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Acoustic Observations

in

Seabed Materials.

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Alasdair G. McKay

Durham

December 1983



13 AUG 1984

Thesis
1983/MAC

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- a) Figures for seismogramme no. 1289.
(fig. 9.24 - fig. 9.44)
- b) Figures for seismogramme no. 1281.
(fig. 9.45 - fig. 9.73)
- c) Figures for seismogramme no. 496.
(fig. 9.76 - fig. 9.123)
- d) More figures for nos. 1289 and 1281.
(fig. 9.124 - fig. 9.131)

INCLUDED IN POCKET

1. OVERLAYS

Overlay for field seismogramme no. 1289.

Overlay for field seismogramme no. 1281.

Overlay for field seismogramme no. 496.

2. PUBLISHED PAPERS AND OTHER REPORTS.

McKay and Peacock (1970)

McKay (1974)

McKay (1980a)

McKay (1980b)

McKay and McKay (1982)

McKay (1983)

Howells and McKay (1977)

ABSTRACT

It is current practice to investigate surficial seafloor sediments by the acoustic technique of vertical-incidence reflexion profiling whether the investigation is done for the purposes of foundation engineering, of underwater sound propagation or pure sedimentological research. Such a technique gives no information about the variation with depth of such properties of the seabed materials as the acoustic velocity and attenuation. Gross changes in these properties can be expected if sediments are gasified or ice-bonded. Comprehensive reviews of what is known about these phenomena are given. Briefer summaries are included of the significance to the fields of engineering and of underwater acoustics, of the acoustic properties of seabed materials. The long-established seismic techniques for measuring velocities on the basis of travel time measurements can be applied to seabed sediments, but require adaptation, particularly in deeper water. The necessary amendments to field equipment and data-reduction technique are described, with results from areas of Maritime Canada and from Lake Erie. A description is given of equipment built to gather 12-channel acoustic data on a scale suitable for the investigation of surficial sediments and results are presented from the Beaufort Sea (Arctic Canada). Application of synthetic seismogramme modelling to these records shows that the acoustic technique when analysed in this way is capable of discovering properties of seabed sediments which are not revealed by either the vertical incidence profile or the travel-time methods of velocity analysis. In particular, a substantially different velocity profile results from this analysis. Suggestions are made for optimising the future gathering of field data for treatment in this way.

Note on spelling conventions

A university thesis remains one of the few places in which an author can retain a preferred spelling without editorial intervention. Accordingly, in this work, words derived from classical roots normally retain the root spelling. This gives rise to "refraction" from 'refrangere' but to "reflexion" from 'reflectere'. The form "seismogramme" suggests direct derivation from the Greek, with the picturesque meaning of "earthquake writing", rather than a term of undoubted technical utility but of questionable etymological hybridisation (which is indicated by the more common spelling).

1. INTRODUCTION.



The investigation of seabed sediments has been largely dependent on laboratory testing of core samples and on extrapolations from sampling sites by geological argument based on vertical incidence subbottom profiles.

It is difficult to take from the seabed an undisturbed sample suitable for conventional geotechnical testing, particularly in the case of coarse-grained sediments, and the application of laboratory measurements in assessing the performance under load of the bulk material must always be doubtful.

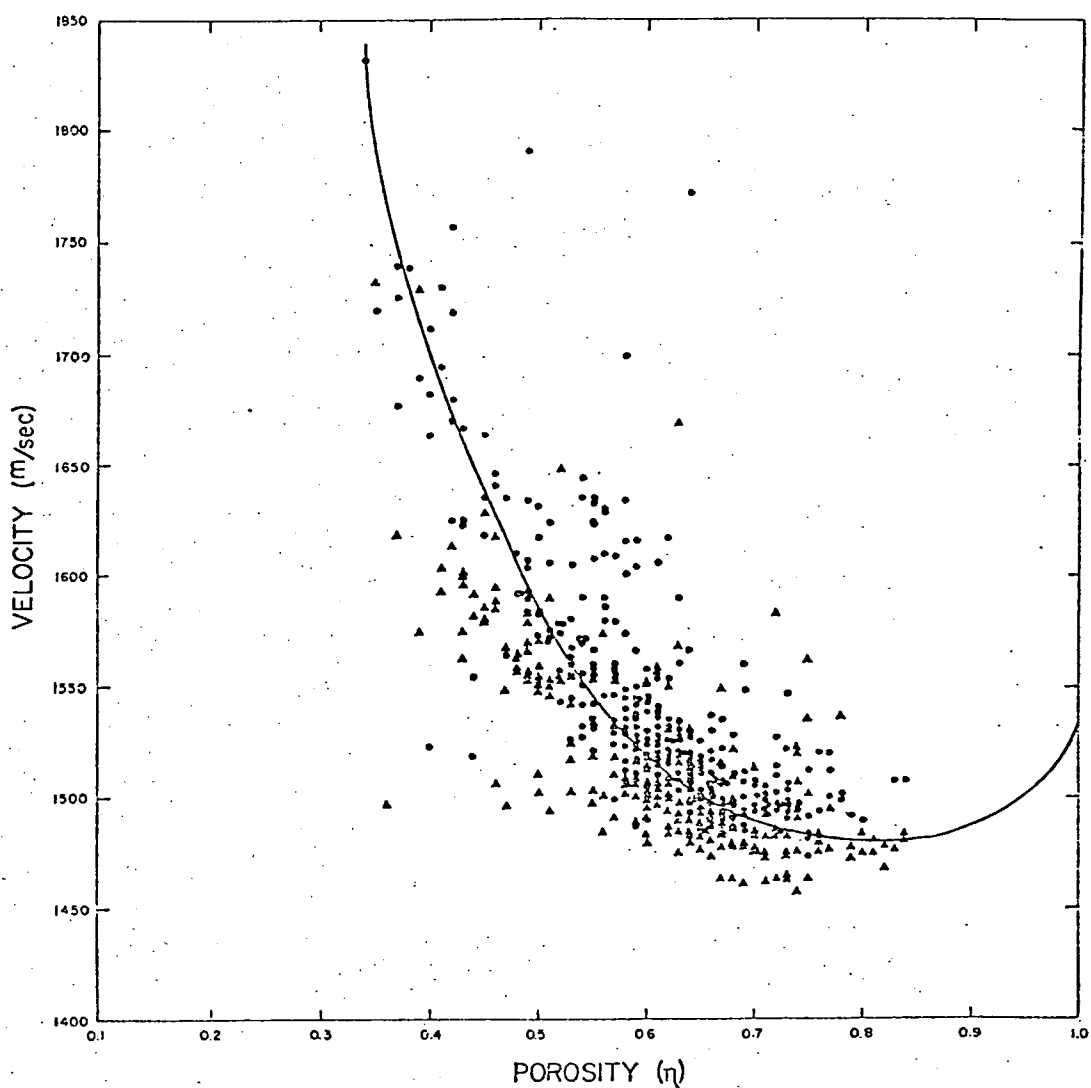
One approach to this problem has been to look for correlations between mechanical and acoustic properties of sediments, with the aim of predicting mechanical behaviour from acoustic parameters measured in the field. Sediment acoustic properties are also required in the field of underwater communications (see chapter 5). The simplest acoustic parameter to measure is the compressional-wave velocity. Empirical relationships have been established between velocity and such variables as density, grain-size and porosity.

Early work on sediment cores by Laughton (1954), Nafe and Drake (1957) and Shumway (1960) suggested that the application of Wood's (1931) emulsion equation (see section 3.4) to water-saturated sediments was appropriate,

although some scatter could be expected as shown in fig. 1.2 (Horn et al. 1968). Simpkin (1976) and Taylor Smith (1976) have assembled data from cores taken in the deep ocean (N. Atlantic) and for shelf sediments (fig. 1.3) and find better agreement with Wood's equation for the finer than for the coarser sediments.

Correlation of velocity with density has been discussed by Hamilton (1978) who notes problems with the generalised curves published by Nafe and Drake (1957, 1963) and instead presents a more complex set of relationships in which he classifies rock and sediment types and gives a curve for each of his classifications (fig. 1.4). Hamilton (1976a, 1979b) has also correlated compressional with shear-wave velocity in water-saturated sands (fig. 1.5). Simpkin (1976) has compiled data for both crystalline rocks and unconsolidated sediments and compared sound speed with the static Young's modulus (fig. 1.6). This shows a trend for surficial sediments which is different from that for crystalline rocks. Gross effects on compressional-wave velocity can arise from sediment gasification (chapter 3) and ice bonding (chapter 4).

Most of the relationships described above have been established on the basis of laboratory measurements on samples, and the importance of more field observations has



Porosity versus velocity. On this and subsequent figures data from the Norwegian basin are represented by triangles, data from the Mediterranean Sea by circles.

Fig. 1.2

Correlation between sediment porosity and acoustic velocity observed by Horn et al. (1968).

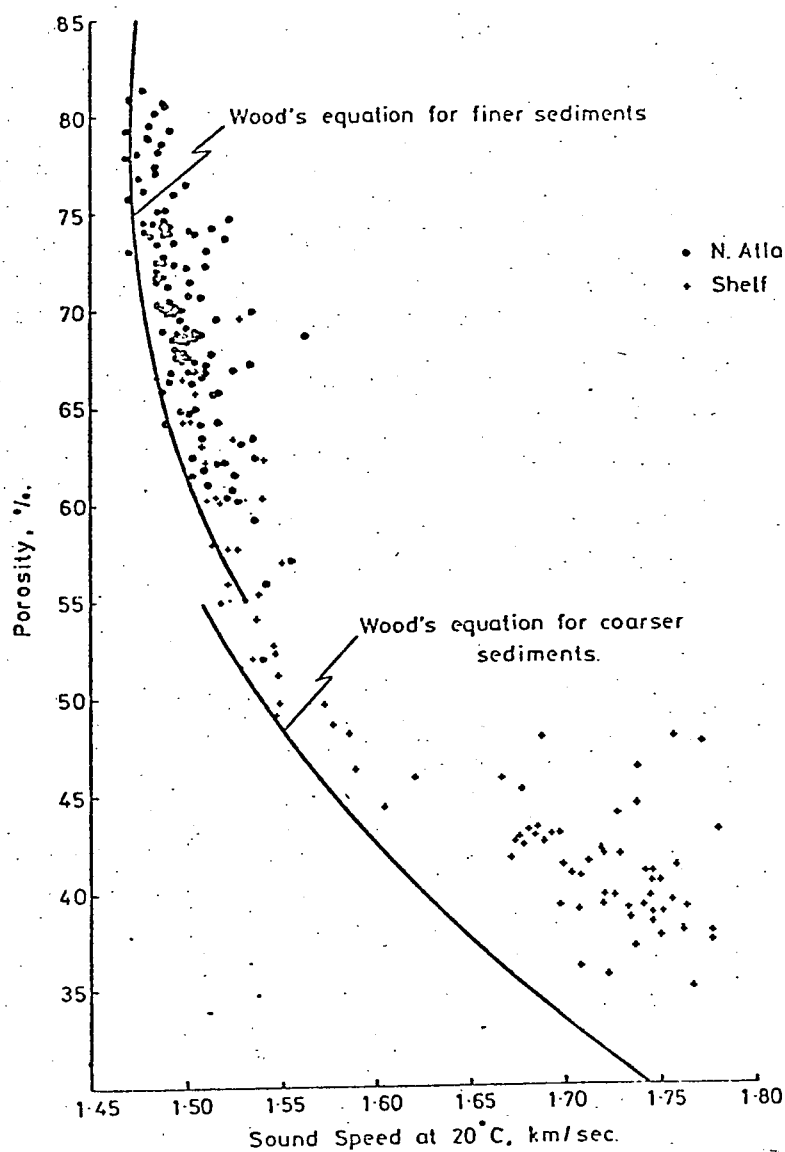
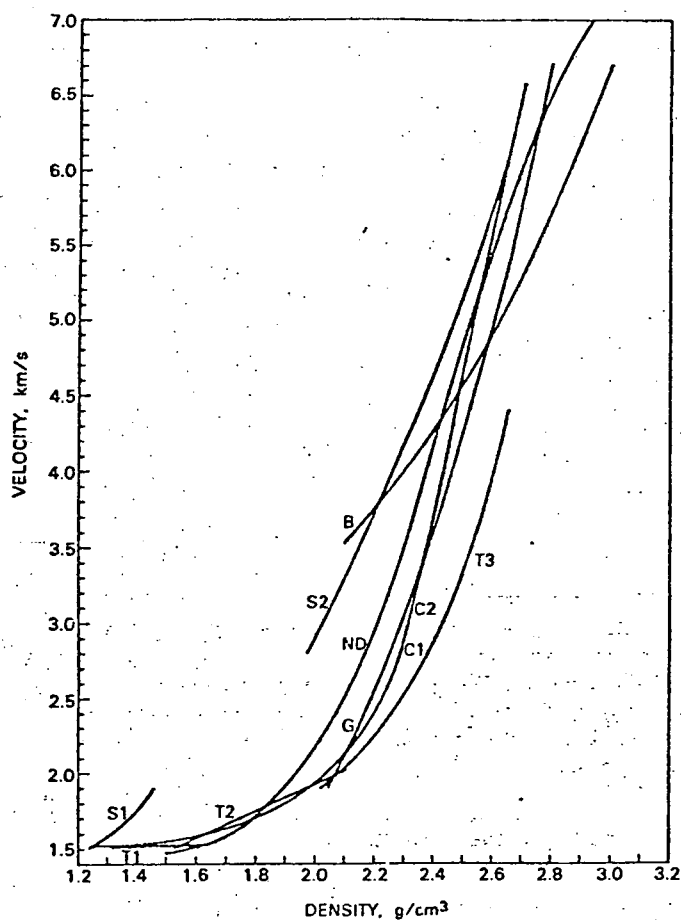


