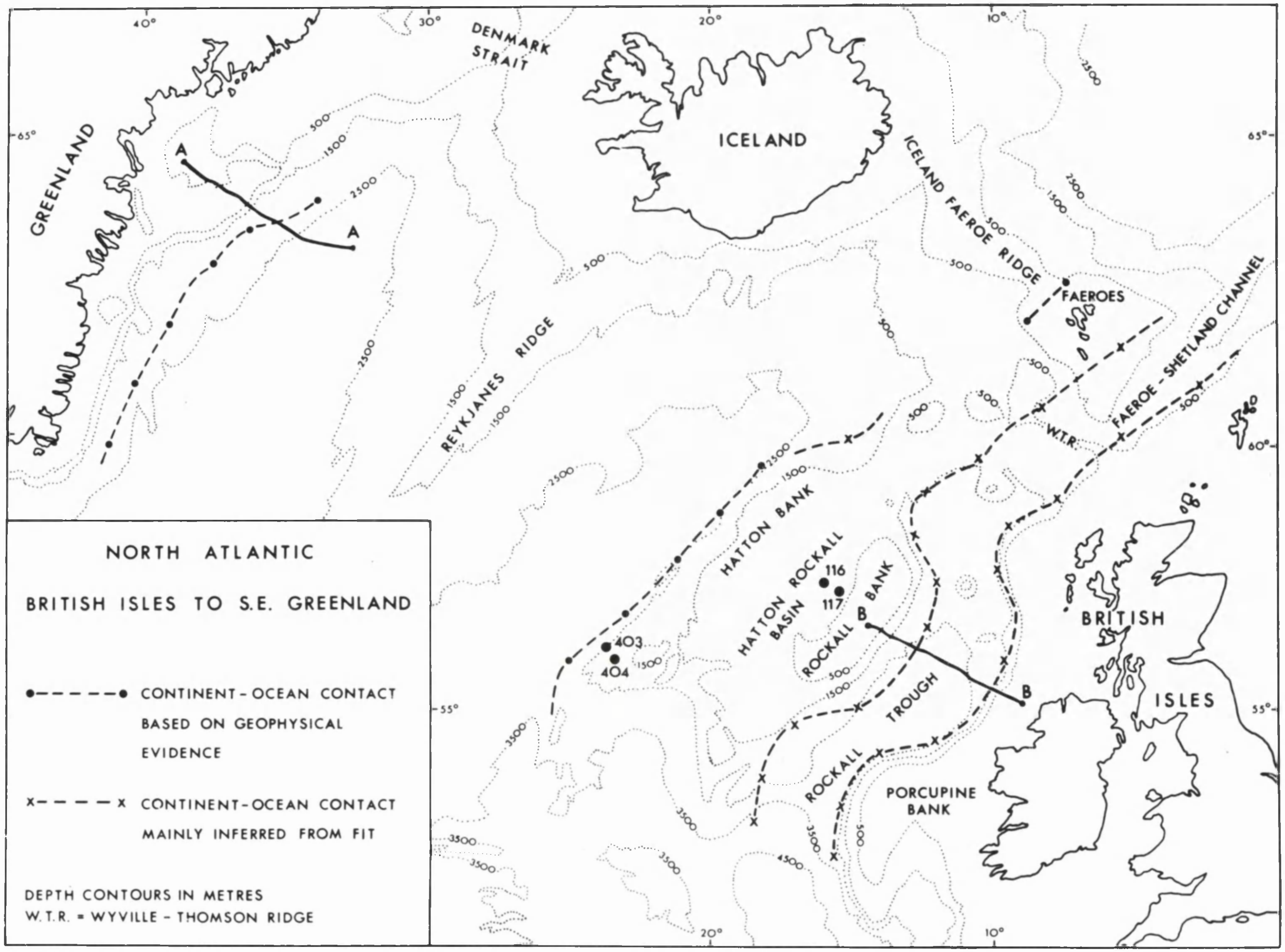


Fig 1·13. Bathymetric map of the Atlantic Ocean between the British Isles and Greenland, showing interpreted continent–ocean crustal contacts and D.S.D.P. holes 116, 117, 403 and 404. Depth contours based on unpublished data kindly provided by Dr A. S. Laughton.
(from (Bott 1978)

The Thulean vulcanism and the base Tertiary unconformably coincident with the onset of the last phase of spreading indicates the presence of a thermal anomaly in the Upper Mantle. Vann (1978) proposes that this anomaly originates in the outer core increasing heat flow into the base of the Upper Mantle by as much as ten times. Active convection commenced at approximately 65 my ago, as a result of this heat flow, raising the temperature in the asthenosphere to levels where partial melting could occur in a few million years.

Hallam (1972) believes that the formation of sedimentary basins such as the Minch Basin is due to faulting parallel to and as a result of spreading from the Reykjanes Ridge, by comparison to present day rifting observed at the Red Sea. However, this view is not shared by Hall and Smythe (1973) who suggest that the formation of these basins pre-dates spreading.

1.5 Objectives and Work Done

This study summarises the work done by the University of Glasgow (cruises RRS Challenger 1977 and RRS John Murray 1979) in the sea area around Mull and aimed at the general problems of the evolution of the western sea board in post Caledonian times. An understanding of basement features, such as major fractures, e.g. Great Glen Fault, is essential in developing viable models concerning the evolution of this area.

The following geophysical data was collected to enable the drawing of a geological map showing the three dimensional distribution of rocks in the area:

(1) 3,000 km sparker, magnetometer and echo sounder traversing

(2) 10 seismic refraction sites investigated

(3) 16 km deep seismic reflection and refraction line

CHAPTER TWO

TECHNIQUES

2.1 Introduction

The data presented in this thesis were obtained during two Glasgow University cruises in the Firth of Lorne area aboard N.E.R.C. research vessels; RRS Challenger, May 1977 and RRS John Murray, February 1979.

The work was divided into continuous reflection profiling, largely undertaken at night, and seismic refraction surveys using daylight hours. The aim was to elucidate the geological structure of this complex area.

The accuracy of the data, as of any geophysical survey, is dependent on the precision with which the geophysical parameters are observed and the location of their measurements determined. The quality of marine geophysical data, seismic in particular, is weather dependent; for the duration of the Challenger cruise a sea state of 2 or 3 was usual and it never exceeded 5, whereas during the John Murray cruise sea states of 5 or 6 and higher were common.

2.2 Navigation

Position fixes were taken at five minute intervals when an automatic buzzer sounded and, synchronously, data recordings were marked. The positions were plotted immediately on 1:50,000 U.T.M. base maps supplied by I.G.S.

The principal navigational technique was that of using a Decca Scientific Radar Type MT626 to obtain bearings and ranges from suitable local distinctive features of the coastlines. The reading of range is accurate to ± 20 m and bearings to $\pm .1^{\circ}$.

At a maximum range of 13 km this defines an error quadrangle of approximately 45 m x 40 m (Fig. 2.1), at closer range this quadrangle becomes narrower. However, it is likely that the

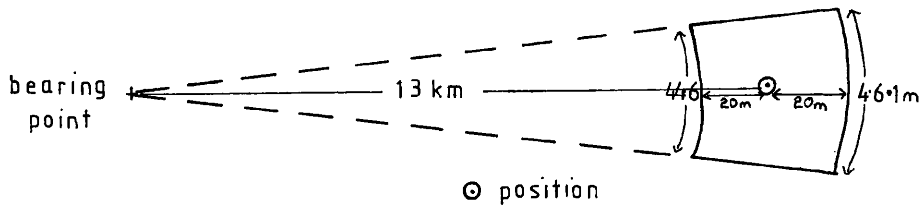


Fig. 2.1

final error is greater than this due to errors in identifying the target. In comparing the water depths (and sometimes sediment thicknesses) at crossovers on the sparker records it was found that the maximum difference between the plotted crossover position on one line and the point where the water depth equalled that at the plotted crossover position on the other line was 75 m and that the standard deviation of the differences was 26 m.

Readings were also taken from the Deccometer of Decca Navigator MK.21 system. This was tuned in 1977 to Decca Main Chain 8 E MP and in 1979 to Decca Main Chain 3 B MP. The errors incurred using this system are greater than those of radar positioning in this area, particularly at night due to skywave interference. Hence this system was used only as a back-up facility.

The Decca track plotter was utilized for the seismic array-line (1977) and some of the sono-buoy lines (1979). In this method the desired sail line is plotted on graph paper in terms of two sets of Decca lane co-ordinates e.g. red and purple, at convenient intervals (NB. a true straight line will plot as a curve due to the hyperbolic nature of the Decca lanes). This is then fitted to the track plotter where a pen marks the ship's position. The ship is then maintained on the correct course by steering so that the pen follows the line.

As a further check, the ship's heading and distance log were also recorded.

2.3 Seismic Techniques

Variations on seismic reflection and refraction techniques were employed to satisfy geological requirements that no one technique alone could achieve.

2.3.1 Continuous sub-bottom, single-channel seismic reflection profiling (Sparker)

This method is used to discriminate between and determine the spatial relationships of different rock units down to depths of 400 - 500 m with a resolution of about 10 m.

The method is summarised in Fig. 2.2. An E.G. & G. 3 candle spark array operated by discharging a capacitor bank across a submerged spark gap, is the acoustic source. Generally a spark of 1 kJ was used though 3.5 & 5 kJ sparks were also available. The seismic signal was detected by an E.G. & G. 263 C Array hydrophone. This comprised an 8m long, neutrally-buoyant, liquid-filled hose containing 16 individual, equispaced hydrophones, having a band width of 20 Hz - 5 kHz.

The seismic signal was recorded unfiltered onto magnetic tape in analogue form using a Hewlett Packard 3960 instrumentation recorder. This allowed subsequent playback at different filter settings. A 6.4 kHz reference signal, supplied by the E.P.C. chart recorder, was recorded on one channel and used as a synchronizing control on playback. On a third channel the 5 min. navigational fix mark was recorded.

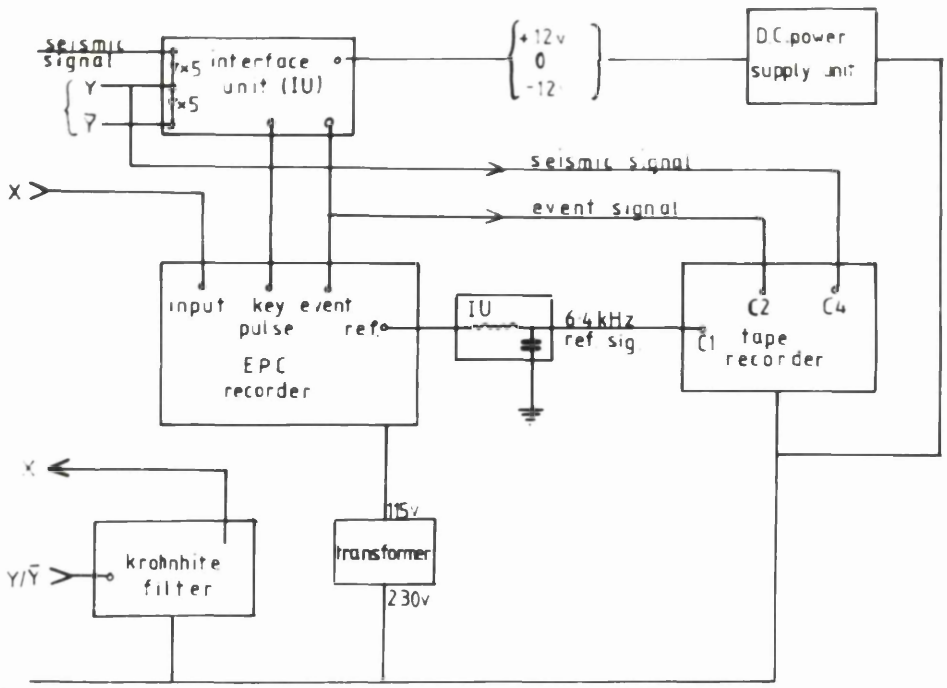
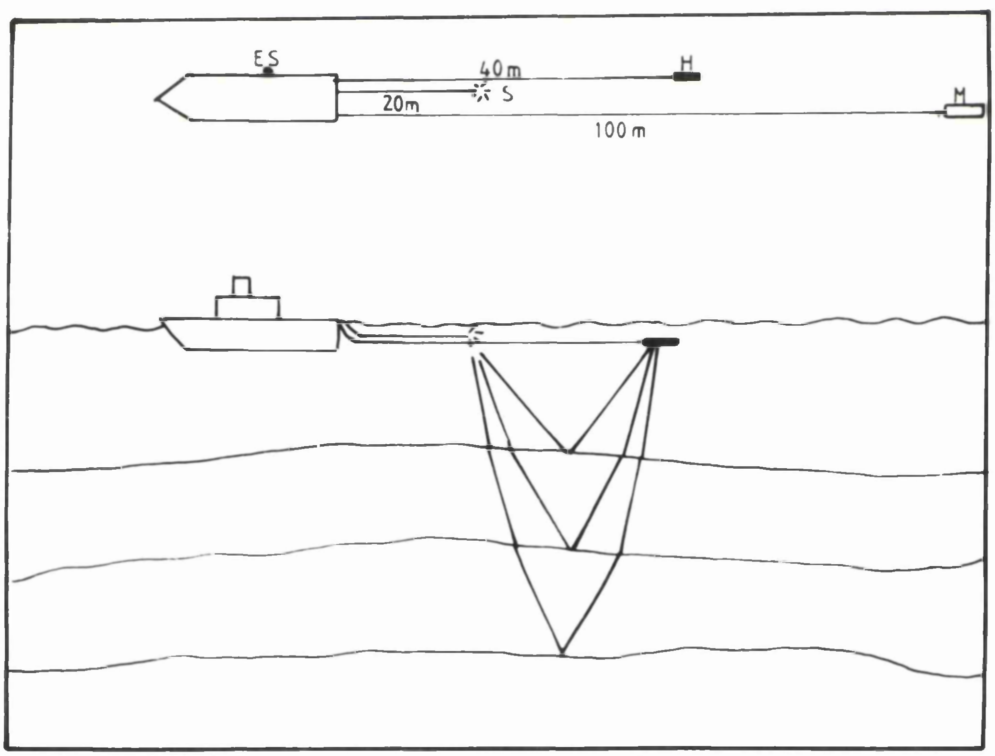


Fig 2.2a Schematic circuit diagram for recording seismic data.



H-hydrophone S-sparker M-magnetometer ES - echo sounder

Fig 2.2b Sparker and magnetometer profiling.

The seismic signal was also recorded in graphic form on an EPC 4600 chart recorder after passing through a Krohnhite 3550 analogue filter. The filter was used in the bandpass mode with bandpasses in the range 80 - 800 Hz adjusted for optimum recording quality dependent on conditions. The chart recorder supplied the trigger pulse for the spark chosen to be at 1s intervals. The stylus swept the width of the paper every 0.5s printing for the first 0.5s after the spark only. Timing lines were marked at 0.1s intervals, these and the seismic signal were scorched on the paper by an electric current passing through the stylus.

The majority of the seismic reflection data was acquired using a 1 kJ spark. Two lines were shot in 1979 using different sources: one a 3.5 kJ spark with a shot interval of 2s, with the recorder printing the first 1s; the other a 40 cu. inch air gun firing at 4s intervals, with the recorder printing the first 1s. These were used in order to increase penetration by increasing the energy and lowering the frequency content of the seismic signal, thus reducing attenuation (see Fig. 2.3).

2.3.2 Seismic refraction profiling using Sono-buoy

These experiments were carried out in areas where reflection profiling did not provide sufficient basis for determining the geology. Refraction data provide information on the seismic velocity and also indicate the layering of the rock units to a depth greater than that obtainable by sparker profiling.

Previous seismic refraction data obtained in the area had involved the use of 4 free-floating Bradley sono-buoys to each of which a neutrally-buoyant near-surface hydrophone is attached. The buoys were deployed in a straight line and explosive charges, used as the acoustic source, were detonated

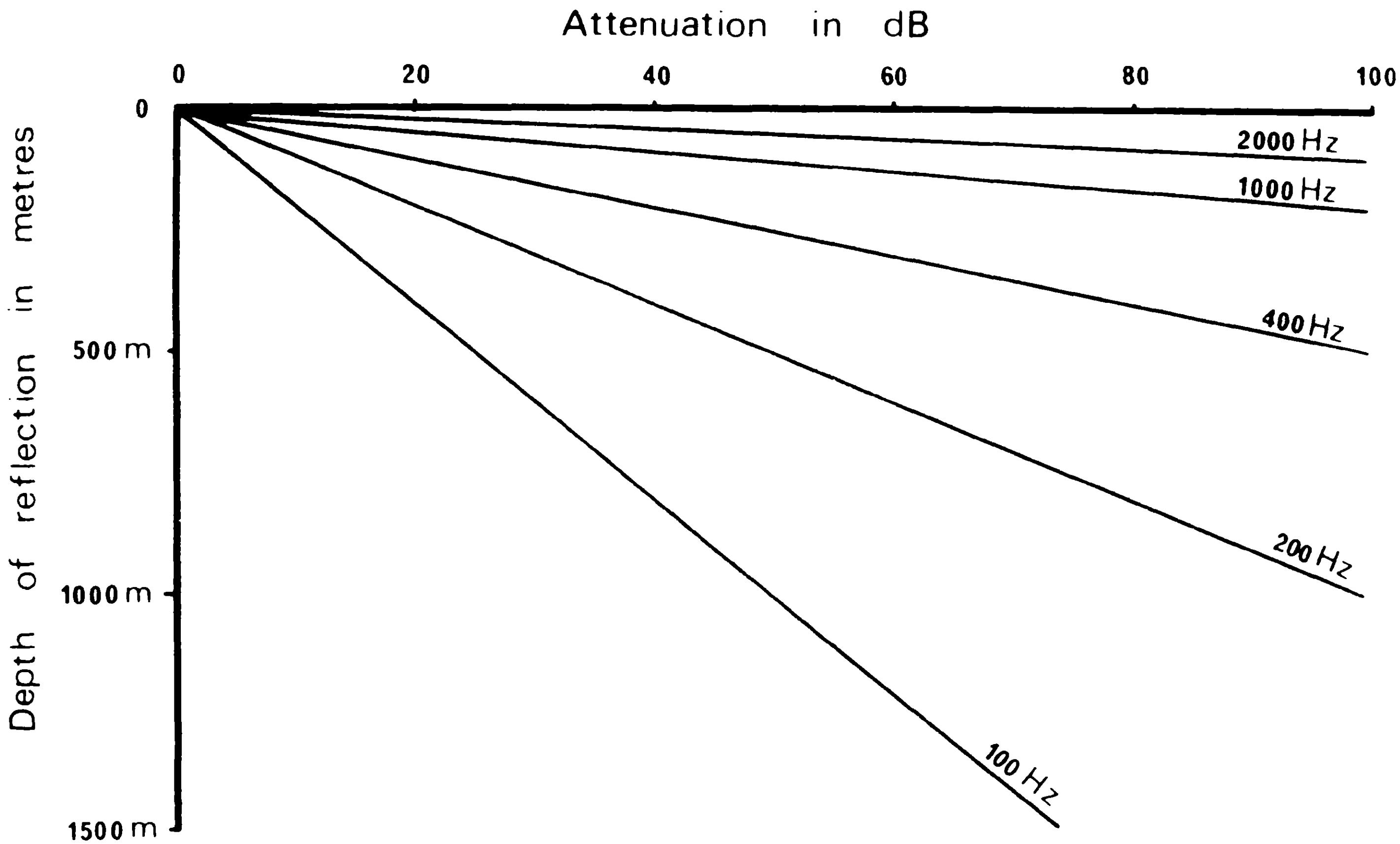


Fig 2·3 Attenuation of seismic waves for an absorption of 0.5dB per wavelength plotted as the effect of absorption on the amplitude of a sound wave reflected at a range of depths 0–1500m.
from McQuillin & Ardus (1977)

at various distances from one end of the line. Several disadvantages are inherent to this method:-

(1) The buoys drift in the strong currents prevalent in this area, therefore (a) they can easily drift out of line, (b) they can drift off the outcrop under investigation, and (c) their absolute position is uncertain.

(2) Infrequent shots lead to a paucity of points on the time distance graph.

(3) The range over which first arrivals from a low velocity top layer may occur is short, resulting in poorly-defined refracted arrivals (Faruqee 1972).

These problems can be solved by anchoring one buoy at a suitable site, using a rapid repetition rate acoustic source, e.g. air gun or sparker, and by placing the hydrophone on the sea bed. Fig. 2.4 illustrates the difference in range over which the low velocity arrival, refracted along the sea bed, occurs as a first arrival. With the hydrophone near sea level the low velocity arrival is the earliest over the range 245 - 360 m from the hydrophone, whereas with the hydrophone anchored the low velocity arrival is earliest over the range 122 - 360 m i.e. double the range. When the shot interval of the two methods is also considered approx 300 m using explosives approx 20 m (for an air gun firing at 4s intervals and ship speed of 5 knots) the probability of observing a low velocity arrival is minimal using explosives.

Line lengths of 6 km either side of the buoy are sufficient to determine the apparent velocities of sediment and underlying rock head (for a sediment thickness of 1 km overlying rock head of 4 kms^{-1} the first arrival from rock head would arrive first at a range of approx 2 km from the buoy).

The anchored sono-buoy technique is summarised in Fig. 2.5.

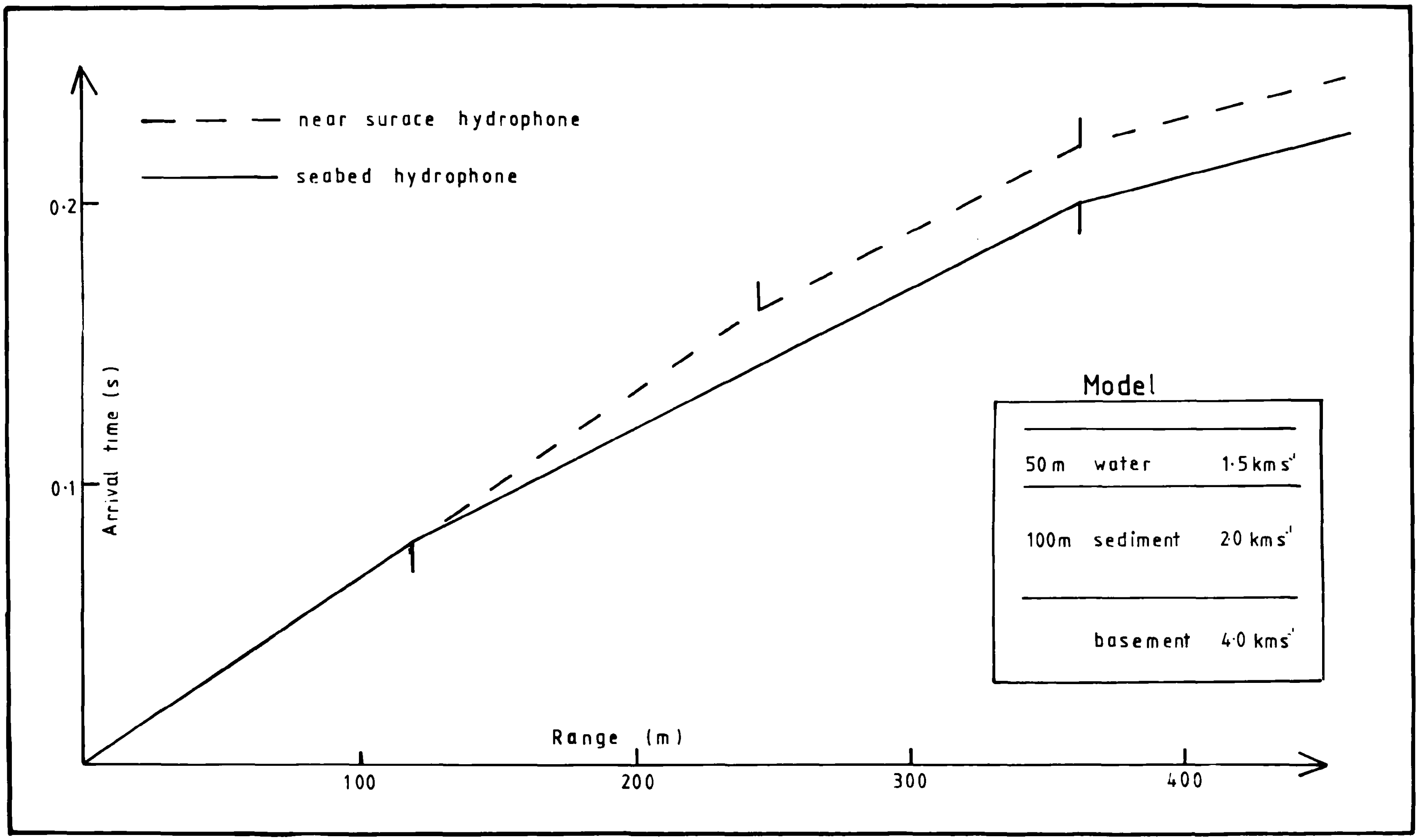
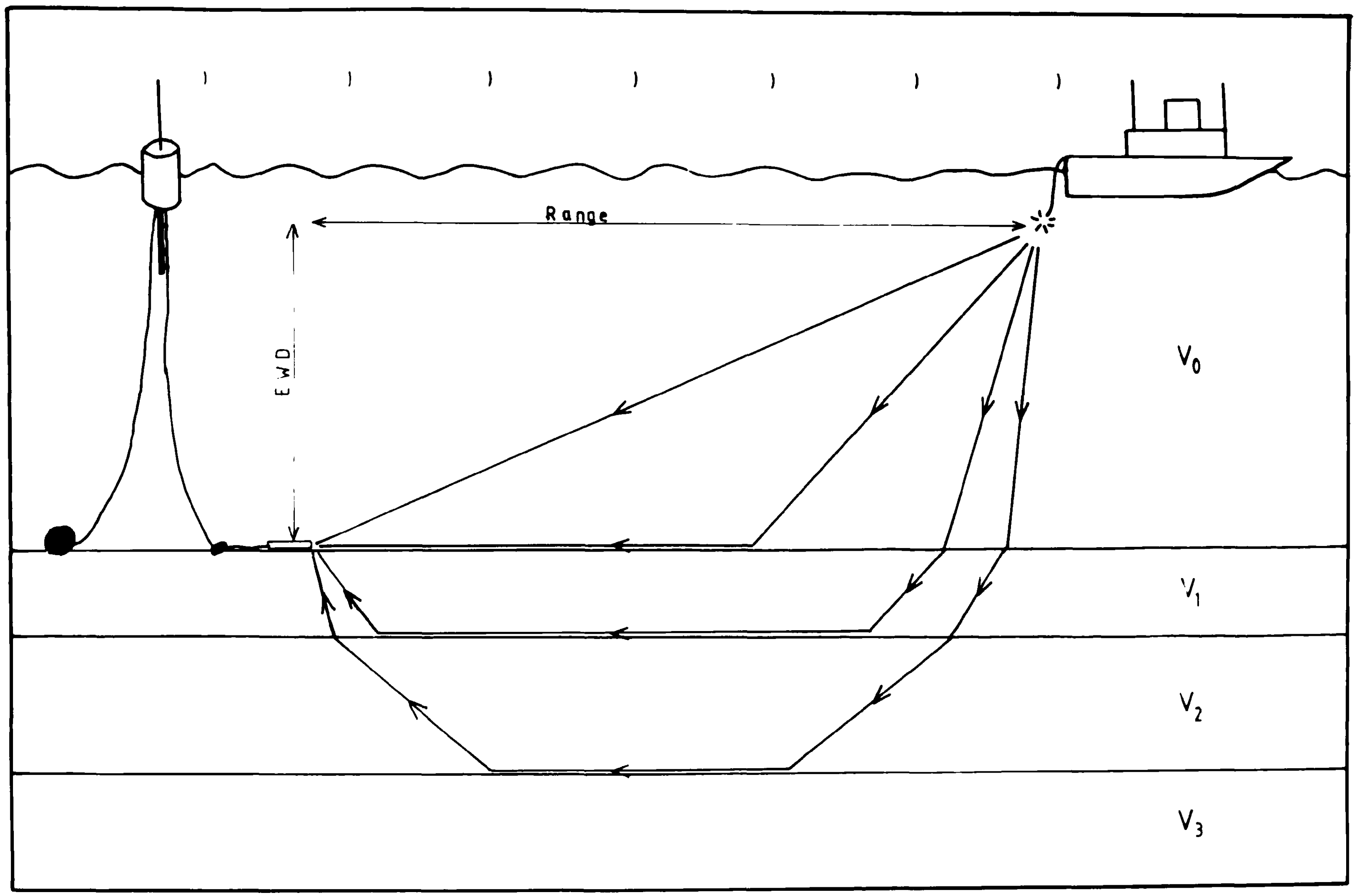


Fig 2·4 Comparison between near surface and seabed hydrophones



$V_3 > V_2 > V_1 > V_0$

Fig 2·5 Anchored sono buoy technique

Bolt Par air guns with 40 cu. in. and 150 cu. in. barrels operating at 2000 psi. were used as acoustic sources. These discharge compressed air on receipt of trigger pulses from the EPC recorder. The maximum fire rate is dictated by the capacity of the compressor to recharge the air gun chamber. In this work the 40 cu. in. air gun was fired every 4s and the 150 cu. in. air gun every 16s. 1 kJ and 5 kJ sparkers were also used along one line in an attempt to increase the resolution of the low velocity arrivals by using a source with a higher frequency content and a rapid repetition rate (1-2s).

A hydrophone built at Glasgow University was lightly anchored (approx 2 kg) to the sea bed and attached by a length of cable, at least $1\frac{1}{2}$ x water depth, to a sono-buoy also anchored with a greater weight ($\approx$15 kg). The seismic signal is pre-amplified in the hydrophone and fed into the buoy where it is amplified again. The signal then modulates the frequency of a 3.75 kH subcarrier which amplitude modulates a radio transmission at a frequency close to 27 MHz.

On board ship the signal is detected on one of two $\frac{1}{4}$-wave whip aerials, fed into the receiver rack where the subcarrier is recovered. This is fed into the demodulater rack where the original seismic signal is recovered.

Apart from the hydrophone the equipment described comprises the Geophysical Telemetry System, manufactured by G. & E. Bradley.

The same recording system employed for the sparker data was used, except that occasionally an EPC 4100 recorder was used to permit a greater number of firing and printing sequences.

The operational procedure was as follows:-
On arrival at the selected site the sono-buoy was deployed

astern the ship in the following sequence (1) buoy anchor, (2) buoy, (3) hydrophone cable played out from buoy end, (4) hydrophone and anchor last. The acoustic source to be used was then streamed. The ship then sailed away from the buoy along a prescribed line until the signal strength diminished to the level of the background noise or the geological objective of obtaining refracted arrivals of sufficient length to permit their apparent velocities to be determined was achieved. At some sites the line was repeated using different acoustic sources for comparison. At other sites a cross line perpendicular to the first was run. On completion of the programme the sono-buoy and hydrophone were retrieved.

2.3.3 <u>The seismic array experiment</u> (1977 only)

The aims of this experiment were (i) to obtain a complete subsurface seismic reflection coverage of depths greater than that obtained by sparker profiling and (ii) to gain first arrival refraction data along the traverse. This was carried out along a line approximately parallel to sparker and sono-buoy lines.

The method is summarised in Fig. 2.6.

A 40 cu. in. air gun was used as an acoustic source. A Geomechanique multi-channel seismic hydrophone array of 11 active sections (50 & 60 m long), spring sections were placed between the array and the tow cable to decouple the array from much of the ship noise. Depth controllers were fitted to maintain the array at a constant depth of 6 m (see Fig. 2.7). Due to the considerable length of this streamer (1146 m) operation was restricted to open water.

The seismic signals from the array were recorded on a T.I. 8000B 24-channel seismic recorder (Glasgow owned), a

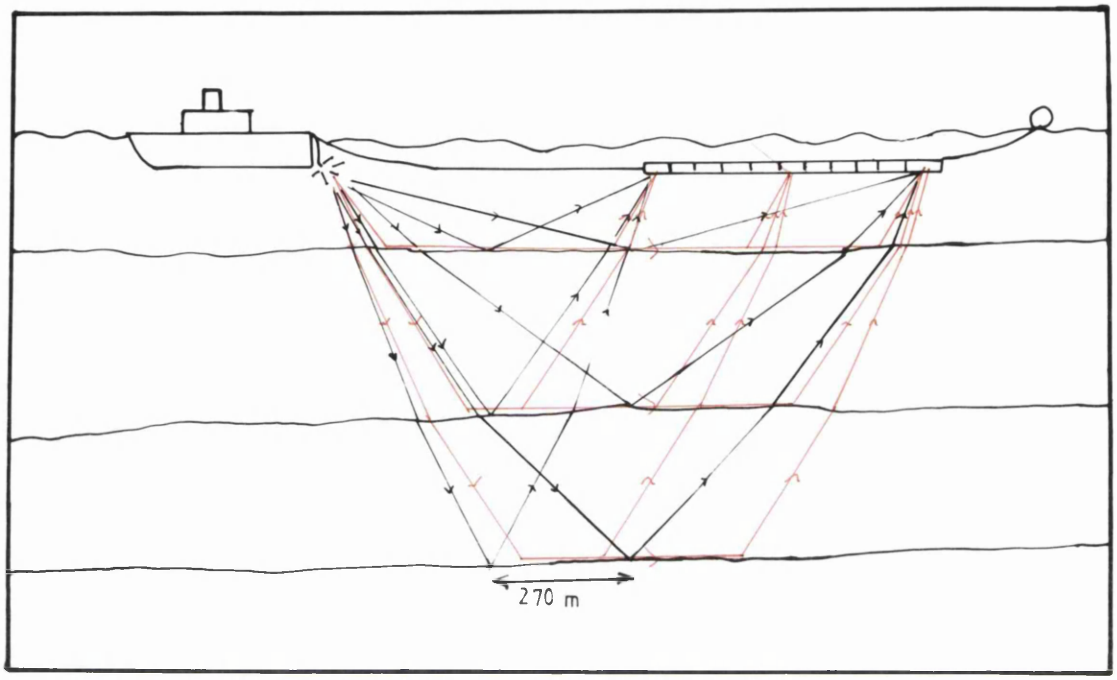


Fig 2·6 Seismic array profiling

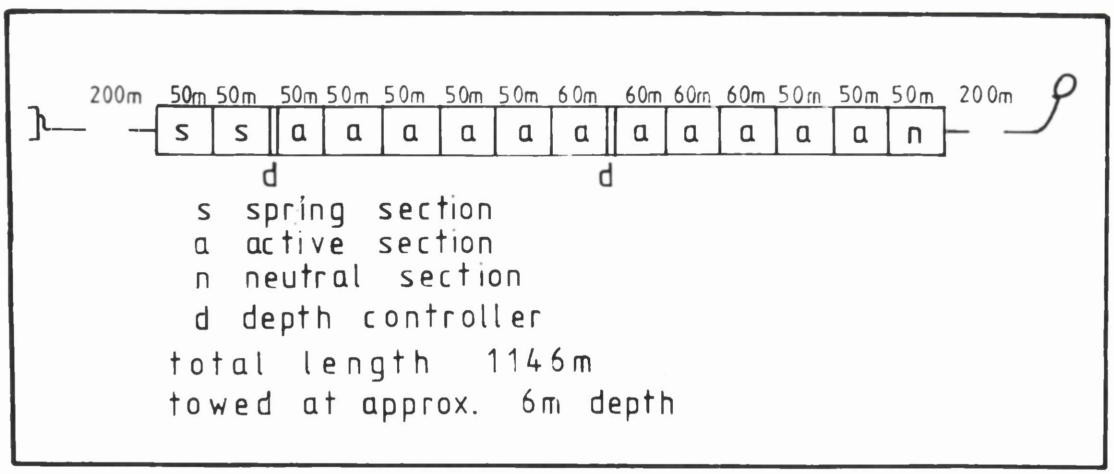


Fig. 2·7 Hydrophone array

detailed description is given by Hall (1971, appendix 3). This system consists of three major components: an amplifier and filter unit, a multihead, analogue, magnetic-tape-drum recorder and a squiggle-trace oscillograph. The seismic signal is fed into the amplifier and filter unit. The processed signal is output in parallel to the recorder and oscillograph. The magnetic tape recording is of 6s duration and the single sheet tape used for each recording can be replayed back through the amplifier and filter unit into the oscillograph, allowing further adjustment of the filter and gain settings to be made. As the recorded signal is filtered care was taken not to over-filter initially.

The operational procedure was as follows:-
The ship followed a pre-plotted course fitted to the bridge Decca track plotter. Shot points at 272 m intervals were marked along the course to give 100% subsurface coverage. The shot was triggered by a pulse from the recorder 700 ms after it was started. After each shot a fresh tape was fitted to the recording drum and the monitor squiggle record stowed in a light proof box.

2.3.4 Echo Sounding

The operational procedure is described by McQuillin & Ardus (1977). A P.E.S. Mk.3 echo sounder was used in 1977. This was used principally to measure water depth, particularly useful when deploying sono-buoys. The full scale deflection of this instrument was 100 m and depths could be read off to an accuracy of ± 0.5 m using an overlay grid. The depths were corrected for the depth of the transducer below the sea surface but not for tidal variations (tidal range approx 6 m), thus reducing the accuracy to ± 3.2 m with respect to mean sea level. In addition to water depth, some information regarding the nature of the sea bed could be obtained using this instrument by examining the form of the echo and multiples (if any).

During the 1979 cruise, a Simrad echo sounder was used which supplied water depth information only. The full scale deflection on this record was 125 fm (228 m) and depths can be read off to $\pm \frac{1}{2}$ fm (1 m). So the overall accuracy of this method is $\pm$ 3.2m

2.3.5 Side-scan sonar

A Kelvin Hughes MS47 side-scan sonar fish was used scanning the starboard side of the ship only, theory of operation described by McQuillin & Ardus (1977, pp.39-45). This was used to pick out topographical features of the sea bed in two dimensions. Due to limited success, this was used during the 1977 cruise only.

2.3.6 Velocity meter

A Plessey Type TL031 sound velocity meter was deployed over the side at one site during the 1977 cruise. The velocity of sound in water was determined at depths of 9 m and 18 m and found to be 1485.00 $\pm$ 0.12 ms^{-1} and 1484.50 $\pm$ 0.15 ms^{-1} respectively. These lie in the range 1489 $\pm$ 8 ms^{-1} as derived using the formula (McGuinness 1966)

$$C = 1449 + 4.6\,t - 0.055\,t^2 + 0.000\,3t^3 + (1.39 - 0.012\,t)\,(s - 3s) + 0.017\,d$$

where C - velocity of sound in water (ms^{-1})

t - temperature (^{o}C)

s - sea water salinity (p.pth.)

d - water depth (m)

values for t & s were obtained from a Hydrographic department publ. 1970 and taken to be 10 ± 2°C and 34 ± 2 p.pth. respectively.

2.4 Magnetometer Profiling

The strength of the earth's total magnetic field was measured simultaneously with sparker profiling and the style of variations observed used as an aid to the interpretation of the sparker data.

A Varian proton magnetometer was used in 1977 and a Barringer proton magnetometer used in 1979. Both magnetometers were towed approx 100 m astern the ship in order to clear the magnetic effects of the ship. The principal and operation of proton magnetometers is discussed by McQuillin & Ardus (1977, Ch.6).

During both surveys the output was recorded in analogue form on a Servoscribe recorder with a full scale deflection of 1000 nT and the record can be read to ± 1nT. In addition to the analogue record, a digital display is continuously monitored and was noted in the log at fix times.

CHAPTER THREE

DATA PROCESSING AND INTERPRETATION

3.1 Introduction

This section describes the quality of the data obtained during 1977 and 1979 Glasgow cruises (Fig. 3.1). The geological map (Fig. 3.2) is the primary result of the work done in this thesis.

3.2 Results and Processing

3.2.1 Sparker

The quality of the shipboard graphic records obtained in 1977 was generally better than those obtained in 1979 due to better weather. However, the quality of both is generally good enough that filtered playbacks were not required.

These records are used in the production of:

(i) The Geology Map (Fig. 3.2);

(ii) The Depth to Rockhead Map (Fig. 3.3); and

(iii) The Sediment Isopach Map (Fig. 3.4);

where possible the apparent dip of clearly-bedded rock was calculated.

Preparation of Maps of Depth to Rockhead and Sediment Isopachs

The water depth and sediment thickness were measured directly from the records at fix intervals. The sediment thickness was plotted on one map (Fig. 3.4) and the sum of water depth and sediment thickness plotted on another map (Fig. 3.3). These maps were then contoured manually using a 50 m contour interval.

Transparent templates were prepared to allow direct reading of water depths and sediment thicknesses from the records. Vertical incidence reflection was assumed. Interval velocities of 1.5 kms^{-1} and 2.0 kms^{-1} were assumed for water and sediment respectively. In assuming vertical incidence a slight over-estimate in the measurements is indicated (Table 3.1).

Two-way Travel Time (ms)	Water Depth			
	Assuming vertical incidence	Shot to hydrophone separation		
		15ms(22.5m)	20ms(30m)	30ms(45m)
100	75	74.15 (1.1%)	73.48 (2.1%)	71.5 (4.0%)
200	150	149.58 (0.3%)	149.25 (0.5%)	148.3 (1.1%)

Table 3.1 Comparison of water depths determined by assuming vertical incidence and with true travel path depths. % over-estimate assuming vertical incidence is given in parenthesis.

Generally the separation of hydrophone from shot is 15 - 20ms (22.5 - 30 m) for all the sparker data. The grids can be read to ± 1 m. A tidal range of 6 m reduces the accuracy to ± 3.2 m (4.3% for a depth of 75 m).

Similar errors are incurred by the measurement of sediment thickness, though the error due to tidal range does not apply.

The overall accuracy of these contour maps is not likely to be worse than ± 5 m.

In some areas (e.g. around N56°13' W06°30') it was not possible to identify rock head. In these instances a minimum sediment thickness is given, that measured from the seabed to the first seabed multiple. Likewise a minimum depth to rock-head is given as the sum of the minimum sediment thickness and water depth. In other areas (eg axial trough S of Mull) rockhead can be traced to the limit of the record - 500 ms. Where it passes beyond this limit, minimum thicknesses are again given.

In the upper Firth of Lorne there is no detectable sediment over most of the area. The depth to rockhead contours for this area were redrawn from the Bathymetry map based on Admiralty soundings (Rashid 1978), as a denser coverage was obtained by that survey than by ours.

Determination of apparent and true dips in rockhead

In areas of well-bedded rocks (e.g. North and South of Ross of Mull) the apparent dip along the traverse can be determined. If bedding is visible on two lines at intersection the true dip can be calculated.

In order to measure apparent dip, a suitable reflector is selected (i.e. one that is continuous for at least ½ a fix interval). Depth to seabed, sediment thickness and thickness from rockhead to the reflector are measured, using templates, usually at 1/5 fix intervals (an interval velocity of 3.0 kms^{-1} was usually assumed -- rock with the low dips observed on the records are believed to be Mesozoic and interval velocities of this magnitude have been measured in the area (Smythe & Kenolty 1975)). The distance between fixes is measured on the map. The section is then plotted at natural scale allowing the dip of the reflector to be measured directly from the section.

The accuracy of this estimate is affected by the accuracy of measurement, the assumption of interval velocities, the range over which the measurement is made and the actual value of the dips as the reflectors are not migrated.

True dips can be obtained from intersecting profiles using the equations as given in McQuillin and Ardus (1977)

$$\tan C_t = \frac{\tan B_2.\cos C_1 - \tan B_1.\cos C_2}{\tan B_1.\sin C_2 - \tan B_2.\sin C_1}$$

$$\tan B_t = \frac{\tan B_1}{\cos(C_t-C_1)} = \frac{\tan B_2}{\cos(C_t-C_2)}$$

B_1, B_2 - apparent dips along lines 1 and 2

C_1, C_2 - bearings of lines 1 and 2 relative to North

B_t - true dips

C_t - true bearing of dip direction relative to North

Apparent thickness (Ta) can be estimated from the outcrop width (Wo) and apparent dip (θ).

$$Ta = Wo \sin\theta$$

<u>Preparation of the Geological Map</u>

The sparker data combined with a semi-quantitative assessment of the magnetic record provided the basis of identification of the rock types occurring in the area. Different reflection and magnetic characters were distinguished; units were identified with reference to the known land geology, marine seismic refraction data and other seismic velocity data (Rashid 1978). (Table 3.2, Figs 3.5 - 3.13).

As can be seen in Figs. 3.9, 3.10, 3.11, 3.12 and 3.13 the seismic characters of Dalradian, Moine, Granite, Torridonian and Lewisian rocks are similar in that these rocks form acoustic basements from below which no coherent arrivals return. Careful inspection of the records reveals some differences. The Dalradian and Torridonian have smoother surfaces. The Torridonian occurs at an almost constant depth of about 50 m. The Dalradian has an irregular topography partly smoothed in places by overlying sediment.

	Sparker	Magnetic character	Velocity measurements (kms^{-1}) Hammer seismic	Laboratory (mean of 3 measured along perpendicular direction)	Land Array unpublished G.U. data	Marine refraction
Soft Sediment	Even well bedded very low dips following topography of rockhead which can often be observed clearly. Fig. 3.5	-	-	-	-	1.75-2.5 Fig.3.54
Tertiary Basalt	Smooth surface forms plateaus. Little or no overlying soft sediment. Can see layering below surface Fig. 3.6	edged by lows of moderate wavelength (600m) amplitude 800	4.3 *	5.6 *	4.4	-
Mesozoic Sediments	Smooth surface usually overlain by soft sediment dipping reflection observed below the surface. Fig. 3.7	-	3.1-3.8 * (sandstone)	4.09 * 4.4 (limestone) 3.9 * 3.6 (sandstone) 4.7 (conglomerate)	3.3-3.5 3.93	2.7-3.7 Fig.3.35-4
LOR Sediments	Form an acoustic basement. No coherent reflectors from below surface.	-	3.9 * (sandstone) 4.1 * (conglomerate)	4.8 * 5.1	- -	- -
	Irregular surface, no subsurface layer-					

LOR Lava	Irregular surface, no subsurface layering generally forms shoals Fig. 3.8	short wave-length (300m) moderate amplitude(25)	4.8*	4.6* 5.4	-	5.0 (unreversed) Fig.3.34
Dalradian Quartzite	Smoother surface than Moine and granite. No internal reflection Some hyperbolic Fig. 3.9	-	4.8*	5.3*	-	-
Dalradian Phyllite		-	4.5*	5.0*	-	-
Moine	Rugged surface plus hyperbolic reflection Generally no over-lying soft sediment Fig. 3.10	-	(schist) 4.3*	(quartzite) 5.7* (schist) 5.7*	(schist) 5.1	4.4-6.2 Figs.3.42,43 44,48 5.3 Fig.3.50
Ross of Mull Granite	Less rugged than Moine plus some soft sediment Fig.3.11	-	-	-	-	-
Torridonian Sandstone	Smooth surface. No coherent subsurface reflection Fig. 3.12	-	-	-	-	-
Lewisian	Very irregular sur-face plus hyperbolic reflections Fig. 3.13	broad high	-	-	-	-

TABLE 3.2 Rocks identified in the area (Results from Rashid (1978))

FIGURES 3.5 - 3.13

Sparker profiles

Timing lines at 100 ms intervals

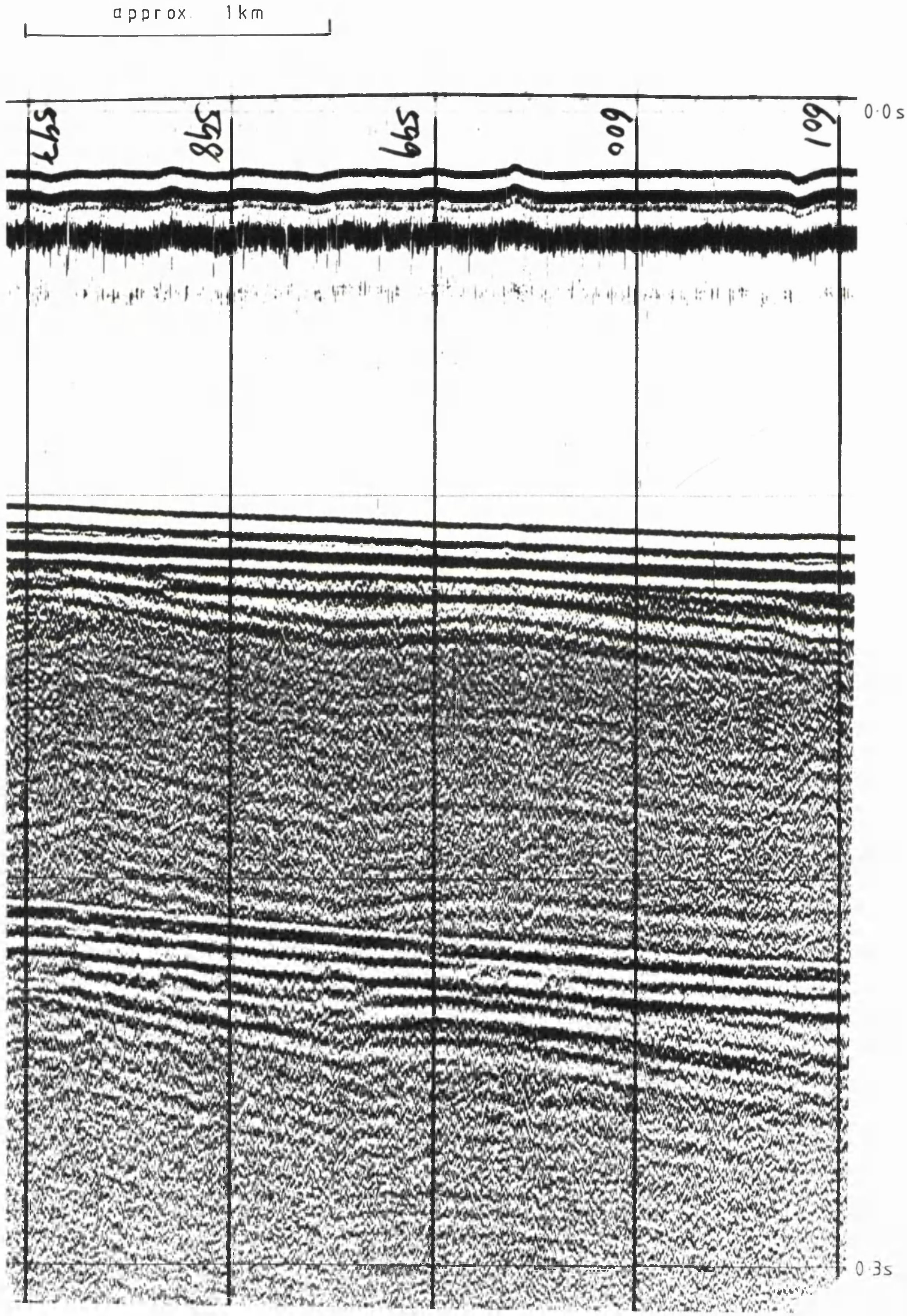


Fig 3.5 Sparker profile of recent sediment

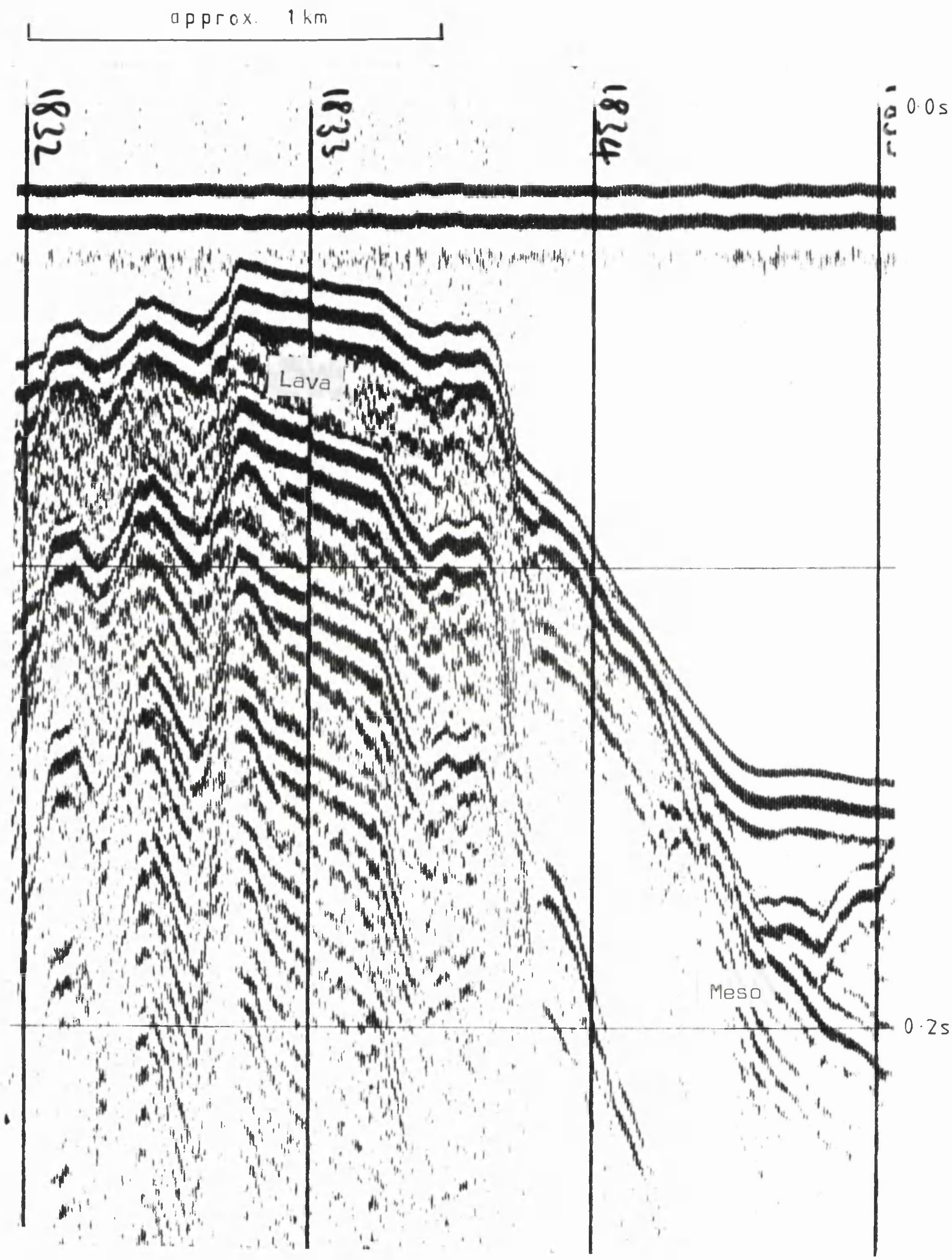


Fig 3.6 Sparker profile of Tertiary lavas and Mesozoic sediments

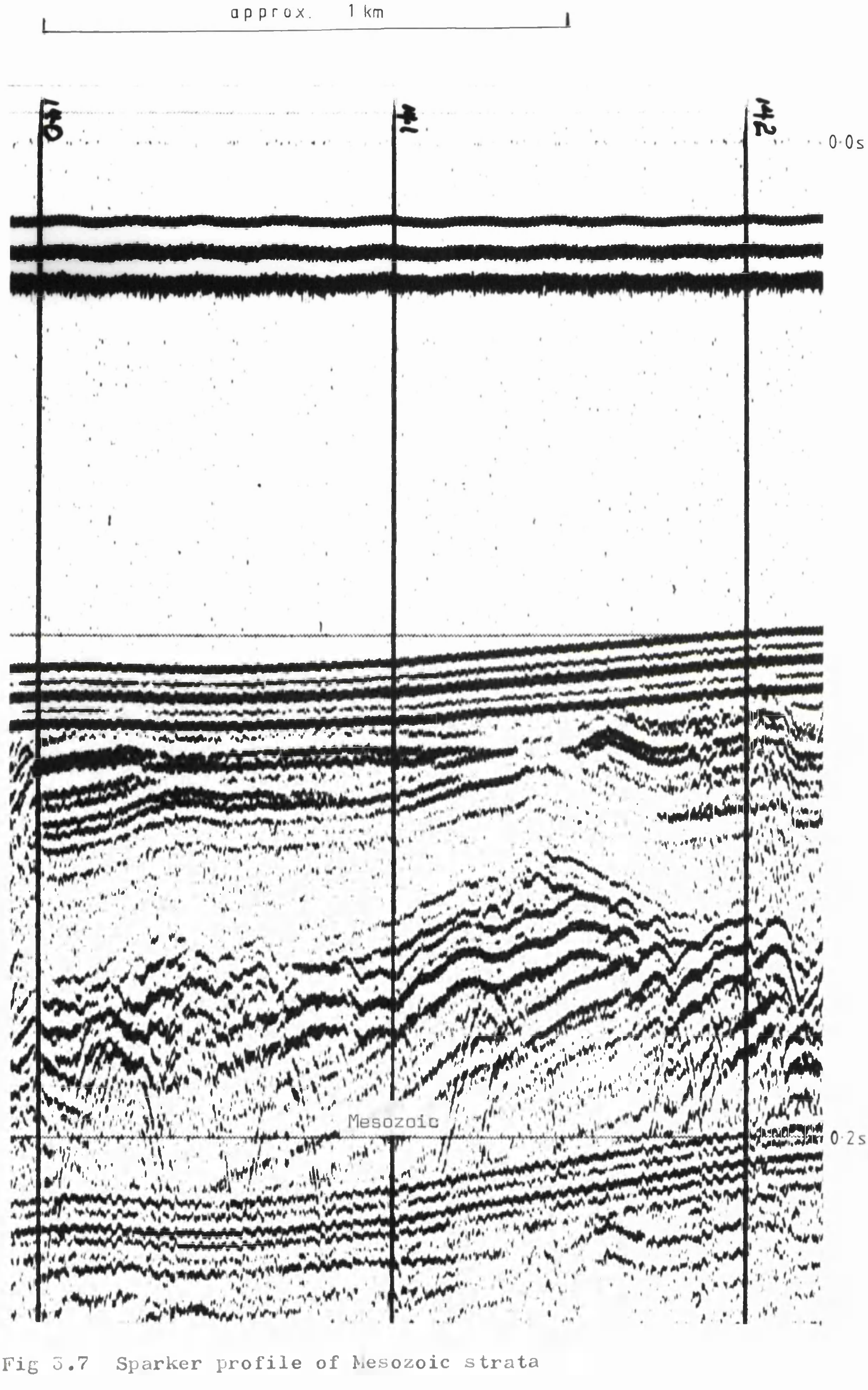


Fig 3.7 Sparker profile of Mesozoic strata

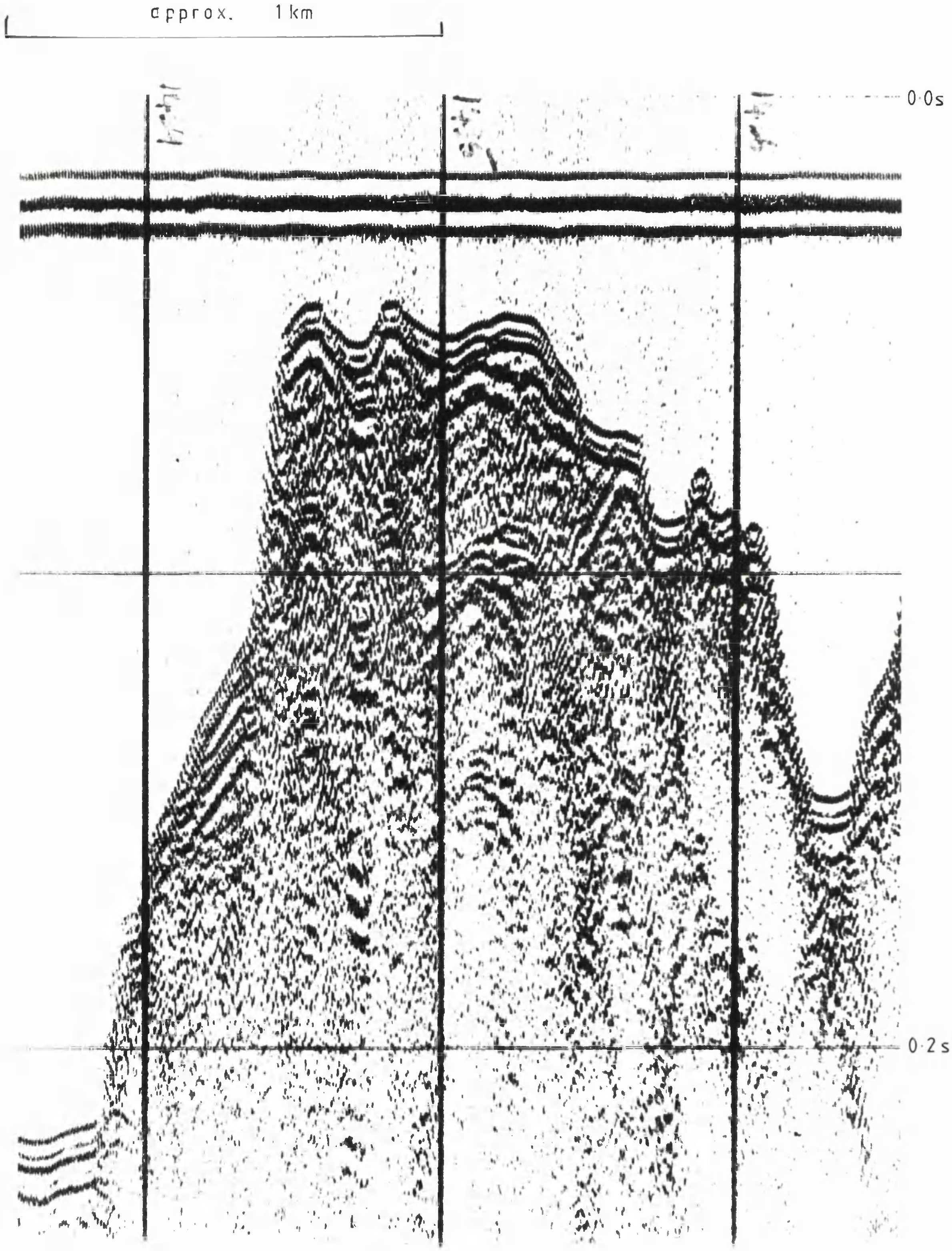


Fig 3.8 Sparker profile of L.O.R. lava

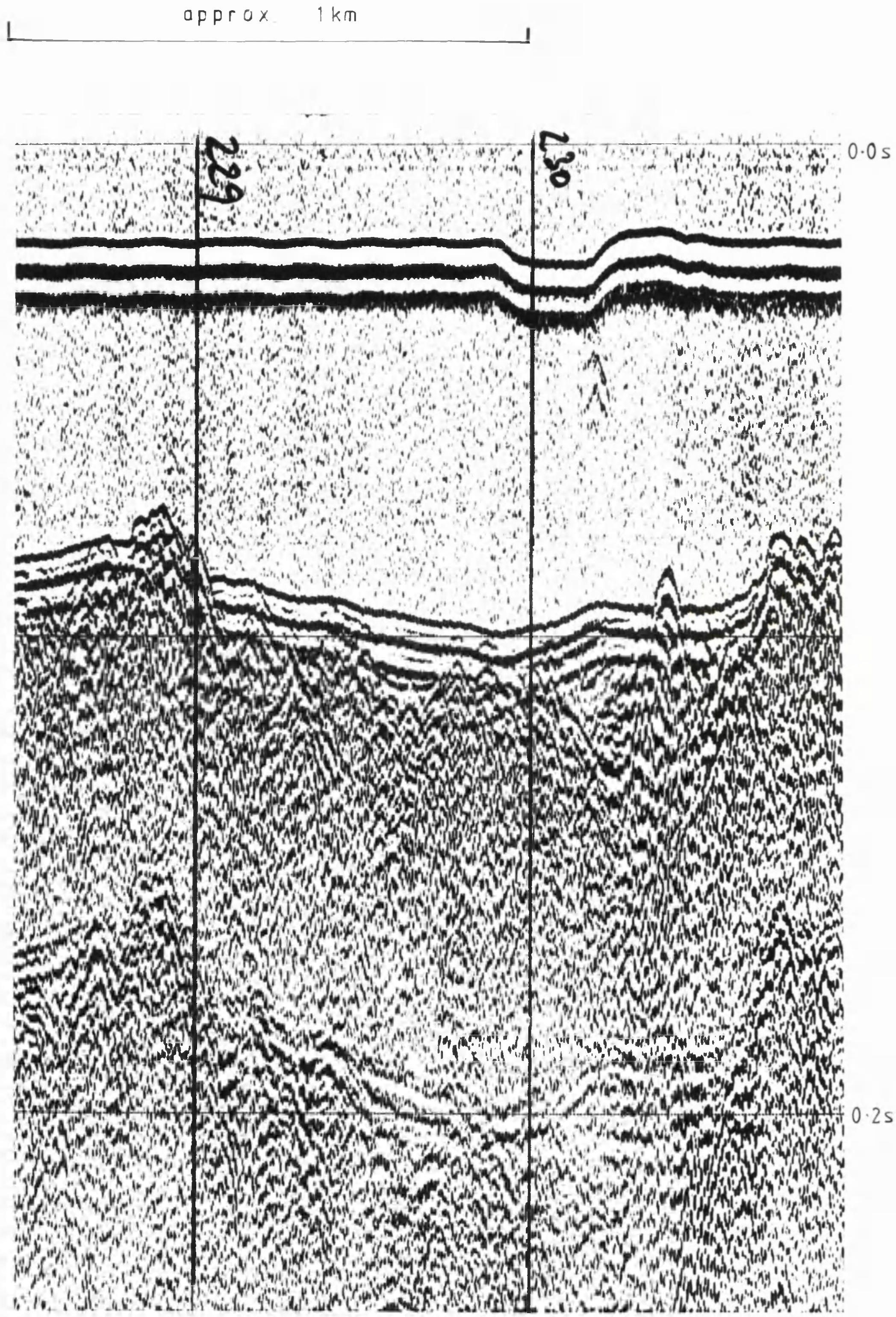


Fig 3.9 Sparker profile of Granite

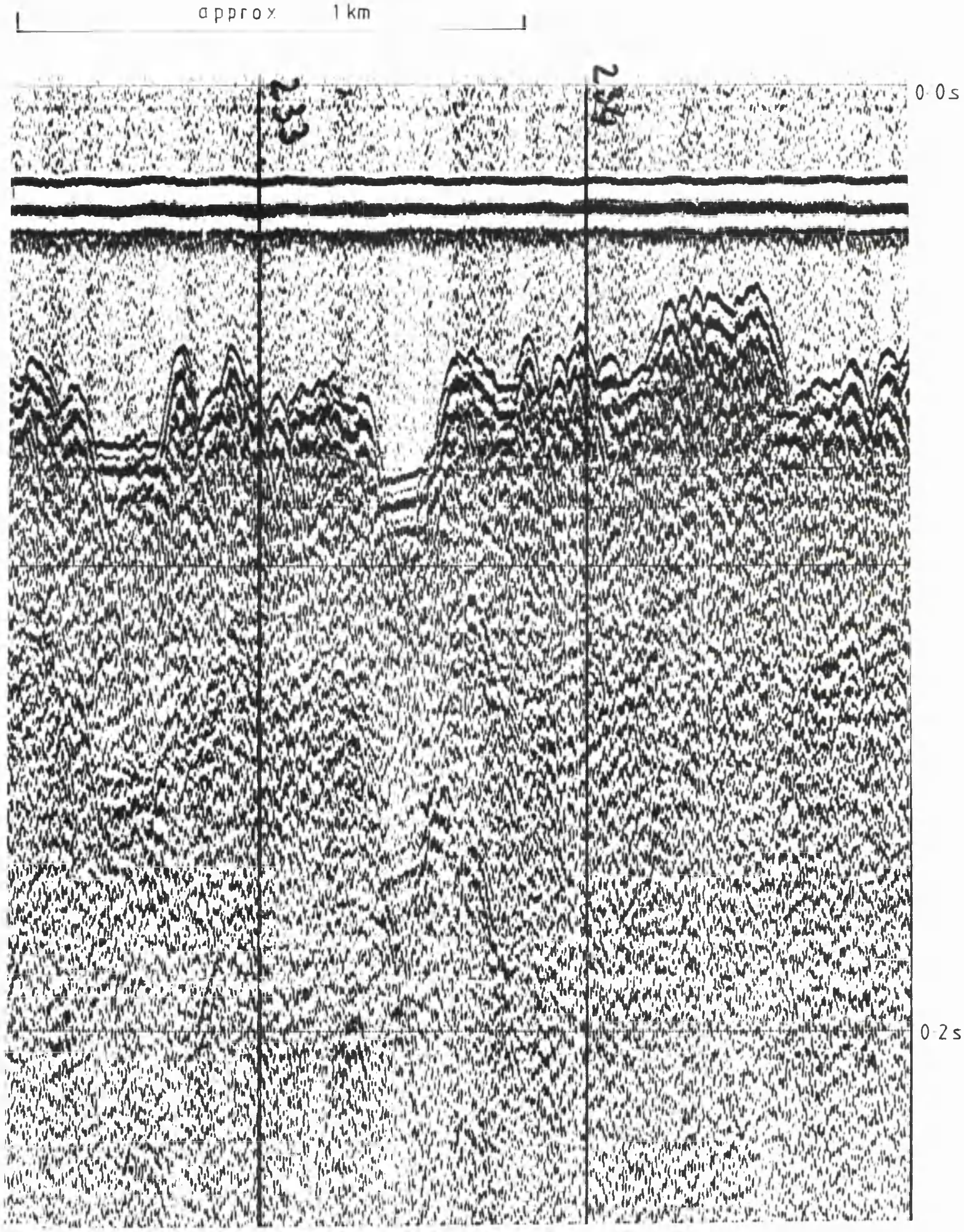


Fig 5.10 Sparker profile of Moine rocks

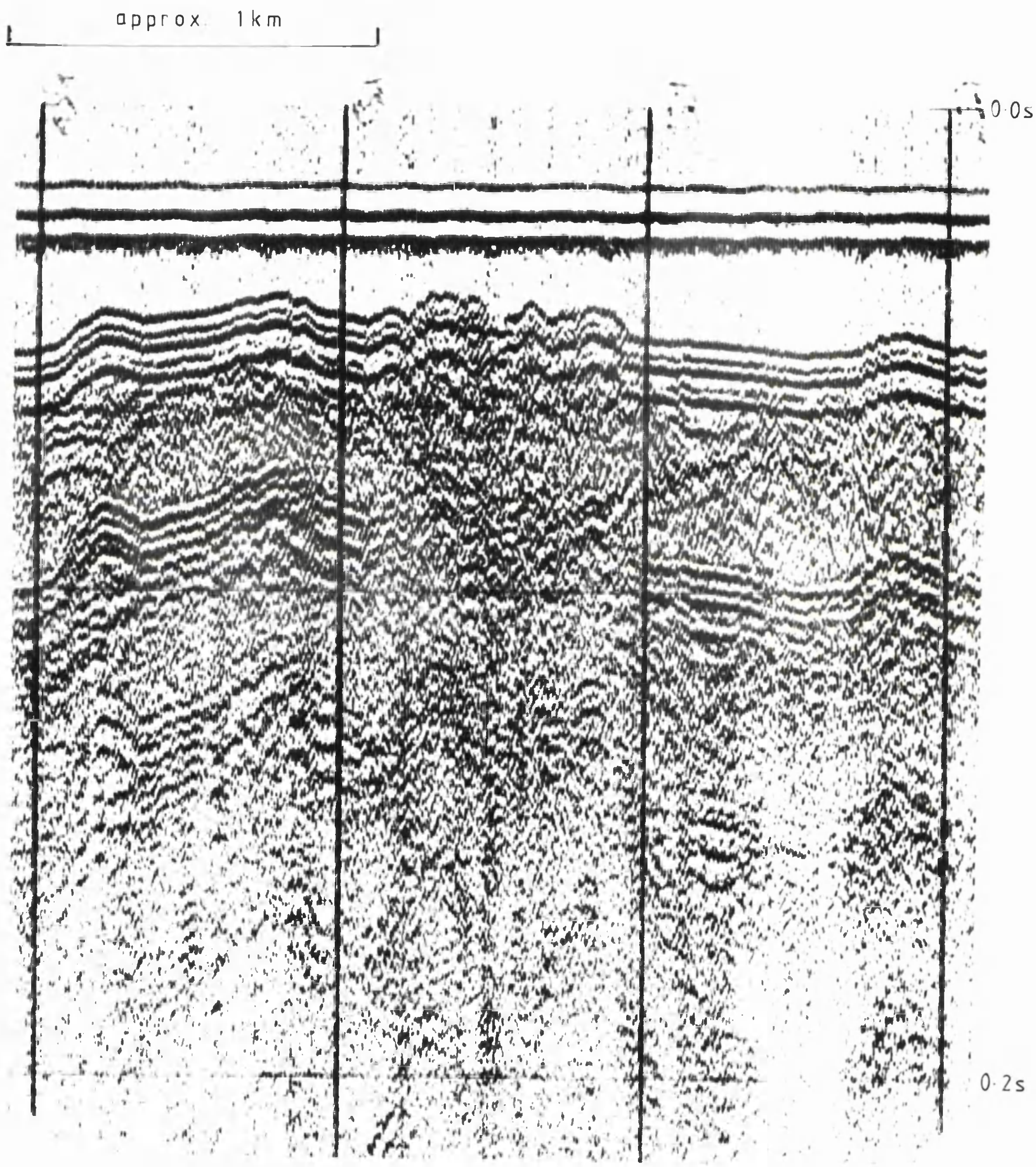


Fig 3.11 Sparker profile of Torridonian rocks

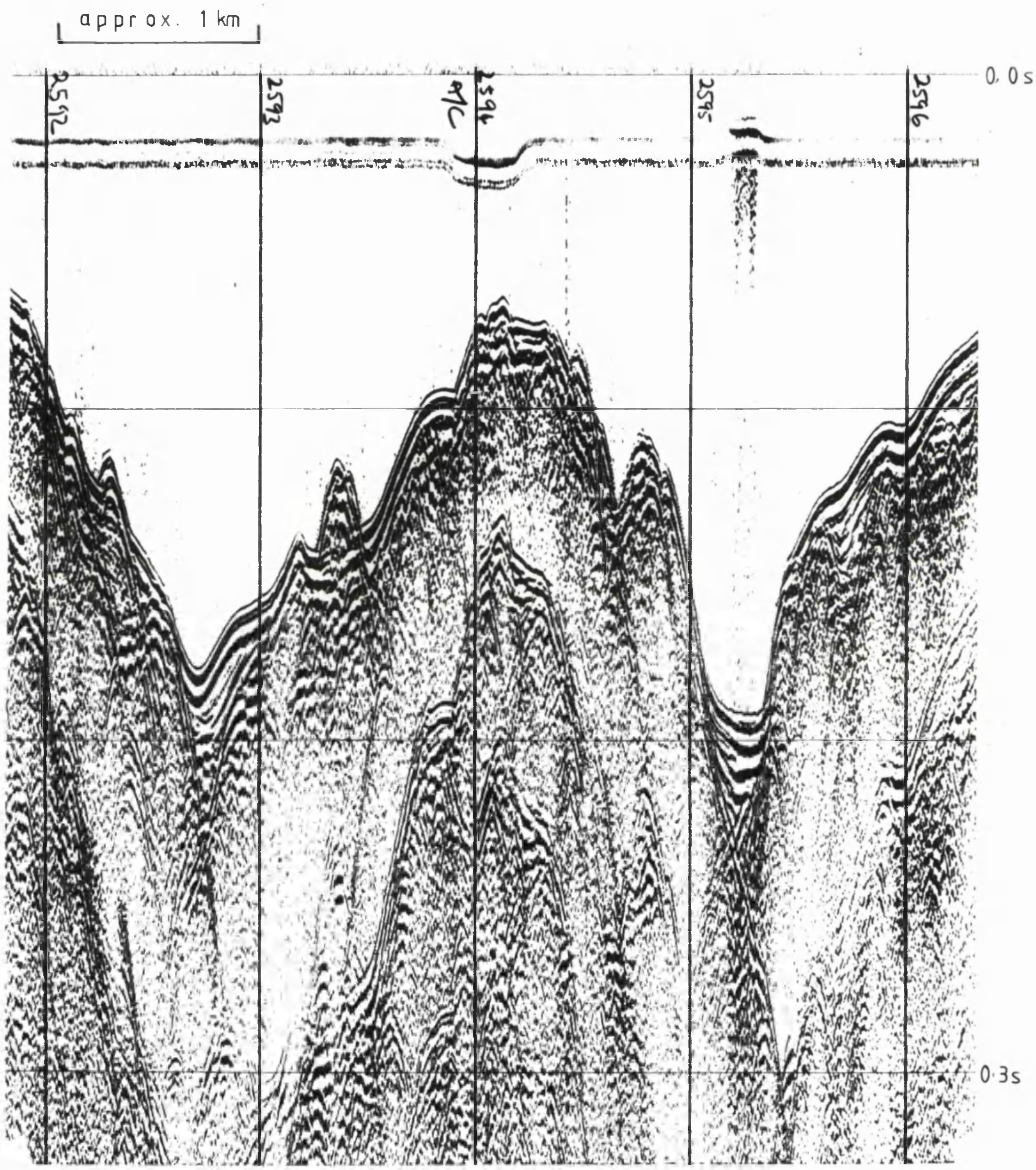


Fig 3.12 Sparker profile of Dalradian rocks

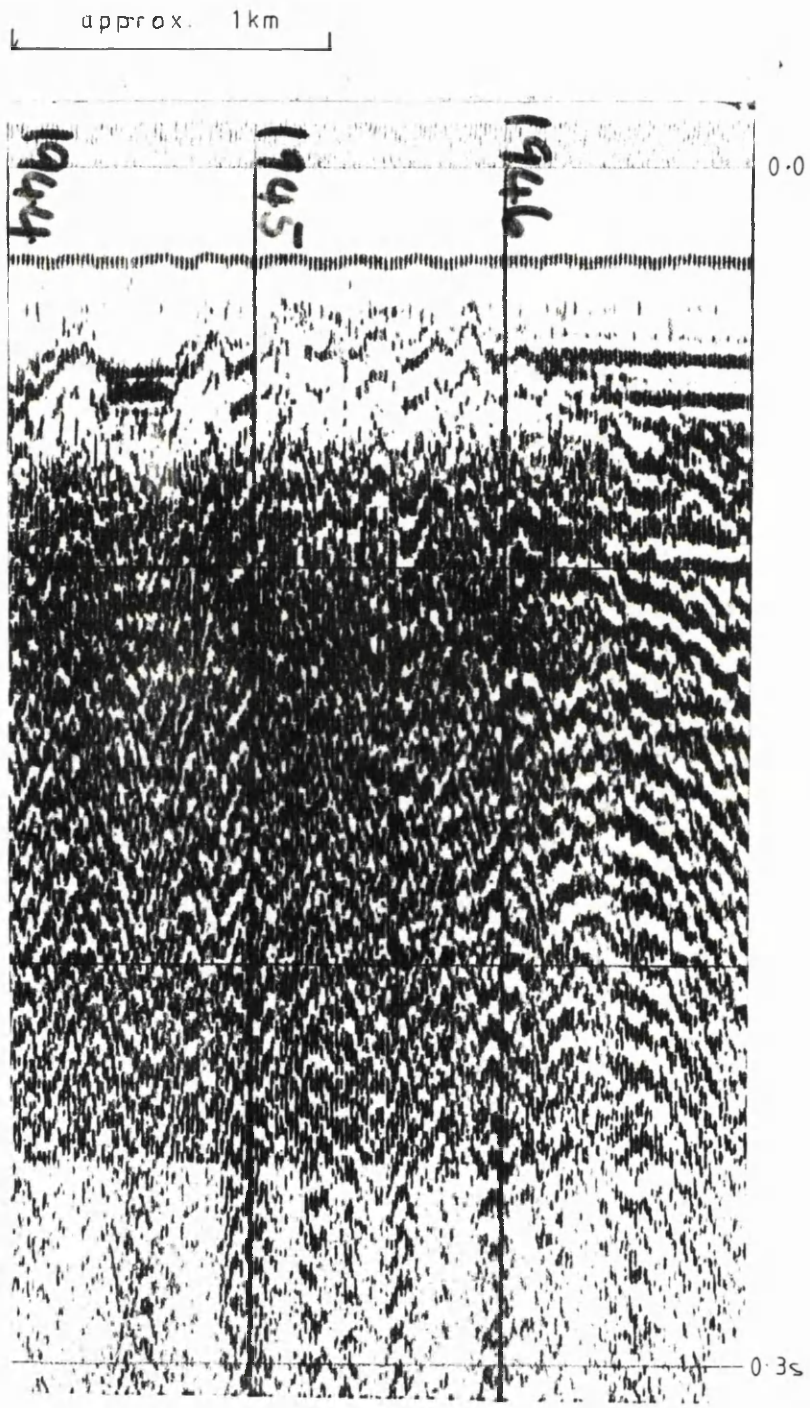


Fig 3.13 Sparker profile of Lewisian rocks

The Moine, Granite and Lewisian basements can be grouped as their surfaces are irregular with many hyperbolic reflections. The Lewisian can be separated by its associated magnetic character. The irregularities on the Moine surface occur with greater frequency than those of the Granite surface. The average depth ($\approx$50 m) of the Moine surface is less than that of the Granite surface ($\approx$75 m).

3.2.2 Sonobuoy data

The data collected in 1977 was of good quality, i.e. the onsets of first arrivals were clear, and an acoustic range of 7 km was achieved. In general analogue playbacks of these data were of limited use though they illustrated well that the dominant frequency of the refracted arrivals lies well below 40 Hz. Band passes wider than this however must be used for shipboard recordings in order to define the direct arrival (used to determine range) which has a dominant frequency about 150 Hz.

The data collected in 1979 is of inferior quality: considerable background radio noise, apparently not originating on the ship, dogged the experiments. The onsets of first arrivals were not clear and an acoustic range of 4 km max (generally less than 3 km) was achieved. Analogue playbacks of these data were not usable due to reference signal recording level problems.

Good results in terms of acoustic range and record resolution were obtained using a 40 cu.in. airgun firing every 4s, with the graphic recorder stylus sweeping the paper every 4s. Maximum acoustic range was attained using the largest gun available (150 cu.in.) but the long compression time gave a minimum fire period of 16s, the recording stylus sweeping the record twice in the 16s and printing only for the first 8s. This resulted in a compressed recording on which it is difficult to measure the arrival times accurately (see Table 3.3, Figs. 3.14 - 3.31).

Acoustic source	Fire rate	Sweep	Sweeps/Shot	Acoustic range (km)
1 kJ sp	2s	2s	1	-
5 kJ sp	4s	4s	1	1
40 cu.in airgun	4s	4s	1	6
150cu.in airgun	16s	8s	2	8

Table 3.3 Comparison of acoustic sources.