Fig. 1.3

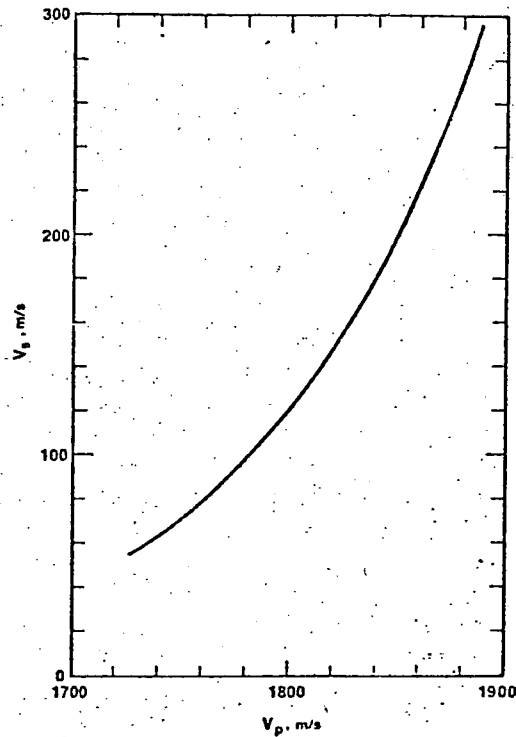
Correlations between sediment porosity and acoustic velocity compiled by Simpkin (1976) from Marine Science Lab. (North Wales) data.



A summary of compressional wave velocity versus density in various sediment and rock types of the sea floor, plus general curves from the literature. The letters beside the curves denote the sediment or rock types: S1—diatomaceous sediments (0–500 m); S2—siliceous rocks; T1—ter-rigenous surface sediments; T2—turbidites (0–500 m); T3—mudstones, shales; C1—chalk, limestone; C2—limestone; B—basalt. The general curves of Nafe and Drake⁸⁵ (labeled ND) and Gardner *et al.*⁸⁷ (labeled G) are included for comparisons. See Hamilton⁸⁴ for references, discussions, and regression equations.

Fig. 1.4

Hamilton's (1980) compilation of correlations between acoustic velocity and density for different sediments and rocks. The references he cites are: Nafe and Drake (1963), Gardner *et al.* (1974) and Hamilton (1978).



Compressional wave velocity, V_p , and shear wave velocity, V_s , in selected water-saturated sands. The regression equation is: $V_s = 21.05 - 24.617V_p + 7.215V_p^2$; where V_p and V_s are in km/s; see Hamilton for discussion.

Fig. 1.5

Hamilton's (1980) correlation between compressional and shear wave velocities in water-saturated sands. The reference cited in the above caption is Hamilton (1979b)

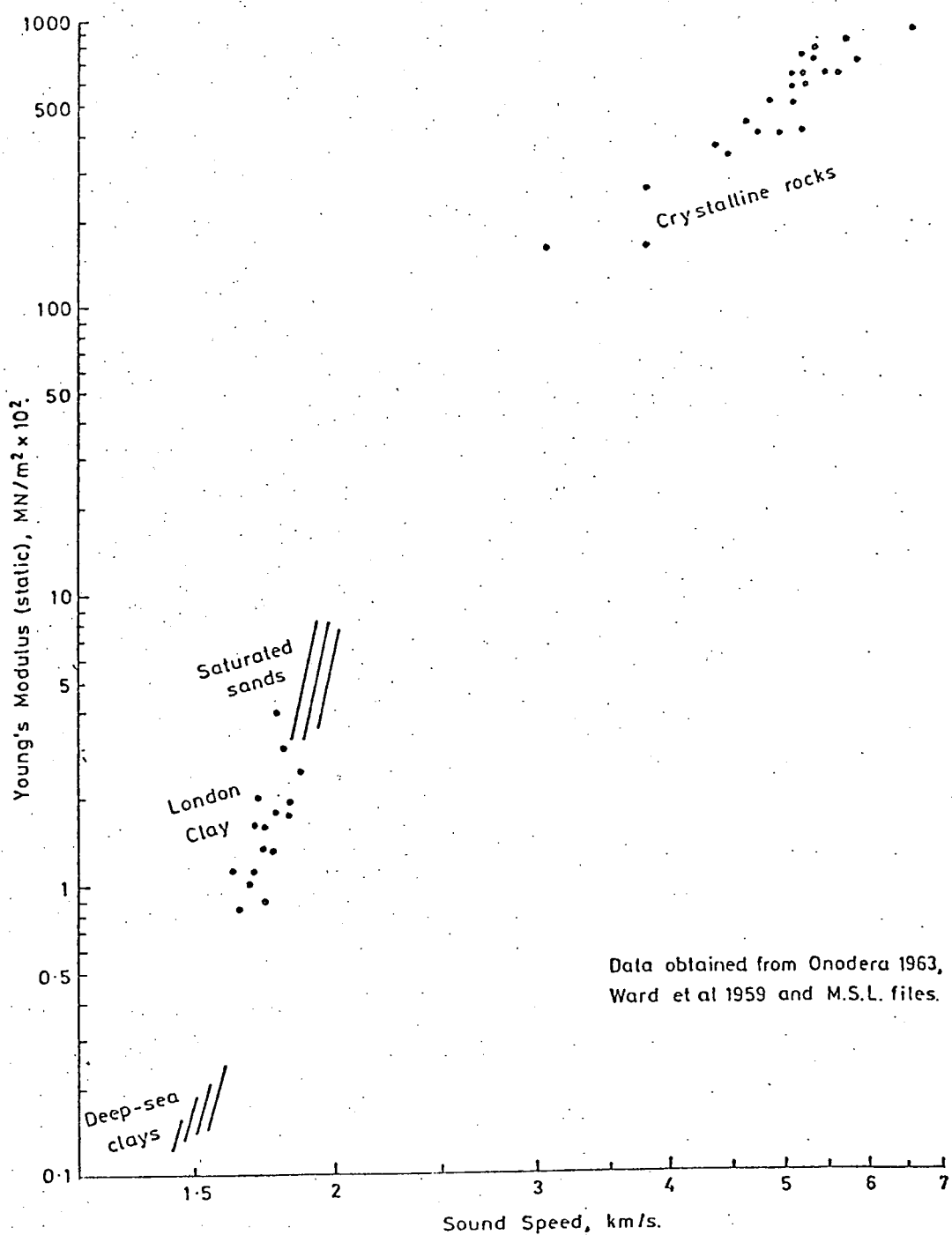


Fig. 1.6

Simpkin's (1976) compilation of correlations between acoustic velocity and the static Young's Modulus of various sediments and rocks from data in the archives of the Marine Science Labs. (North Wales) and from other sources.

often been pointed out (e.g. by Simpkin, 1976). Part of this study (chapter 7) is concerned with methods for the field measurement of compressional-wave velocities and with the results which have been obtained from different geological environments, to serve as an improvement over the conventional subbottom profiling method currently used for the investigation of surficial materials, and to allow, without interrupting conventional survey progress along pre-determined tracks, the possibility of adding to the knowledge of geoacoustical parameters (Hamilton, 1980) required for sound propagation studies (see section 5.5) or of gleaning geotechnical information through such correlations as shown in fig. 1.6 and such relationships as have been set up (Table 1.1; Murphy, 1974) to relate velocity directly to the requirements of construction engineering. The ability to make these measurements without interference with conventional profiling practice (sections 7.3, 7.4 and 7.5) is important. Engineers are rightly conservative in approaching construction problems and sedimentologists generally do not understand geophysical techniques. To pursue the gathering of field data suitable for the determination of velocity in the quantity that would be required before confident predictions could be made on its basis is unlikely to be accomplished by field practices which require dedicated ship time or station work.

Symbol	Velocity (ft./sec.)	Probable Excavation Characteristics	Anticipated Proportion of Normal* or Easy Excava- tion to Difficult or Hard Excavation
V ₁	6,000	Mostly easy or normal dredging	$\frac{100}{0}$
V ₂	6,500	Normal excavation with very little hard dredging	$\frac{95}{5}$ to $\frac{85}{15}$
V ₃	7,000	Little normal to some normal and some hard	$\frac{70}{30}$ to $\frac{65}{35}$
V ₄	7,500	Little normal and some hard or mostly hard	$\frac{35}{65}$
V ₅	8,000	Very hard dredging	$\frac{0}{100}$ to $\frac{5}{95}$
V ₅ -V ₆	8,000 to 10,000	Assume Blasting	

*"Normal" = Normal for Port Hedland for the dredging equipment used there (probably not normal for other geologic areas where there are thick deposits of loose alluvial materials).

NOTE: Dredge referred to is "The Alameda" of Utah Construction Corporation.

Table 1.1

Empirical relationships noted by Murphy (1974) between velocities he measured by the method of Porter et al. (1974) and excavation characteristics at Port Hedland in Western Australia.

Although the measurement of compressional-wave velocity is important in itself, further sediment parameters might be obtained if field seismogrammes of sufficient quality could be recorded to allow modelling of the field data by synthetic seismogramme techniques as is the current practice in analysing seismic data gathered for other purposes. Appropriate sediment models to use for this approach, together with the method of constructing the synthetic seismogrammes is given in chapter 8. Results of the modelling process applied to data from an area of ice-bonded sediments in the Beaufort Sea (chapter 9) appear consistent with laboratory observations of frozen sediments.