The Sparker sources were unsuccessful in producing any refracted arrivals.

Preparation of Time-Distance graphs from sonobuoy records

Using a template the onset times of the direct and refracted arrivals is measured (± 1 ms for 40 cu.in. records) at 1/5 fix intervals. The shot range from the buoy is obtained from the direct (water wave) arrival time (t) (Fig. 3.32). Water depth is obtained from the echo sounder record running at the time of buoy deployment accurate to ± 1 m thus range can be determined to within 1%.

The times for the refracted arrivals are plotted at the determined ranges. The bathymetric profile, from the echo sounder record obtained at the same time as the refraction data, is plotted beneath the time-distance graph. Where sparker/magnetometer lines ran close to the refraction lines the magnetic profile and sediment thickness are also plotted.

In spite of attempts to run lines over level seabed, the seabed is irregular or sloping in most cases. A correction has been applied to project the shots onto the seabed (i.e. removing the travel time through the water). The ray path is

FIGURES 3.14 - 3.31

Sono-buoy records

Timing lines at 10 ms intervals

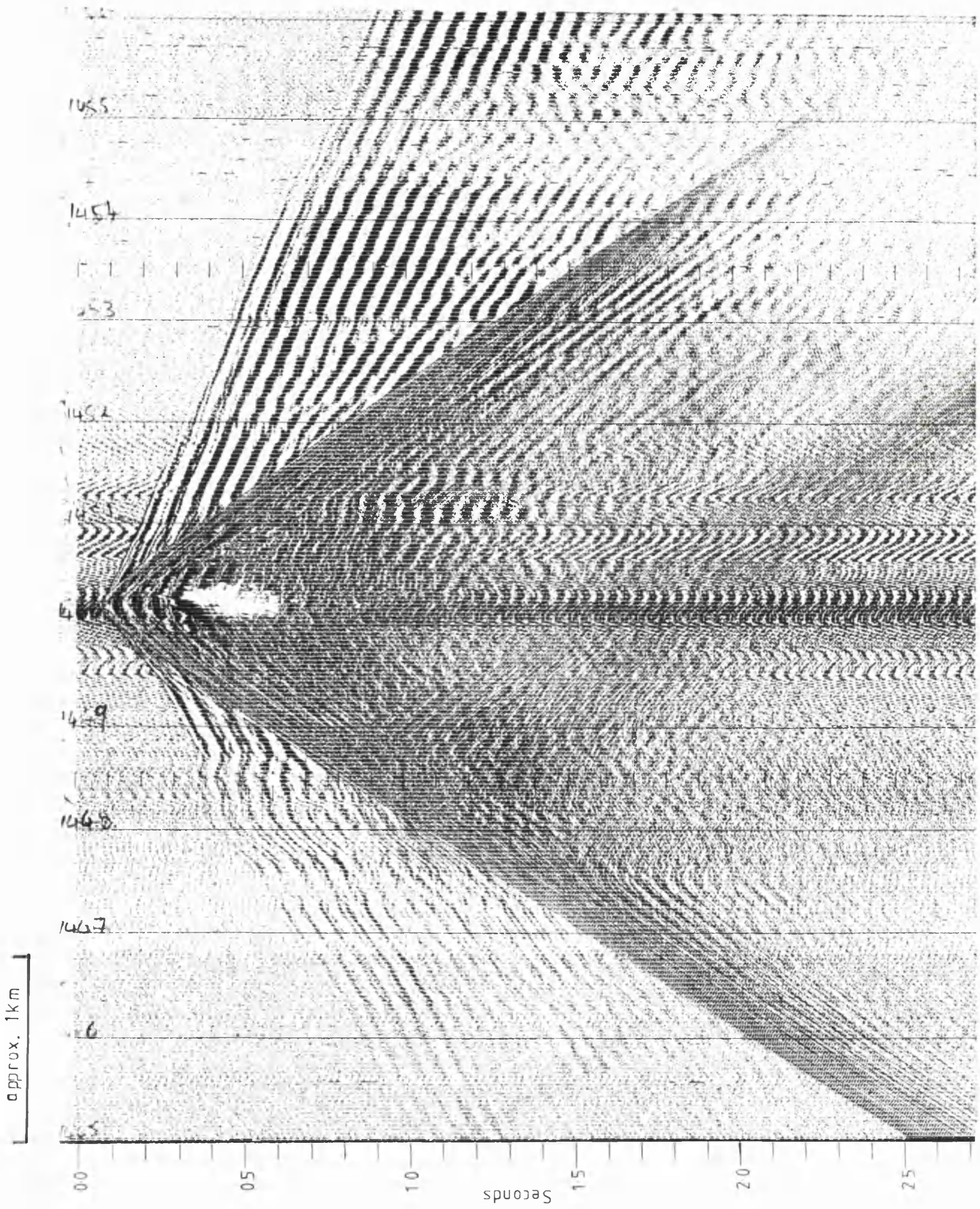


Fig 3.14 Line 2A/1977

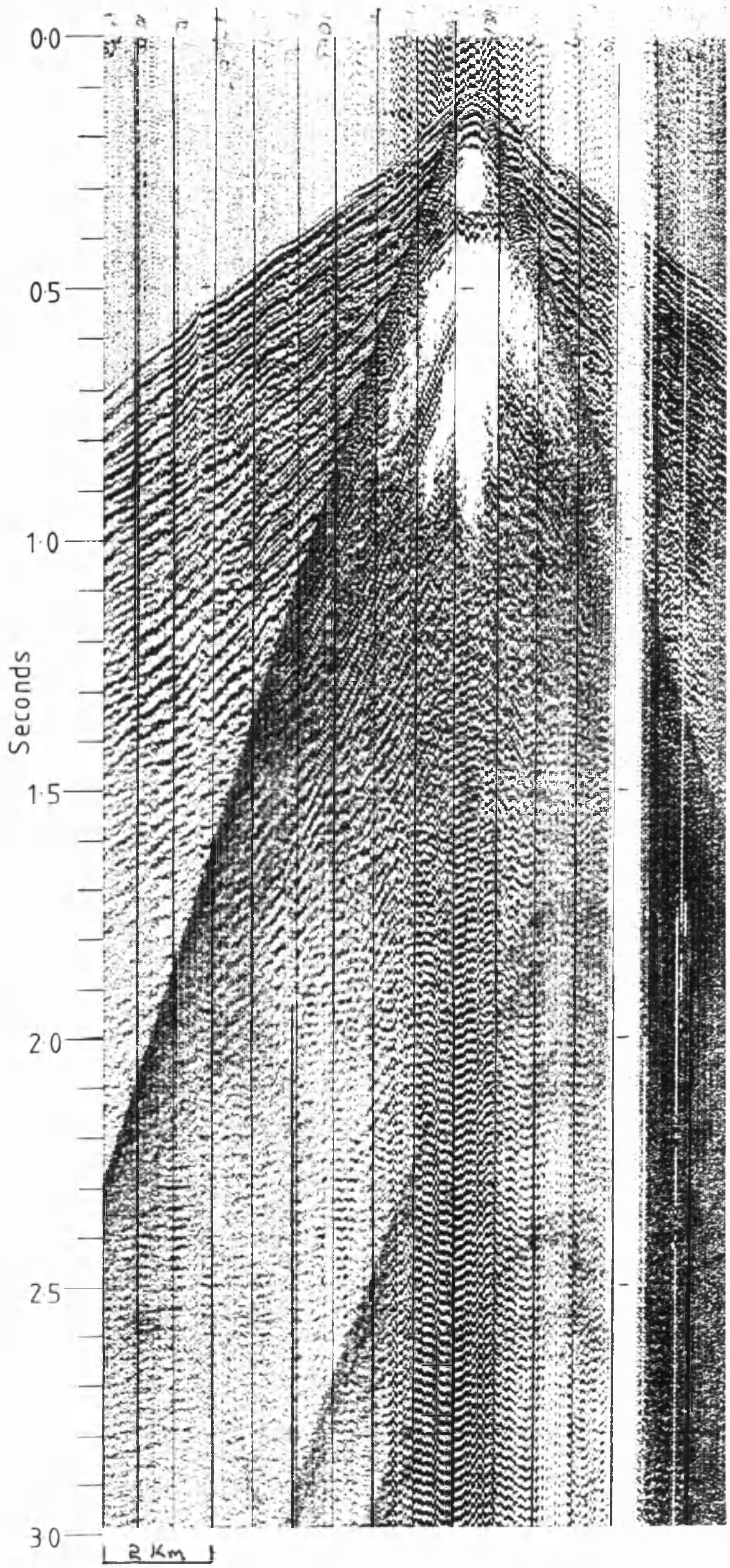


Fig 5.15 Line 5/1977

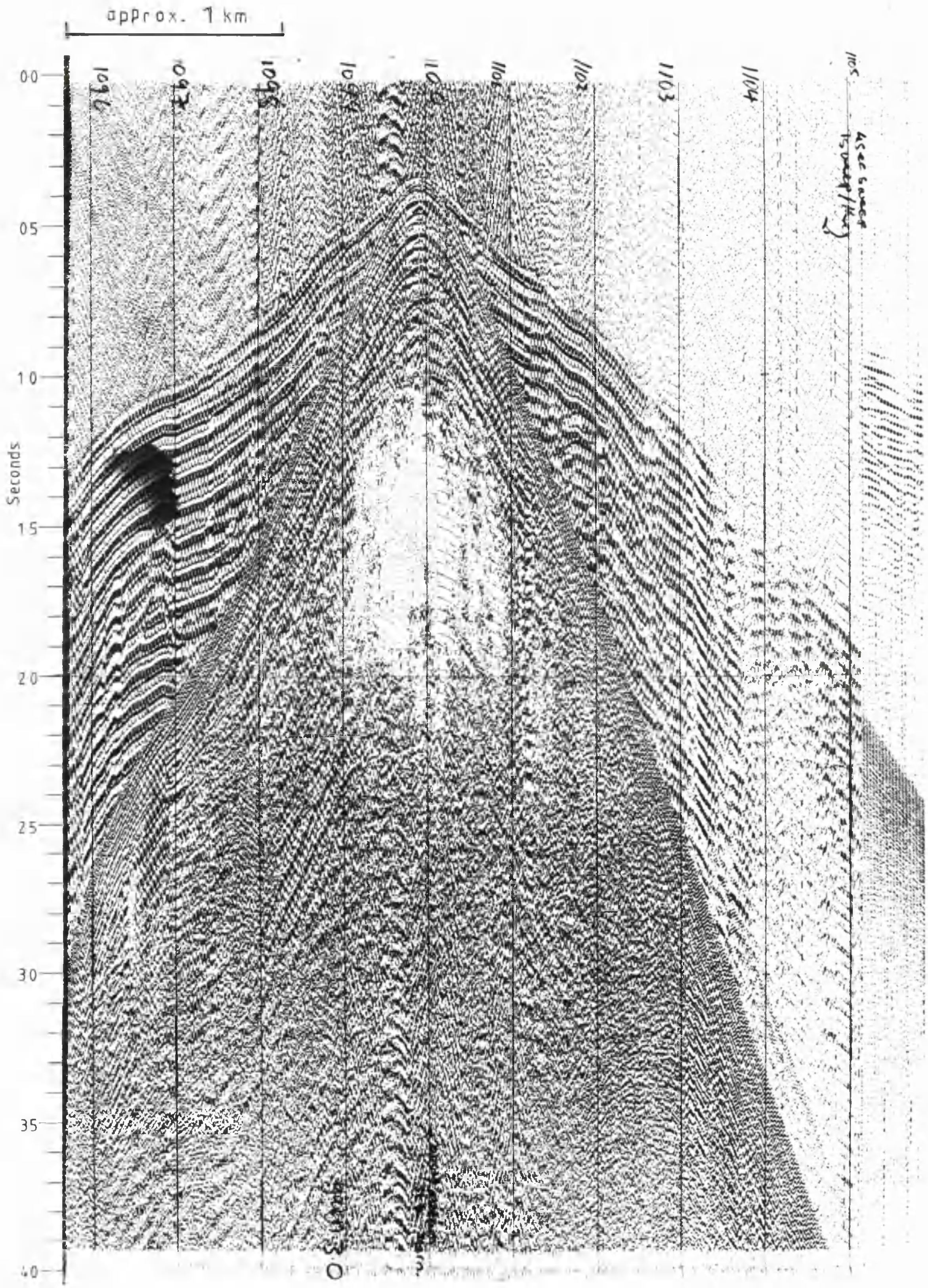


Fig 3.16 Line 5X/1977

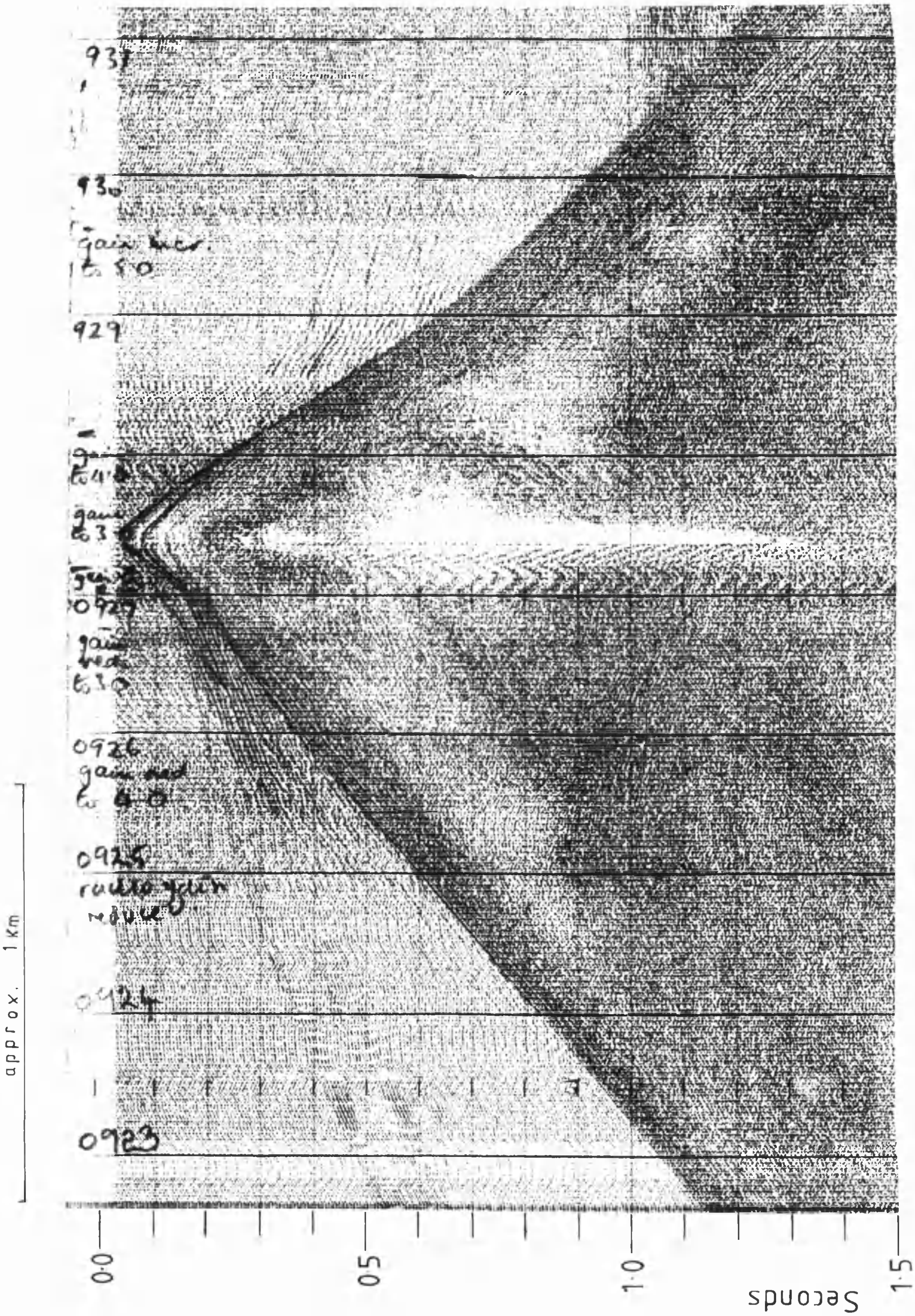


Fig 3.17 Line 6$_{sp}$/1977

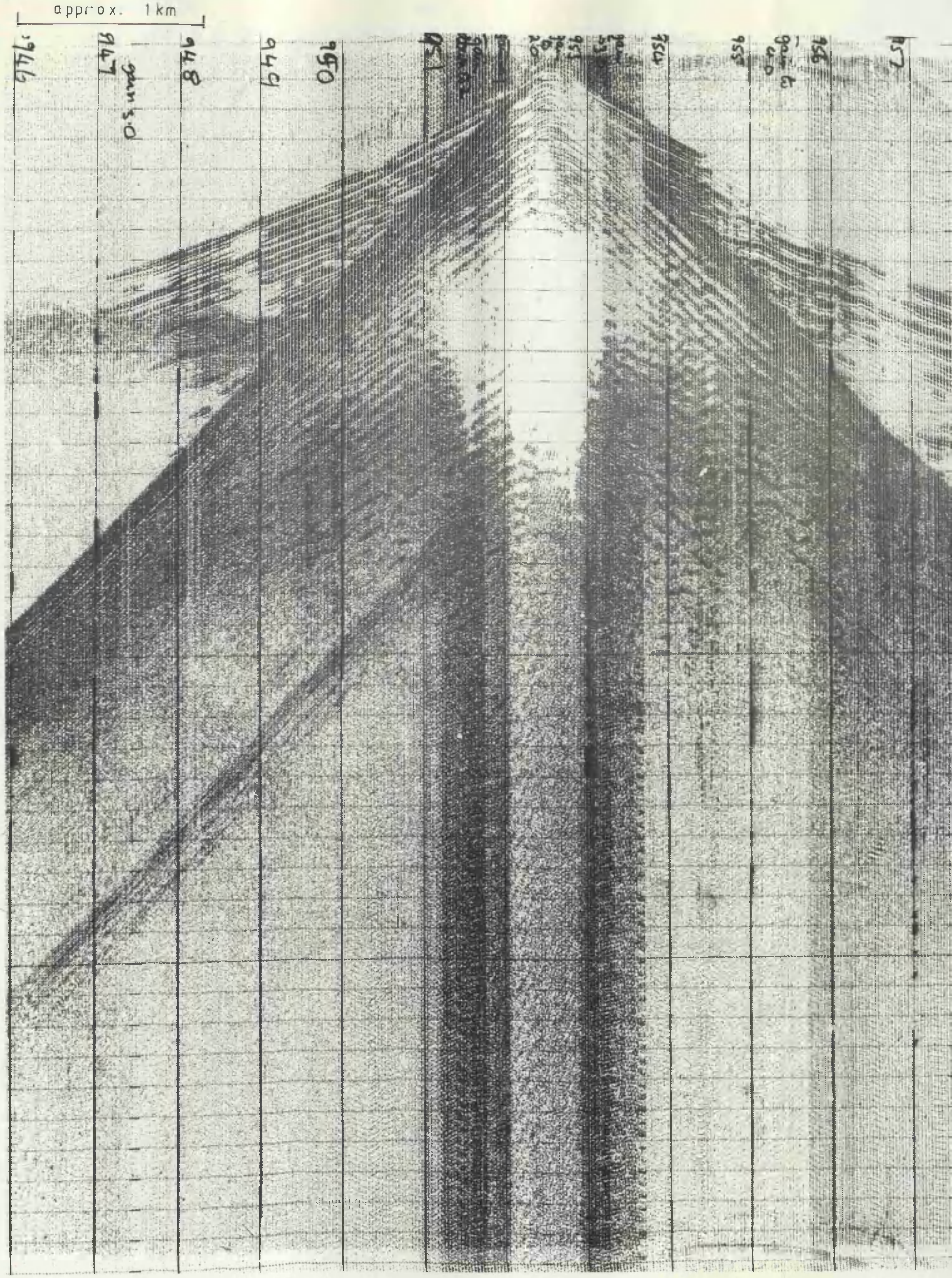


Fig 3.18 Line 6_{40}/1977

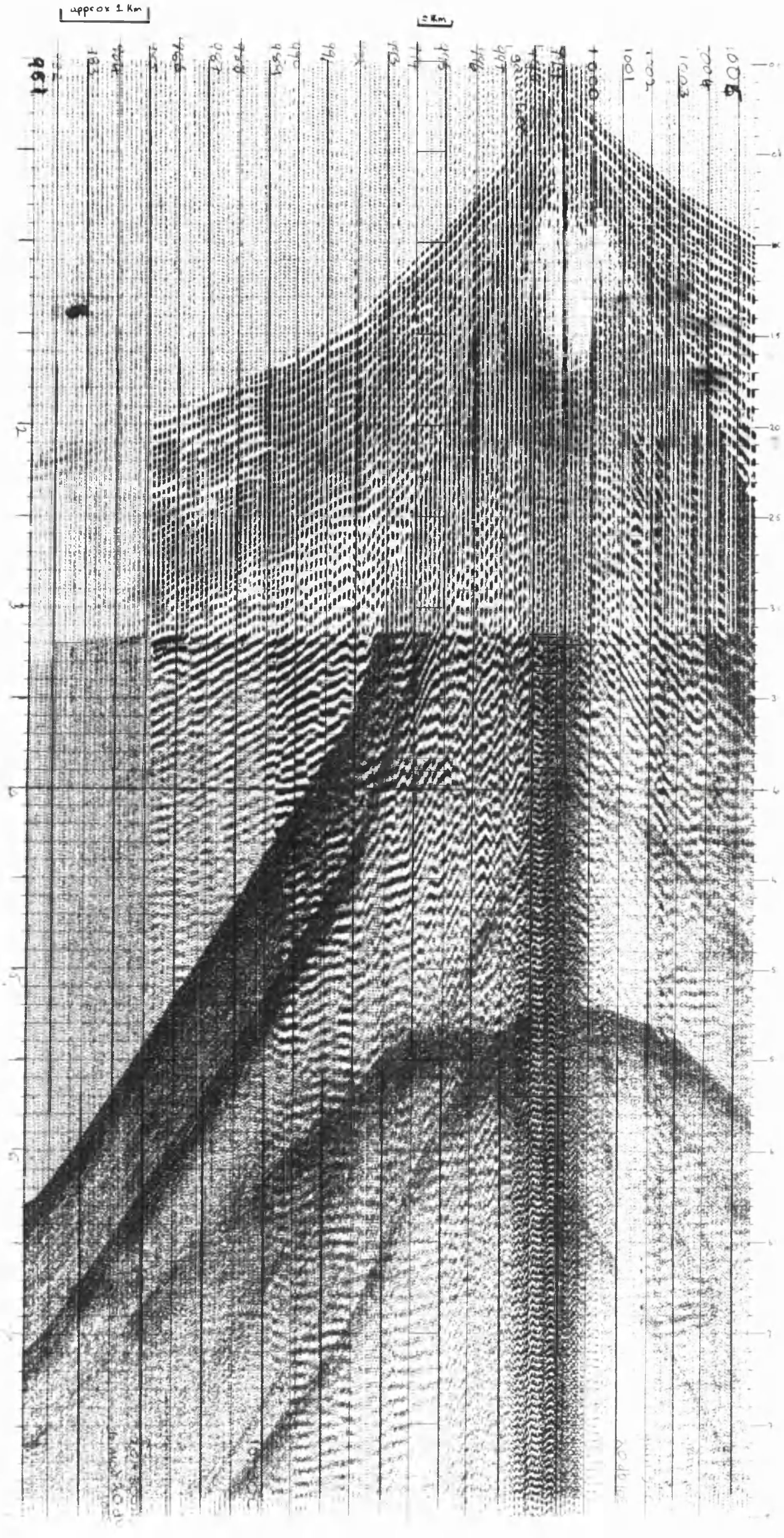


Fig 3.19 Line 6_{150}/1977

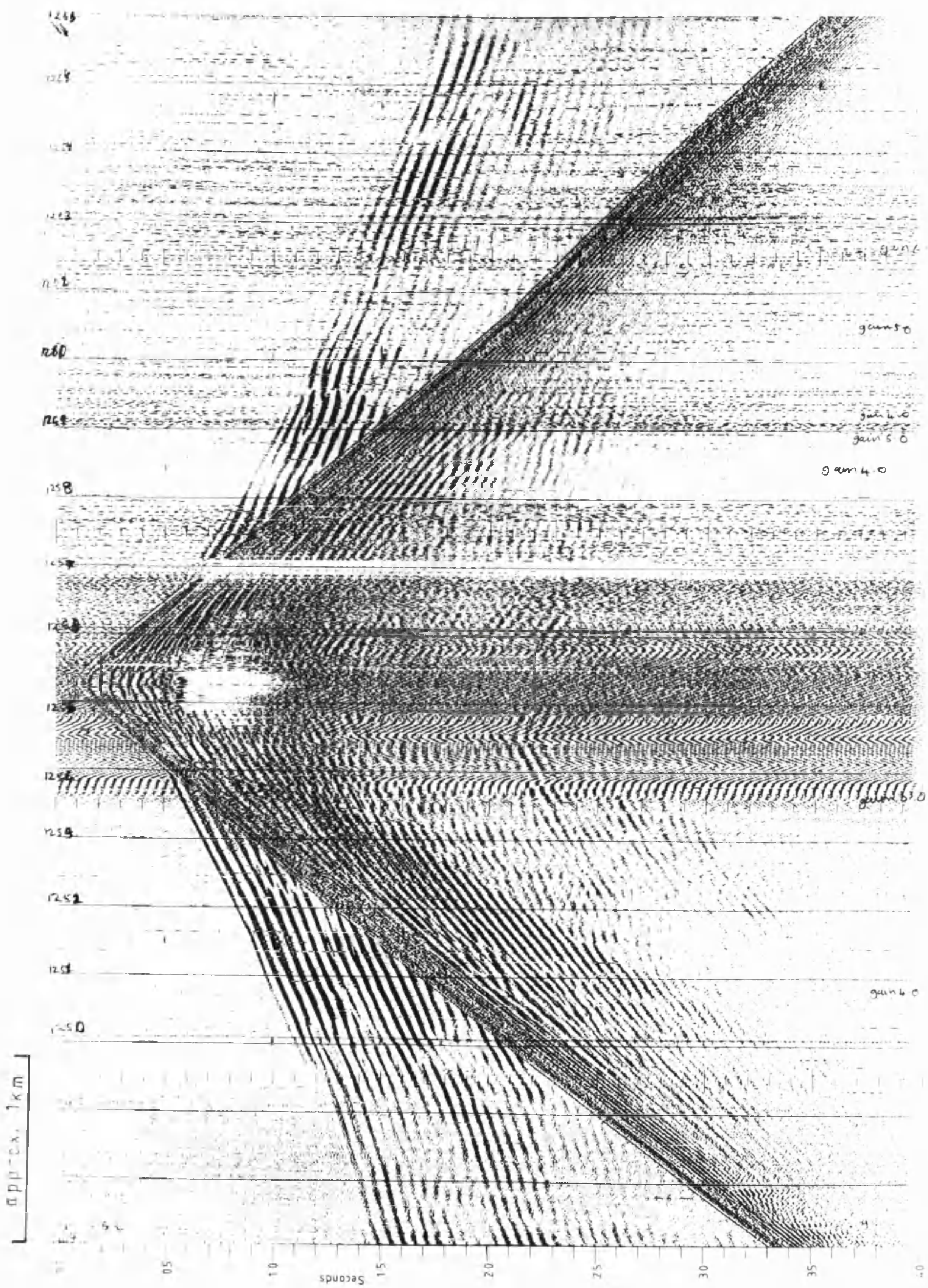


Fig 3.20 Line 9/1977

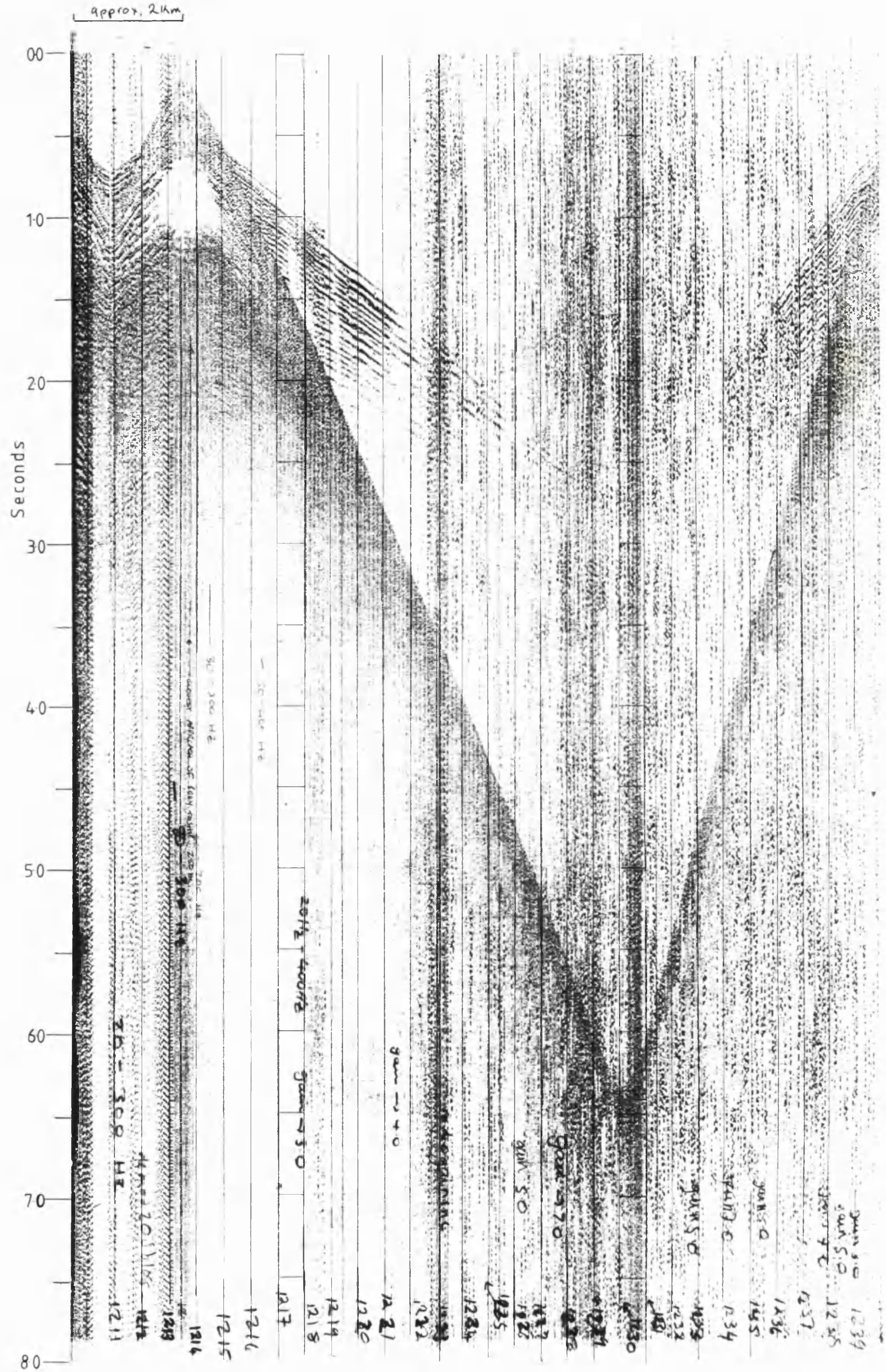


Fig 3.21 Line 9_{150}/1977

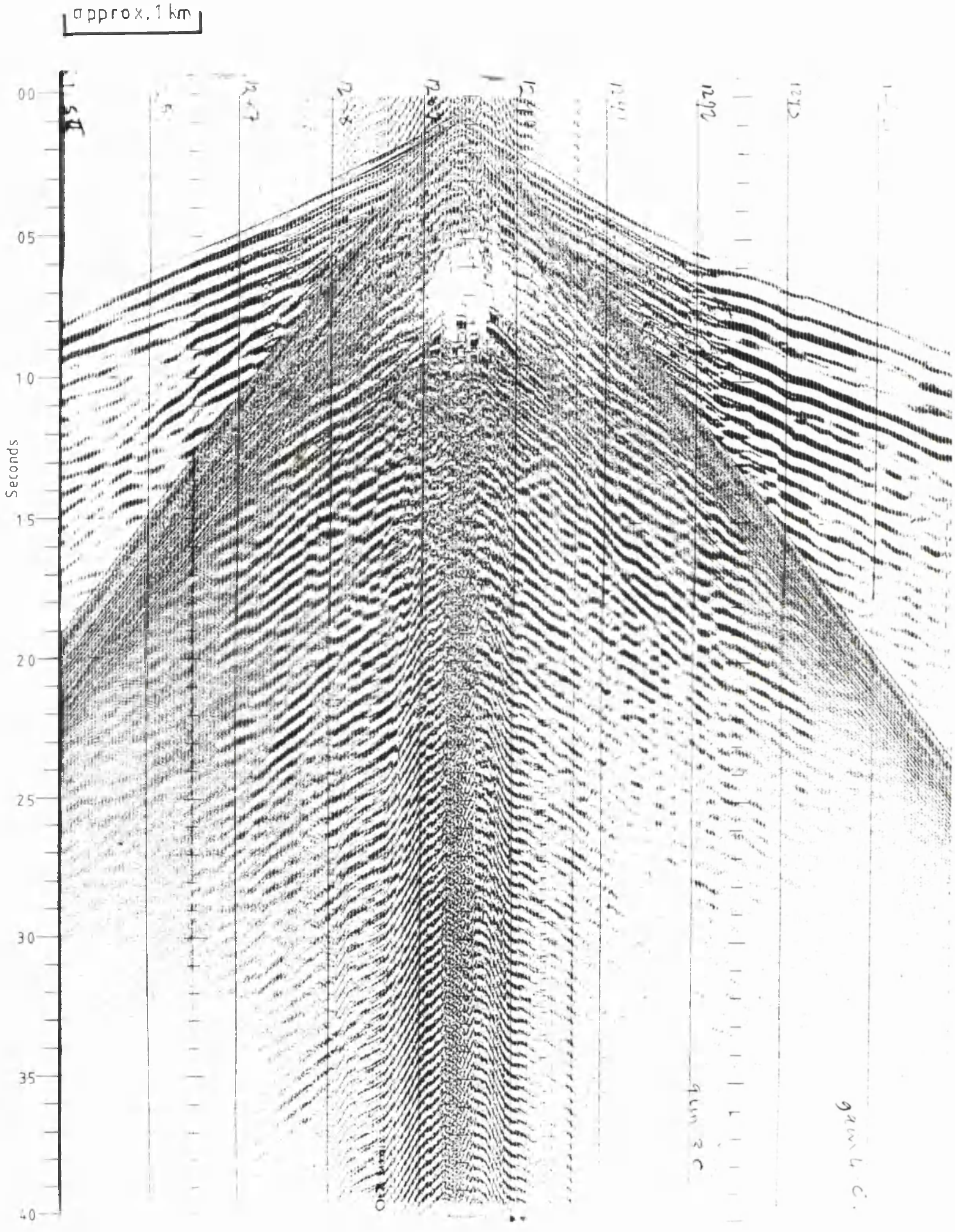


Fig 3.22 Line 10/1977

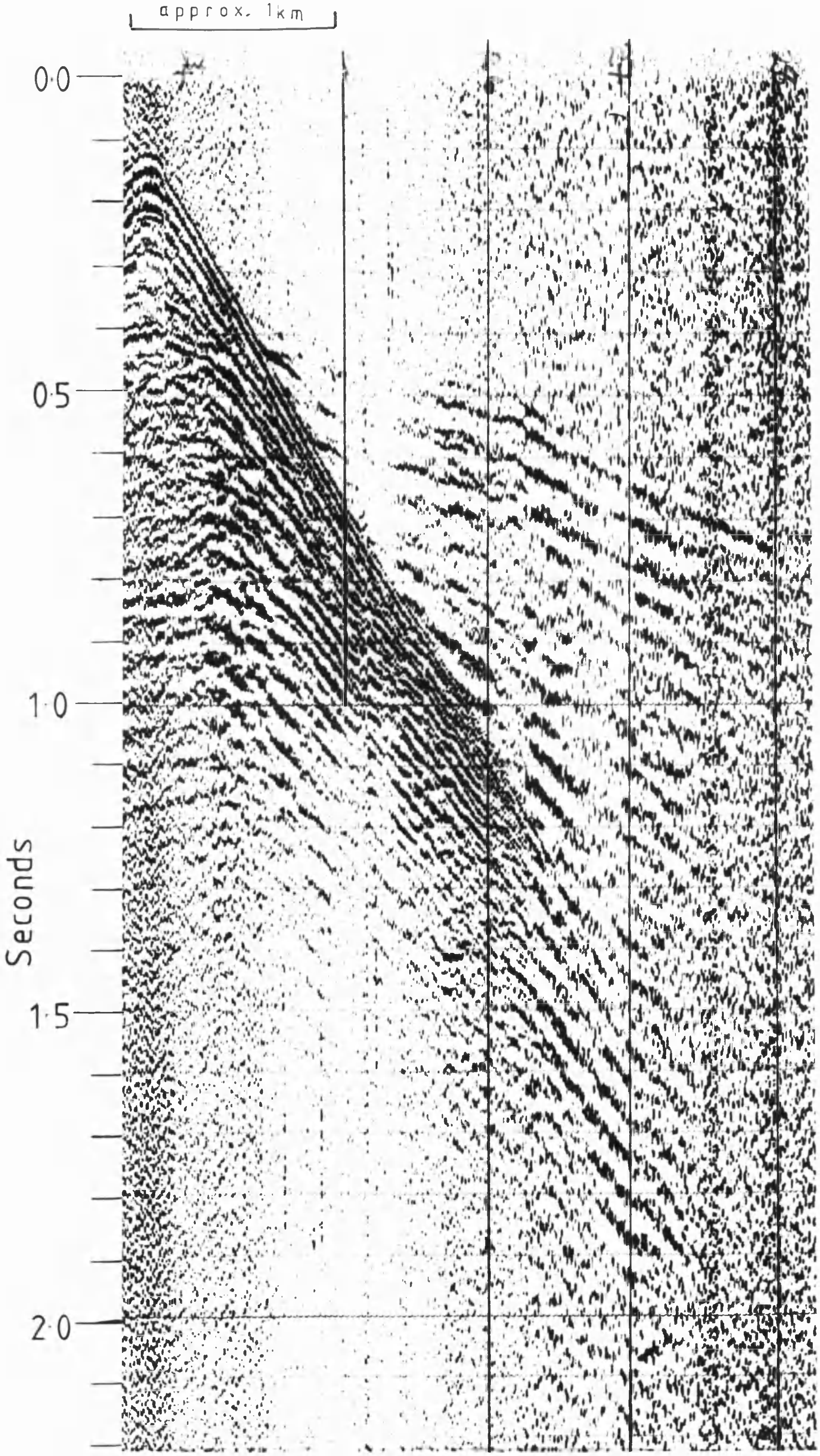


Fig 3.23 Line 1/1979

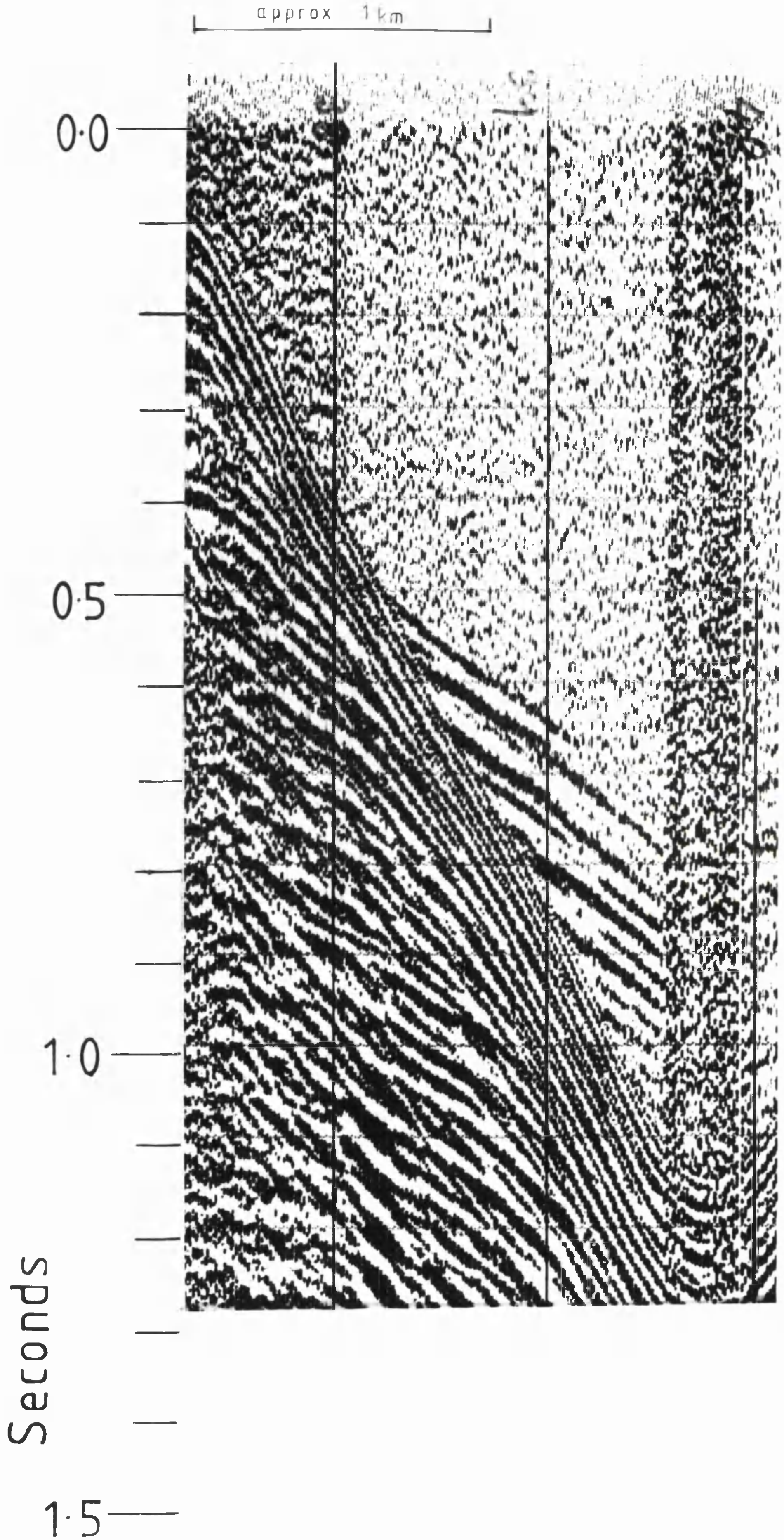


Fig 3.24 Line 1/1979

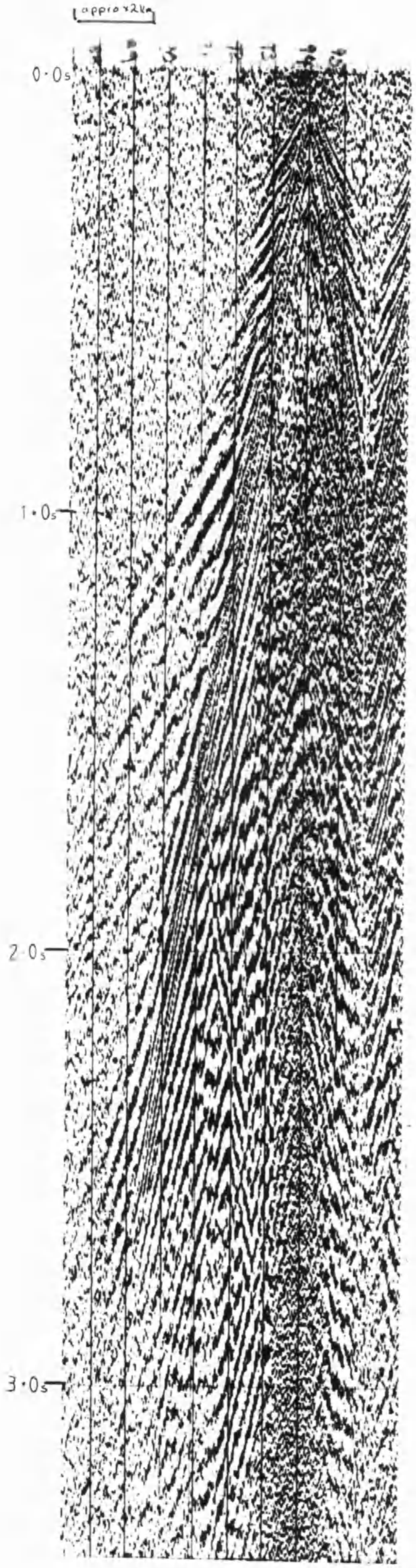


Fig 3.25 Line 2/1979

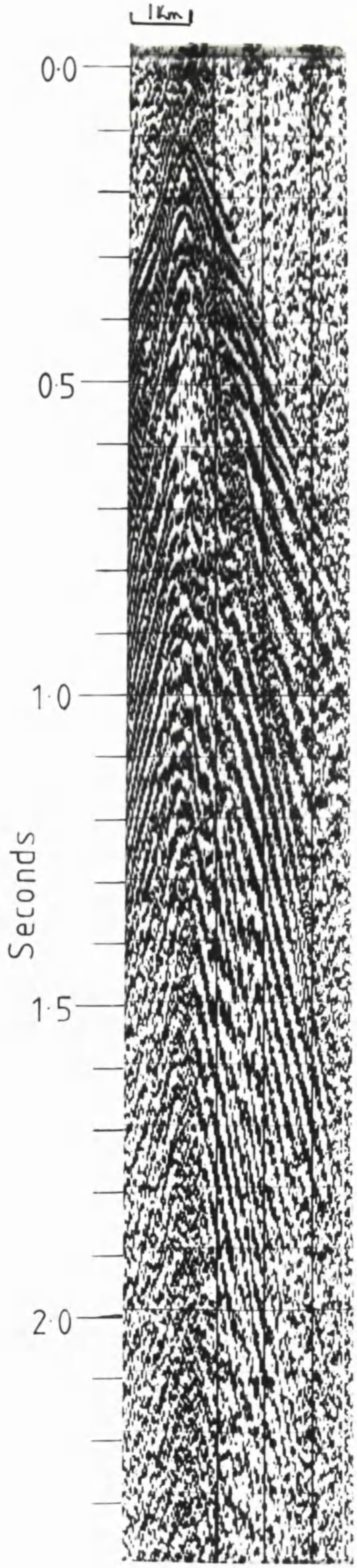


Fig 3.26 Line 3/1979

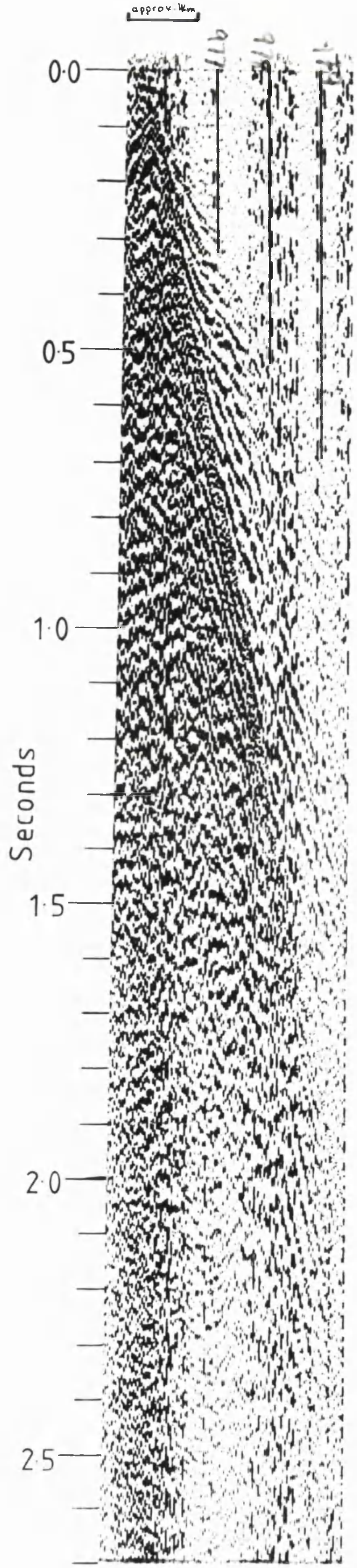


Fig 3.27 Line 6/1979

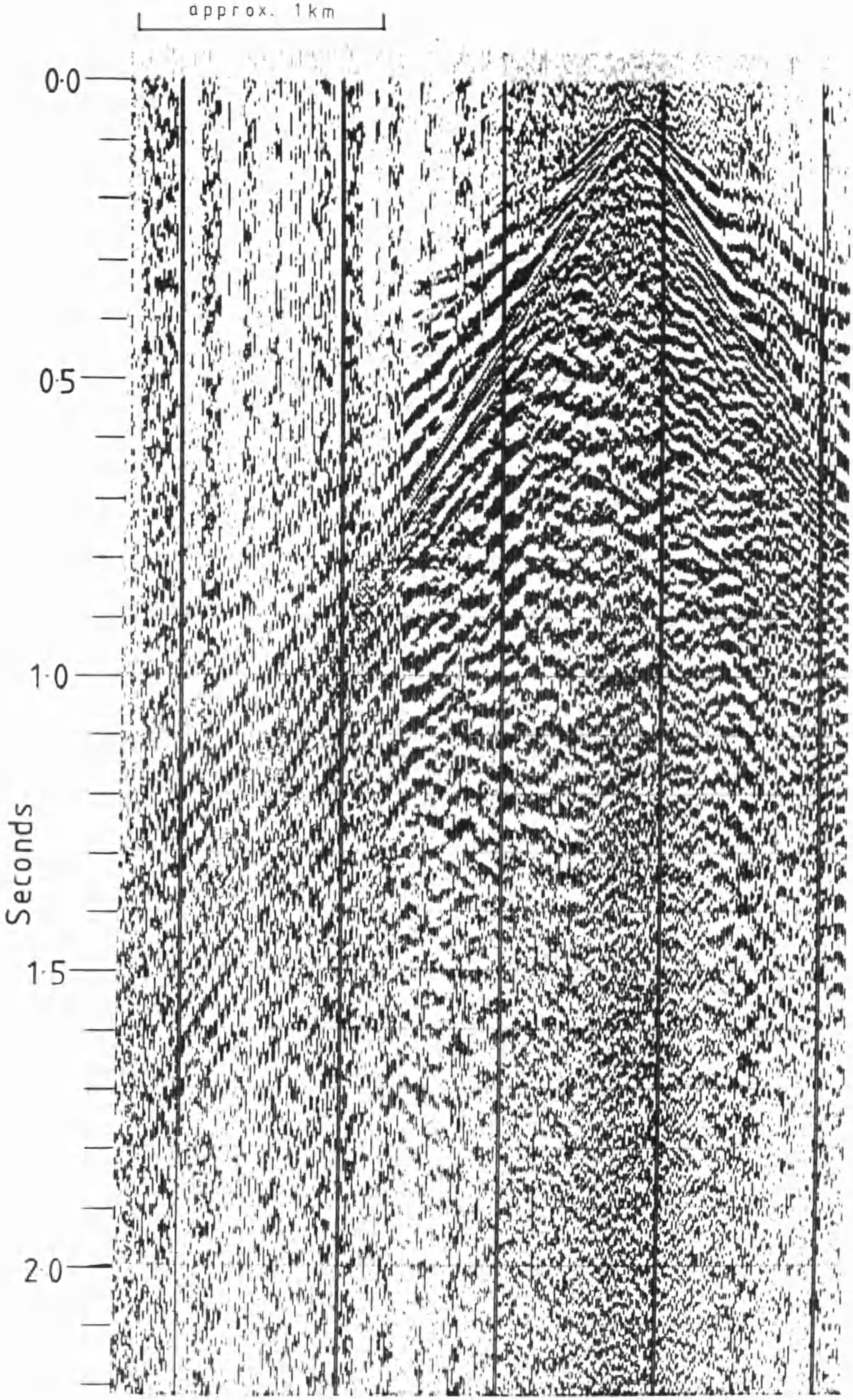


Fig 3.28 Line 7/1979

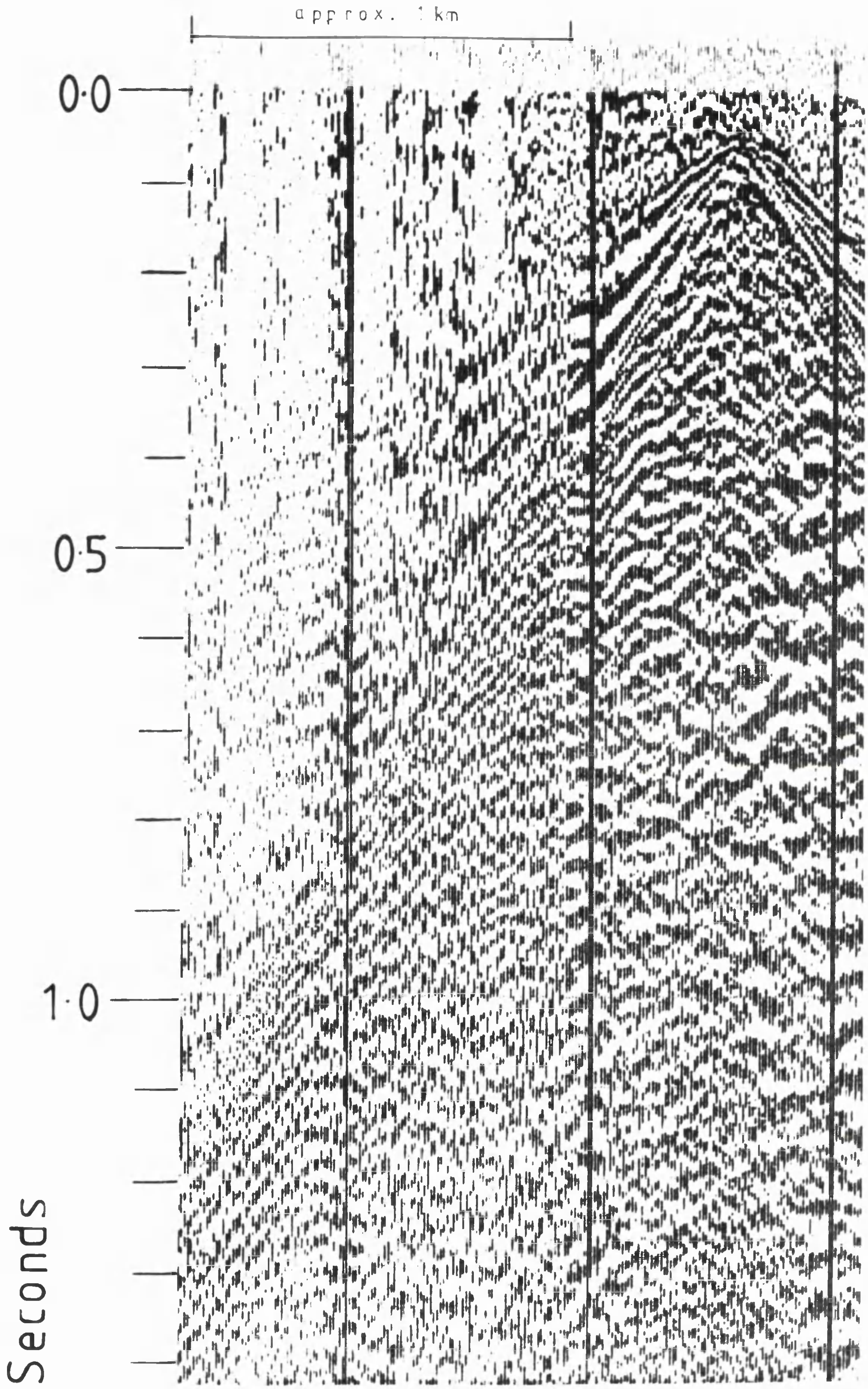


Fig 3.29 Line 7/1979

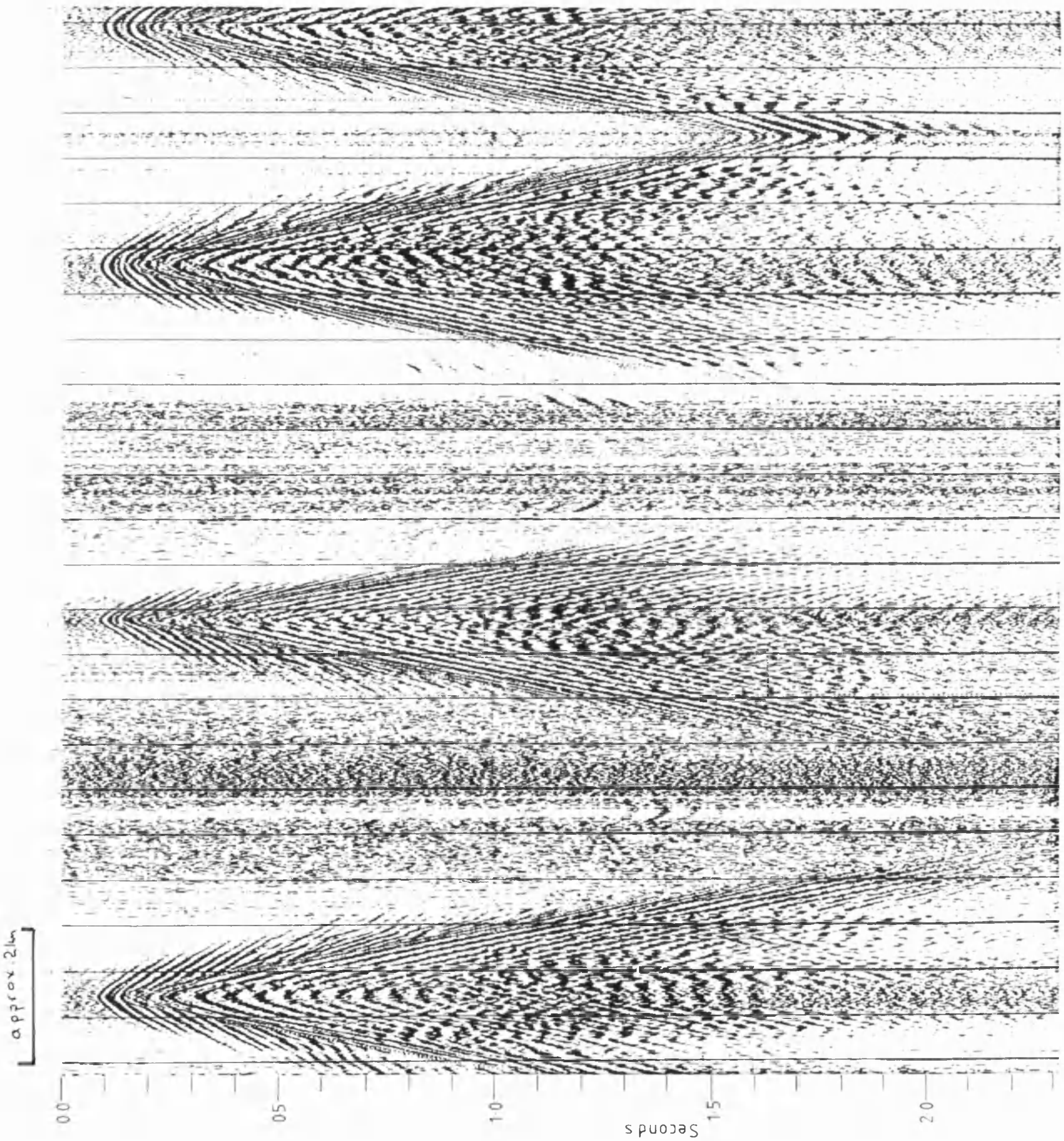


Fig 3.50 Line 8/1979

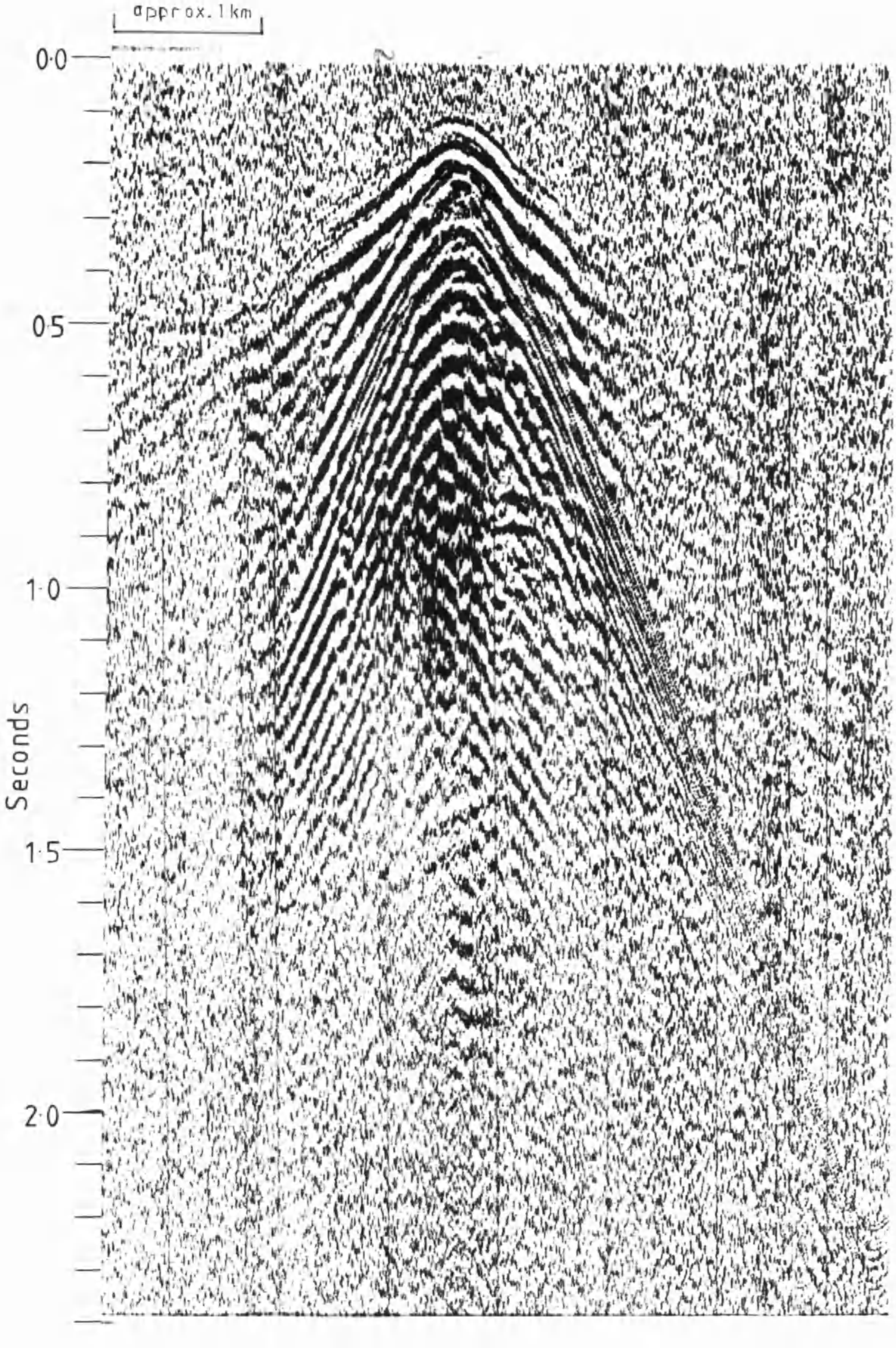


Fig 3.31 Line 9/1979

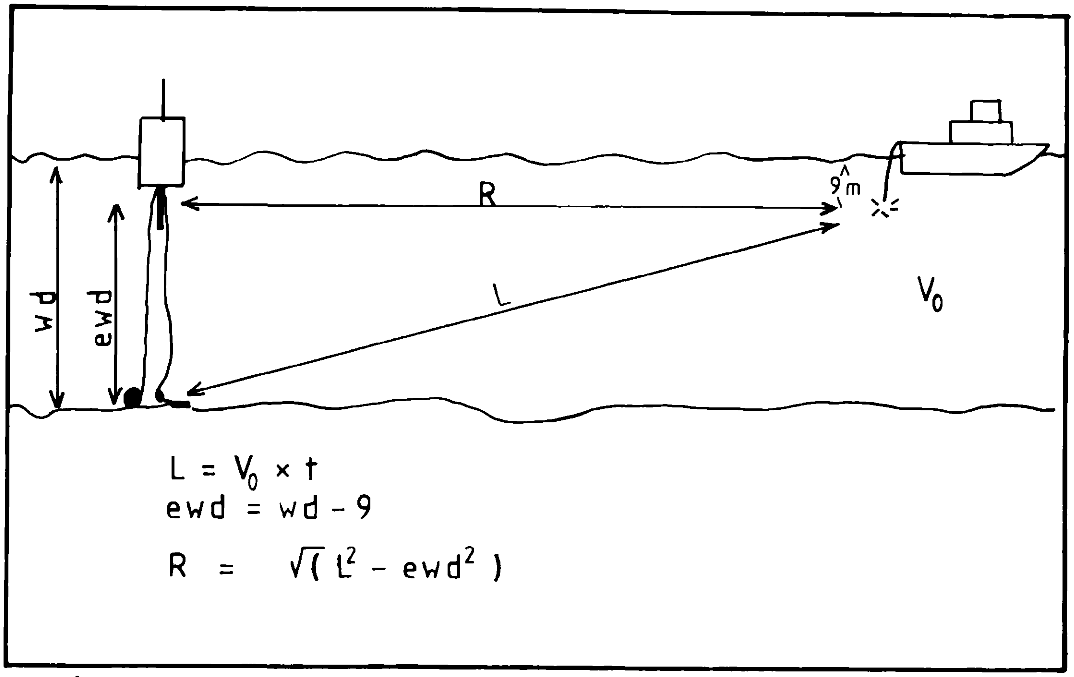


Fig 3·32 Range calculation

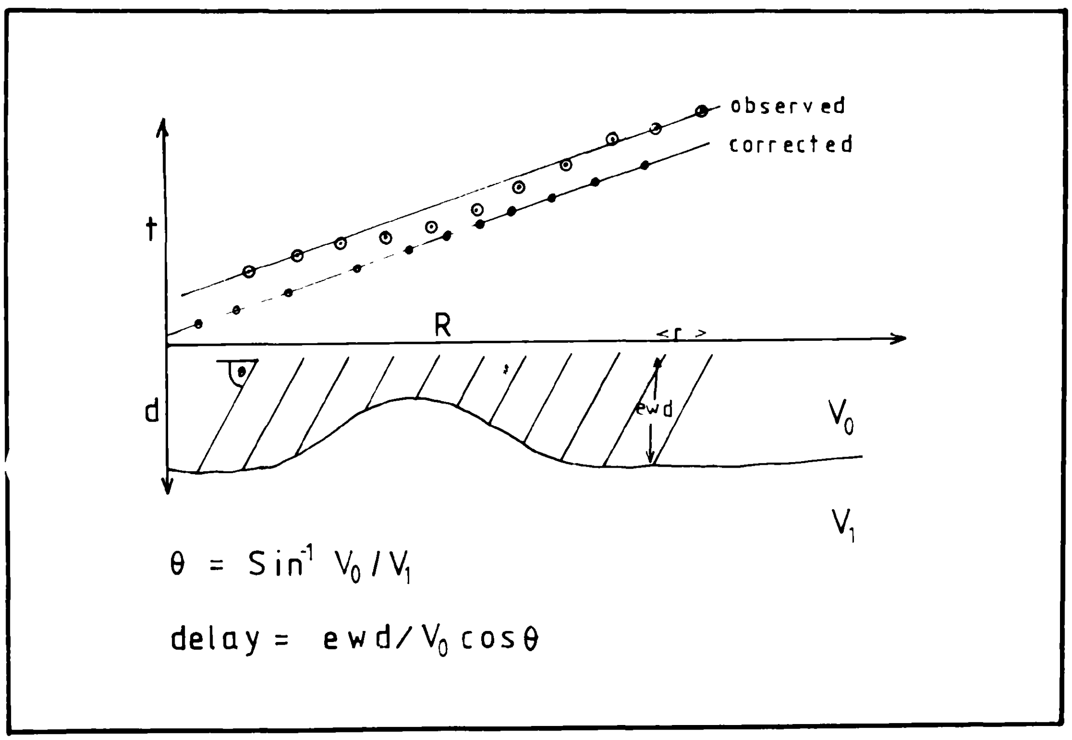


Fig 3·33 Water layer correction

drawn on the section (Fig. 3.33) and the effective water depth is measured at the point where the ray hits the seabed. The delay = EWD/($\cos\theta_c$ x Vo)

where EWD - effective water depth

θ_c - critical angle ($\sin^{-1}$ Vo/Vapp)

Vo - velocity of sound in water

Vapp - velocity of refracted arrival (found from a by eye fit through the points)

The lateral shift towards the buoy is = EWD x $\tan\theta$.

The corrected ranges and arrival times are used to compute the apparent velocities by a least squares fit to a straight line (using a Hewlett Packard programme supplied by I.G.S.).

It was only possible to determine sediment thickness along the length of Line 5 / 1977. For this line only a similar correction was applied to remove the delay time due to travel through the sediment layer.

The time-distance graphs obtained are given (Figs. 3.34 - 3.50).

Composite T - X graphs were drawn for Lines 2, 3 & 8 / 79 (Fig. 3.51) and Lines 6 & 7 / 79 (Fig. 3.52).

Estimates of layer thicknesses have been made for the 1977 data where the onset of first arrivals could be picked with reasonable confidence. However, for the 1979 data the quality was such that it was not possible to pick the onset of the refracted arrivals, thus preventing estimates of layer thicknesses to be made.

These results show that the method employed on these cruises produces better-defined arrivals which can be picked with confidence as compared to previous refraction work in the area (Faruquee 1972). Also the experiments are relatively quick and simple to carry out as it is only necessary to deploy

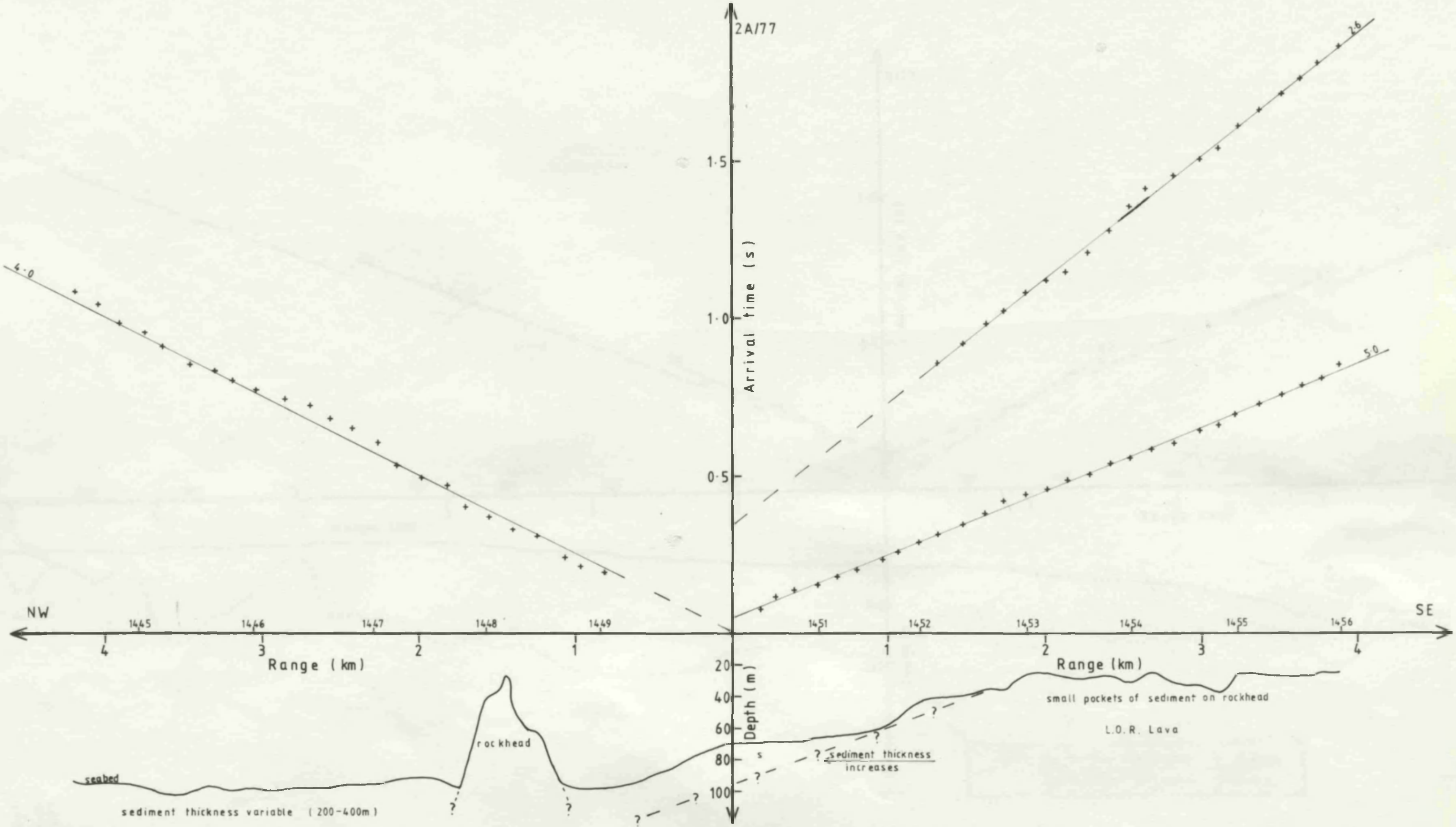


Fig. 3·34

Time-distance graph for sonobuoy line 2A/1977. Arrival times corrected for delay through the water layer. The 2.6Kms segment is not a plot of the onset times for this event; it is a plot of the first continuous pickable cycle. (see pp. 42,43)

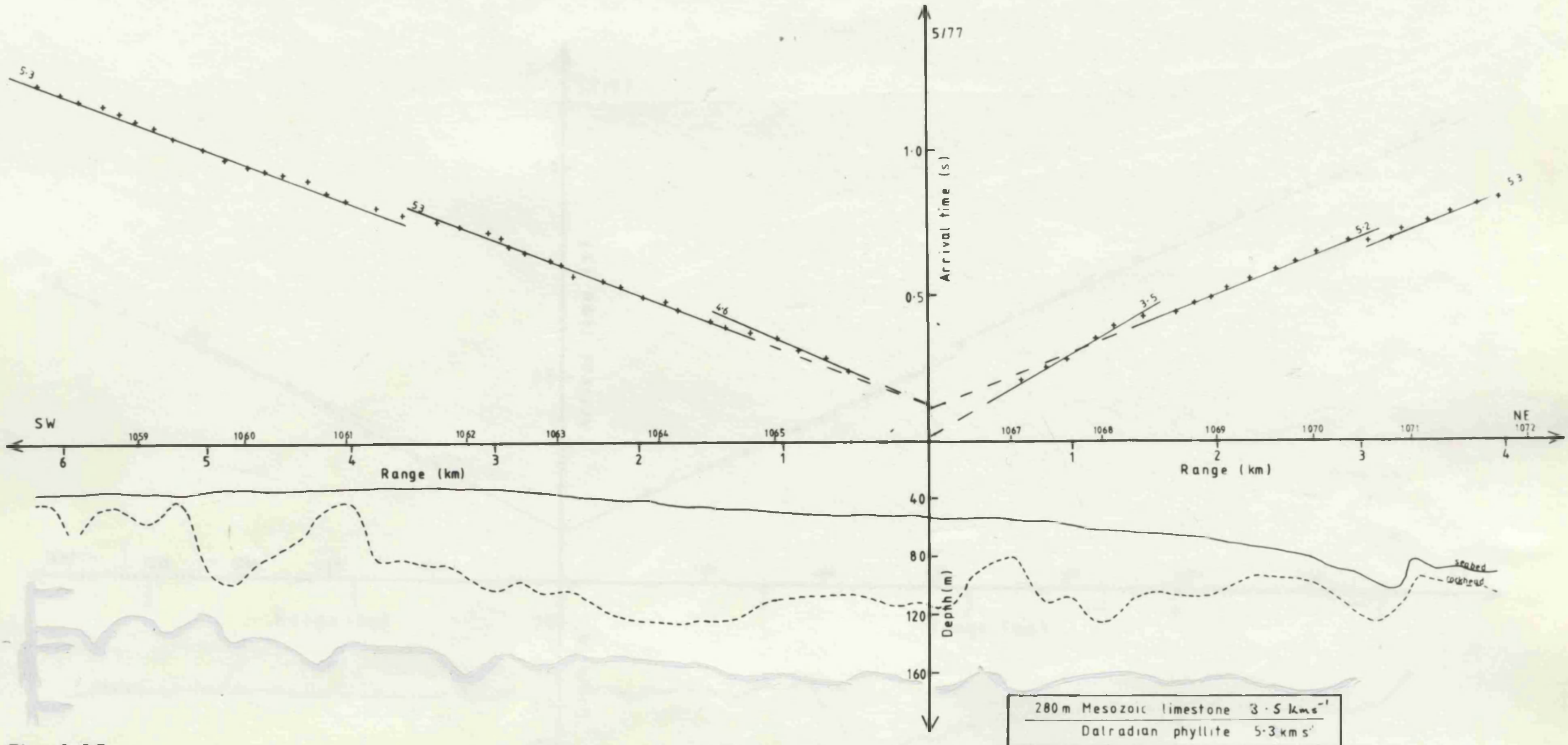


Fig 3·35

Time-distance graph for sonobuoy line 5/1977. Arrival times corrected for delay through the water and superficial sediment layers. Interpretation (in box) is for northeast section (see pp. 38-40; 46-48) Smooth magnetics.

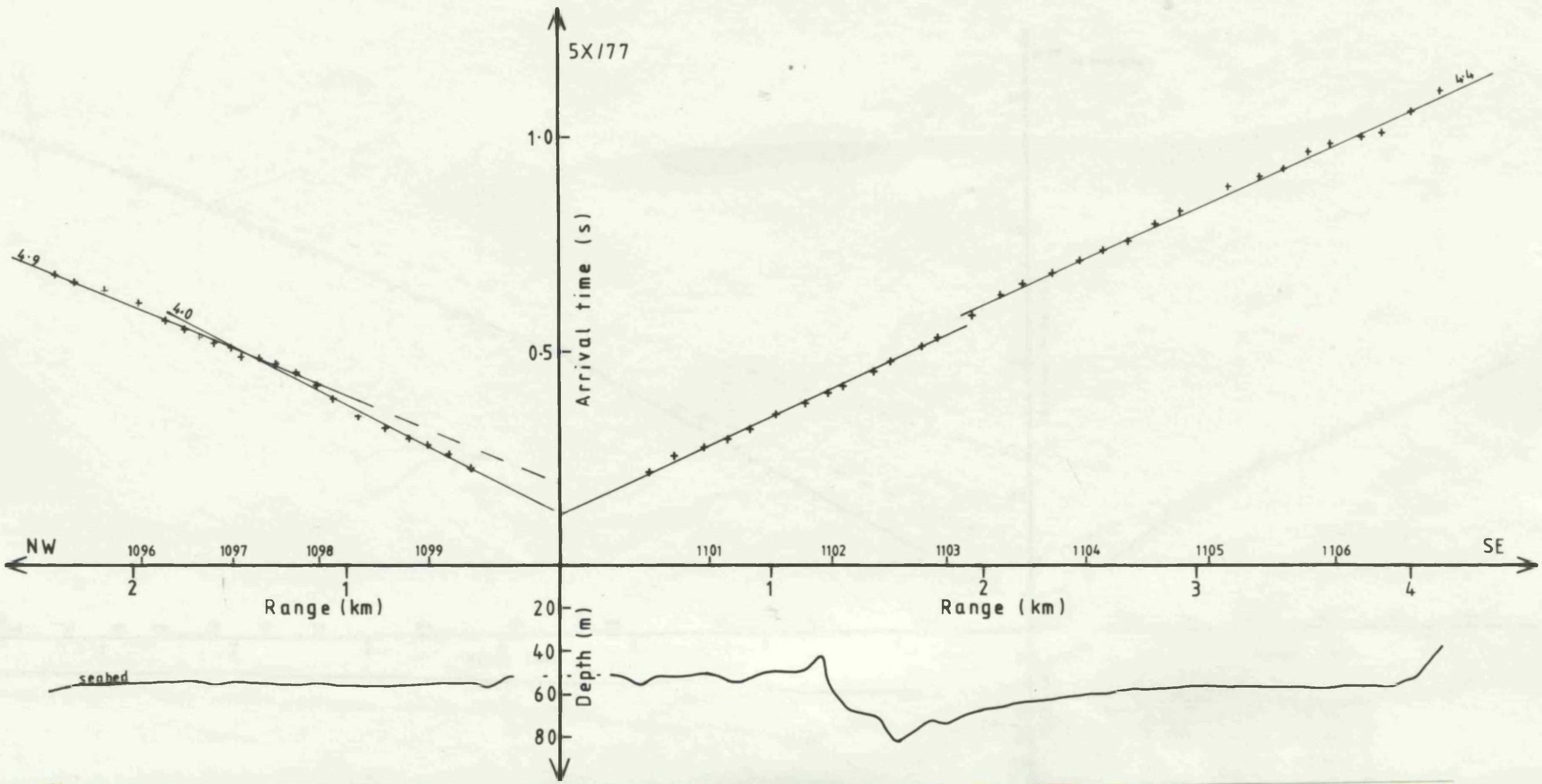


Fig 3.36 Time-distance graph for sonobuoy line 5X/1977. Arrival times corrected for delay through the water layer. (see pp. 47, 48) .