Maps which may be of help in locating specific areas referred to in various parts of the following text are given in figs. 1.7 and 1.8.



Fig. 1.7

Map of Canada.

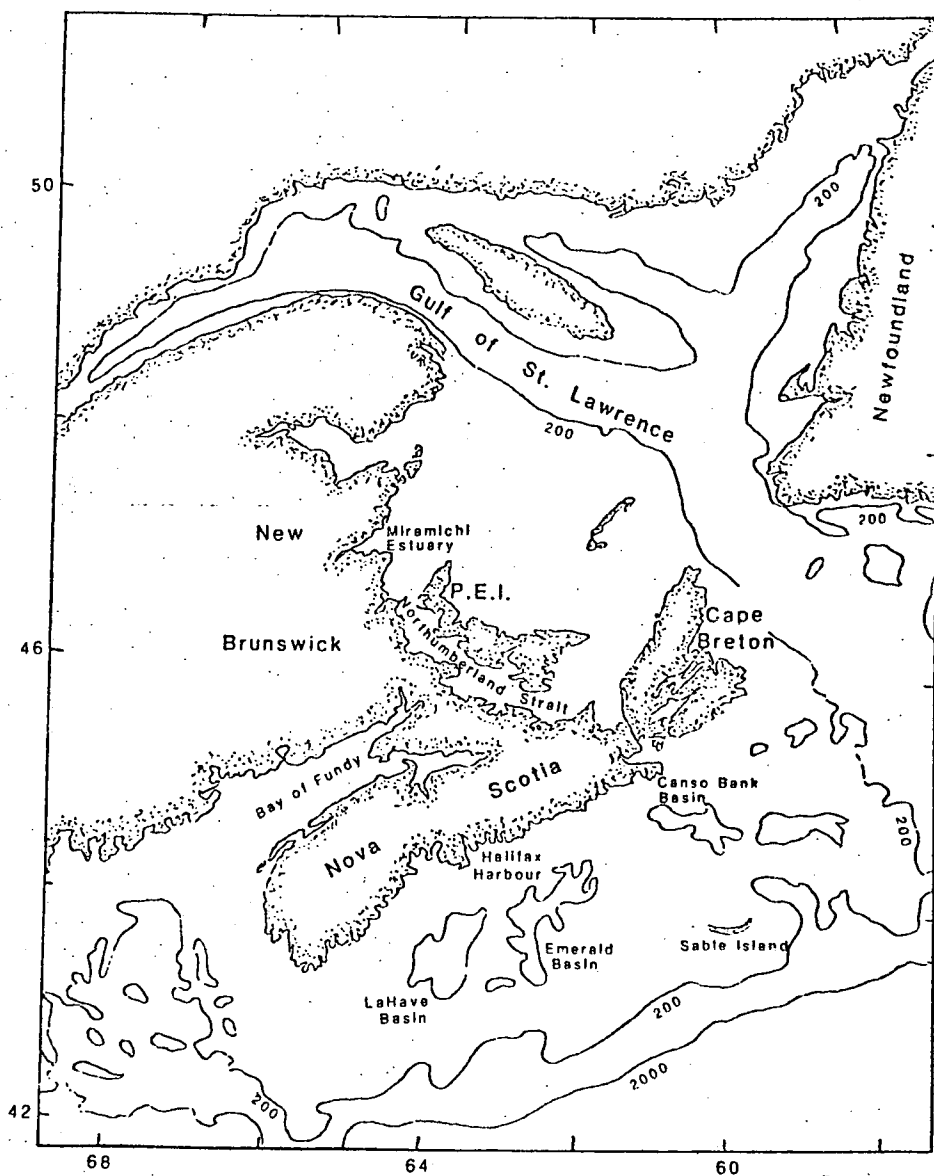


Fig. 1.8

Map of the Maritime Provinces.

2. CONSTRUCTION AND ADAPTATION OF FIELD EQUIPMENT.

2.0 INTRODUCTION.

Multichannel systems have long been commonplace for use in the conventional land and marine techniques of seismic structural exploration. Equipment comprising many hundreds of channels exists and forecasters such as Hewitt (1983) predict that the number of channels will soon reach thousands. Two factors help in making possible such elaborate equipment. One is economic and reflects the large amounts of effort which can be expended in geophysical work for delineating potential oil-bearing structures. The other is technical and relates to the relatively low frequencies at which such equipment operates. Frequencies even in the "vibroiseis" techniques do not generally exceed 250 Hz (Edelmann and Werner, 1982) and are normally lower.

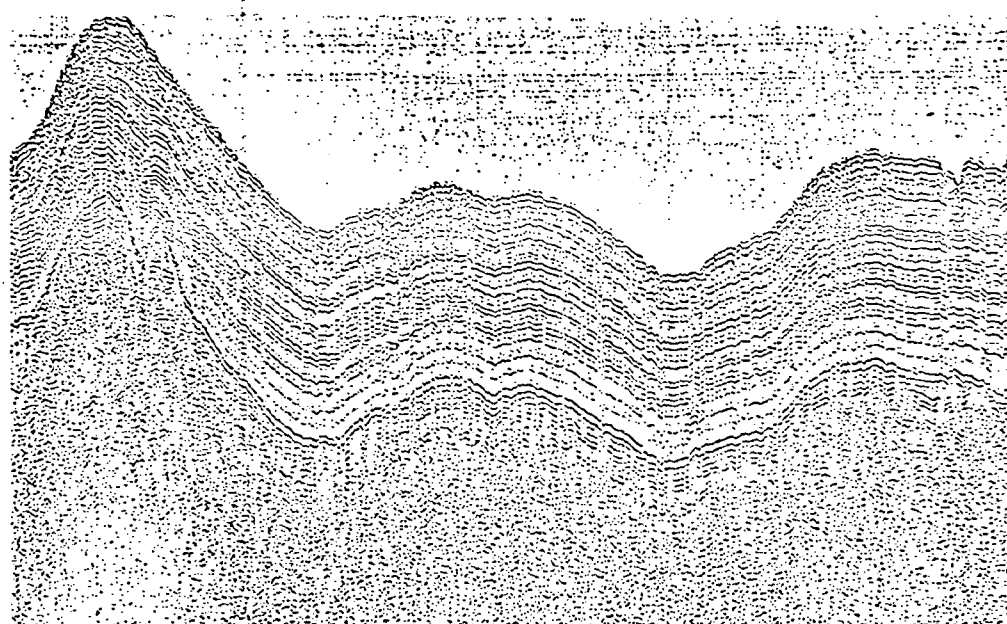
By contrast, seismic equipment designed for surficial sediment or soils investigations operates at higher frequency (1-10 KHz) and is made on a much lower budget. Of these constraints the latter is the more dominant, but both are important and combined to keep most high frequency seismic equipment as single channel analogue recorders until the late 1970's when small portable 6,12, and 24-channel fixed point digital stacking seismographs began to be manufactured, making use of the increasing power and availability of microprocessor

technology. Such devices have now become established for use on land, but for aquatic use many single-channel analogue machines are still accepted as "state of the art". There seems to be no very good reason for this other than inertia: the small land seismic recorder can be taken to sea and the problems associated with making appropriate towed hydrophone arrays are not enormous.

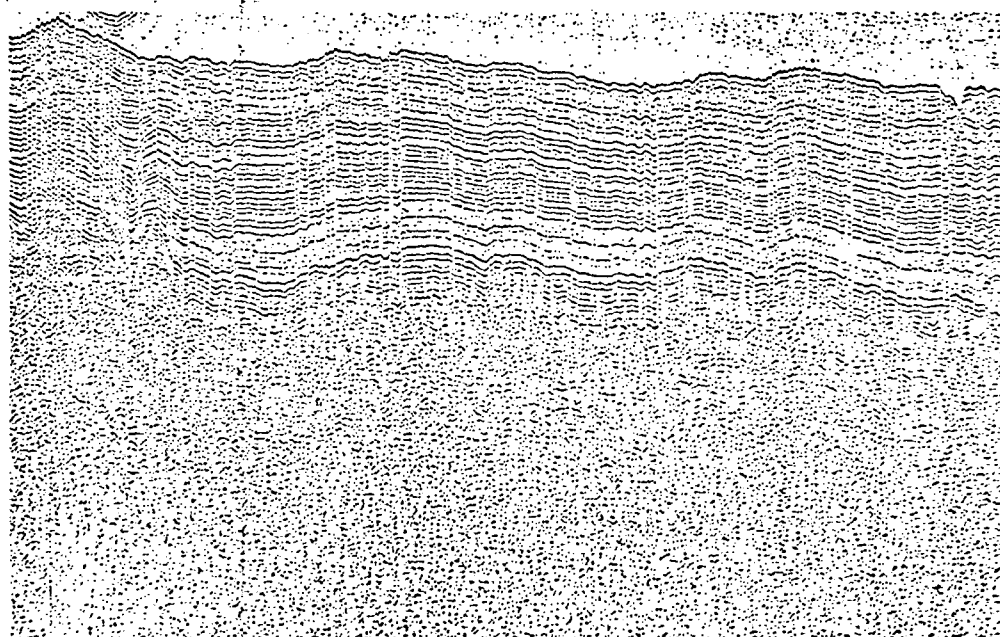
This section describes adaptations of single channel systems to observe a reflected sound field at varying offsets; continues with further modifications which were made to produce a rudimentary multichannel system; and concludes with the description of a 12-channel towed array. To begin with, a general description is given of the Nova Scotia Research Foundation deep tow seismic equipment.

2.1 DEEP TOW SEISMIC EQUIPMENT.

The deep tow seismic system which was eventually adapted for two to six channel operation is that of the Nova Scotia Research Foundation (N.S.R.F.) (Bidgood 1974). This system comprises an acoustic transmitter and receiver which can be towed near the seabed on the end of a 600 metre cable. The cable is a composite of a high voltage coaxial cable with solid polyethylene dielectric



UNCOMPENSATED RECORD



DEPTH COMPENSATED RECORD

Fig. 2.1

Subbottom profiles made with the NSRFC deep tow system to show the effect of depth compensation. The vertical extent of each record is 50 msec. and the horizontal corresponds to 500 metres.

(RG8/U) together with a seven conductor signal cable. Each of these cables is armoured and the two are bound together with steel seising. The coaxial cable feeds high voltage pulses to a sparker, while the other cable is used for signals from the receiving hydrophone and from a pressure transducer. The pressure transducer provides a signal giving the depth at which the acoustic transducers are towing in the water and this is used to advance or retard the firing of the acoustic source. Thus is obtained a reflexion record which is compensated for these vertical movements and we have a true profile of the seabed (see fig 2.1). An important effect of lowering the acoustic source (sparker) through the water column is that the output pulse increases in frequency with depth (see fig. 2.2). This has the benefit of improving resolution, but there is a decrease in the amplitude of the output signal accompanying the increase in frequency. This effect is summarised in fig. 2.3, which also gives an indication of the efficiency of transduction of the input electrical energy. Measurements of transient voltages during the sparker firing indicate that more than half of the input power is lost in the tow cable. The remainder is lost in transduction, presumably in vaporisation and ionising processes to give an overall conversion efficiency of ca. 0.03% near the surface and 0.015% at depths of 300 metres.