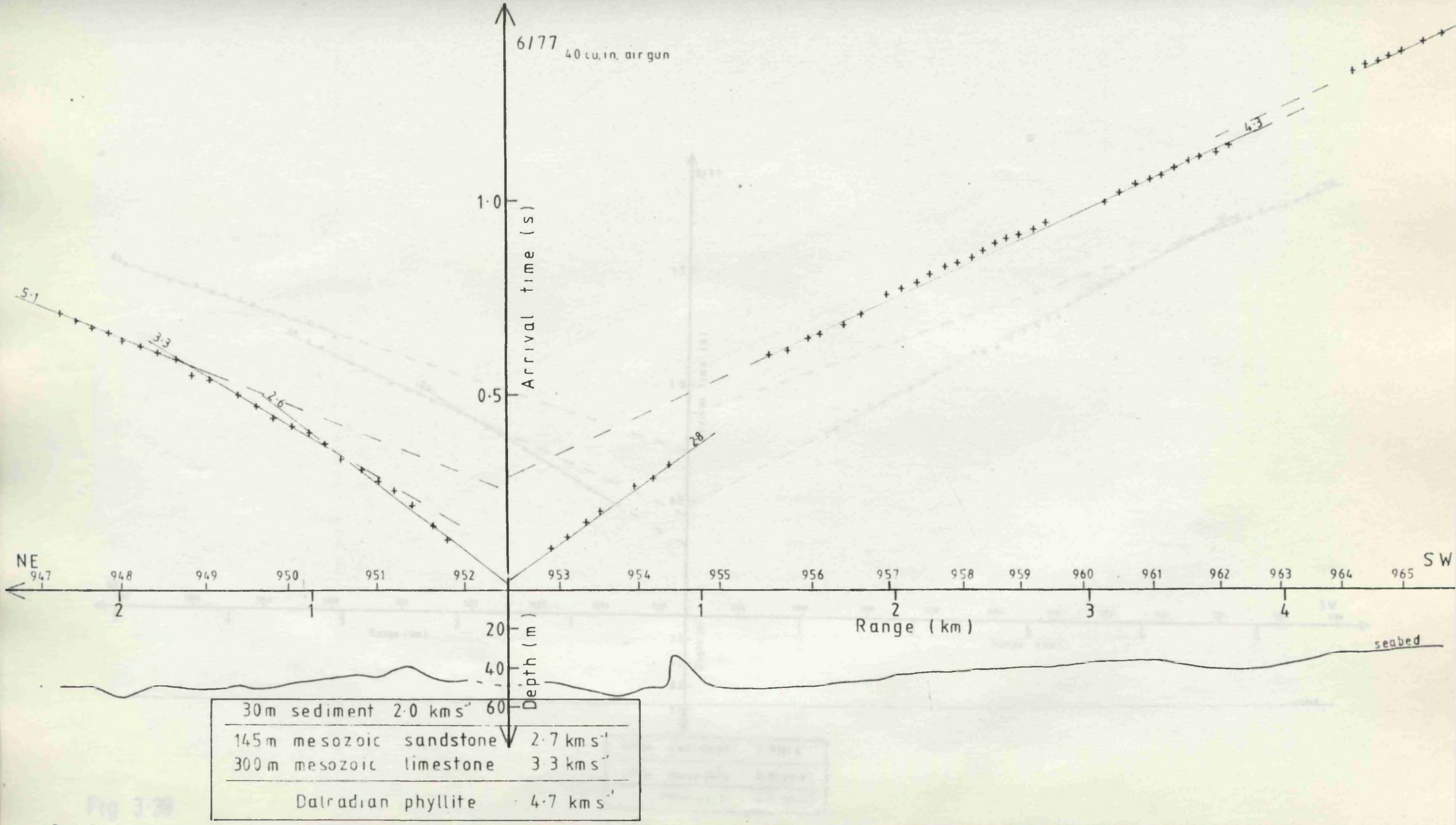


Fig 3·38 Time-distance graph for sonobuoy line 6/1977 obtained using a 40 cu.in. airgun. Arrival times corrected for delay through the water layer. Interpretation (in box) of thicknesses pertain to buoy position (see pp.38-40, 46-48). Smooth magnetics.

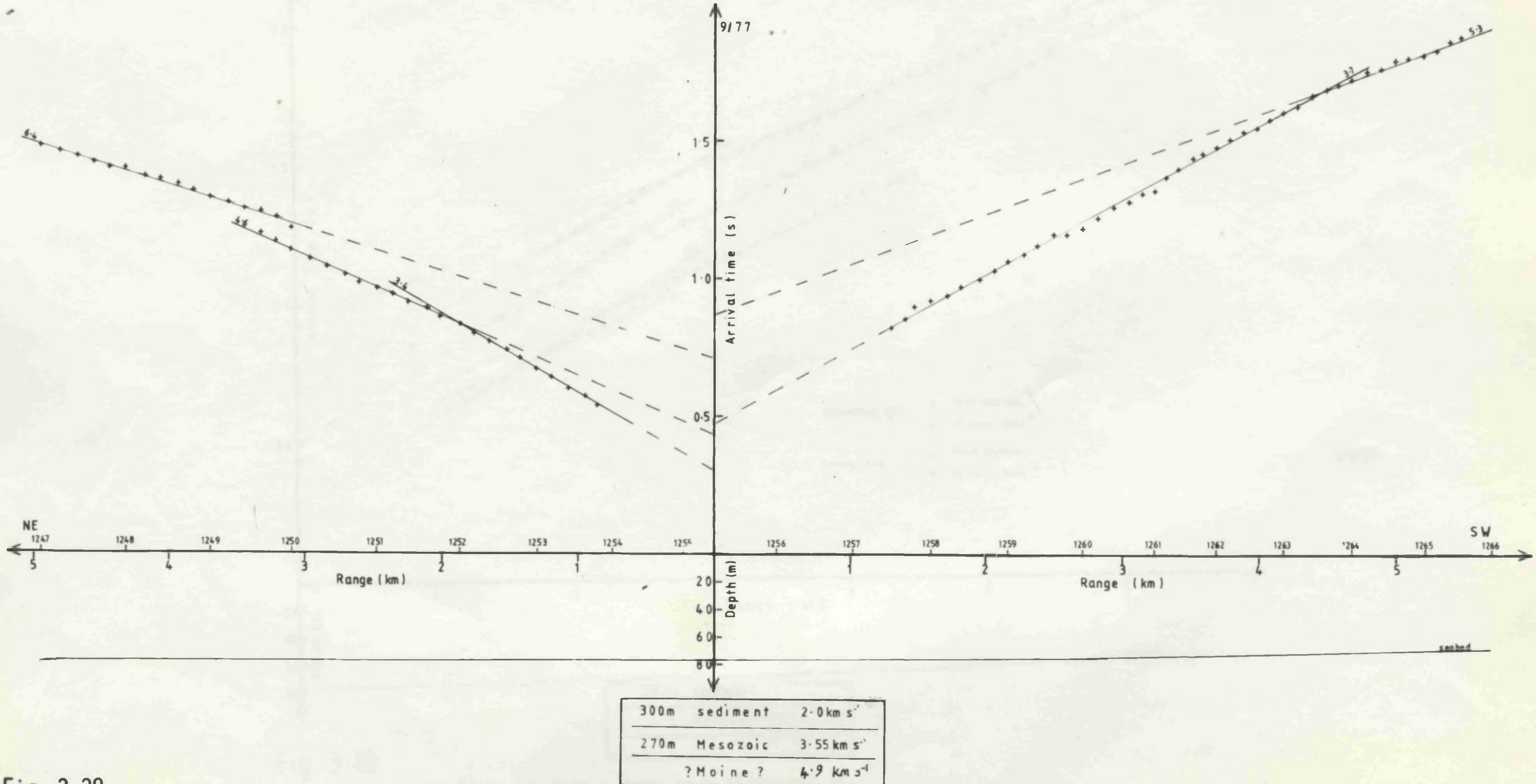


Fig 3·39

Time-distance graph for sonobuoy line 9/1977 obtained using a 40 cu.in. airgun. No water layer correction applied. Interpretation (in box) of thicknesses pertain to buoy position. (see pp. 38-40; 50-51)
Smooth magnetics.

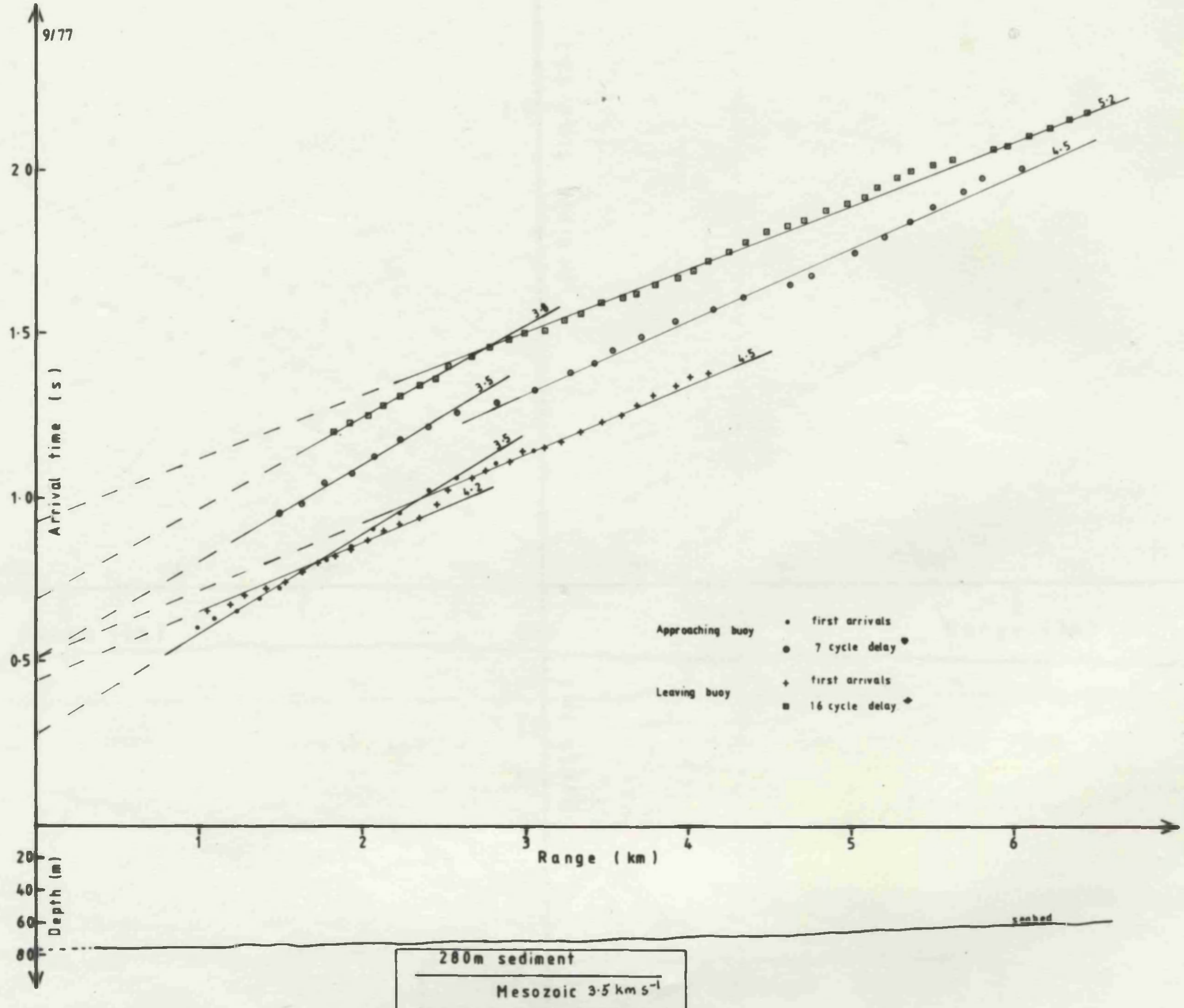


Fig 3·40

* When the first cycle of an arrival fades a later cycle is followed.

Time-distance graph for sonobuoy line 9/1977 obtained using a 150 cu.in. airgun (see pp. 50-51). Smooth magnetics.

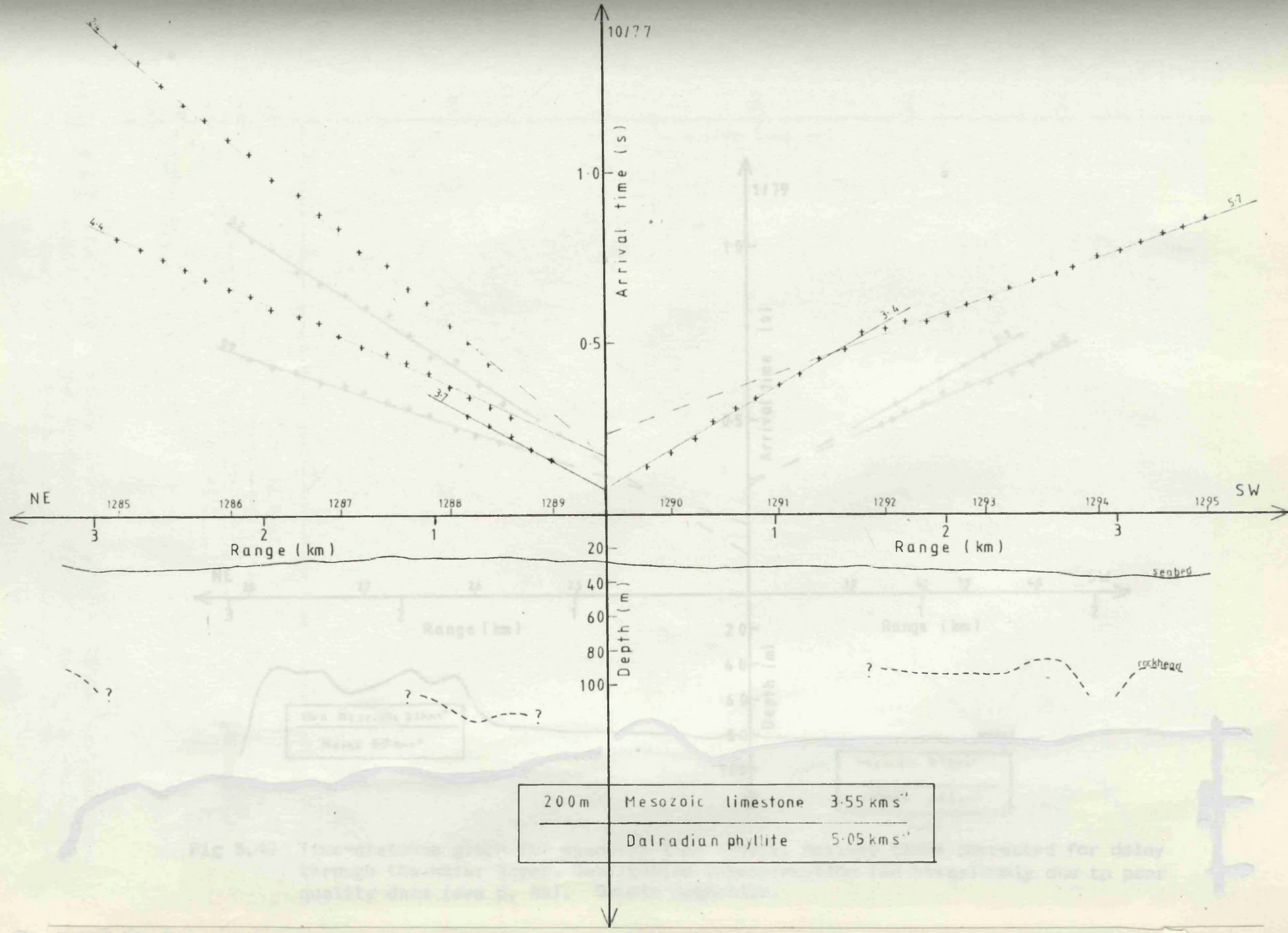


Fig 3.41 Time distance graph for sonobuoy line 10/1977. Arrival times corrected for delay through the water layer. Interpretation (in box) of thicknesses pertain to buoy position. Smooth magnetics.

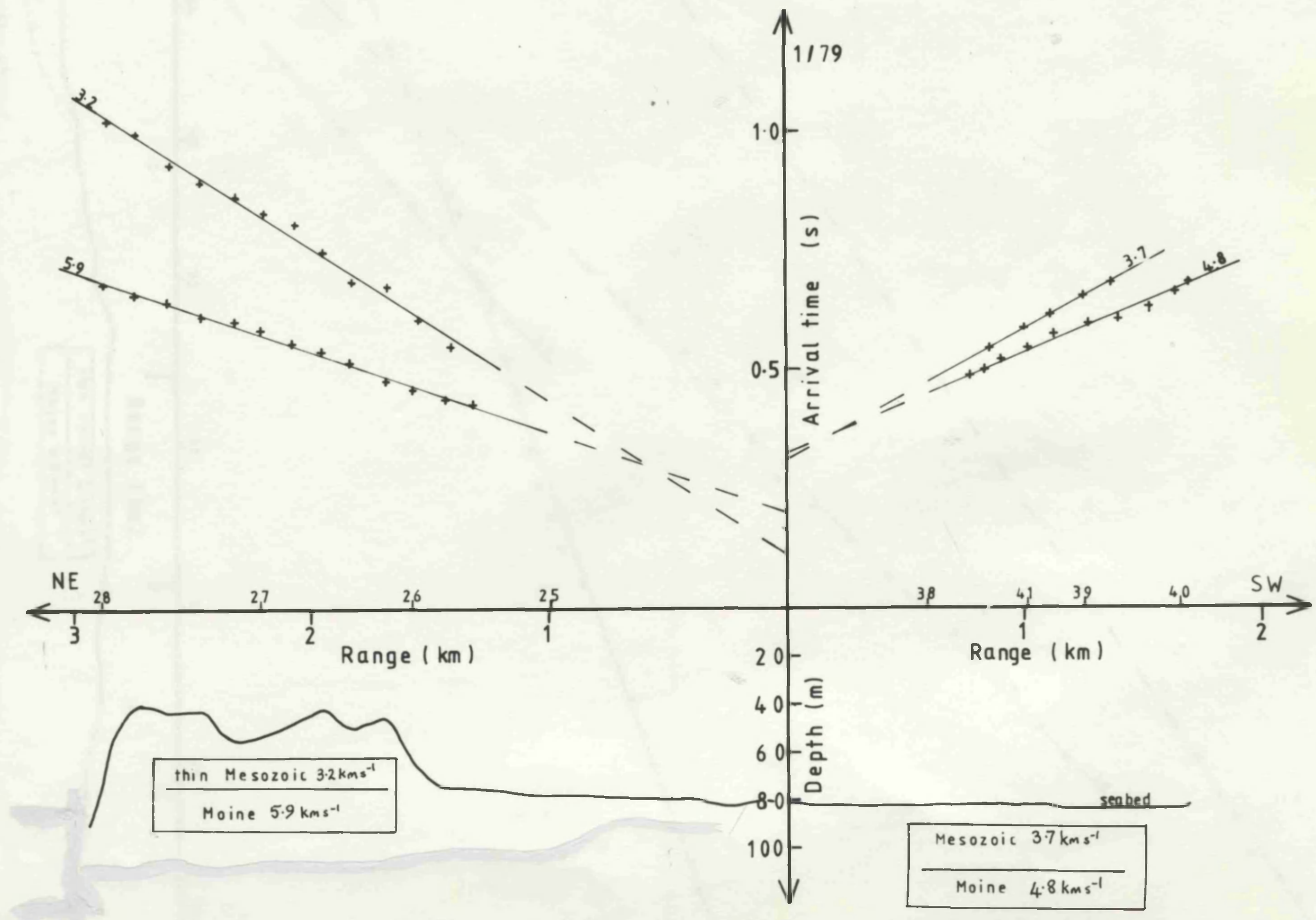


Fig 3.42 Time-distance graph for sonobuoy line 1/1979. Arrival times corrected for delay through the water layer. Qualitative interpretation (in boxes) only due to poor quality data (see p. 49). Smooth magnetics.

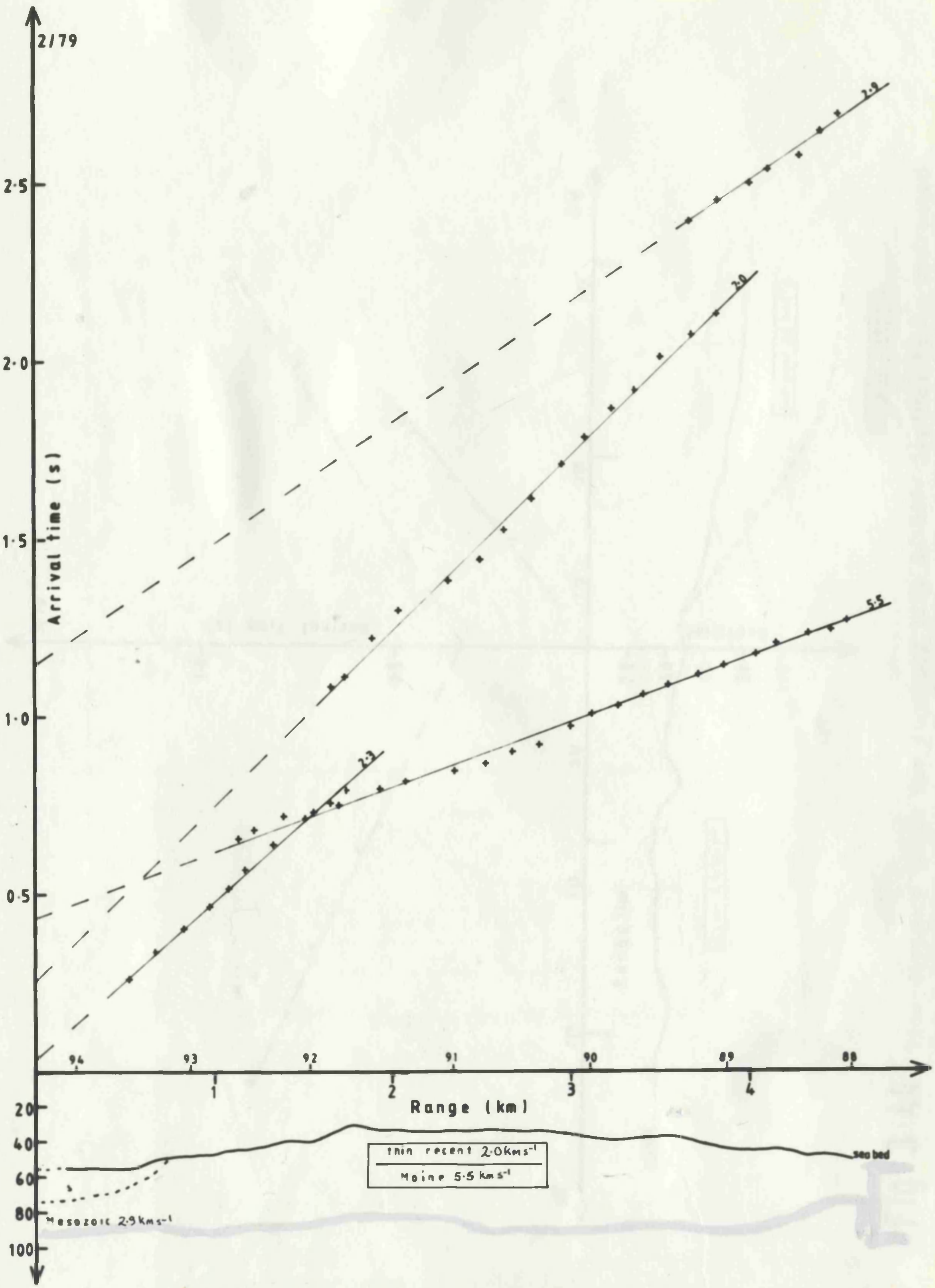


Fig 3·43 Time-distance graph for sonobuoy line 2/1979. Arrival times corrected for the delay through the water layer. Qualitative interpretation (in box) only due to poor quality (see p. 49). Smooth magnetics.

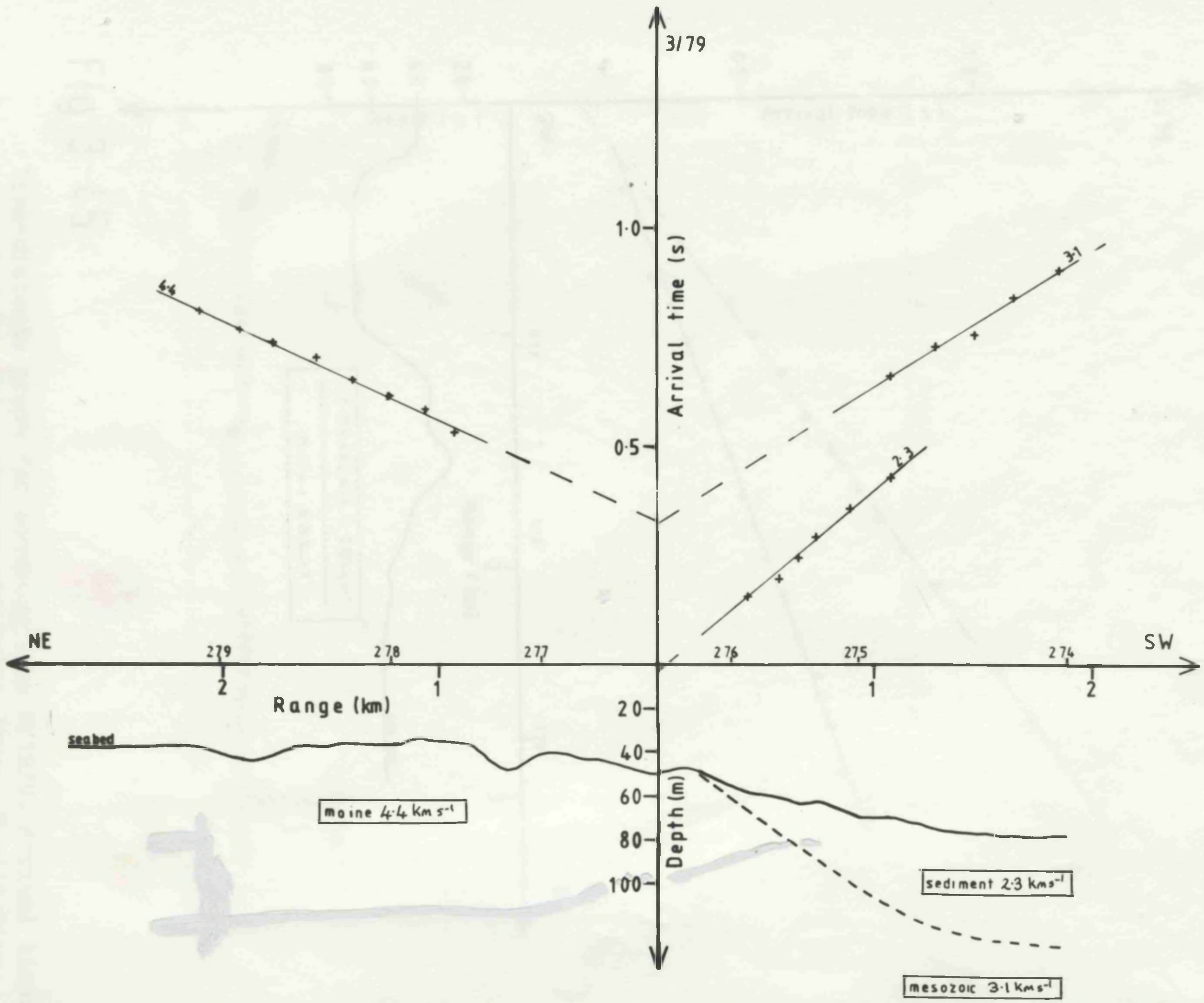


Fig 3·44

Time-distance graph for sonobuoy line 3/1979. Arrival times corrected for delay through the water layer. Qualitative interpretation (in box only due to poor quality data (see p. 49). Smooth magnetics.

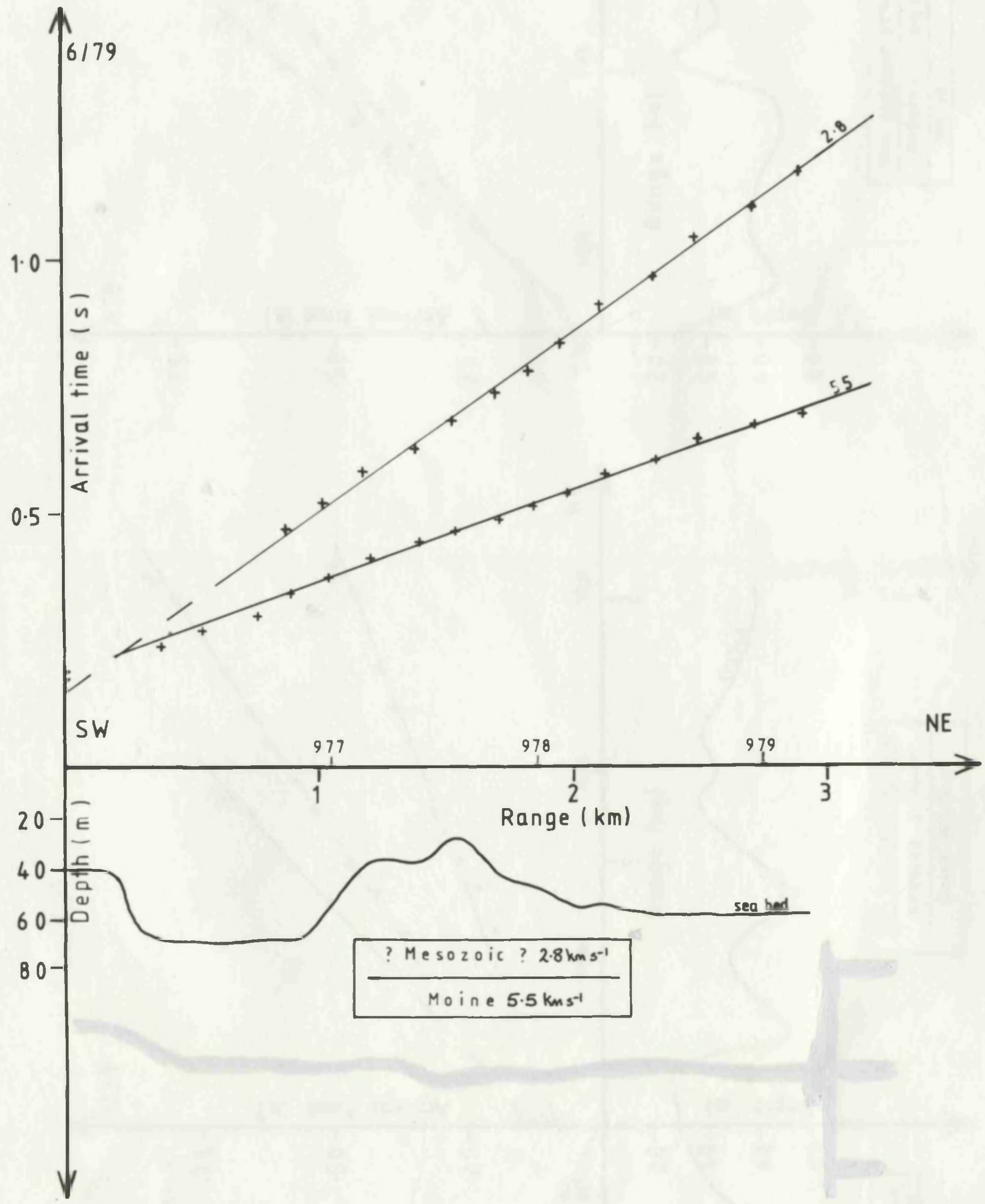


Fig 3·45

Time-distance graph for sonobuoy line 6/1979. Arrival times corrected for delay through the water layer. Qualitative interpretation (in box) only due to poor quality data (see p. 50). Smooth magnetics

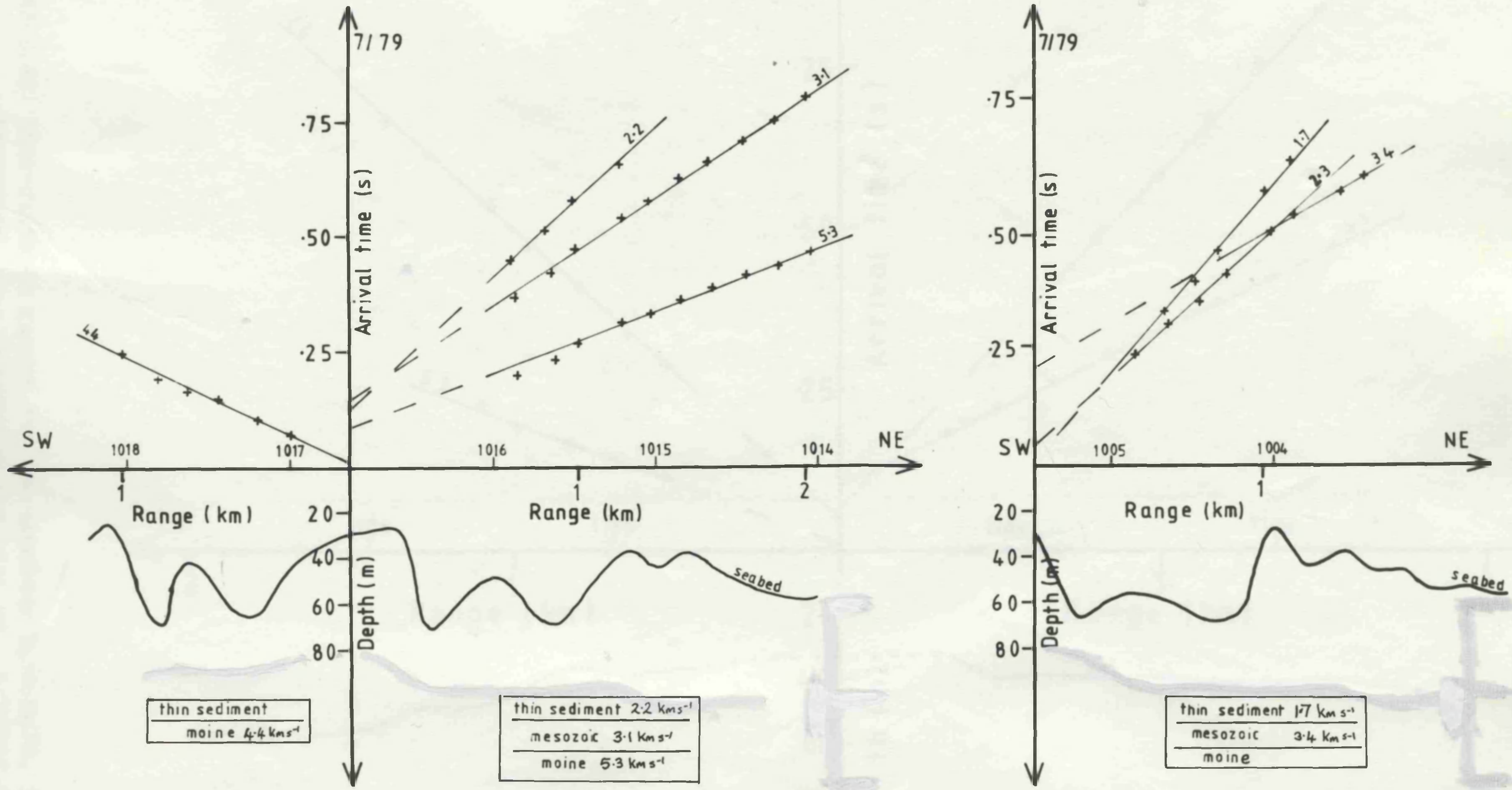


Fig 3·46

Time-distance graph for sonobuoy line 7/1979, repeated traverses shown separately for clarity. Arrival times corrected for delay through the water layer. Qualitative interpretation (in boxes) only due to poor quality data (see p.50). Smooth magnetics.

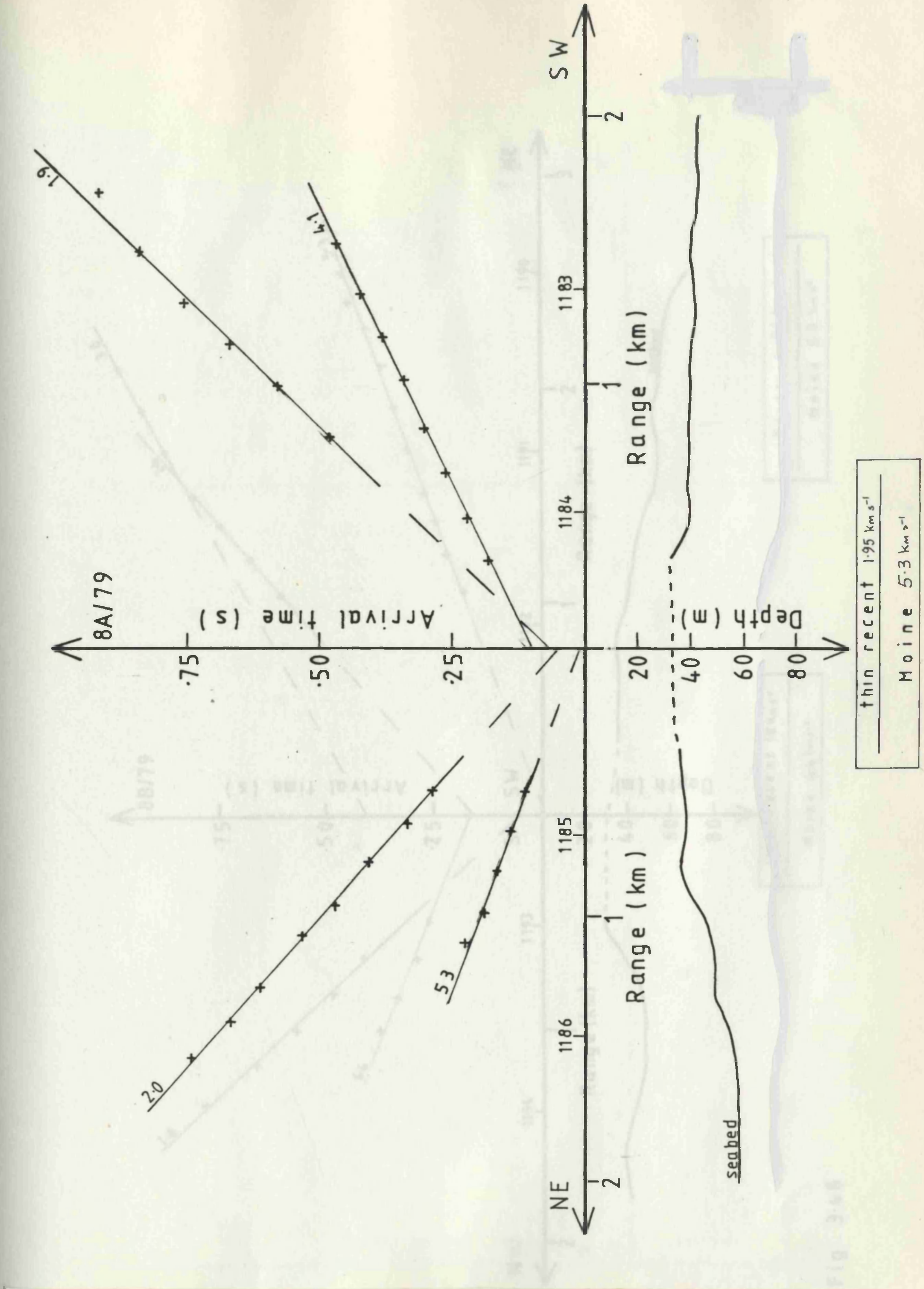


Fig 3·47

Figs 3.47-3.49 Time-distance graphs for the sonobuoy line 8/79, individual traverses shown separately for clarity. Arrival times corrected for delay through the water layer. Qualitative interpretation (in boxes) only due to poor quality data. Smooth magnetics.

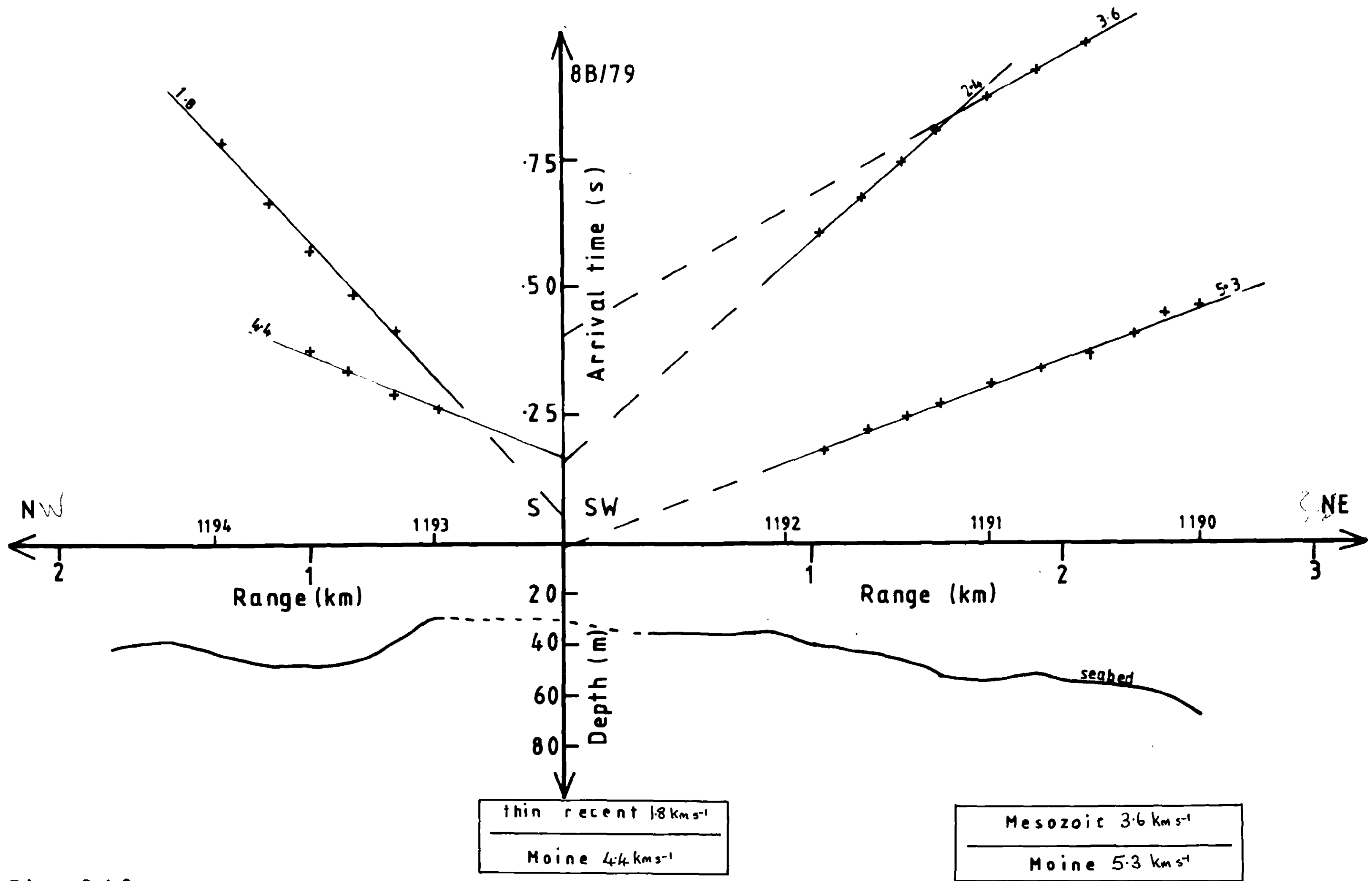
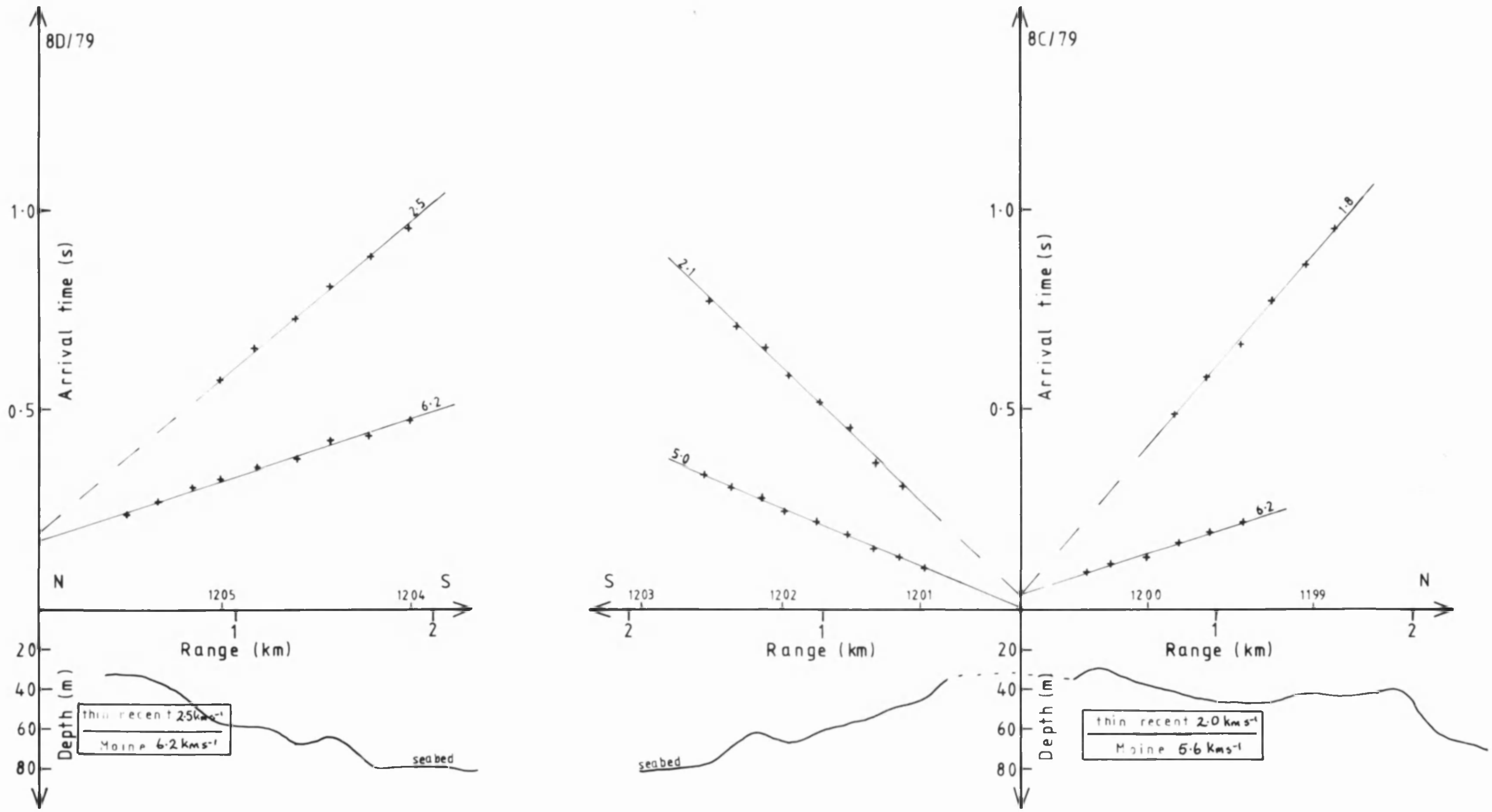

8B/79
Arrival time (s)
·75
·50
·25
1·8
4·4
3·6
2·4
5·3
NW
S
SW
NE
1194
1193
1192
1191
1190
2
1
1
2
3
Range (km)
Range (km)
20
40
60
80
Depth (m)
seabed
thin recent 1·8 kms-1
Moine 4·4 kms-1
Mesozoic 3·6 kms-1
Moine 5·3 kms-1

Fig 3·48



8D/79
Arrival time (s)
1·0
0·5
2·5
6·2
N
S
1205
1204
1
2
Range (km)
20
40
60
80
Depth (m)
thin recent 2·5 kms⁻¹
Moine 6·2 kms⁻¹
seabed
8C/79
2·1
5·0
1·8
6·2
1203
1202
1201
1200
1199
thin recent 2·0 kms⁻¹
Moine 5·6 kms⁻¹

Fig 3·49

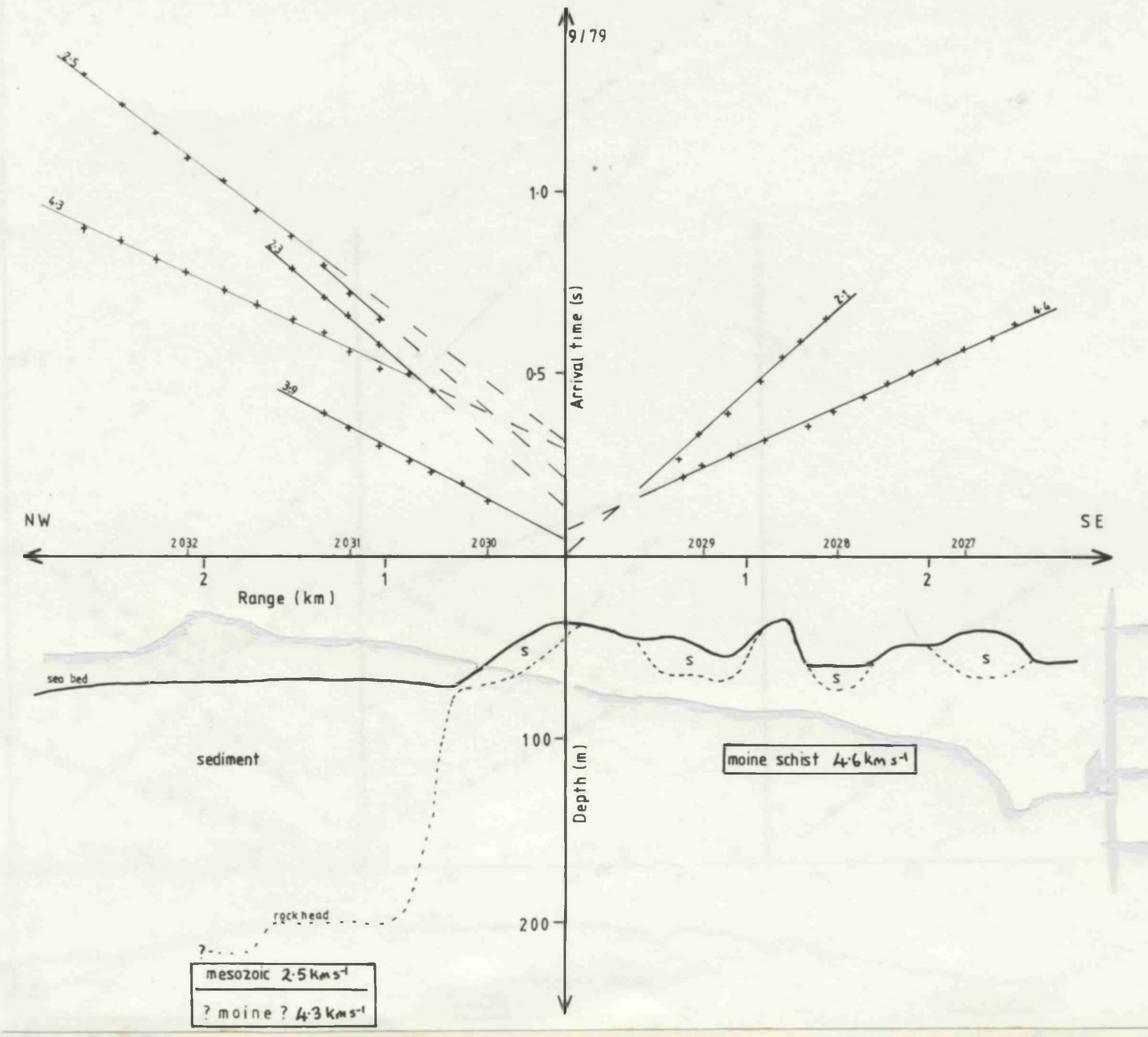


Fig 3.50 Time-distance graph for sonobuoy line 9/1979. Arrival times corrected for delay delay through the water layer. Qualitative interpretation (in boxes) due to poor quality data.

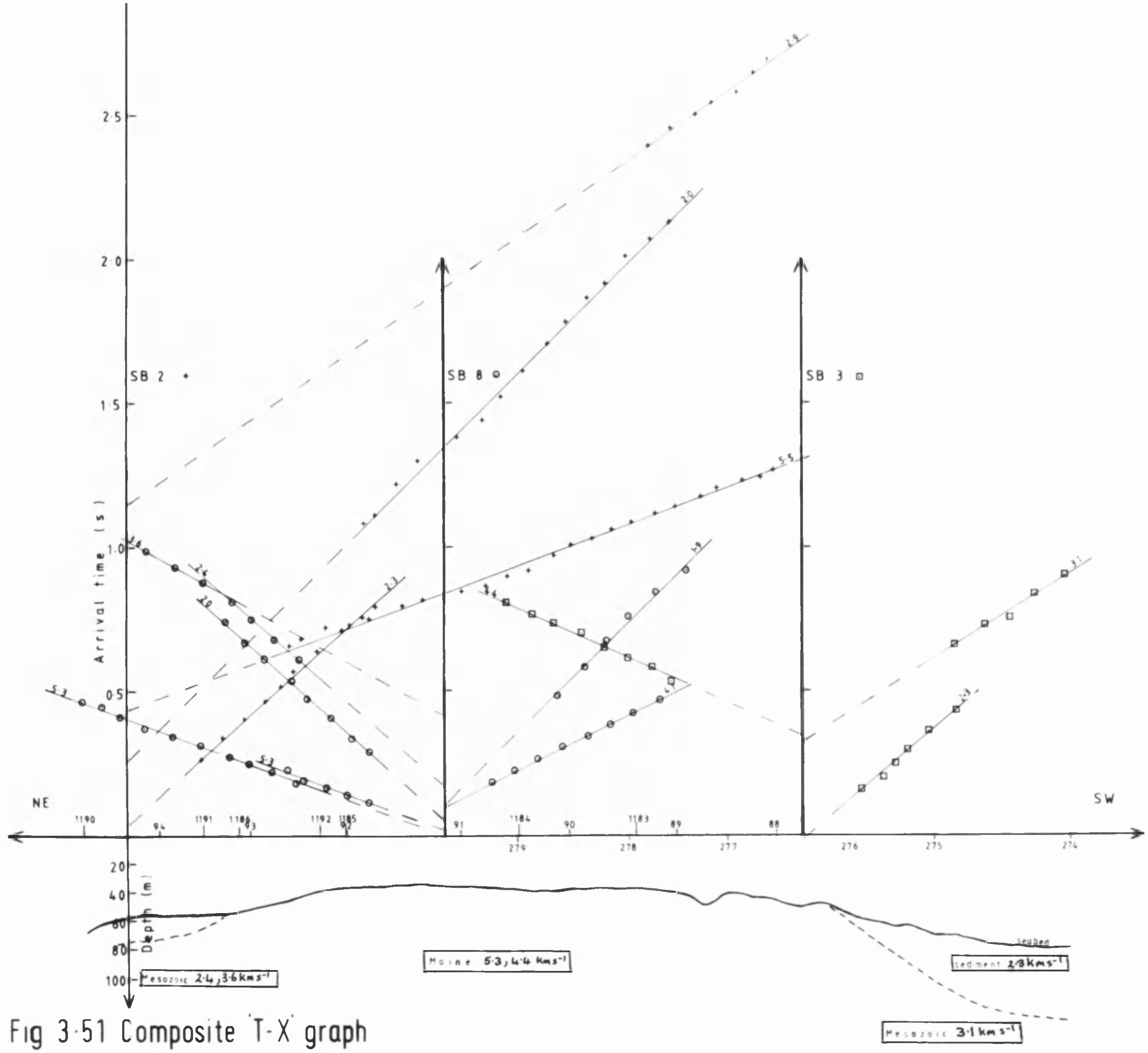


Fig 3.51 Composite 'T-X' graph

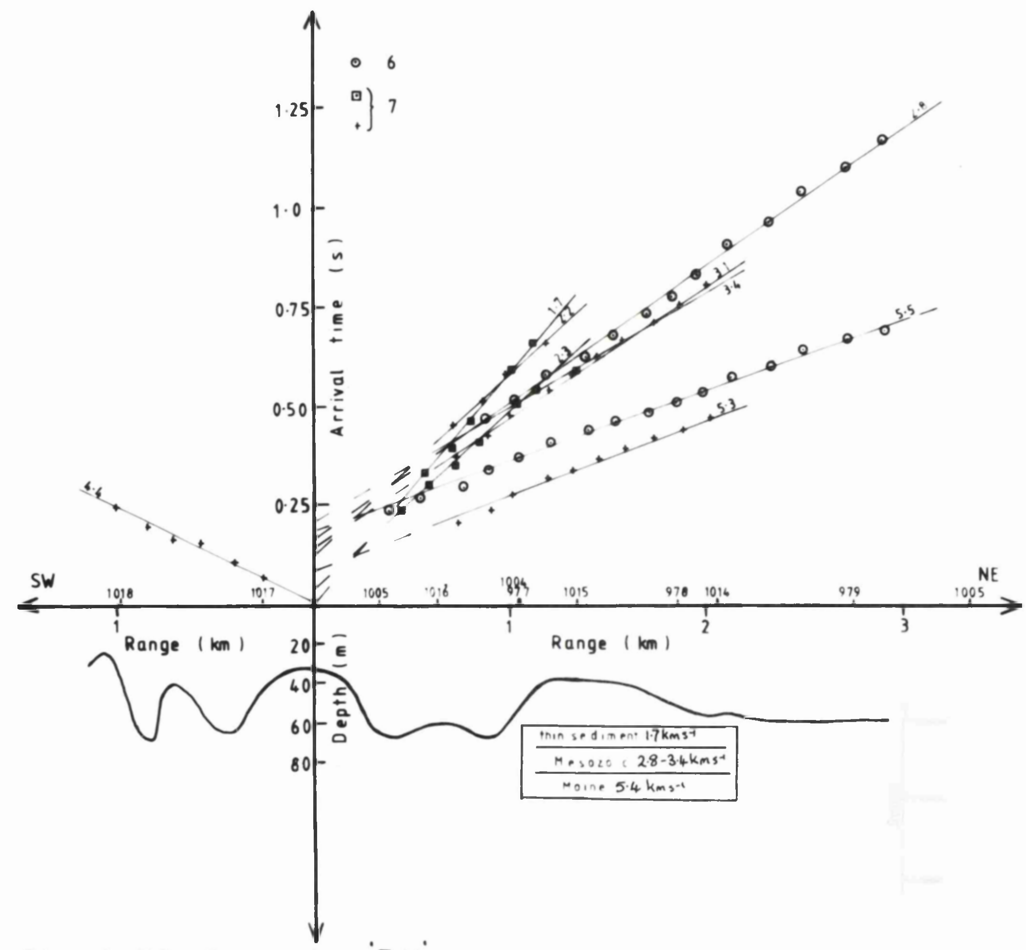


Fig 3·52 Composite 'T-X' graph.

one sonobuoy at a time. From a safety point of view there is no need to use explosives, i.e. this technique is well suited to shallow water refraction profiling.

True velocities may be estimated by taking the mean of apparent velocity segments correlated from opposite sides of the buoy by common intercept times. This involves assuming constant dip along the line. Better estimates could be made if reversed profiles were obtained especially with longer line lengths. This would require the line to be resurveyed with the buoy displaced approx. 4 km along the line from the initial position. However, due to the limitations of position fixing (see Chapter 2) it is difficult to cover precisely the same bit of ground (to better than 100 m).

It has been noted that the velocities obtained for the same rock by different methods vary, i.e. that laboratory measurements are faster than first arrival refraction times which in turn are faster than hammer lines (Hall 1970). This is largely due to the crack distribution. Laboratory specimens are small and relatively crack free, whereas hammer lines are run over exposed rock surfaces which can be heavily fractured.

3.2.3 Seismic Array

Analogue playbacks of the 50 records obtained were made. The band pass was narrowed and the attenuation of the signal amplitude was increased in order to enhance reflection line ups and/or first breaks. Also the option of straight or composited (adjacent channels added) paybacks was available. The composite mode can be helpful in enhancing reflection events.

The records were then examined for reflections. This was unsuccessful. The lack of recognizable reflections is probably due to (a) insufficient acoustic energy and (b) reverberation in the source signal and multiplicity of the seabed reflections serve to mask any weak reflection events.

It is possible to pick first breaks on consecutive records (sh. 36, Fig. 3.53).

On some records a second refracted arrival may also be picked. These results were plotted as T - X graphs for each shot. Fig. 3.54 shows the composite T - X graphs for the whole line. Bathymetric and magnetic information is also shown. No corrections for water depth have been applied to these data, as the depth remains approximately constant for the distance covered by each shot.

3.2.4 Magnetic Results

These records were examined with the sparker records and were a valuable aid in mapping certain geological boundaries and in discriminating between differently magnetised rocks (Table 3.2). They were also of use in determining the probable subcrop of the Moine Thrust in the area (see section 3.34, pp 52-53).

Lewisian, L.O.R. lava and Tertiary basalt outcrops can be identified by their magnetic character. The Lewisian outcrop North of Iona is accompanied by a broad smooth high. This high is also observed on the I.G.S. aeromagnetic map (Fig. 3.64) near the Lewisian outcrop of western Iona.

Both lavas possess distinctive magnetic characters. L.O.R. lavas are characterised by anomalies of wavelength 300m and amplitude 25 nT. Tertiary basalts are edged by lows of wavelength approx. 600m and amplitude 800 nT.

Magnetic models (Figs. 3.55 & 3.56) have been produced to illustrate the differences in anomaly form and magnetic parameters for these two lava types (Hall, pers.comm).

The essential differences being: a) the magnitude and direction of the intensity of magnetisation vector (J); for Tertiary lavas J = 400 nT, dip -80 to N170^{o} and for L.O.R. lavas J = 40 nT, dip 72^{o} to N350^{o}; b) geometry of magnetic bodies; shallow dipping sheets outcropping at both sides of the Knoll are used for the Tertiary lava model (Fig. 3.55) whilst steep dipping sheets (vertical in model for simplicity) with only one edge outcropping are required for the L.O.R. lava model (Fig. 3.56). Dips of 10^{o} - 30^{o} are observed in the L.O.R. lavas of the Oban district (Lee et.al. 1925).

Extensive use of the magnetic records was made when interpreting the Firth of Lorne geology (see section 3.3.1 p42). It was found that the magnetic character attributed to L.O.R. lavas occurred over the rocky ridges and that flat sections in the magnetic record corresponded to the troughs which are believed to be floored by L.O.R. sediments.

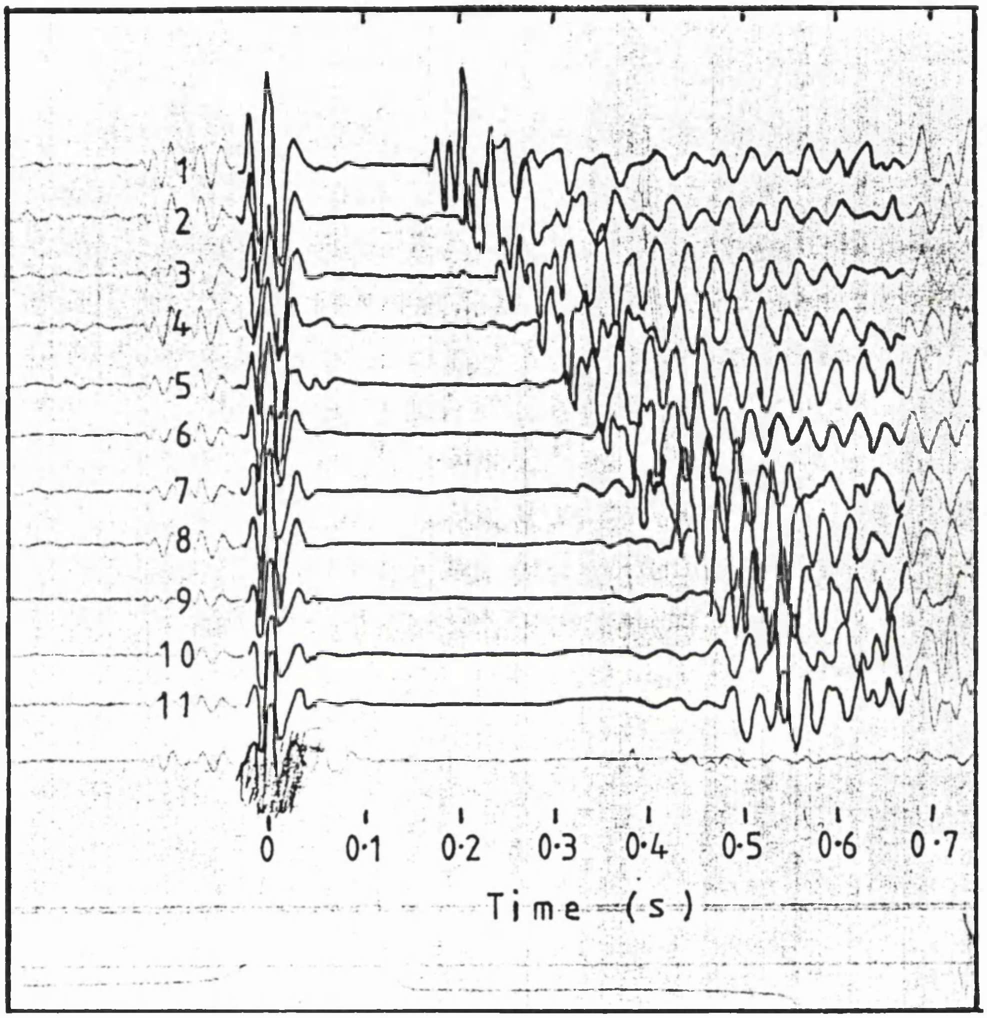


Fig 3.53 Shot 36 , seismic array line

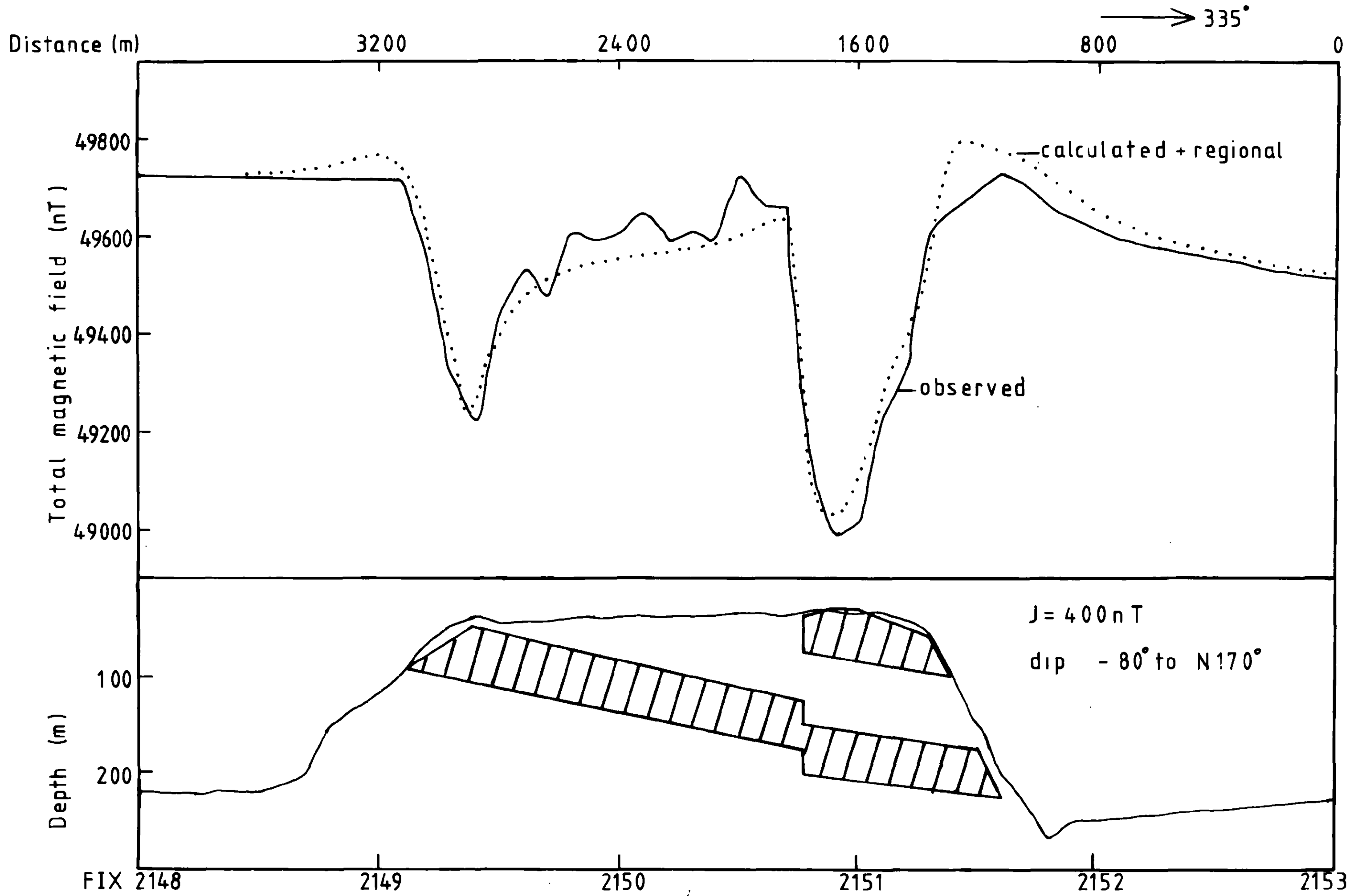


Fig 3·55 Magnetic model for Tertiary lava plateau, Staffa. (JM 3/79) (Hall, pers. comm.)

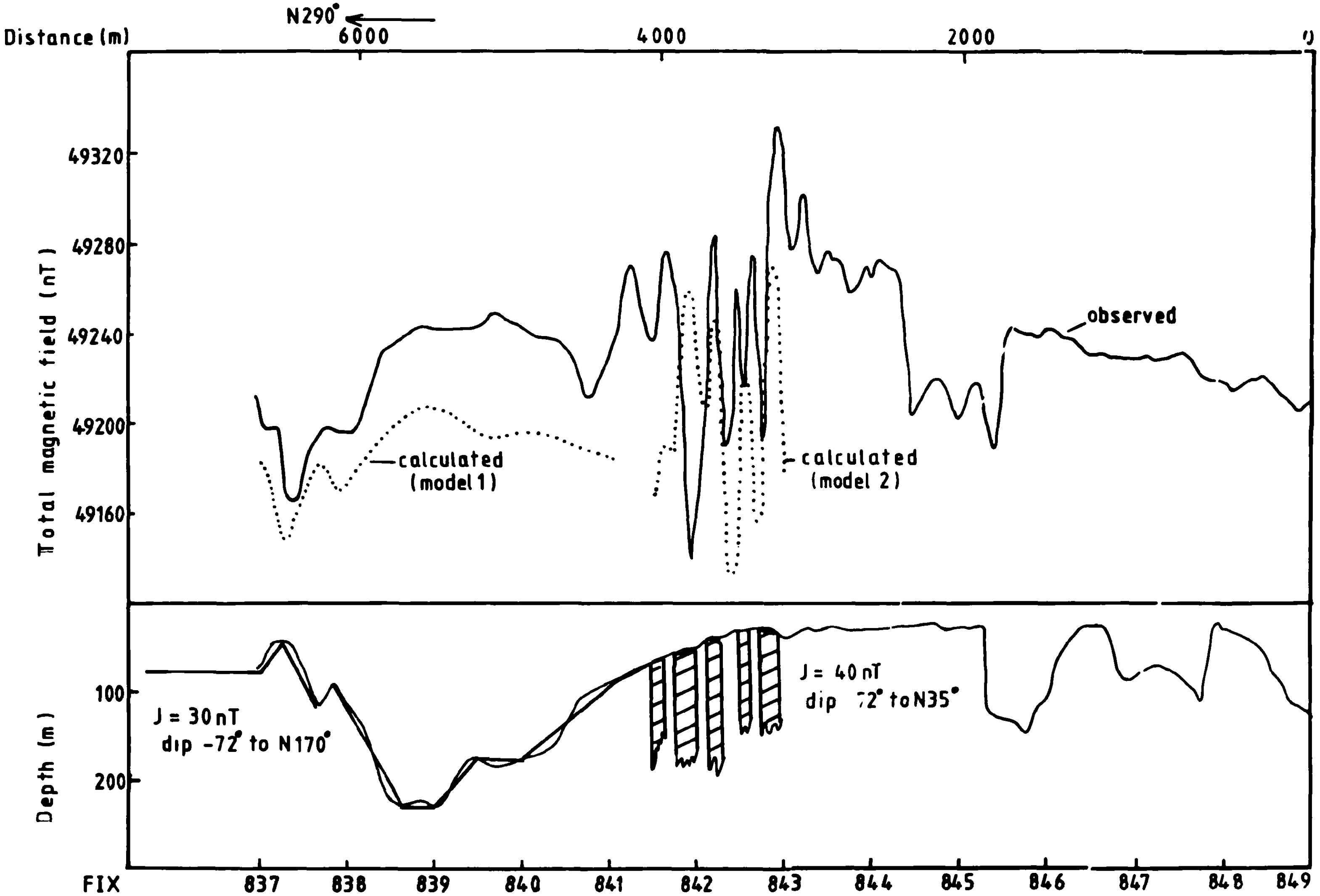


Fig 3·56 Magnetic model for L O R lava plateau, S of Mull (Challenger 77) (Hall, pers. comm.)

3.3 The Geological Interpretation

The map has been divided into 5 areas (Fig. 3.57) to be discussed in turn.

3.3.1 Area 1, Firth of Lorne and the Garvellachs

This area is characterised by the strong bathymetric relief of two groups of narrow, deep troughs, one trending NNE-SSW and the other NE-SW (Fig. 3.58). A broader trough (A on Figs. 3.3 & 3.4) between Lismore and Kerrera contains at least 100 m of recent sediment thinning southwards. South of Kerrera (B on Figs. 3.3 & 3.4) there is a smaller sediment-filled basin.

Minor pockets of recent sediment occur between the Garvellachs and Scarba (C, D & E on Figs. 3.3 & 3.4). Apart from these localised occurrences the area is devoid of recent sediment.

The rocks outcropping at the seabed can be separated into those which are magnetised and those which are not. Generally the rocks which are magnetised form the ridges in the area at around 30 m below sea level. However, some do occur at greater depths, the magnetic record being correspondingly attenuated (Fig. 3.59, F-F' on Fig. 3.2).

The magnetised rocks can be divided on the basis of magnetic and seismic character (Table 3.2, Figs. 3.6, 3.8, 3.55 & 3.56) into lavas of LOR and Tertiary age. The Tertiary lavas occurring at the east coast of Mull extend a short distance off shore. LOR lavas, which are recognised on land as far west as the Loch Don anticline on Mull, form the remainder of the shoals.

The most westerly submarine LOR lava outcrop is crossed by the SE side of sonobuoy line 2A/77 (Figs. 3.14 & 3.34). The 5.0 kms^{-1} arrival observed is consistent with other velocities

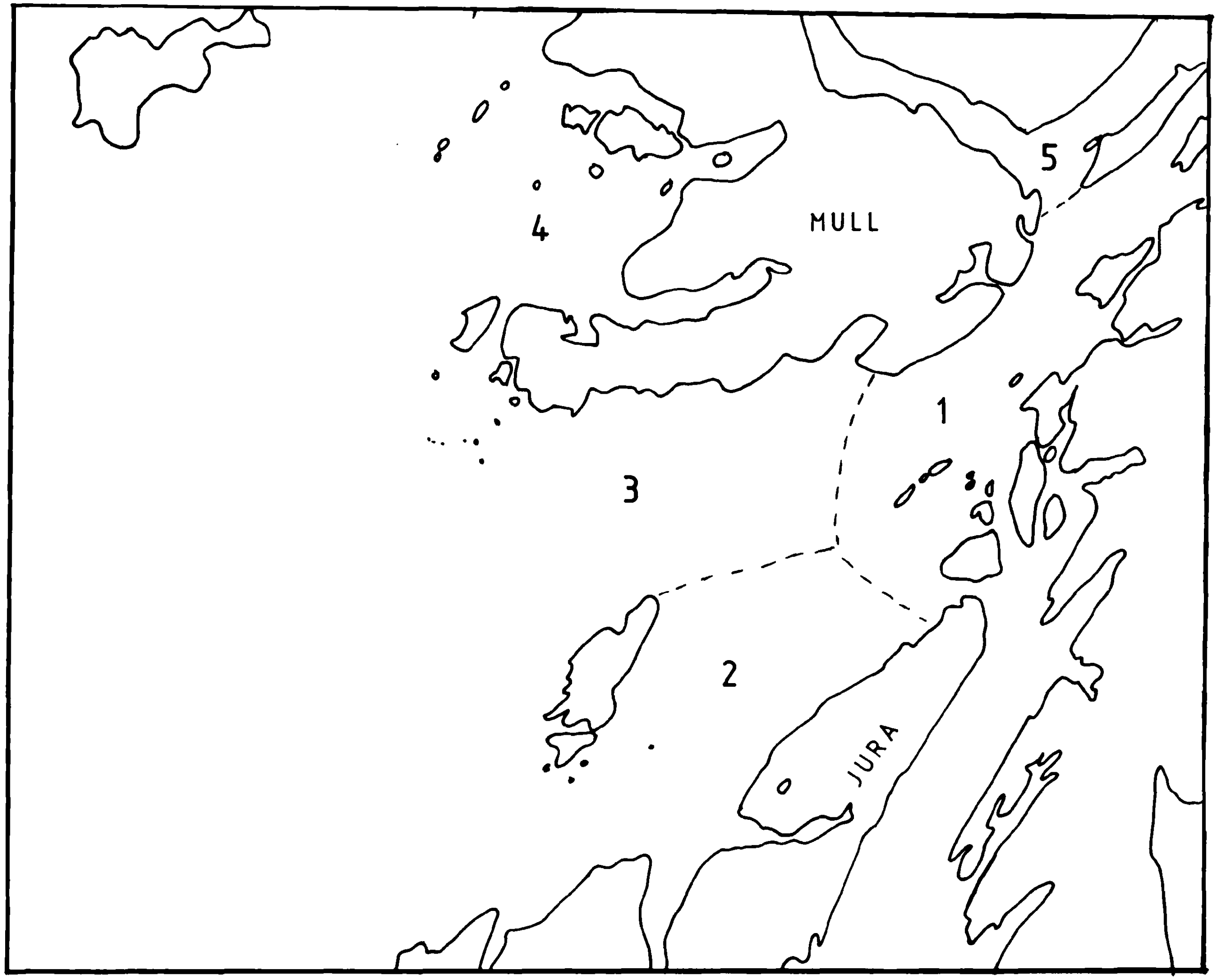


Fig 3·57 Areas used in geological discussion.

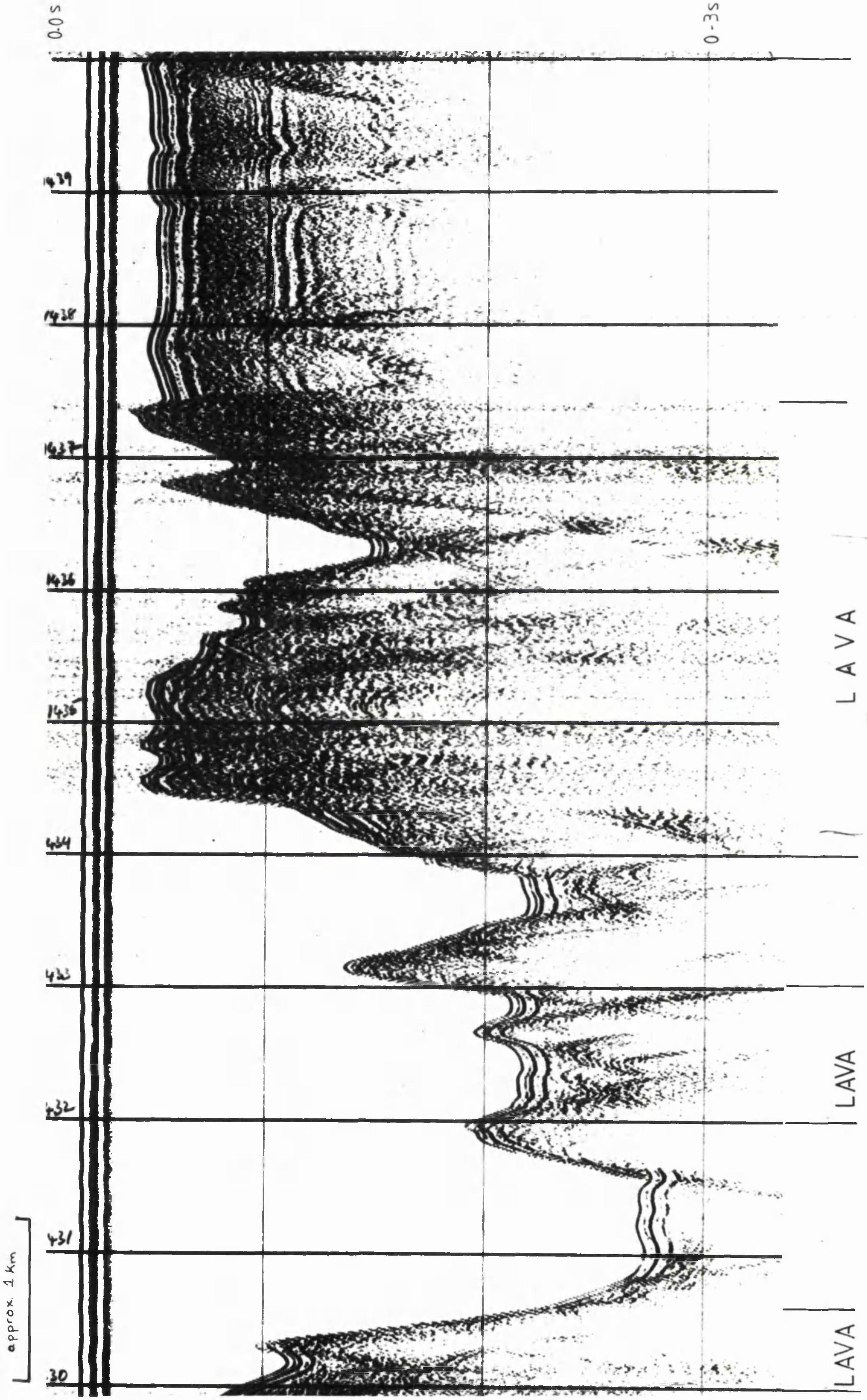


Fig 3.59 Profile across faulted sequence

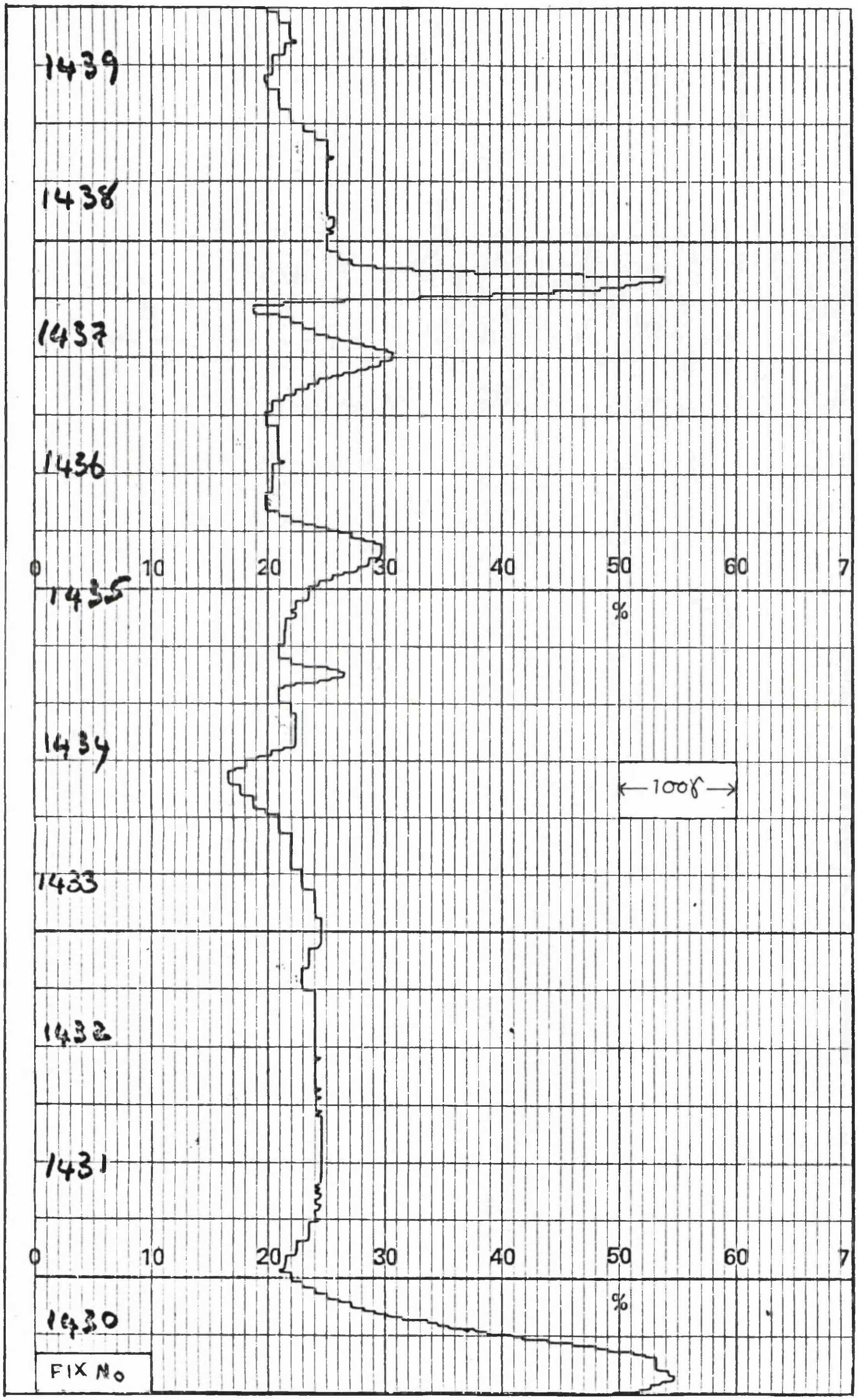


Fig 3.59 Magnetic profile across faulted L.O.R. sequence

measured on LOR lavas (Table 3.2). From sparker records crossing this line very little recent sediment is observed on the shoal (only small pockets) therefore suggesting that the 2.6 kms^{-1} arrival also observed on this side cannot be due to refraction along a seabed layer of recent sediment. It is possible that this may be a shear wave arrival resulting from conversion at the water - seabed interface. This would give a V_p/V_s ratio for LOR lavas of 1.9 and a Poisson's Ratio (σ) of 0.31. The prevalent lava type in the Oban and Kerrera district is a fine-grained basalt (Lee and Bailey 1925). Measurements on a large number of basalt samples obtained during D.S.D.P. leg 37 yielded Poisson's Ratios in the range 0.27 - 0.32 (at 0.07 kbar) (Hyndman 1974). Laboratory measurements on rocks of similar composition have also produced similar values; altered diabase (Vp/Vs 1.91 - 1.93, σ 0.31 - 0.32), partially serpentinised peridotite (Vp/Vs 1.88 - 1.93, σ 0.30 - 0.32) and metagabbros (Vp/Vs 1.83 - 1.93, σ 0.29 - 0.32). Rocks of different composition have different Vp/Vs and Poissons Ratios, e.g. quartzite Vp/Vs 1.54, σ 0.136) (Christensen 1965).