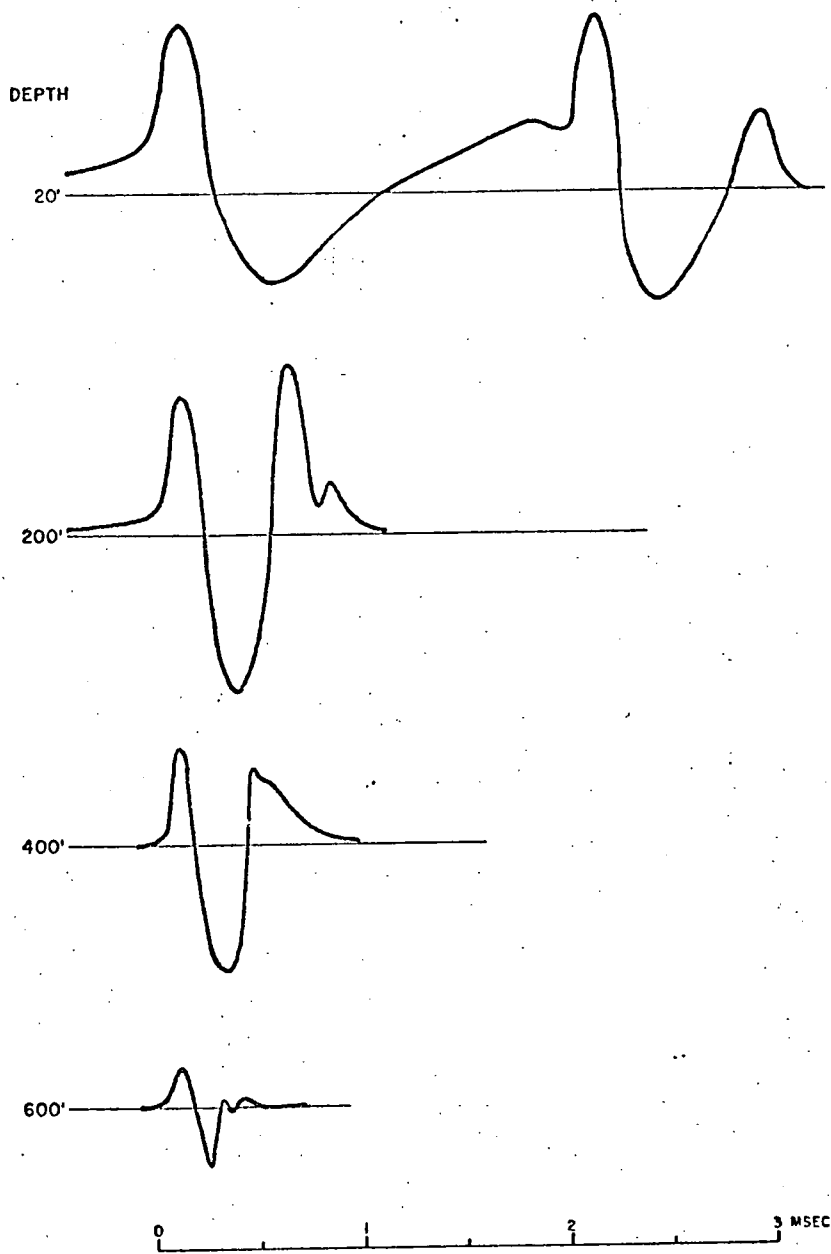


Fig. 2.2

The variation of pulse shape with depth. The patterns are from oscilloscope photographs of the output of a single-element hydrophone trailed 30m. astern of the single-tip sparker source.

Radiated energy v. depth for 3 different sparker types
operated at 3 different levels of electrical energy input

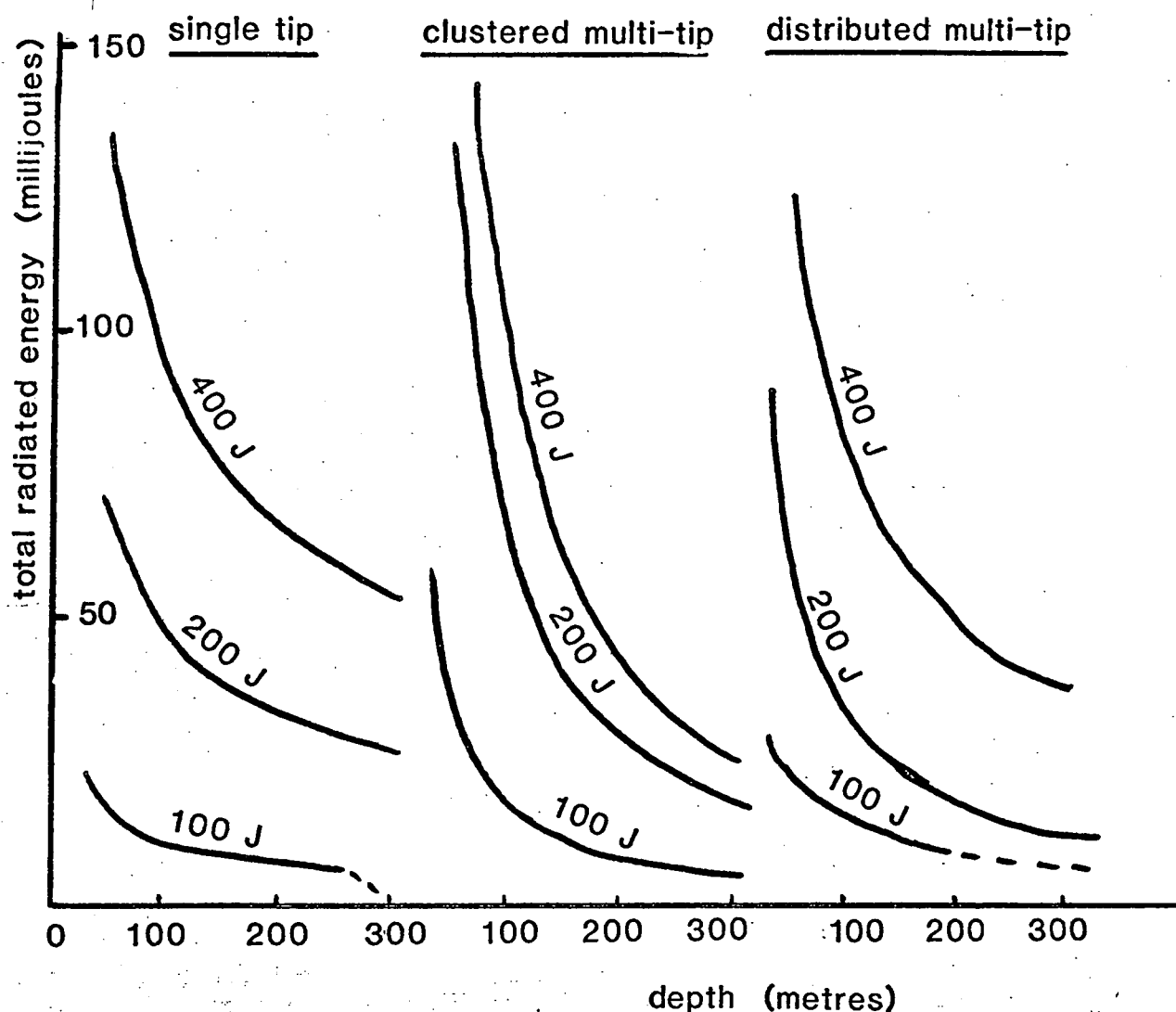


Fig. 2.3

The total radiated acoustic energy was obtained from the signal received by 2 calibrated hydrophone positioned 30m. above each of three different sparker types. The electrical energy input was 100, 200 and 400 Joules.
(Prentiss, 1983)

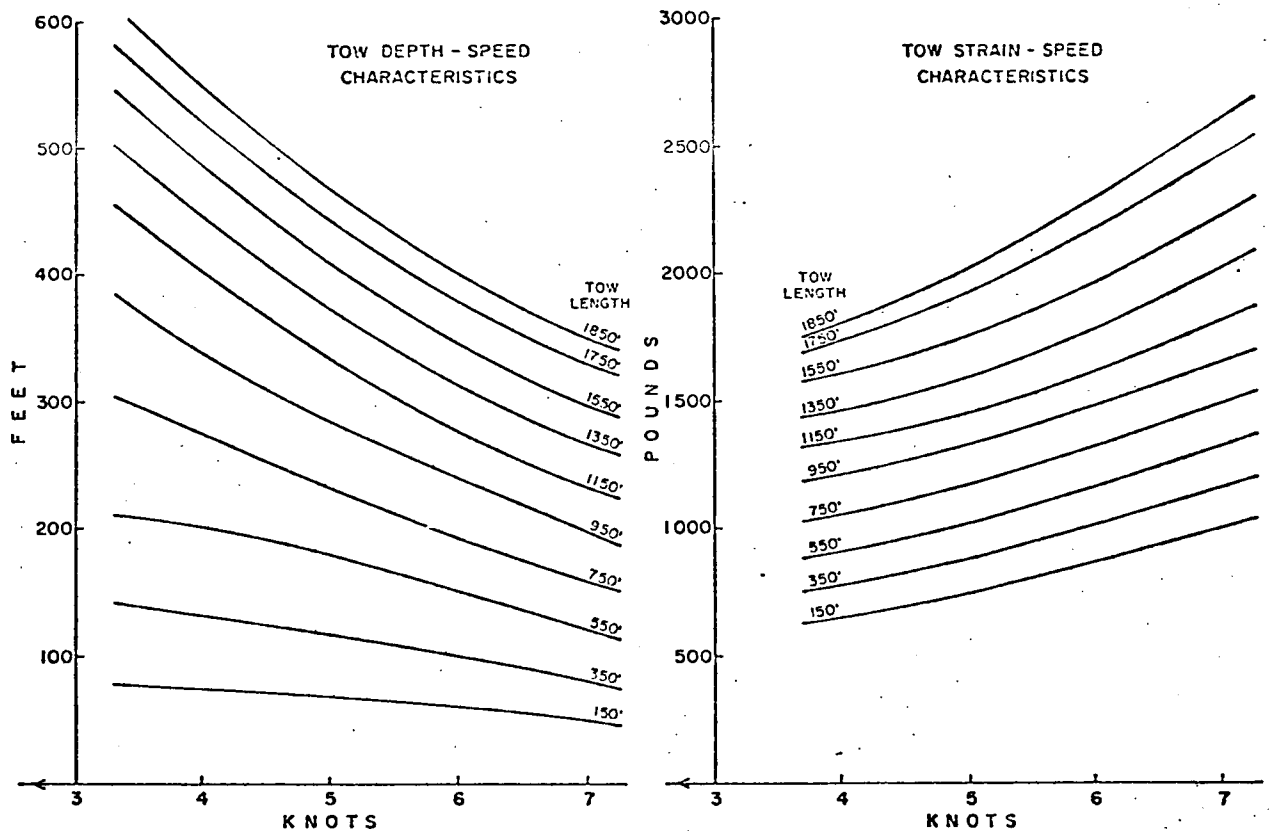


Fig. 2.4

The depth attained by the fish and the stress in the main tow cable vary with tow speed as shown above.

Other characteristics of the basic N.S.R.F. system are shown in fig 2.4. Both the depth-speed and the tow strain-speed characteristics are of importance in considering modifications and in designing more advanced systems.

2.2 SONOBUOYS USED IN CONJUNCTION WITH THE N.S.R.F. DEEP TOW PROFILER.

The use of a radio-transmitting sono-buoy either free floating or anchored as a way of observing a sound field at a fixed point while a vessel underway tows a pulsed acoustic source towards or away from the buoy is an established method of producing wide angle reflexion and refraction profiles (Le Pichon et al. 1968, McKay and Peacock, 1970). Such profiles are made for interval velocity determination both in the deep ocean (Ludwig et al., 1975; Houtz, 1977; and Bachman and Hamilton, 1980) and in upper slope and shelf seas (Houtz 1978; Houtz 1980). The scale of overall moveout extends to several tens of kilometers and airguns or explosives are used as the acoustic source, the usual aim being to transmit frequencies in the range ca. 5 to 50 Hz.