The troughs are probably occupied by LOR sediments separated from the LOR lavas by either normal or faulted boundaries. Both types of boundary are observed on Kerrera (I.G.S. sh. 44), the faults having either NNE-SSW or NE-SW trend.

Frank Lockwood's Island is probably of LOR sediments (I.G.S. sh. 35) and, if this is so, comprises the most westerly land outcrop of these strata. Barber et al. (1979) have recovered samples (3341, 3352, 3363) of conglomerate believed to be LOR age from shoals S of L. Buie, indicating the most westerly extent of these sediments in the area examined.

It must be noted that an alternative floor to the troughs

ESE of Mull could be Dalradian slates and phyllites which are also non-magnetic and more easily eroded than LOR lava. However, as the LOR sediments, which underlie the lavas, thicken westwards (Lee & Bailey 1925) it is considered here more likely that they occur in the troughs.

Minor NW to SE trending faults (GG', HH' and II') are proposed to account for the difficulties in correlating the bathymetric and magnetic profiles of successive lines up the Firth. Similar trending faults are observed on Mull in the Tertiary lavas and central intrusive rocks, also on the mainland at 56^{O} 20'N 05^{O} 30'W in the L.O.R. lavas.

In the southern part of this area Dalradian rocks form the Garvellach Islands (Portaskaig Boulder Bed) and Scarba (Jura quartzite).

The Portaskaig Boulder Bed and the Jura Quartzite are probably brought into outcrop in these areas by a culmination on the Islay Anticline. The bathymetric trough immediately west of the Garvellachs is probably occupied by Islay Limestone which underlies the Portaskaig Tillite on Islay (Table 1.2). Spencer (1971) proposed that minor sinistral faults between the Islands of the Garvellachs offset horizons in the Tillite. He observed faults of similar trend and disposition on the islands. The sygmoidal trend of the troughs West and East of these islands probably offsetting the more easily eroded horizons would appear to support this hypothesis.

The deep trough between the Garvellachs and Scarba J-J' (Fig. 3.58) may mark the line of a fault apparently downthrowing to the east the Jura Quartzite succession. The line of this trough extends southwards, into area 2, where it affects Mesozoic sediments.

Structurally this area is dominated by NNE-SSW and NE-SW grains attributed to faulting and to the geological strike of the rocks outcropping in the area. For the submarine structures it is not possible to discriminate between the two boundary types. Bailey (1960) suggests that the volcanic activity giving rise to the Lorne Plateau Lavas may be genetically related to fault activity along the Great Glen Fault line. Morton (1976) suggests that this faulting may have influenced the nature and location of the margins of the basins in which the LOR sediments were deposited and comprise a wide fault zone. Horizontal and minor vertical movements probably took place along these faults during Lower Old Red times, bringing lavas and sediments into contact with each other. Similar fault patterns are observed on Kerrera (Fig. 3.2).

The NW-SE faults, G-G', H-H' & I-I' (Fig. 3.2) are probably later than the WNE-SSW and NE-SW faults which terminate against them. These faults are probably later than the Tertiary central igneous complex on Mull as similar faults affect rocks of the complex (I.G.S. sh. 44).

The curved shape of the Mull coastline is controlled by rim folds related to the intrusion of the Mull igneous centre.

3.3.2 Area 2, Between Colonsay and Jura

The north-east of this area contains two NE-SW trending (J-J' and K-K', Fig. 3.58) bathymetric troughs which are probably fault lines. The most westerly trough is an extension of the deep trough (J-J') running east of the Garvellachs. The more easterly trough is much broader and contains two parallel narrow troughs both containing recent sediment (see Fig. 3.4).

The west-east depression extending from the Gulf of Corryvreckan is probably due to intense tidal scour from the

Great Race -- the tides associated with the Gulf of Corryvreckan which can reach speeds of 12 knots. These currents also serve to sweep the area clear of recent sediments.

Much of this area is occupied by Mesozoic sediments as indicated by a smoother rockhead with shallow dips below (Table 3.2, Fig. 3.60). Apparent velocities (3.7 kms^{-1}, 3.4 kms^{-1}, 3.5 kms^{-1}, 3.6 kms^{-1}, 3.2 kms^{-1}, 2.6 kms^{-1} and 2.8 kms^{-1}) consistent with Mesozoic sediment velocities obtained elsewhere (Table 3.2) have been recorded on sonobuoy lines 10/77, 5/77, 6_{40}/77 and 6_{150}/77 (Figs. 3.41, 3.35, , 3.37, 3.38). Red sandstone of possible LOR or Permo-Trias age has been recovered from I.G.S. borehole 72/9 (Chester *et al*. 1972). The present work would favour a Permo/Triassic age for these sediments on the evidence of apparent seismic refraction velocities and the occurrence of shallow dipping reflectors in the rockhead of this area.

The western margin of the Mesozoic outcrop is faulted against the Torridonian sediments. This fault is expressed by minor bathymetric scarp (L-L', Fig. 3.58) to the east of Colonsay. The scarp becomes less well defined northwards. A magnetic anomaly also coincides with the position of this fault (Westbrook & Borradaile 1978). This fault has been correlated with the Loch Gruinart fault on Islay (Dobson *et al*. 1975, Durrance 1976, Westbrook & Borradaile 1978).

A minimum downthrow to the east of 1 km has been proposed for the Loch Gruinart Fault.

The eastern margin of the Mesozoic is also believed to be a fault. The fault line runs parallel to, and just off, the west coast of Jura.

Seismic refraction lines in the area:

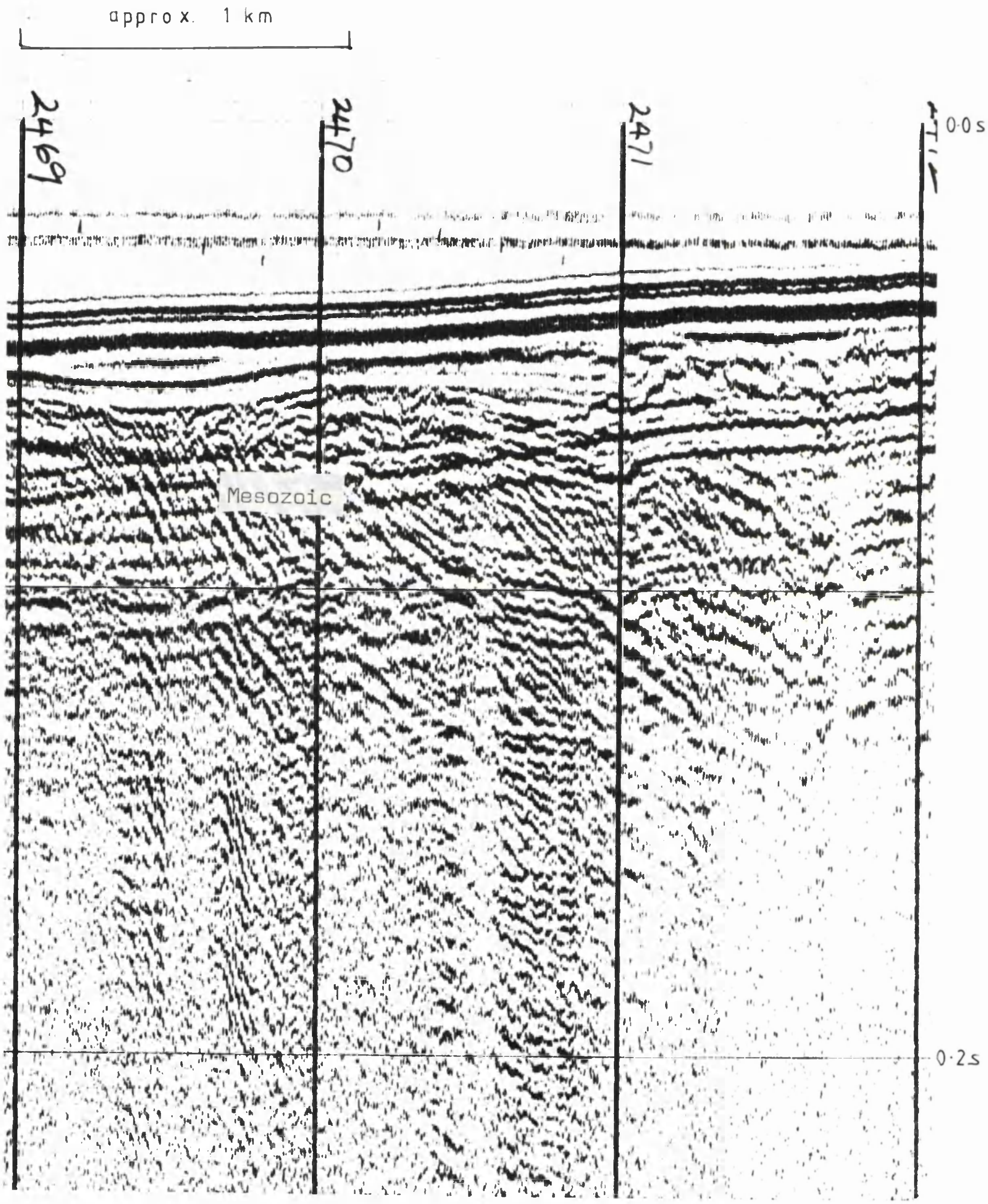


Fig 3.60 Mesozoic strata from Area 2

Sonobuoy Line 10/77

The arrivals with apparent velocities 3.4 km s^{-1} and 3.7 km s^{-1} and intercept times of 0.079s and 0.75s respectively are probably refracted from the same surface, this is considered to be Mesozoic sediment. The difference in apparent velocies can be attributed to a dip of approximately $1\frac{1}{2}^{o}$ SW on the Mesozoic surface.

No arrival from the recent sediment layer forming the sea-bed can be recognised. If a velocity of 2 km s^{-1} is assumed for this sediment layer its thickness would be 72m. This estimate lies within the range of sediment thicknesses (50-90 m) observed on a sparker line running close to the sonobuoy line.

Correlation of the 4.4 km s^{-1} and 5.7 km s^{-1} arrivals gives a mean velocity of 5.05 km s^{-1}. The difference in apparent velocities can be accounted for by an apparent dip of 8^{o} NE on the refractor. The mean velocity of 5.05 km s^{-1} is consistent with the velocities measured (Table 3.2) for Dalradian phyllites and quartzites. The velocities for Torridonian sediments (quartzite, not measured) are also likely to fall in this range thus preventing identifications of this refractor on velocity grounds.

The 2.4 km s^{-1} arrival seen only on the NE side is attributed either to a boulder clay horizon or to an 'S' wave from the 5.05 km s^{-1} refractor which would give a P/S velocity ratio of 1.83 and Poissons Ratio of 0.29. Christensen (1965) has obtained a Vp/Vs ratio at 1 Kb of 1.82 and a Poisson's Ratio of 0.284 from laboratory measurements on a slate specimen ; these values are not close to those measured for quartz (Vp/Vs 1.49; Poisson's Ratio 0.08(Christensen 1965)).

Therefore the determination of P/S velocity ratio and Poissons' ratio is of value by indicating what the refractor is not. The refractor has been interpreted as Dalradian slate on the basis of three lines of evidence:-

1) When the general trend of the Dalradian rocks, from the Garvellachs to Islay is taken into account it is probable that they underlie the sonobuoy line.
2) The values of P/S velocity ratio and Poissons' ratio show that the refractor is not quartzite and is unlikely to be Torridonian metasediment which is predominantly quartz.
3) The Moine thrust has been located to the west of the sonobuoy line thereby separating Torridonian to the west from Dalradian to the east.

The siting of the Moine subcrop was based primarily by consideration of the onshore geology in conjunction with the marine magnetic data. As in the area North of the Ross of Mull (p.53) the Moine Thrust subcrop was placed approximately 2 km east of the maxima observed on the marine magnetic lines. It must be noted that this is only an approximate position as the following assumptions are made:-

1) the shallowest magnetic horizon in the Lewisian is at a constant depth below the Moine Thrust plane.
2) the Moine Thrust plane has a constant dip.

<u>Sonobuoy Lines 5/77, 5X/77 and 6/77</u> (Figures 3.35, 3.36 3.37, 3.38)

Similar apparent velocities are observed along these lines. On line 5/77 approximately 280 m of Mesozoic sediments are estimated to underlie a varying thickness of recent sediments (Fig. 3.35).

which has been corrected for before obtaining the apparent velocities. As like line 10/77, this is believed to overlie Dalradian phyllite (velocity of approx. 5.3 kms^{-1}).

Lower apparent velocities 2.6 kms^{-1} and 2.8 kms^{-1} are recorded on sonobuoy line 6_{40}/77 and could be refracted from either a coarse, porous, Mesozoic sandstone or a very stiff boulder clay, the former favoured. On the NE side only a somewhat faster, probable Mesozoic arrival is observed. This velocity of 3.3 kms^{-1} (3.2 kms^{-1} on 150 cu. in. record) which could be either Mesozoic sandstone or limestone.

For line $6/77_{40}$ a total of 445 m of Mesozoic sediment is estimated: 145 m of seismic velocity 2.7 kms^{-1}, probably sandstone and 300 m of a faster, 3.3 kms^{-1}, rock -- either finer-grained sandstone or limestone. A similar estimate (425 m) for the total Mesozoic thickness is obtained from line $6/77_{150}$ where the low velocity arrival 2.5 kms^{-1} is ill-defined and only recognised on the NE side.

Mean velocities of 4.7 kms^{-1} and 4.9 kms^{-1} from lines $6/77_{40}$ and $6/77_{150}$ respectively are obtained for the basement which is believed to be Dalradian phyllite.

Structurally this area contains a broad basin, with boundary faults running close to the shores of Colonsay and Jura. The basin is floored by Dalradian rocks and contains a fill which dates back at least to the Permo/Trias. The thickness of the Mesozoic fill is of the order of 300 - 450 m.

3.3.3 Area 3, Between Colonsay and Mull

The eastern part of this area is occupied by two broad, shallow bathymetric depressions (M & N, Fig. 3.58), one extending SW from Loch Buie separated from the other to the south by a shallower area. In this eastern area a NE-SW-trending series of shoals (less than 40 m deep) is observed. Both the depressions contain substantial thicknesses of recent sediment > 250 m and so form extremely deep troughs in rockhead (see Figs 3.3 and 3.4).

West of this region the seabed is generally no deeper than 80 m b.s.l. with extensive shoal areas (< 40 m) immediately off the coasts of Mull and Colonsay. The rockhead map and sediment isopach maps (Figs. 3.3 and 3.4) in contrast show a deep trough midway between Colonsay and Mull containing substantial recent sediment (> 250 m along the axis). This trough is continuous with the trough originating from Loch Buie though they have different trends.

The shoal area north of Colonsay which is largely sediment free is occupied by Torridonian rocks. The Torridonian rocks extend to the southern margin of the central trough.

The shoal areas south of Mull are formed from several different rocks. South of Carsaig Bay two inliers of Moine rocks (P & Q, Figs. 3.2, 3.58) have been identified on sparker records (Table 3.2, Fig. 3.9). Sonobuoy lines 1/79, 2/79, 3/79 and 8/79 (Figs. 3.23, 3.24, 3.25, 3.29) all cross the outcrop where apparent velocities (5.9 kms^{-1}, 4.8 kms^{-1}, 5.5 kms^{-1}, 5.3 kms^{-1}, 5.0 kms^{-1}, 6.2 kms^{-1}) consistent with Moine rocks (Table 3.2) have been observed. Velocities consistent with Mesozoic sediments (3.2 kms^{-1}, 3.7 kms^{-1}, 2.9 kms^{-1}, 3.1 kms^{-1}, 3.6 kms^{-1}) have also been recognised, supporting the interpretation that these Moinian inliers are surrounded by Mesozoic sediments and may also have pockets of Mesozoic sediments in them.

Outliers of Tertiary basalt (R & S, Fig. 3.2) have been identified west of the Moine inliers, recognised by their distinctive seismic and magnetic character (Table 2).

The shoal area immediately south of the Ross of Mull is largely the southerly extension of The Ross of Mull Granite. To the south east of this granite an outcrop of Moine standing proud of the surrounding granite is recognised. Apparent velocities of around 5.4 kms^{-1} are observed on sonobuoy lines 6/79 and 7/79 which lie close to the northern margin of this outcrop. Apparent velocities consistent with Mesozoic sediments are also recognised on these lines suggesting that pockets of Mesozoic sediments lie on the Moine.

The remainder of the northern rockhead area (shallower than 100 m b.s.l.) is formed of Mesozoic sediments in which apparent dips of 1° - 9° are observed.

The Seismic Array Line (Fig. 3.54) and sonobuoy lines 9/77 40 and 9/77 150 (Figs. 3.39 & 3.40) were both shot over the central trough (Fig. 3.1). Fig. 3.4 shows the apparent velocities observed along the Seismic Array Line. Shot numbers increase southwestwards. Up to shot 13 apparent velocities less than 3.0 kms^{-1} only are observed, probably from arrivals which have been refracted along the soft sediment forming the seabed. Beyond shot 13 velocities greater than 3.0 kms^{-1} (but less than 4 kms) are also observed and are generally considered to be refracted from Mesozoic horizons. Two refractors with apparent velocities greater than 4 kms^{-1} are observed (sh. 26, 4.1 kms^{-1} and sh. 38, 4.5 kms^{-1}). It is less likely that these can be attributed to Mesozoic horizons; more likely they are from L.O.R. sediments or the underlying basement, the composition of which is in doubt.

Evidence from sonobuoy line 9/77 40 (Fig. 3.40) indicates the presence of approx. 300 m of recent sediment. The base of the soft sediments can be followed to the limit of the record (500 ms) on some of the sparker lines crossing the trough

(Fig. 3.61) indicating a thickness of $\gtrsim$ 400 m sediment at that point.

The recent sediment is underlain by approximately 270 m of Mesozoic sediments. A range of apparent velocities, 4.2 kms^{-1} - 6.4 kms^{-1}, are observed (Lines 9/77 40 & 9/77 150 Figs. 3.40 & 3.41) from horizons below the Mesozoic strata. The composition of the basement rocks which gives rise to these arrivals is problematical. One possibility being that the lower velocities are from shattered basement and the higher velocities from less shattered basement which may be Lewisian, Moinian or Torridonian. Alternatively L.O.R. sediments may overlie the basement.

The seismic character attributed to recent sediment, i.e. that of close evenly spaced parallel horizontal or gently undulating reflections (Fig. 3.61) is constant over most of the area. However, west of 06°26' a horizon of greater amplitude and containing many hyperbolic reflections is observed (Fig. 3.62). Rockhead is not observed below this horizon which may be the surface of a boulder clay layer.

Southeast of Dubh Artach, poorly defined rockhead can be observed breaking through the blanketing horizon in the recent sediment described above (Fig. 3.63). On the corresponding magnetic records low amplitude ($\approx$ 200 nT) moderate wavelength anomalies are observed. The character of these anomalies is unlike those associated with Tertiary or L.O.R. lavas.

It is considered likely that this is a dolerite sill similar to Dubh Artach.

<u>Summary</u>: This area is dominated by a major depression in rockhead (> 300 m) which trends 040° immediately south of Loch Buie swinging round to 060° south of Malcolm's Point (Fig. 3.3). The trough contains a substantial fill of recent and Mesozoic sediments. The course of the trough is believed

to follow the offshore extension of the Great Glen Fault zone. The zig-zag shape of the flanks of the trough can be explained by two sets of en-echelon faults, one with a north-north-east trend exploiting the pre-existing Caledonian grain; the other with a more east-west orientation was probably initiated with the commencement of movements along the Great Glen Fault zone. As the orientation of one of these fault sets (north-north-east trending set) is inherited inferences regarding the sense of displacement along the fault zone cannot be drawn from geometric relationships between these fault sets and the main fault zone.

The interpretation of en-echelon faulting is based on geometric considerations. Unfortunately the seismic data available is limited as the depth of penetration is insufficient to permit the contact between Torridonian and Mesozoic sediments to be viewed, thereby preventing visual confirmation of its nature.

The absence of very steeply dipping (greater than 70^{o}) flanks to the trough as would be expected of fault planes along which major horizontal displacement has taken place is probably due to erosion of the fault scarps.

Down-faulting during and/or post Mesozoic times has caused the Mesozoic interval to be depressed in the trough, though differential erosion may account for all or part of the lowering of the Mesozoic surface.

The following is a sequence of events which may have led to the present day configuration.

1) Major sinistral strike-slip movement was taken up along a number of en-echelon faults orientated parallel to the existing Caledonian lines of weakness. The timing of this movement is uncertain but is believed to have taken place during Lower Old Red sandstone times.

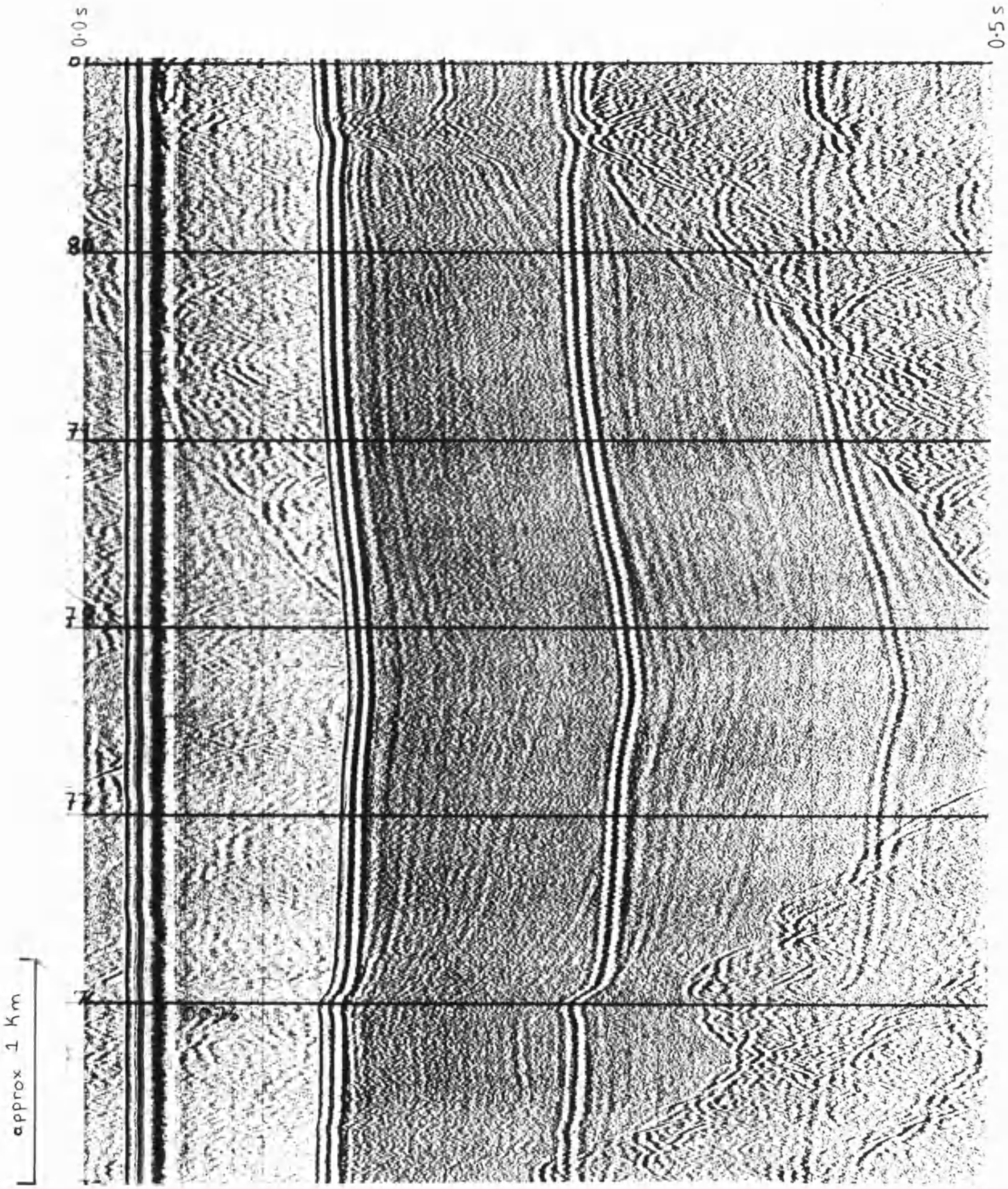


Fig 3.61 Thick recent sediment

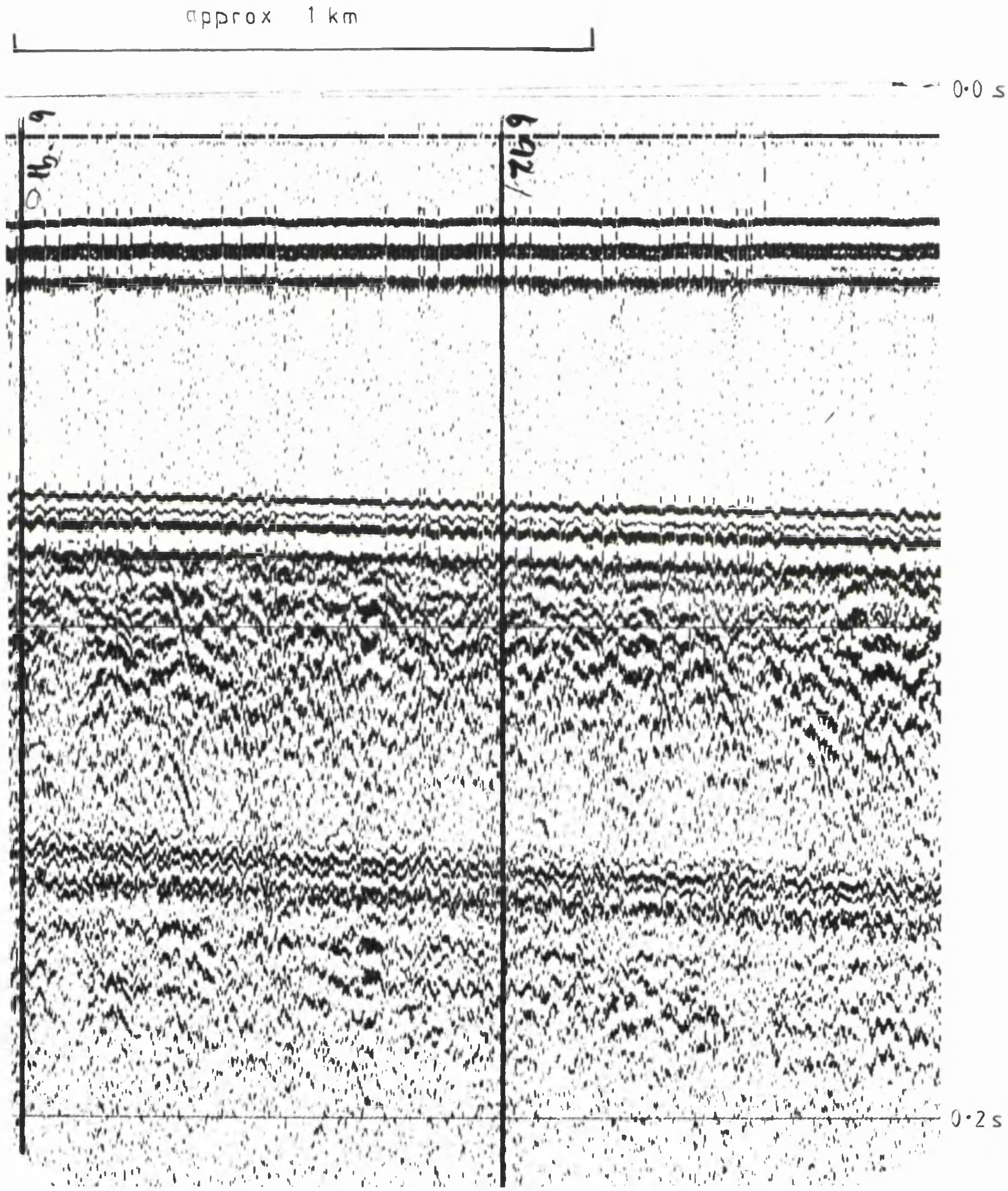


Fig 3.62 Sparker profile of probable boulder clay

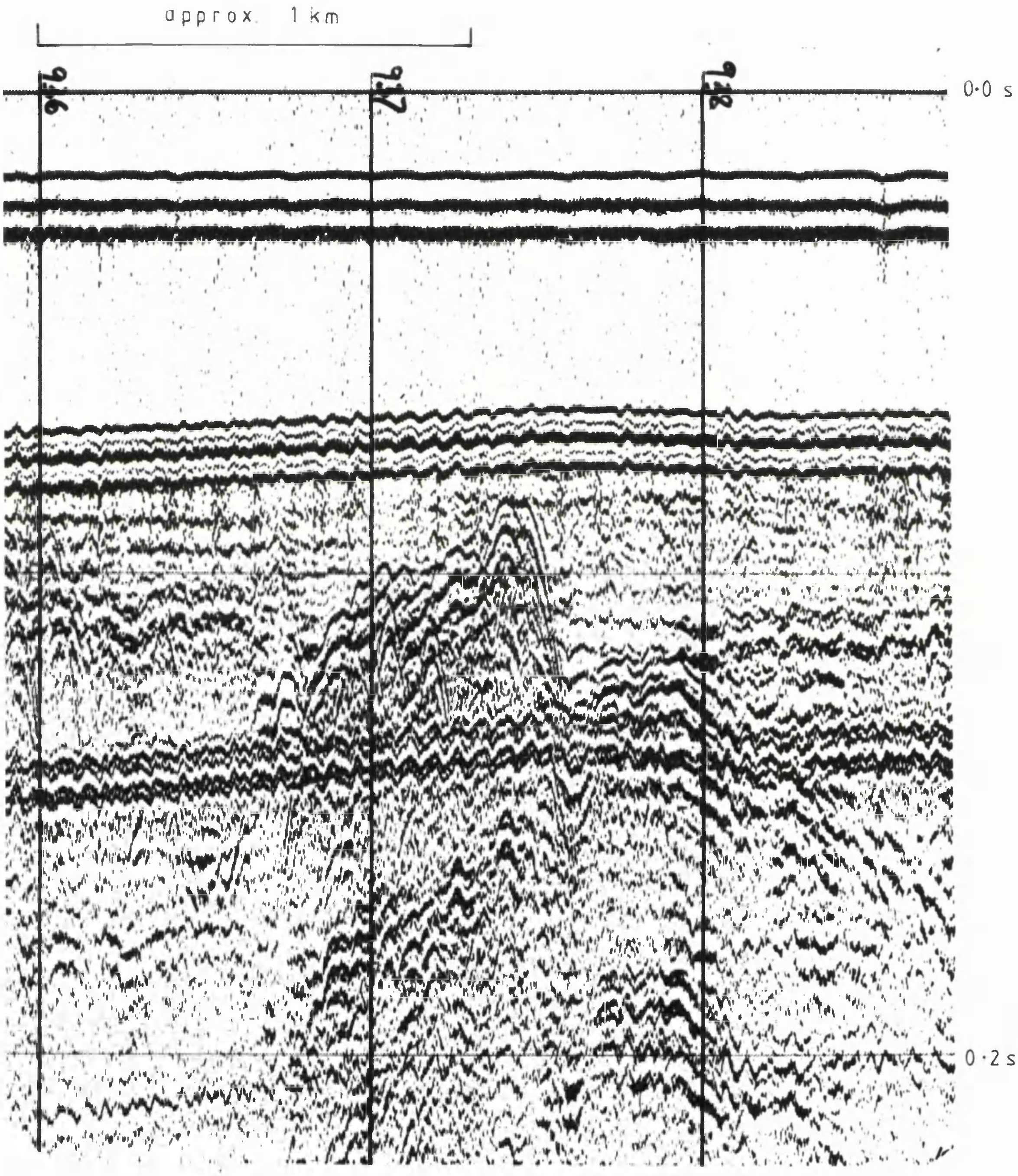


Fig 3.63 Sparker profile of probable dolerite sill

2) Preferential erosion of the fault zone to produce a trough.

3) Widespread deposition of sediment during the Mesozoic smoothing over the steep faulted margins of the trough, accompanied by downwarping - though this may only have been sufficient to cause flexure of the still soft Mesozoic sediments.

4) Deposition of Tertiary sediments.

5) Glacial erosion may have re-excavated the trough.

6) Deposition of post glacial sediments to produce a relatively smooth sea bed.

3.3.4 Area 4, North of the Ross of Mull

In this area pronounced bathymetric scarps are formed by the edges of Tertiary basalt outcrops. The Tertiary basalt is recognised by its distinctive seismic and magnetic character (Table 3.2). These scarps rise approx. 100 m above the seabed (Fig. 3.6) which is usually underlain by a varying amount of recent sediment overlying Mesozoic strata. Apparent dips of 2-6° are observed within the Mesozoic strata. Apparent dips of 1° and 3° are measured along reflections within the Tertiary lavas, i.e. from the surface of lava flows (at N56°12', W06°32'). Sparker and magnetic records show that all the shoals (T, U, V Fig. 3.58) north of 56°10' are formed from Tertiary basalts.

South of 56°10' the character of the sparker and magnetic records crossing shoal areas is unlike that attributed to Tertiary Basalts. In this area several different combinations of seismic and magnetic character are observed. Moine gneiss and granite are recognised by rough surfaces and are distinguished, one from the other, by the greater frequency of the irregularities in the Moine rocks and also by extrapolation from shore geology. The Lewisian outcrop is associated with a magnetic high, also observed on the aeromagnetic map over Iona (I.G.S. Aero Mag. Sh.10, Fig. 3.64)

To the west of the Lewisian outcrop further small outcrops

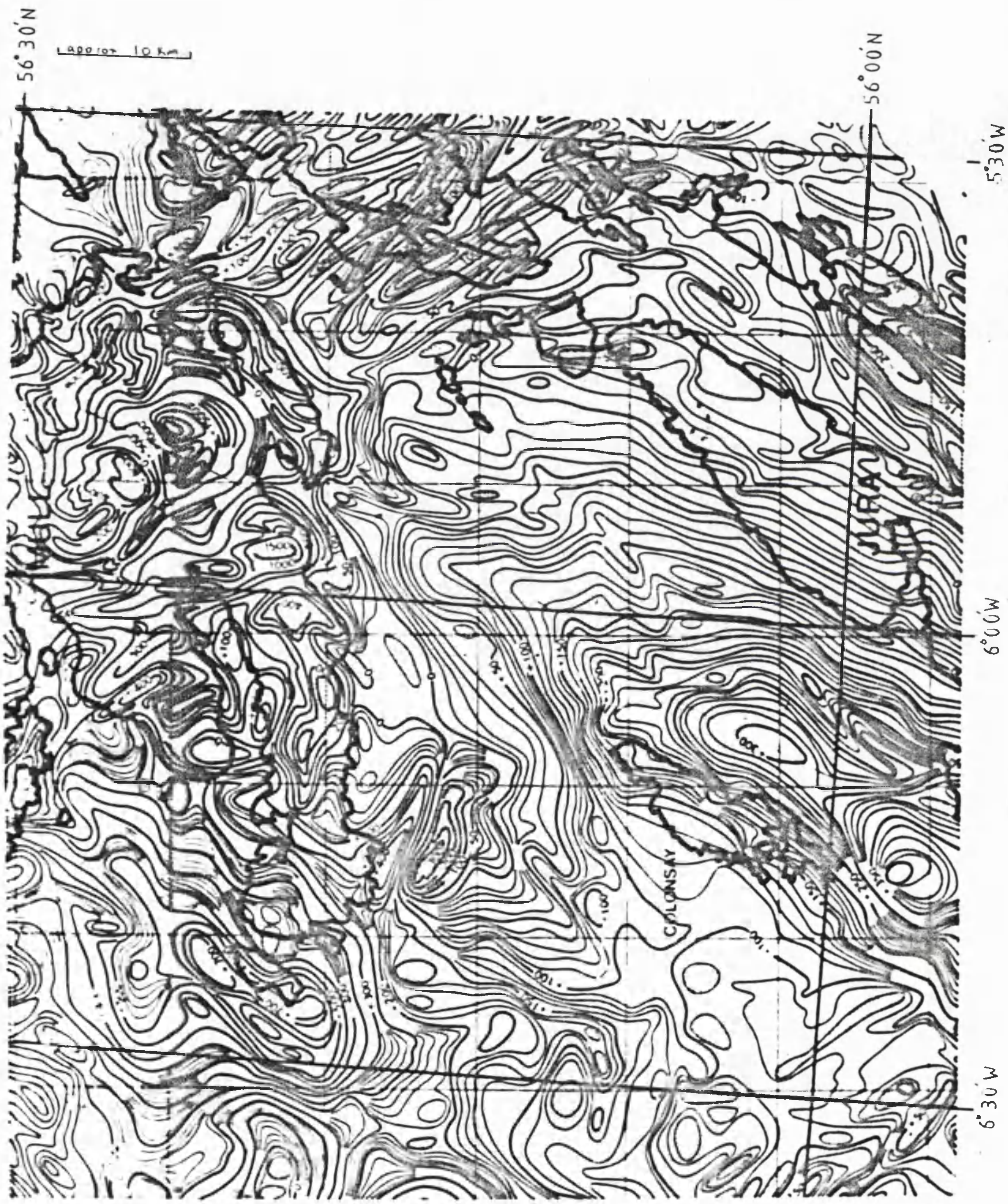


Fig 3.64 Part of IGS sh. 10 Aeromagnetic map

of Tertiary basalt occurs either as small islands or as submarine outcrops.

North of the Lewisian outcrop the magnetic high can be traced trending northeast. It is considered that this marks the trend of Lewisian rocks lying at shallow depth overlain by Mesozoic strata. The high can be seen to decline eastwards with a fairly constant gradient (Fig. 3.64) indicating the increasing depth to the Lewisian surface eastwards. Over much of this area Moine rocks have been shown to form the seabed.

Generally the Moine Thrust separates Moine rocks in the east from Torridonian rocks in the west. Torridonian rocks are found on Iona, suggesting that the Moine Thrust must lie east of this outcrop. It has been suggested that it runs through the Sound of Iona (Binns et al.1973, Barber et al. 1979). This line lies approx. 2 km east of the magnetic high suggesting that the northward extension of the Moine Thrust follows the trend of the magnetic maxima but is displaced eastwards. This is because the magnetic high is attributed to a magnetised horizon within the Lewisian which is overlain by non-magnetised Lewisian and also Torridonian sediments. The magnetic gradient eastwards would therefore correlate with the Moine Thrust plane dipping eastwards beneath the Moine rocks.

The next occurrence northwards of Moine rocks is at the western approach to L. Sunart and on Ardnamurchan. The Moine Thrust must lie to the west of these outcrops which is not incompatible with the course proposed above.

3.3.5 Area 5, West of Lismore Island

The offshore extension of the Tertiary basalts around

Duart Castle is recognised on sparker and magnetic records. The basalt extends approximately 1 km offshore.

West of Lismore a bathymetric scarp trending SW-NE is observed (W-W' Fig. 3.58). This is considered to be the line of a fault separating Lismore limestone to the east from Mesozoic strata to the west. Evidence for Mesozoic strata comes from the sparker records on which apparent dips of approx. $10-15^{\circ}$ to the southeast are tentatively recognised. Further support comes from the nature of rockhead, it being of a smoother nature than that of other candidates for the seafloor as observed elsewhere (Table 3.2).

The presence of Mesozoic strata at Inninmore Bay Morven ($56^{\circ}31'$N, $05^{\circ}42.5'$W) and on Mull separated from the submarine outcrop by a belt of Tertiary basalt would also support this hypothesis.

The Mesozoic outcrop terminates to the NW against the Morven granite -- this contact is not crossed by any of the sparker records.

The deep linear bathymetric depression which forms Loch Linnhe and is believed to mark the shatter zone of the Great Glen Fault is observed to extend across the Sound of Mull (Fig. 3.58) towards Duart Bay, indicating that the path of the Great Glen Fault follows this trend also.

3.4 Discussion

Several differences between this study and that of Barber et al. (1979) (Fig. 3.65) arise:

(1) The course of their (Barber et. al. 1979) Northern splay of the Firth of Lorne Fault is seen to trend discontinuously at 055° as far as 06°W, identified apparently on bathymetric evidence. Inspection of the detailed bathymetric map (Rashid

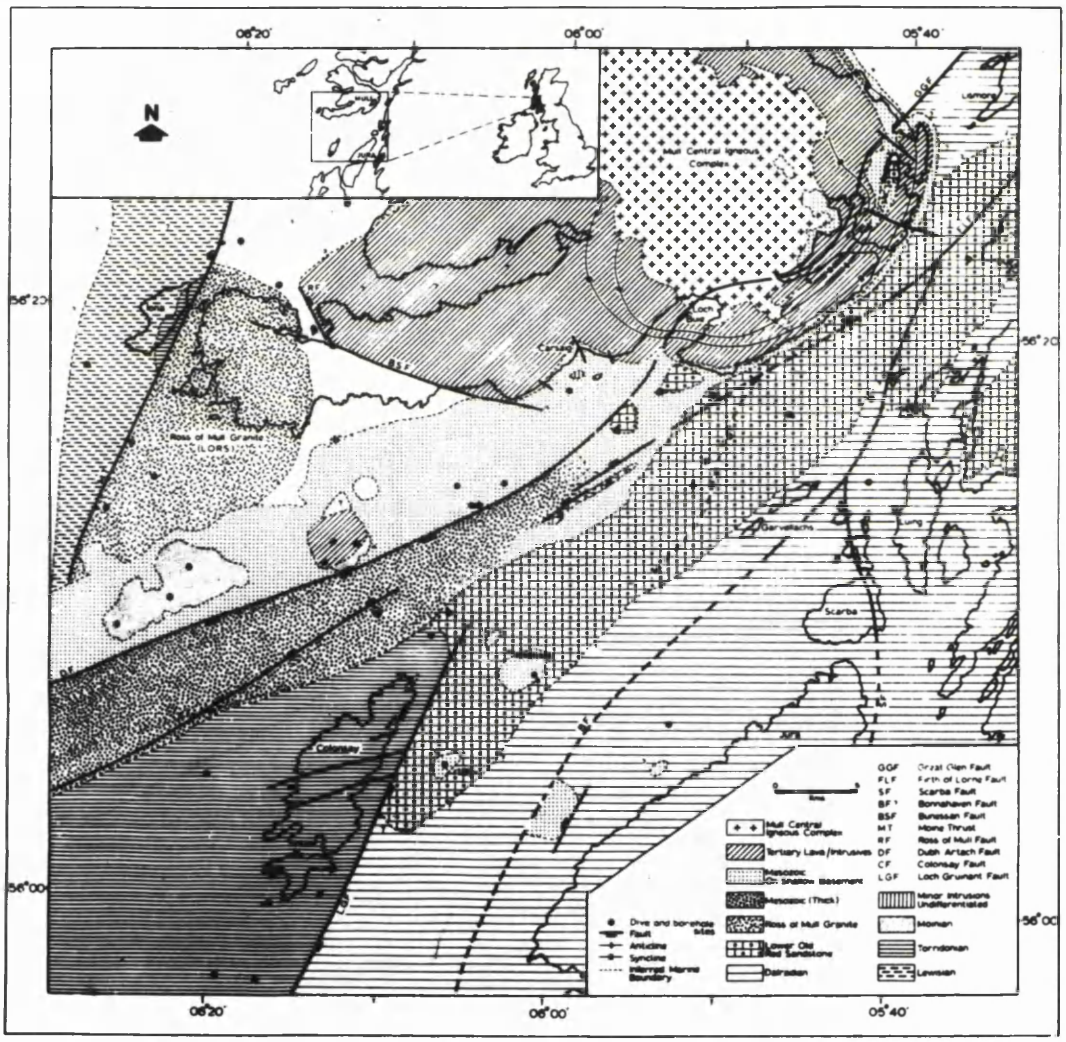


Fig 3.65 Geology of the Firth of Lorne (Barber et. al. 1979)

1978, Fig. 3.58) shows no correlatable bathymetric trend with this supposed fault line, rather the trend of faults (NNE - SSW and NE - SW) observed in the Firth of Lorne continues.

(2) The southwesterly extent of the LOR lavas and sediment. On the admitted tenuous evidence that erratics of L.O.R.S. aspect (Wright & Bailey 1911) found on Colonsay are derived from the seafloor immediately east of Colonsay, Barber et al. (1979) have continued the outcrop of these rocks as far south as 56°4'N. However, it is believed in this study that these rocks extend no further south than 56°10' (the southern tip of Scarba). The distinctive bathymetric character, that of subparallel deep troughs, associated with these rocks in the northeast, does not extend beyond the area of outcrop proposed here.

It is thought that much of the seabed in this area is underlain by Mesozoic strata. Support for this comes from three sources. (a) I.G.S. boreholes 71/9 and 73/25. Red sandstone was recovered from borehole 71/9 at 42 m below the seabed, the age quoted as either LOR or NRS in age (I.G.S. rep. 72/10). Red brown sandstone from 35 m below seabed was obtained from 73/25 which is thought to be NRS in age (I.G.S. rep. 74/7). (b) Apparent velocities (2.6 kms^{-1} - 3.8 kms^{-1}) consistent with Mesozoic strata have been recorded on sonobuoy lines 10/77, 6/77, 5/77 and 5X/77 (this study). (c) Rockhead observed on the many sparker lines traversing this area has a seismic character in keeping with that of Mesozoic strata observed elsewhere (Fig. 3.60).

(3) Acoustic type reflection A of Barber et al. (1979) forms a major part of the trough fill which is believed to be of Mesozoic age (Barber et al. 1979). The evidence for this is based on a dredge sample (3376 Barber et al. 1979) of well

bedded calcareous sandstone obtained from a site close to the southern margin of the trough where the acoustic character described above is observed. This acoustic unit is observed by Barber et al. (1979) to overlie topographically rough acoustic basement where it displays compaction folding and draping effects.

This type of seismic character is considered here to be indicative of Tertiary - recent sediments. Where Mesozoic rockhead is observed - characterised by low angle apparent dips ($3-15^{o}$) - deeper reflections from acoustic basement rock beneath the Mesozoic strata are not observed. In addition, acoustic basement of several types including Mesozoic can be seen to be overlain by such recent sediments.

That the sample collected by Barber et.al. (1979) is of Jurassic age is not doubted. Its correlation with a particular reflection character however is questioned. The use of dredge samples is always hazardous as sample location is uncertain and more importantly one cannot be sure that an 'in situ' sample is obtained. It is not inconceivable that pebbles of this age can be found on the seafloor in this area.

CHAPTER FOUR

DISCUSSION

4.1 The Great Glen Fault

4.1.1 The Great Glen Fault in the study area

The Great Glen Fault forms the well defined trough in rockhead SW from Loch Buie (Fig. 3.3). The morphology of this margin is similar to that of the Great Glen Fault on land. The width of the submarine trough is 2-3 km. The land trough is approximately 3 km wide. Both troughs have strikingly steep margins (11° - 26°).

On Mull the Tertiary lavas obscure the course. The Great Glen Fault is believed to follow the trend of L. Linnhe passing Duart Bay, L. Spelve where a massive thickness (1000m) of Triassic sediments have been recorded (Rast et al. 1968) to Loch Buie. Steel (1971) has suggested that deposition of the Triassic sediments in Loch Spelve was tectonically controlled by NW down-faulting on the Great Glen Fault. It is possible that the course of the Great Glen Fault may have been displaced by the forceful emplacement of the Mull centre (Phemister 1960): the Great Glen Fault may lie below the curved Loch Spelve.

For approximately 14 km southwest of Loch Buie the trend of the Great Glen Fault trough, 040°, is the same as the trend observed in Loch Linnhe. Beyond this point the trend abruptly changes to 060° as the trough continues SW. At the SW edge of of the study area the trough in rockhead is obscured by a cover of more than 50 m of recent sediment.

The margins of the trough are sigmoidal, and are attributed to a series of en echelon faults (Fig. 3.2) inclined obliquely to the main trend of the trough. The orientation of these en echelon faults is consistent with the orientation of gash-fractures (Riedel 1929, De Sitter 1956) caused by sinistral displacement. It is possible these faults may be exploiting pre-existing weaknesses in the basement rocks. A similar pattern of faulting has been observed on the Alpine Fault (Bishop 1968).

A considerable thickness (approx 300 m) of soft sediment, probably post-glacial, fills the trough (100 m at Inverness Horne & Hinxman 1914) so that there is no expression on the seabed. A similar thickness of Mesozoic strata lies beneath the soft sediment. The Mesozoic strata overlies a basement with apparent velocities of uncertain affinities. If as on land the Great Glen Fault is marked by a zone of shattered basement rocks then the velocities observed will be less than those of unaffected rocks. Alternatively, LOR sediments may underlie the Mesozoic strata.

There appears to have been some vertical movement as the surface of the Mesozoic strata is depressed compared to that of the N flank and is considerably younger than the rocks, Torridonian, of the S flank -- however, it is not known which horizon in the Mesozoic is forming rockhead. Evidence from Mesozoic surface alone is insufficient. But, as Mesozoic appears to form normal contacts with the Moine inliers, then the Moine is also displaced to a lower level in the trough.

The age of the fault is uncertain. Evidence from this area suggests that it is younger than the Moine Thrust which it offsets. The southern edge (which is dated at 400 m.y. Brown et al. 1968) of the Ross of Mull Granites terminates abruptly against the Great Glen Fault, indicating that it is older than the Fault. It has been proposed that the Fault formed a conduit for the Granite Magma (Barber et al. 1979).

However, if this was the case then the magma would have exploited the zone of weak rocks associated with the fault and the sharp edge with the fault would not be observed. This suggests then from the age of the granite that movement took place after 391 m.y. ago.

The SW extension of the fault trough beyond the study area is problematical. At the western edge of the area the seismic character of the overlying sediment changes. The acoustic signal does not penetrate to rockhead. Hence the position of the trough is obscured.

A variety of courses has been proposed for the Great Glen Fault across the continental shelf (Riddihough 1964, Ahmad 1967, see Fig. 1.10). It is possible that all these faults and others as yet undiscovered may be correct and that they form a splay system marking the termination of the fault, each fault taking up only a small fraction of the total movement on the fault.

4.1.2 Related faults in the study area

The course of several other faults has been mapped in the area (Fig. 3.2) most of which are thought to be related to the Great Glen Fault.

Several faults trending at approximately 030° are believed to lie between Colonsay and Jura. The most westerly of these runs close to the coast of Colonsay and is considered to be the submarine extension of the Gruinart Fault on Islay (Durrance 1976, Westbrook & Borraidale, 1978; Dobson & Evans 1974; Barber et al. 1979). The nature of movement on this fault is problematical, vertical movement of at least 1 km SE has been proposed and the possibility of minor horizontal movement suggested (Durrance 1976; Westbrook & Borradaile 1978). The possibility of up to 4 km downthrow to the SE, in two stages, has been proposed (Westbrook & Borradale 1978; Durrance 1976; Dobson et al. 1975). Such a movement would displace the trace of the Moine Thrust (assuming a dip of 20°) 11 km westwards.

A pronounced trough in the seabed lying to the east of the Garvellachs extends to the southwest and is believed to be due to a fault which has a similar trend to the L. Gruinart Fault. As sinistral faults offset the Garvellachs and run (trending almost N-S) into the main trough (Spencer (1971) has observed similar faults on the Garvellachs) it is possible sinistral movement may also have occurred on this fault. A fault is also inferred to exist off the west coast of Jura bringing Mesozoic strata into contact with Jura quartzite.

In the Firth of Lorne a set of NE-SW and NNE-SSW faults affect LOR sediments and lavas. These faults are offset by NW-SE faults related to the Mull centre; faults H-H' and I-I' form margins to a small submarine outcrop of Tertiary lava indicating that they are of post Tertiary lava age. It is not possible to determine the nature of movement on the NE-SW and NNE-SW faults. Similar-trending faults observed on Kerrera have both normal and strike-slip movements (Lee & Bailey 1925, Antia pers. comm.).

4.1.3 Significance of the bend

On land the trend of the Great Glen Fault changes from 035° in NE to 040° at Loch Linnhe; a change in trend of 5° over about 160 km. In contrast south of Loch Buie the trend changes abruptly by 20° (from 040° to 060°) over 4 km. The position and abruptness of the bend renders it unlikely to be a superimposed feature attributable to the forceful emplacement of the Mull centre. It is more likely to be an earlier feature of the fault.

It is possible that the fault trend was deflected by the immense thickness ($\approx$4 km) of very competent Jura Quartzite. This deflection resulted in movement being taken up in several splays running between Colonsay and Jura and also on the faults in the Firth of Lorne.

4.1.4 The Great Glen Fault in relation to the other faults of Scotland with a similar trend

Scotland is dominated by faults of approximately NE-SW trend (Fig. 1.10), e.g. Great Glen Fault, Highland Boundary Fault, and the Southern Uplands/Pentland Faults.

Between the Highland Boundary and Southern Upland Faults lies the Midland Valley containing a set of faults with a trend a few degrees east of northeast. Some of these faults are known to have been active during the Carboniferous (Anderson 1951). Movement on these faults is essentially vertical, though of a complex nature.

North of the Highland Boundary Fault a set of faults with rather steeper trends dissect the Highlands. These faults are believed to be older than those of the Midland Valley (Anderson 1951). Sinistral displacement has been described for several of these faults, e.g. Ericht-Laidon Fault (H. H. Read 1923) and Strathconnan Fault (Clough 1910, Hinxman & Crampton 1914) and by implication is thought to apply to the remainder whose sense of movement remains undetermined. The age of the rocks in this area is generally older than those of the Midland Valley being Dalradian and Moinian.

The Great Glen Fault appears to be the principal member of this group of faults. Those faults closest to the Great Glen Fault are almost parallel to it whilst the more southerly fractures have a more NNE trend.

That all these faults have a similar trend would suggest they are the products of the same or similar stress systems.

4.2 Other strike-slip faults

The two faults described in this section were chosen because:-

(1) they are relatively young, active faults and their relationship to the tectonic elements causing the movement is clear; and

(2) they are extensively discussed in the literature (e.g. Phemister 1960; Crowell 1962; Dickenson & Granz 1968; Kovach & Nur 1973; Suggate 1963; Wellman 1964; Bishop 1968).

4.2.1 The San Andreas Fault system (Fig. 4.1)

The San Andreas Fault extends some 1200 km from Cape Mendocino in the northwest to the Salton Trough in the southeast. It is an active dextral strike-slip fault along which a total of 600 km of movement has taken place (Crowell 1979). Other active dextral faults are associated with the San Andreas Fault (e.g. Hayward, Cal averas, San Jacinto and Elsinor faults).

Geological evidence suggests that two phases of movement have occurred. The most recent phase displaced Miocene strata by about 300 km in Central California. The earlier phase of movement believed to have taken place during Late Cretaceous is evidenced by the 600 km offset of the western edge of the sierran basement from southern Great Valley to the Point Arence region and also by the offset of some Palaeocene-Cretaceous sedimentary facies.

The San Andreas Fault is believed to be a transform fault

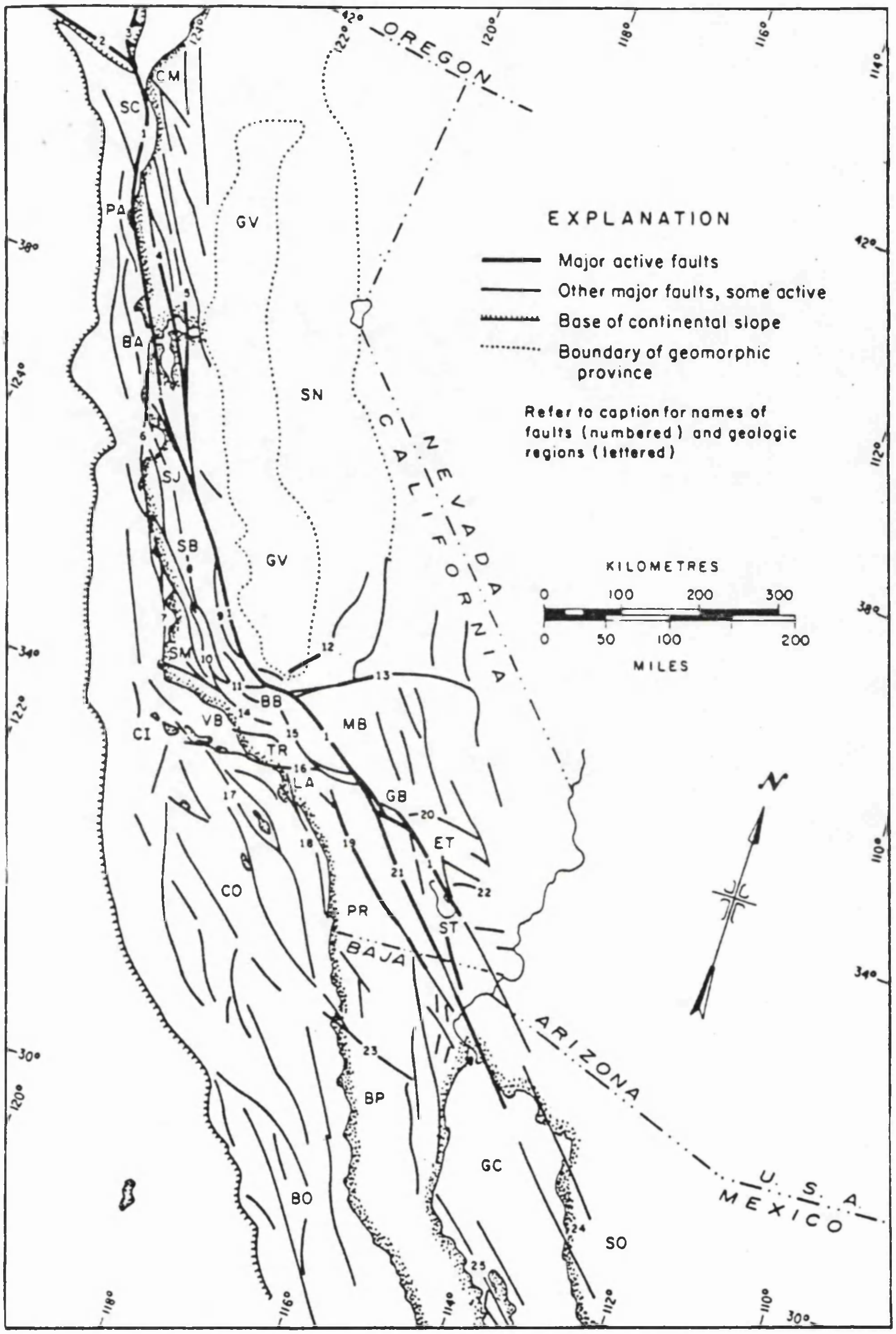


FIG 4.1 Simplified fault map of coastal California and northwesternmost Mexico. CM = Cape Mendocino, SC = Shelter Cove, PA = Point Arena, GV = Great Valley, BA = San Francisco Bay area, SN = Sierra Nevada, SJ = San Juan Bautista Bend, SB = Salinian Block, SM = Santa Maria Basin, BB = Big Bend of San Andreas Fault, SB = Ventura Basin, MB = Mojave Block, CI = Channel Islands, TR = Transverse Ranges, LA = Los Angeles Basin, GB = San Gorgonio Bend, CO = Southern California offshore borderland, ET = Eastern Transverse Ranges, PP. = Peninsular Ranges, ST = Salton Trough, BP = Baja California Peninsula, GC = Gulf of California, BO = Baja California offshore borderland, SO = Sonora. *Faults*: 1 = San Andreas, 2 = Mendocino fracture zone, 3 = Oregon subduction zone, 4 = Hayward, 5 = Calaveras, 6 = San Gregorio, 7 = Hosgri, 8 = Rinconada, 9 = Red Hills–San Juan–Chimeneas, 10 = Nacimiento, 11 = Big Pine, 12 = White Wolf–Kern, 13 = Garlock, 14 = Santa Ynez, 15 = San Gabriel, 16 = Santa Monica–Malibu Coastal, 17 = East Santa Cruz Basin, 18 = Newport–Inglewood, 19 = Elsinore, 20 = Pinto, 21 = San Jacinto, 22 = Salton Creek, 23 = Agua Blanca, 24 = Sonoran faults, 25 = A Gulf of California transform fault. Base from King (1969).

from Crowell (1979)

connecting the Juan de Fuca ridge to the East Pacific Rise (Wilson 1965). However, the tectonic history of this region has been shown to be complex (Atwater 1970). Magnetic evidence suggests that a trench existed offshore from North America during mid Tertiary and that the present episode of strike-slip motion in the San Andreas Fault system originated after subduction stopped, not earlier than 30 m.y. ago.

Movement takes place easily on the straight sections of the fault, stress being released by creep and many small earthquakes. At the bends stress is released by infrequent large earthquakes or the bend is bypassed by splay faults.

In the "Big Bend" region Cholome to Cajon pass the fault exists as a simple strand. The southern two-thirds of this section is oblique by as much as 30° to the Pacific-American plate slip. The Transverse Ranges which trend E-W oblique to the main structural grain of California were formed as a result of the oblique convergence requiring crustal shortening. North of the Transverse Ranges the shortening is taken up by folding and faulting on a complex system of nearly east-west striking right lateral strike-slip and thrust faults.

The "Big Bend" is thought to have developed during the Miocene in response to the expansion of the Basin and Range Province of Nevada and Utah (Scholz et al. 1973).

South of the "Big Bend" region the fault trace is once again nearly parallel to the regional slip vector and splits into several traces, so that the plate boundary in this region is defined by a broad shear zone of several hundred kilometres width.

A second bend is observed in the San Andreas Fault north of the "Big Bend" near the point where the Hayward-Calaveras fault branches off, believed to be another zone of convergence

(Scholz 1977).

4.2.2 The Alpine Fault, New Zealand (Fig. 4.2)

The Alpine Fault can be followed for approximately 1300 km across South Island from Milford Sound to Cook Straight. It is an active dextral fault along which about 500 km displacement has occurred as indicated by the offset of Upper Palaeozoic sediments (Suggate 1963).

The fault is believed to transfer motion between the Hikurangi trench in the north and the Puysegur Trench to the south (Fig. 4.3). This movement is shared between a complex system of dextral faults, Marlborough Fault system, branching away from the Alpine Fault in the north (Fig. 4.1). It has been suggested that these faults developed as a response to a change in the slip vector during early Miocene (Scholz et al. 1973).

In addition ot horizontal movements, considerable vertical movements ($\approx$20 km) elevating the Southern Alps has taken place (Suggate 1963).

4.2.3 Comparison of San Andreas Fault and Alpine Fault

The similarity between the San Andreas Fault and Alpine Fault systems have been noted by several authors (e.g. Hatherton, 1968; Scholz, 1973 + 1977)

In comparing the San Andreas and Alpine Fault systems Scholz (1977) suggests that the region of the Alpine Fault south of Arthurs Pass and the "Big Bend" area of the San Andreas Fault have a similar seismic style, i.e. long periods of quiescence interrupted by great (> 8) earthquakes. In both regions the fault is oblique to the regional slip vector.

The Marlborough system of faults of the Alpine system is compared to the part of the San Andreas system south of the

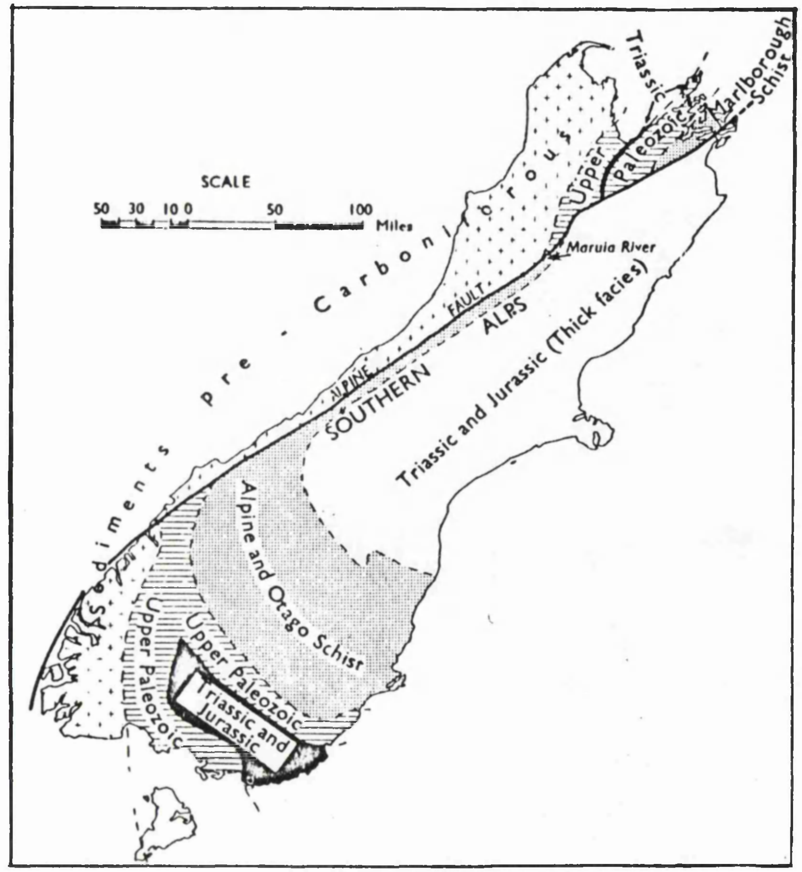


Fig. 4·2 Sketch map showing the Alpine fault and pre-Cretaceous rocks of South Island, reproduced from Wellman (1952). The 300-mile lateral shift is based on the correspondence of regional sequences on opposite sides of the Alpine fault in the north and south of the island. *(From Suggate, 1963.)*

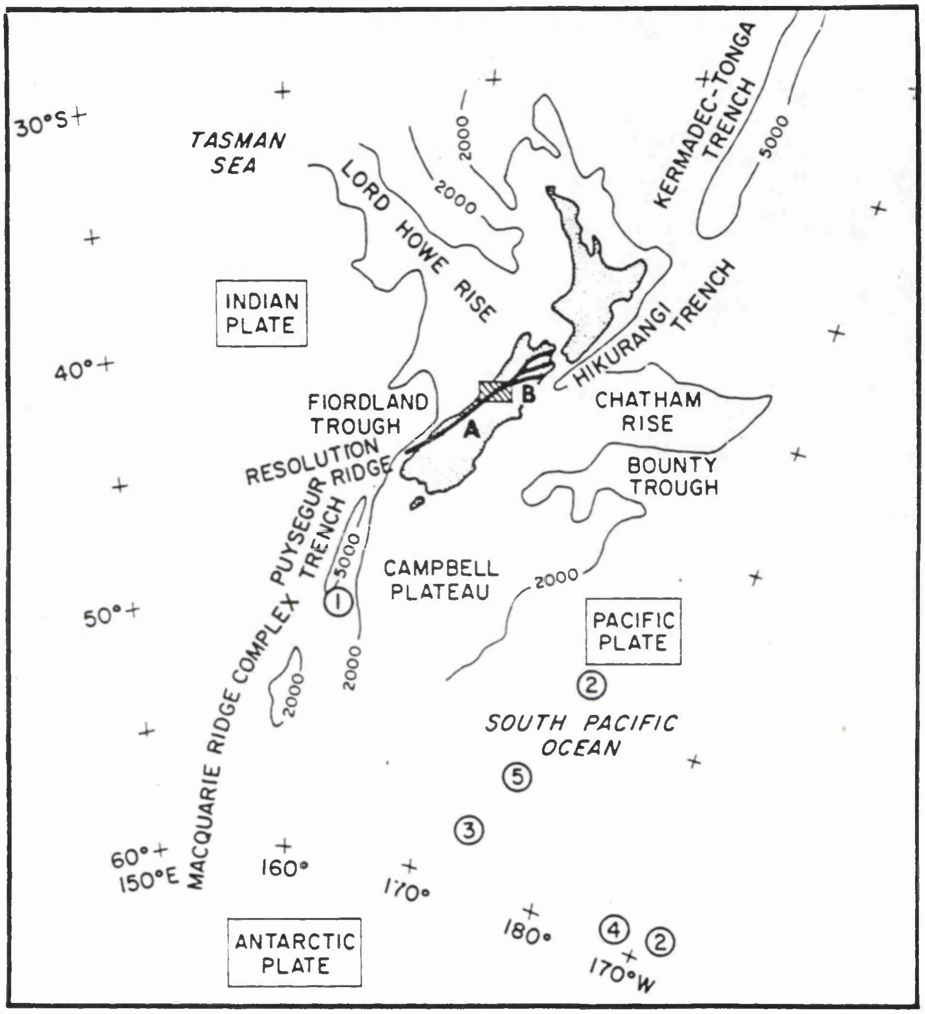


Figure 4·3 Regional tectonic setting of New Zealand. Dark rectangle is location of Arthur's Pass region. A = Alpine fault; B = Hope fault; C = Marlborough–East Coast fault system. Number in circles represent various computed positions of Indian-Pacific pole of rotation: 1, Le Pichon (1968); 2 (two positions), Christoffel (1971); 3, Chase (1972); 4, Weissel and Hayes (1972); 5, Falconer (1973).

from Rynn & Scholz (1978)

"Big Bend". In both these systems plate motion is divided between a group of faults approximately parallel to the regional slip vector.

By orienting the two systems so that their slip vectors were parallel (Fig. 4.4) Scholz (1973, 1977) demonstrated that the tectonic elements of the two systems occupy similar positions.

4.3 The Tectonic setting of the Great Glen Fault with reference to the San Andreas and Alpine Fault Systems

Whilst it can be said the Great Glen Fault and other faults of the system are still active on the basis of the seismic activity recorded in the last 100 years or so, it cannot be considered an active fault in the same terms as the San Andreas and Alpine Faults. The tectonic systems that were in existence during the formation of the Great Glen Fault are no longer operative.

Much of the evidence concerning the nature of tectonic situation prior to and during the formation of the Great Glen Fault has been obliterated. It is known that for much of the Precambrian and Lower Palaeozoic a significant ocean, the Iapetus, separated NW Europe from SE Europe. The orientation of the mid ocean ridge and various associated subduction zones is believed to have been approximately parallel to the NE direction of today. It is likely that the formation of the Great Glen Fault is related to the final suturing of the Iapetus. There is no evidence that the Great Glen Fault is a transform fault linking either trench or ridge segments as is observed at the ends of the Alpine and San Andreas Faults.

Some similarities between the Great Glen Fault, the San Andreas Fault and the Alpine Fault may be noted.

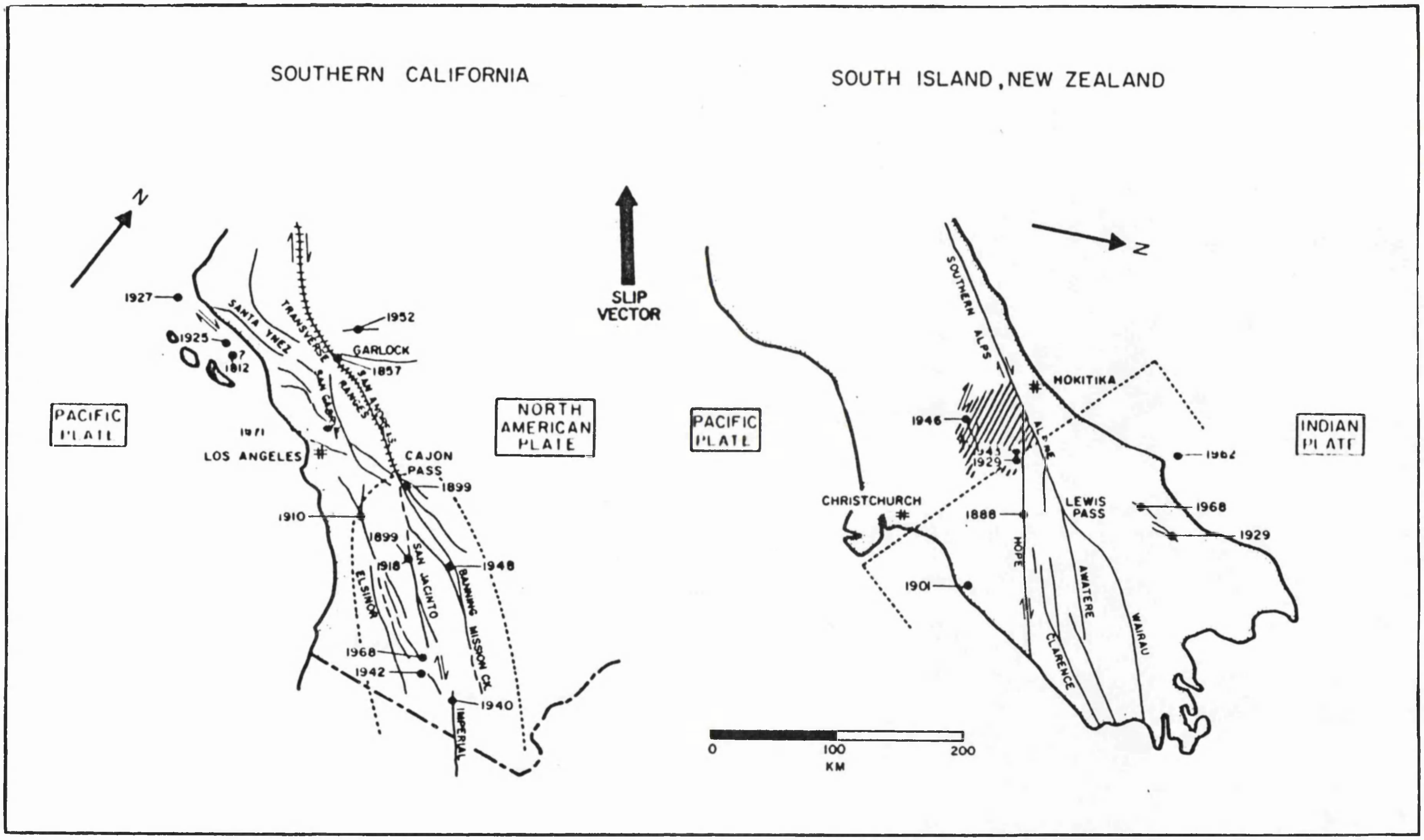


Figure 4.4 Comparison between northern South Island, New Zealand, and southern California. Maps are based on work by Scholz (1973). Fault patterns, major earthquakes with dates of occurrence, and extent of seismically active areas ("below" dotted lines) in both regions are taken from Allen (1965) for southern California and Evison (1971) for New Zealand. Maps are drawn to same scale and are rotated so that slip vectors for both regions are parallel. Hachured area in South Island indicates Arthur's Pass region. Dashed-line part of San Andreas fault indicates part of "locked" area. From Scholz et. al. (1973)

(1) All these faults formed close to plate margins, as segments of the two recent faults are parallel to and oblique to the regional slip vector. It is likely that one of the segments of the Great Glen Fault was parallel to the regional slip vector prevalent at the time of movement, the other oblique.

(2) They all have straight sections, bends, splay systems and associated parallel faults. Where systems of approximately parallel faults are observed on the San Andreas Fault and Alpine Fault the faults are oriented parallel to the regional slip vector. This would suggest that south of the bend in the Great Glen Fault the trend is parallel to the regional slip vector as this trend is approximately parallel to other faults in Scotland (some shown to be sinistral).

(3) All these faults appear to have formed close to subduction zones implying they are the result of convergence.

4.4 Later movements in the Great Glen Fault

Several authors have proposed more recent movements on the Great Glen Fault (Holgate 1969; Donovan et al. 1976; Chesher et al. 1975). Holgate & Donovan proposing later dextral movements (see ch.1). Chesher et al (1975) from deep seismic evidence obtained in the Moray Firth demonstrate that normal movements, downthrowing to the east, were taking place during the Mesozoic.

It may be that these later movements are the result of a different stress system to that which formed the Great Glen Fault. During the Mesozoic the North Mid Atlantic Ridge began to separate Europe from North America. Many sedimentary basins with faulted margins began to subside during this time also. The normal movements on the Great Glen Fault observed in the

Moray Firth may be the result of this new stress regime exploiting old weaknesses and fracture planes.

CHAPTER FIVE

CONCLUSIONS

(1) It has been demonstrated that the technique of using an anchored sonobuoy with a seabed hydrophone for seismic refraction studies produces better results than free floating sonobuoys previously employed (Faruquee 1972).

(2) Using facilities currently available from the NERC equipment pool, a 40 cu.in. airgun firing at 4s intervals produces the best combination of resolution and acoustic range of refracted arrivals from rockhead.

(3) The courses of the Moine Thrust and the Great Glen Fault which sinistrally offsets the Moine Thrust by 55 km have been traced over the area. It is possible the offset may originally have been 66 km and was reduced by 4 km downthrow on the Loch Gruinart Fault.

(4) A system of faults trending N 10° E and NE-SW related to the Great Glen Fault observed in the Firth of Lorne are described and interpreted as a shatter belt accommodating the change in direction of the Great Glen Fault.

(5) By comparison with the geometry of the San Andreas and Alpine Faults it can be inferred that the Great Glen Fault was formed in a similar tectonic setting, i.e. close to a continental plate margin and oriented parallel to the margin south of the bend south of Loch Buie and oblique to the margin north of the bend.

Comments on conclusions

Many factors contribute to the accuracy and reliability of the geological map (Fig.3.2).

The identification of specific rock units is dependent on correlation of onshore geology with geophysical properties - i.e. seismic and magnetic characters. Refraction velocities provide a further aid, though in spite of accurate determination they do not provide an unequivocal answer on their own as a given rock is not associated with a unique velocity but rather a range of velocities.

The identification of geological boundaries is dependent on there being a contrast in the physical properties of the adjacent rocks. Where there is a strong contrast and, or where more than one method is used to identify a boundary then its reliability is high e.g. the boundary between Mesozoic sediment and Tertiary lavas (e.g. around Staffa) can be recognised on sparker and magnetic records and also forms bathymetric scarps. Whereas the boundary between the Ross of Mull granite and the Moinian roof pendant (similar to the smaller example observed onshore near the northeastern margin of the granite) south of the Ross of Mull is of low reliability as it can be determined only by a subtle change in seismic character.

The actual position of the identified boundaries is dependent on the accuracy of navigation ($\pm$ 75 m) and on the frequency the boundary is crossed by survey lines. (Larger dots indicate where survey lines cross boundaries on Fig. 3.2).

Future research

It is unlikely that further work of the nature described here would yield additional information in this area, though extension into adjacent areas would.

What is required in this area is further sampling by a variety of methods.

1) Sampling of the shoals in the Firth of Lorne (Area 1) could be achieved by divers or possibly the use of a simple "over the side" drill such as the I.G.S. midi drill.

2) Modest drilling, similar to the I.G.S. MV Whitethorn work. (Chesher et al 1972), to a depth approximately 50 m could test and date the proposed Mesozoic outcrop in the area.

3) Deep Drilling. One well drilled to a depth of greater than 600 m in the Great Glen trough zone would confirm the thicknesses of recent and Mesozoic sediments and identify the underlying rock. A second well east of Colonsay and proposed location or Moine thrust subcrop would test for Mesozoic thickness and nature of underlying rock to test the proposed position for the Moine Thrust and thereby determining the sense of displacement of the Great Glen Fault.

Regretfully only method 1, is within the scope of academic research and that is limited to shoal areas with little or no recent sedimentary cover.

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