The possibility was investigated of using a similar technique to obtain such profiles with the deep-towed

acoustic source by using standard SSQ-41A sonobuoys (military specification MIL-S-22793C(AS), 1971) given by a local manufacturer (Hermes Electronics, Dartmouth, N.S.) together with an Eddystone 990R radio receiver loaned by Bedford Institute of Oceanography. The buoys were unmodified, and thus gave the acoustic frequency response which is set up for submarine detection. This response cuts out frequencies below about 100 Hz and peaks at about 1.5KHz. It is suited to the deep-tow sparker source at all operational depths (see fig. 2.2).

Profiles made in this way were obtained during a cruise on C.N.A.V. Sackville in 1974, but proved unsatisfactory for interpretation because of the difficulty of maintaining good geometrical relationships between source and receiver. The method which has become usual for generating wide angle profiles with an airgun source (where a buoy is thrown overboard at a time when the vertical incidence record shows that sediment layering is simple and uniform) is impracticable for use with the deep tow source for three reasons:

(i) Insufficient time elapses after the buoy is launched for its hydrophone to deploy to operational depths beyond about 50 metres before it is overtaken by the deep tow source.

(ii) There is a very good chance of entangling it with the deep tow cable.

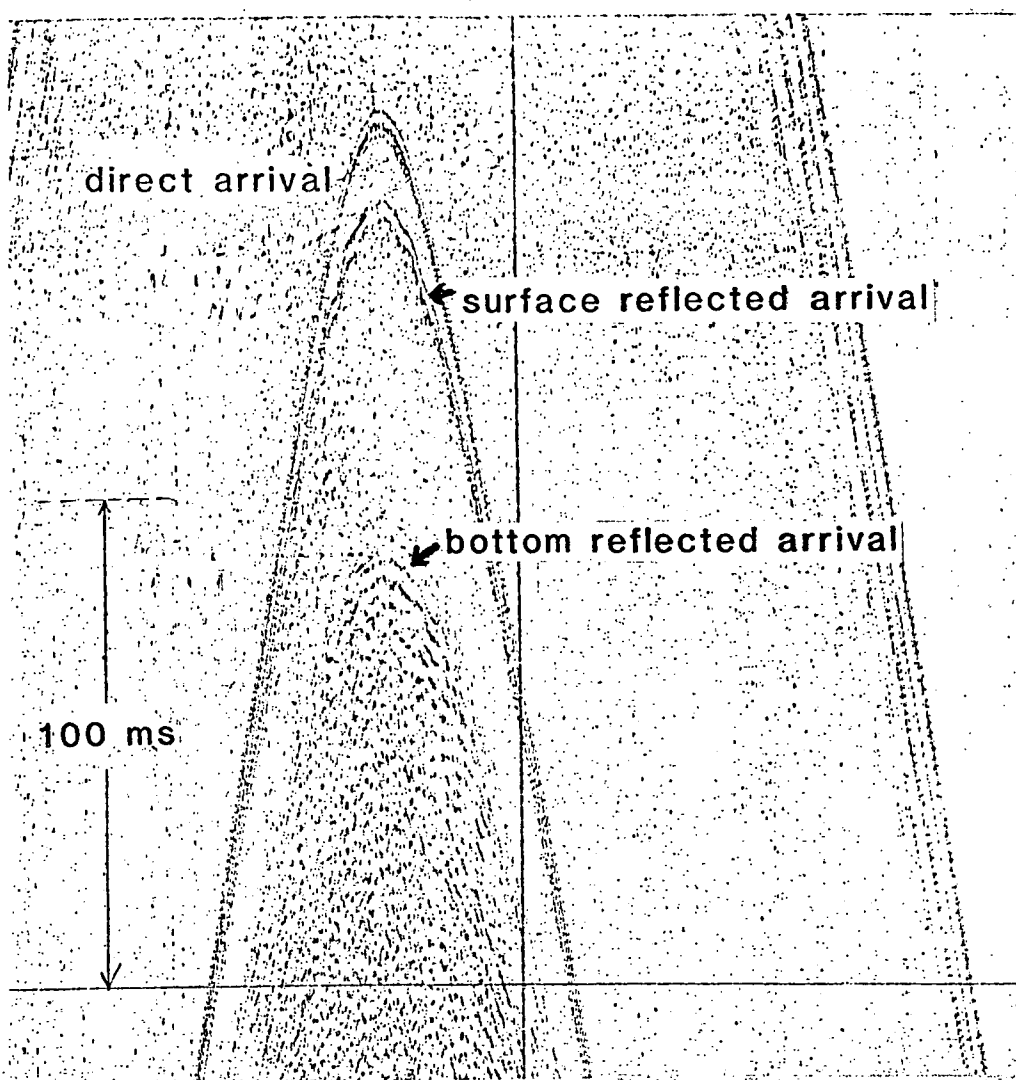
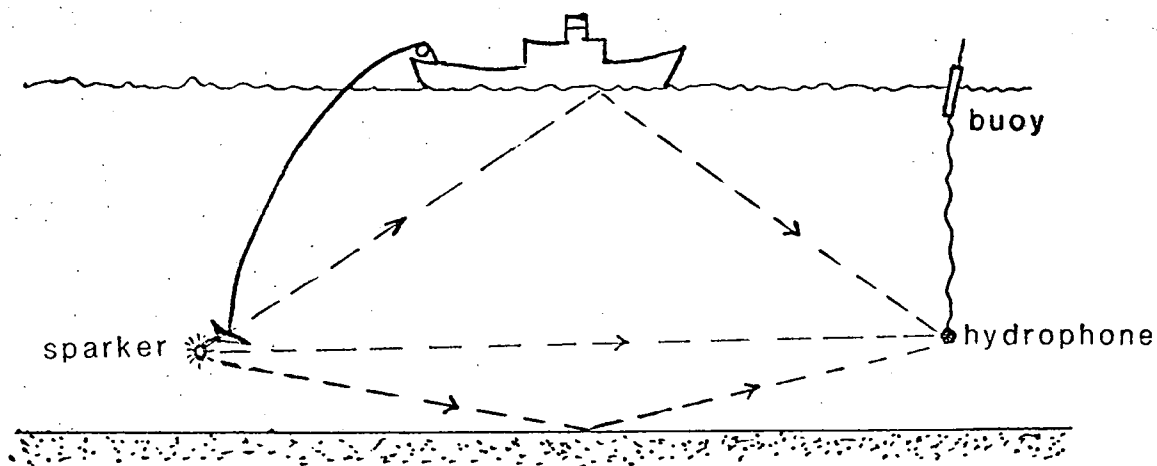


Fig. 2.5

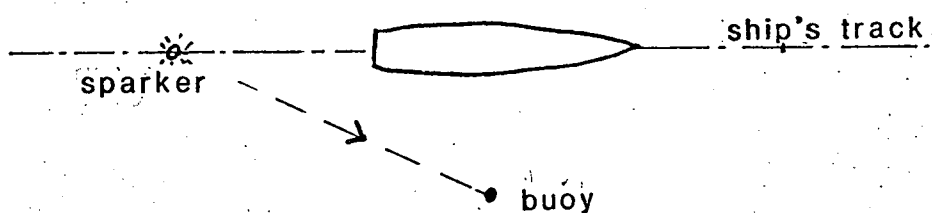
A profile obtained with deep-tow sparker and sonobuoy.

(iii) In many areas, surficial sediment layering varies over short distances so that picking an appropriate launching time can be difficult.

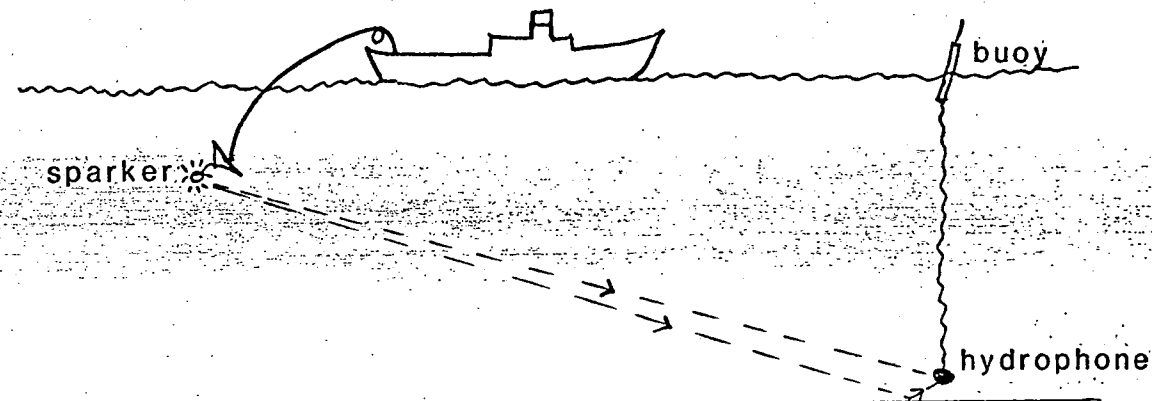
As an alternative approach, the buoy was first launched and then the ship's course was altered to sail back past it. This technique yielded profiles of the type shown in fig. 2.5. All such records which were obtained suffered from the problems of insufficiently tightly controlled geometry, as illustrated in fig. 2.6 and from rather poor time resolution of the acoustic arrivals. It is likely that the latter problem might have been overcome by attention to hydrophone suspensions and optimising the radio link to give improved signal to noise ratios. The severe difficulty of realising good geometrical relationships between source and receiver together with the obvious logistical problem of taking a ship off normal survey lines to sail backwards and forwards past a buoy made it appropriate to abandon this approach and to look for alternative means of observing the sound field at different angles.



A



B



C

Fig. 2.6

The disposition of submerged sound source and receiver in making wide angle profiles.

- (a) an idealized situation, seldom realized because of:-
- (b) buoy's being well to the side of ship's track, or
- (c) hydrophone's being ill-placed in the water column.

2.3 HYDROPHONES DEPLOYED ON CABLES OF VARYING AND FIXED LENGTH.

The method of using a hydrophone on a long cable which can be let out astern while a survey vessel is underway was described by Maries and Beckmann (1961) as a way of making refraction profiles for velocity determination. The principal constraints on doing this is the need to have the hydrophone maintain a constant water depth while it is being let out astern so that no spurious static aberrations are introduced into the data. The hydrophone is best kept at less than half a metre depth below the water surface to avoid introducing much interference from surface ghosting effects, when a typical low-power (100-400 joule) sparker is used as the source of acoustic frequencies between about 500Hz and 1.5KHz. This constraint requires careful buoying of the hydrophone tow cable. Heavy cables require more floats and more floats lead to more drag when the cable eventually becomes taut. This makes it difficult to retrieve after each cast is made. Thinner cables require little or no buoying as they tend to be held more nearly horizontal by the water drag. Over the years, several cables of this type have been made up with lengths up to about 150 metres. Since the technique involves recoiling the cable by hand after each cast, there is a potential problem of tangling the cable. This problem becomes more severe the thinner the cable. A

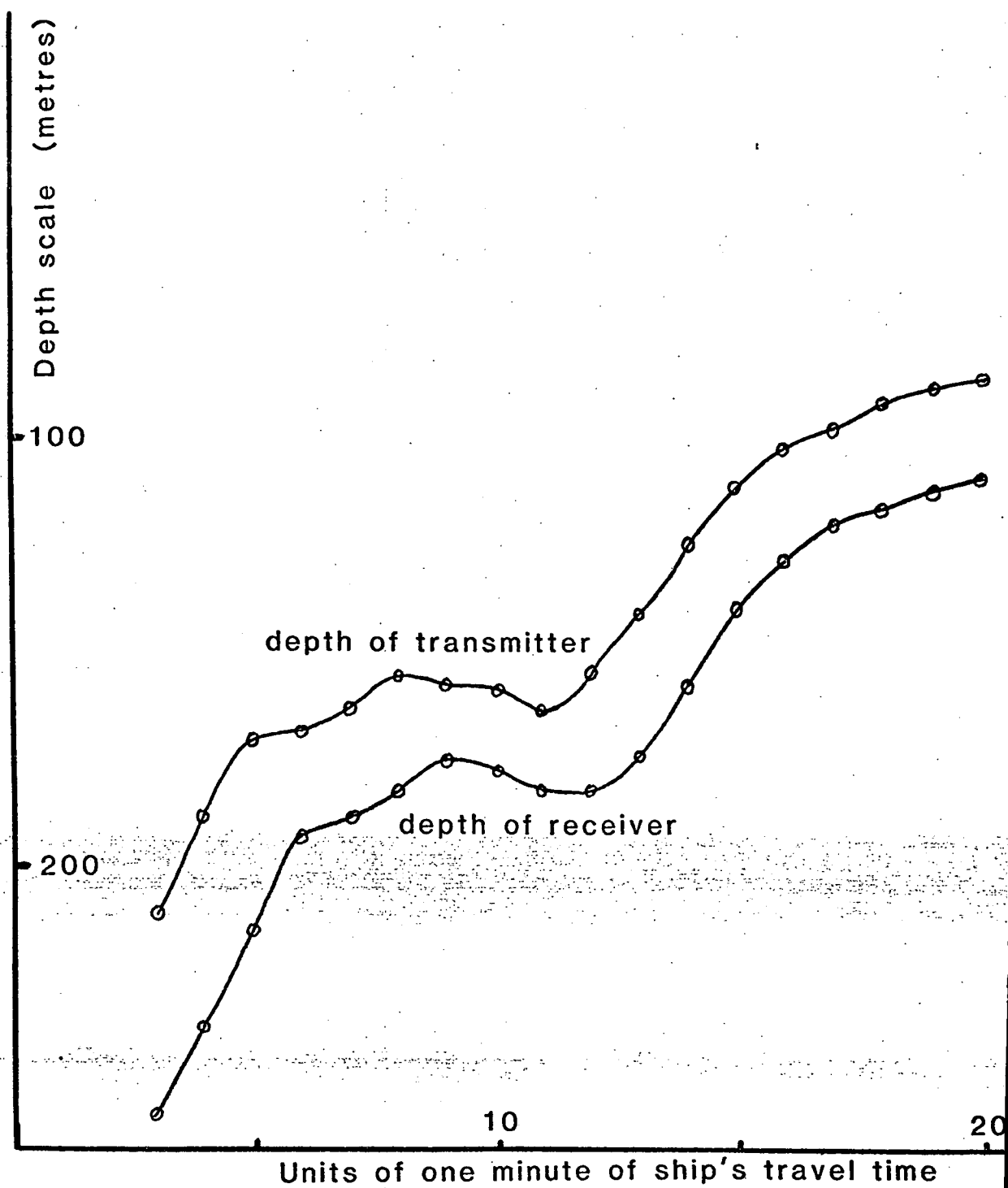


Fig. 2.7

Towing characteristics of hydrophone on a 150m. long tow.

reasonable compromise is an 8mm diameter neoprene-sheathed shielded wire-pair, but a 3mm diameter coaxial cable has been used successfully. . An earlier assembly with a 14mm cable required so many floats that survey boats had to be stopped to allow retrieval. The hydrophone assembly to be used with these cables consists of either a single ceramic element, or a compact array of three or four elements, wired in parallel and tightly clustered within ca. 0.3m. The elements have a sensitivity of ca. 0.01mV per microbar. A f.e.t. impedance matcher with voltage gain of about 1.0 is located close to the receiving elements and transmits on a two-wire system to inboard amplifiers.

2.4 LONG-TOW HYDROPHONES FOR USE WITH DEEP-TOW SEISMIC EQUIPMENT.

Cable-towed hydrophones similar to the above were used in conjunction with deep-tow seismic equipment. The principal difference in concept is that in the deep-tow mode, the hydrophone or hydrophones must be set at a fixed distance from the source on launching. An initial experiment with a hydrophone towed on a 150 metre long land-seismic cable of 5mm diameter gave an indication of the towing characteristics and the geometrical relationship between source and receiver to be expected with such a deployment. Figure 2.7 shows the variation in

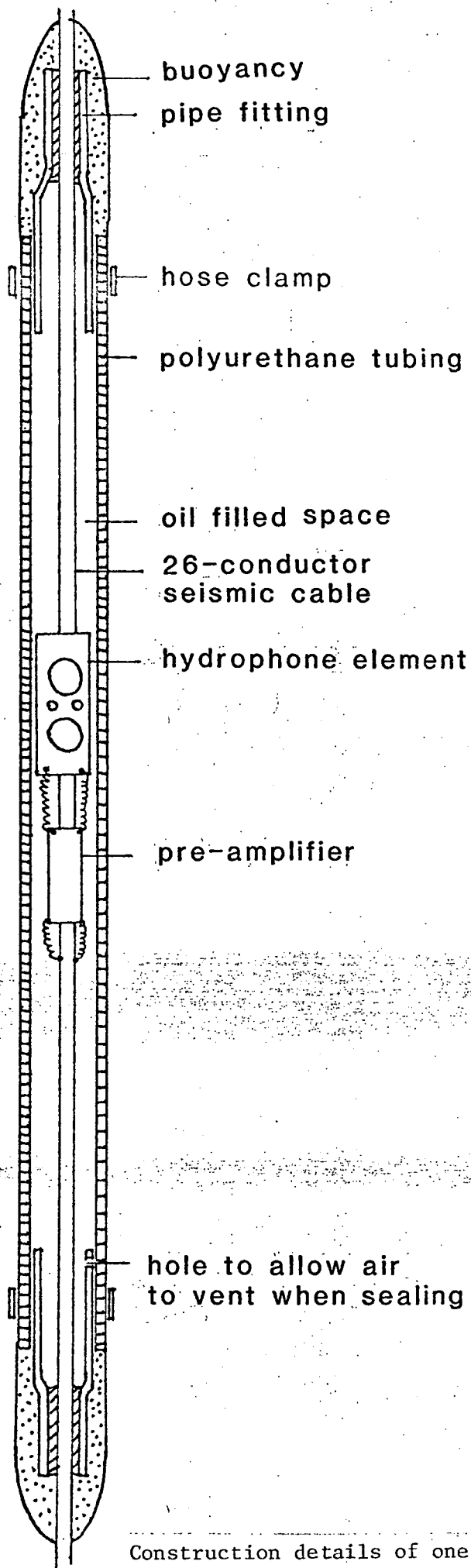


Fig. 2.8

Hydrophone assembly

Construction details of one of the hydrophones of the 5-channel deep-towed array.

depth of the acoustic source and receiver, which was observed as the main towfish was hauled in from a depth of 250 metres to one of less than 100 metres. The increased drag on the towfish was about 30kg at 3 knots, as measured on deployment of the long tow hydrophone at the surface, before the towfish was launched. As can be seen from fig.2.7, the vertical separation between the source and receiver varied from ca. 30 metres when the towfish was in level flight to 50 metres when the fish was climbing steeply.

Following on this experience, it was proposed (McKay, 1975) to build a twenty channel buoyant oil-filled hydrophone array. This proposal did not find support, but it did prove possible to put together another towed assembly comprising 5 hydrophones spaced at 40ft (12.2 m) intervals along a seismic cable similar to that described above. Each hydrophone element was housed in a 1 metre length of polyurethane tubing of 2cm internal diameter, as shown in figure 2.8, and accompanied by a f.e.t. impedance converter. The tube was oil filled and clamped to hose fittings bonded to the seismic cable. This bonding had the effect of making the assembly very difficult to take apart for servicing, a shortcoming which led to problems when it became apparent that the gain of the f.e.t. devices was not well matched between hydrophone assemblies. An attempt to overcome this problem involved

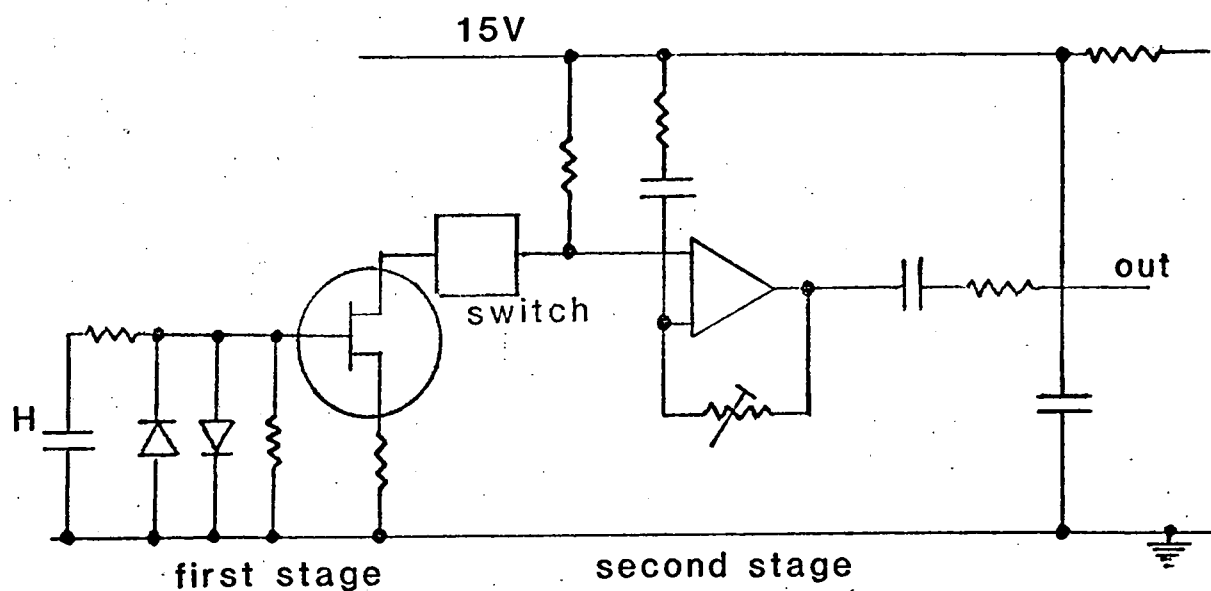


Fig. 2.9a

The pre-amplification applied to the signal of the hydrophone, H, is accomplished in two stages by field effect transistor and operational amplifier. The 5 hydrophones, each with its own first-stage pre-amplifier were fed via the switch to a single second-stage preamplifier. (Circuits after Prentiss, 1973 and Bidgood, 1975).

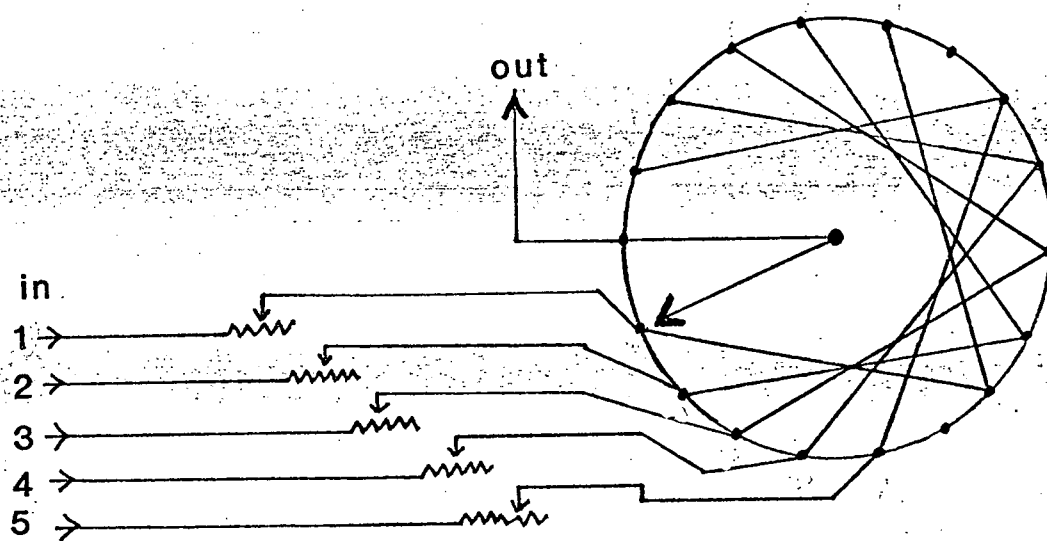


Fig. 2.9b

The hydrophone switch assembly, showing the location of variable resistors inserted to balance the hydrophone outputs.

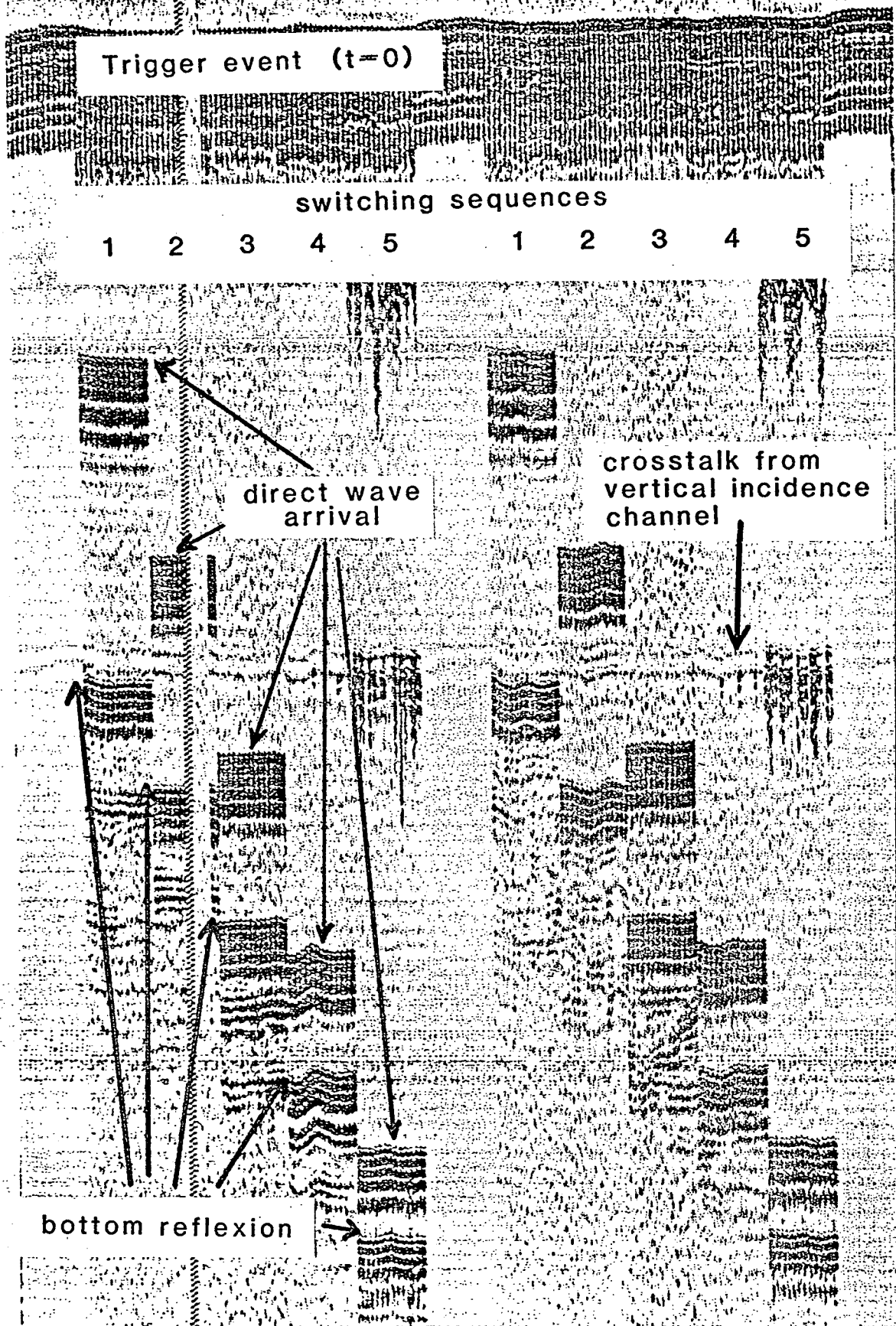


Fig. 2.10

Output of 5-channel hydrophone array.

the use of variable resistors between the f.e.t. and the second stage of the amplifier, (see fig. 2.9) and by this means it was possible to balance the output of the channels to within about 3dB as can be seen from the density of background noise in fig. 2.10. Channel 3 has the highest amplitude and channel 5 the lowest. Stimulation of the hydrophones in a test tank by means of an underwater loudspeaker showed inter-channel crosstalk at 1 KHz to be better than -60dB. This is not surprising, because the switch selector mechanism selects only one channel at a time. Crosstalk from the vertical incidence hydrophone which records continuously is much poorer, at about -12dB. This crosstalk can be seen (fig. 2.10) as a faint arrival corresponding to the vertical incidence seabed reflexion. This problem is inherent in the N.S.R.F. deep tow system and appears to be associated principally with coupling between conductors in the main 600m long tow cable. The limited number of channels, the sequential sampling pattern and the relatively poorly matched dynamic characteristics of the 5 hydrophones made this equipment unsuitable for investigating the character of reflected sound fields other than on the basis of travel time measurements. A facility for making travel time measurements was what had been sought, however, so that this equipment achieved its aim and results obtained with it have been discussed by McKay and McKay (1982).

An elaboration of the switching hydrophone array was made by combining it with an array of 8 sparkers which could be activated sequentially. The sparkers were separated one from another at 5ft. (1.5m.) intervals so that they covered a spread equal to the separation of the hydrophones. A suitable combination of switching between sparkers and hydrophones would then allow a total of 40 different moveout distances and give spatial resolution which would approach what is required by differential techniques such as the ray parameter method of Bryan (1974) for velocity determination in thin layers, as well as providing a more complete profile for interpretation by conventional methods. The production of the necessary switching assembly was technically successful, but was regarded as unsafe for use at depth in case of flooding of the underwater housing which might result in damage to the main tow cable of the deep-tow system, if high voltages appeared on the signal wires. A test of this equipment was carried out with the switching mechanism submerged, but not connected through the deep-tow seismic system. Further details of the construction of this equipment are given in Appendix 1.

Proposal to build 12 channel array

Proposal for the Design, Construction and Testing of a 12-Channel Hydrophone Array for Experiments in the Measurement of the Acoustic Velocity in Seabed Materials

General Specifications

Length of array 90 metres.
Outside diameter of array 3.2 cm. Oil filled polymer tubing.
Length of lead-in cable (from head of array to inboard connector) 180 metres.
Number of hydrophone groups 12.
Spacing between centres of hydrophone groups 7.5 m.
Number of hydrophone elements in each group 4.
Spacing between centres of hydrophone elements within groups 20 cm.
Sensitivity of individual hydrophone elements -103 ± 0.5 dB re 1V/microbar from 50Hz to 3000 Hz.
Pre-amplifier gain $+40 \pm 3$ dB from 50 Hz to 3000 Hz.
Maximum towing stress 150 kg in oil-filled parts.
Maximum stress in lead-in cable 300 kg.
Signal outputs 12 channels with common ground.

Notes on Proposed Design

1) Electrical

Groups of 4 hydrophone elements would be wired in parallel and each group would be connected to a pre-amplifier and line driver circuit located nearby. Transmission to the output end would be by a bundle of twisted pairs, one pair for each hydrophone group. At the output end of this transmission cable, differential amplifiers, one for each channel, would restore signals to a common ground. Such an arrangement would offer low crosstalk (-40 dB) between channels and good suppression of electrical noise.

2) Mechanical

There are different ways in which this seismic array could be put together mechanically. It is perhaps desirable that it should be sectional so that it can be shortened for some applications. Sectional construction would also aid servicing.

Fig. 2.11

Specifications for a 12-channel array designed for seismic work over surficial sediments at acoustic frequencies between 50 and 3000 Hz.

2.5 CONSTRUCTION OF THE 12-CHANNEL ARRAY.

As a result of interest by the Geological Survey of Canada (G.S.C.) in the methods of velocity determination subsequently published in McKay and McKay (1982), it was arranged to construct a 12-channel hydrophone array for the G.S.C., to be used in the Beaufort Sea with the aim of detecting by the refraction technique anomalously high velocities which would be indicative of ice-bonding in the sediments. In order to tow a few metres off the seabed, the array was to be capable of operating at depths to 50m, although the means of depressing it to these levels was to be the responsibility of G.S.C.. The characteristics required were agreed upon as in fig. 2.11. The pre-amplifier gain of 40dB was intentionally made rather higher than the 20 to 30dB which would be more usual in reflexion equipment employing the type of ceramic hydrophone element used, in order to assist in the detection of low amplitude headwaves. The idea of high gain near a sensing element has more recently been proposed by Klaassen and Peppen (1983).

The method of construction (fig. 2.12) was traditional in essence, but effected economies by using land seismic cable as both strain member, signal carrier, and pre-amplifier power line. The various sections were connected together by sub-miniature connectors enclosed in

Schematic diagramme of array

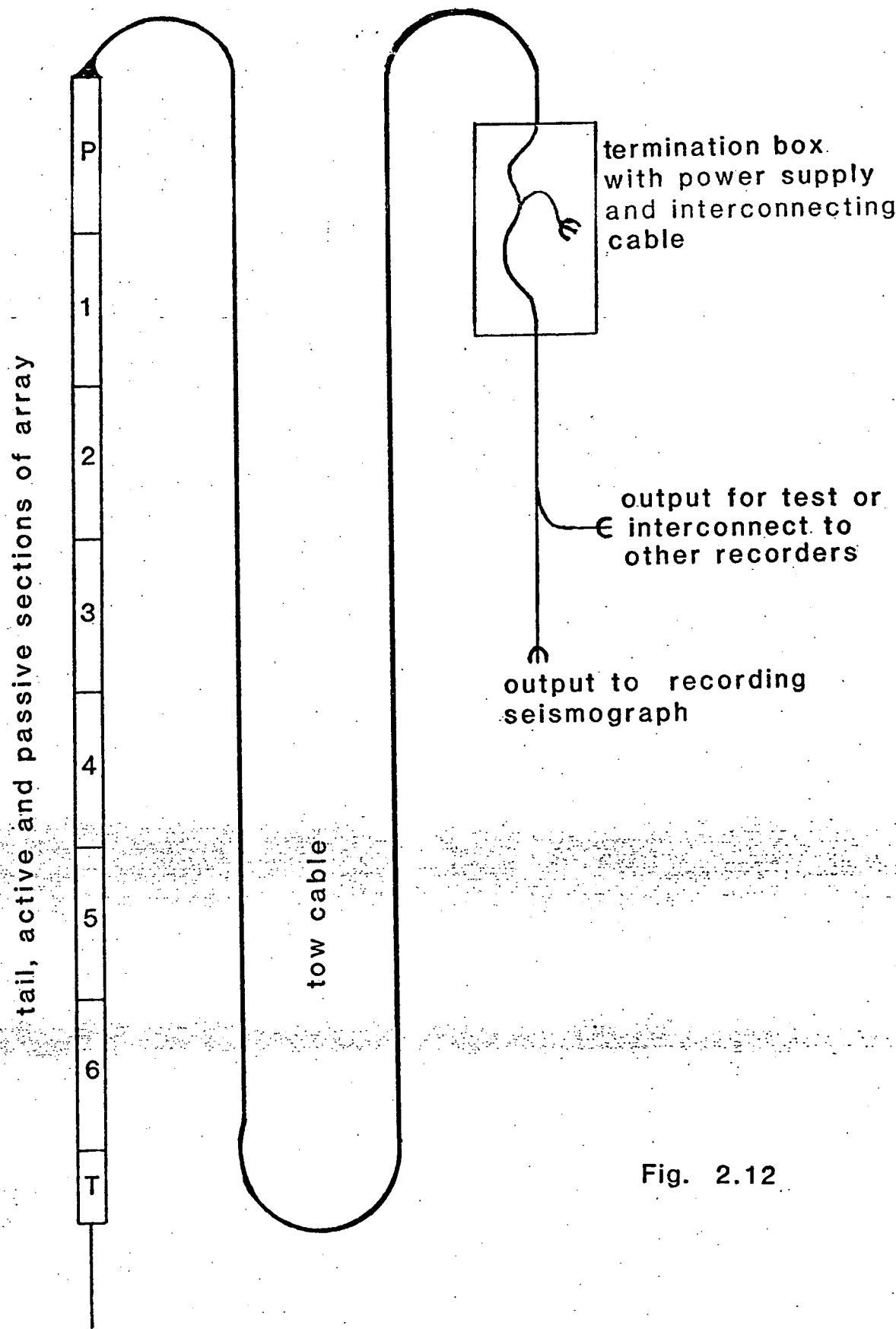


Fig. 2.12

TABLE 2.1

OBSERVED TOWING STRESS ON SECTIONS OF THE 12-CHANNEL
ARRAY.

BOAT		TOW STRAIN (kg.)			
Revs.	Approx Speed (knots)	Number of Array Sections.			
		1	2	4	5
600	2	4	6	8	10
1000	4	7	8	11	10
1300	6	10	10	14	16
1600	8	10	12	19	21

an outer nylon connector designed to be watertight and to carry the strain from one section to the next. Actual towing stresses measured in towing the completed array in the surface water are shown in table 2.1. At the operational speed of 4 knots, there is a safety factor of about 10 between this actual stress and the breaking stress of the seismic cable. Hydrophone elements in groups of four with a pre-amplifier were attached at takeout points, two groups of elements for each active section. The gain characteristics of the pre-amplifier assemblies are shown in table 2.2. The additional 6dB of gain comes from an active line-driver stage which follows the f.e.t. pre-amplifier. The main tow cable was a 56 conductor seismic cable having a breaking stress of over 300kg. and protected with a thick polyurethane sheath. A short cable within the inboard termination box carried the signal wires straight through, but allowed the connexion of the 15v d.c. power for the pre-amplifiers. Finally an interconnect cable allowed connexion to the Nimbus seismograph which was to be used and gave a take-off point for signal display on graphic or other recorders. The conductor connexions in these various cables are shown in figs. 2.13 to 2.17. Illustrations of the connectors used between sections are given in figs. 2.18a and b. The method of attachment of hydrophone elements is seen in fig. 2.19a and a completed section appears in fig. 2.19b.

Active section

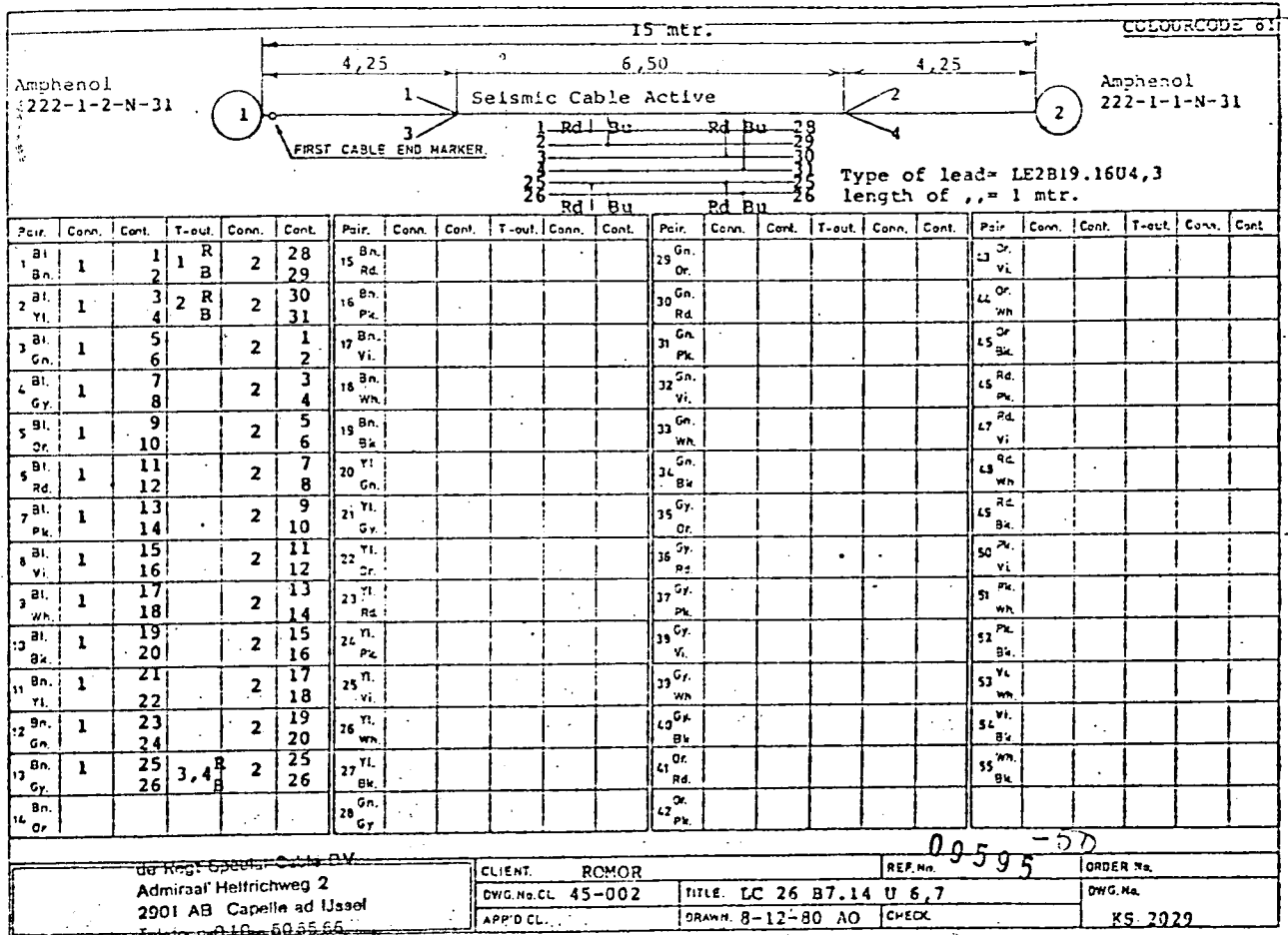


Fig. 2.13

Wiring connections for active section of array.

Passive section

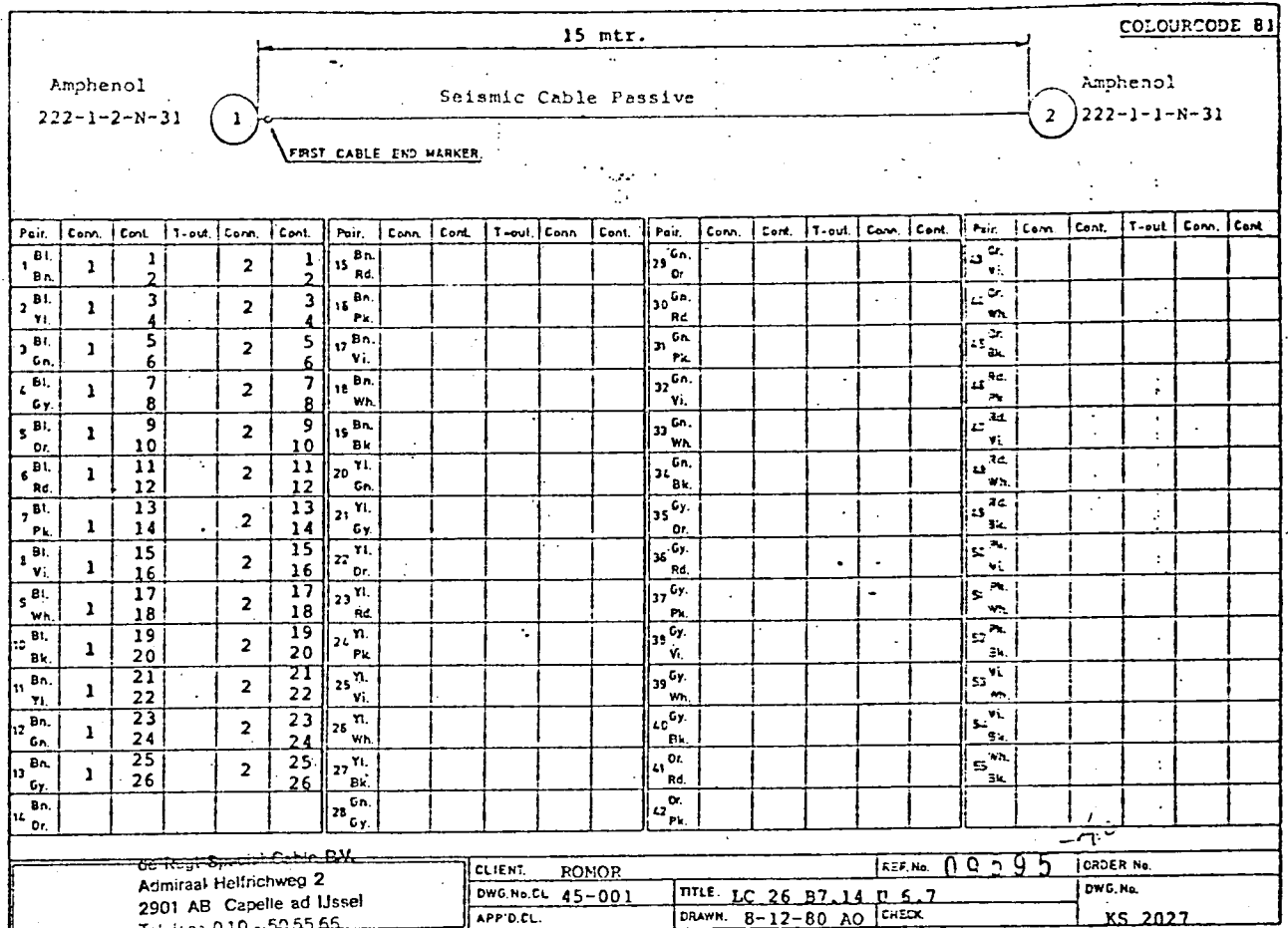


Fig. 2.14

Wiring for passive section of array.

Tow cable

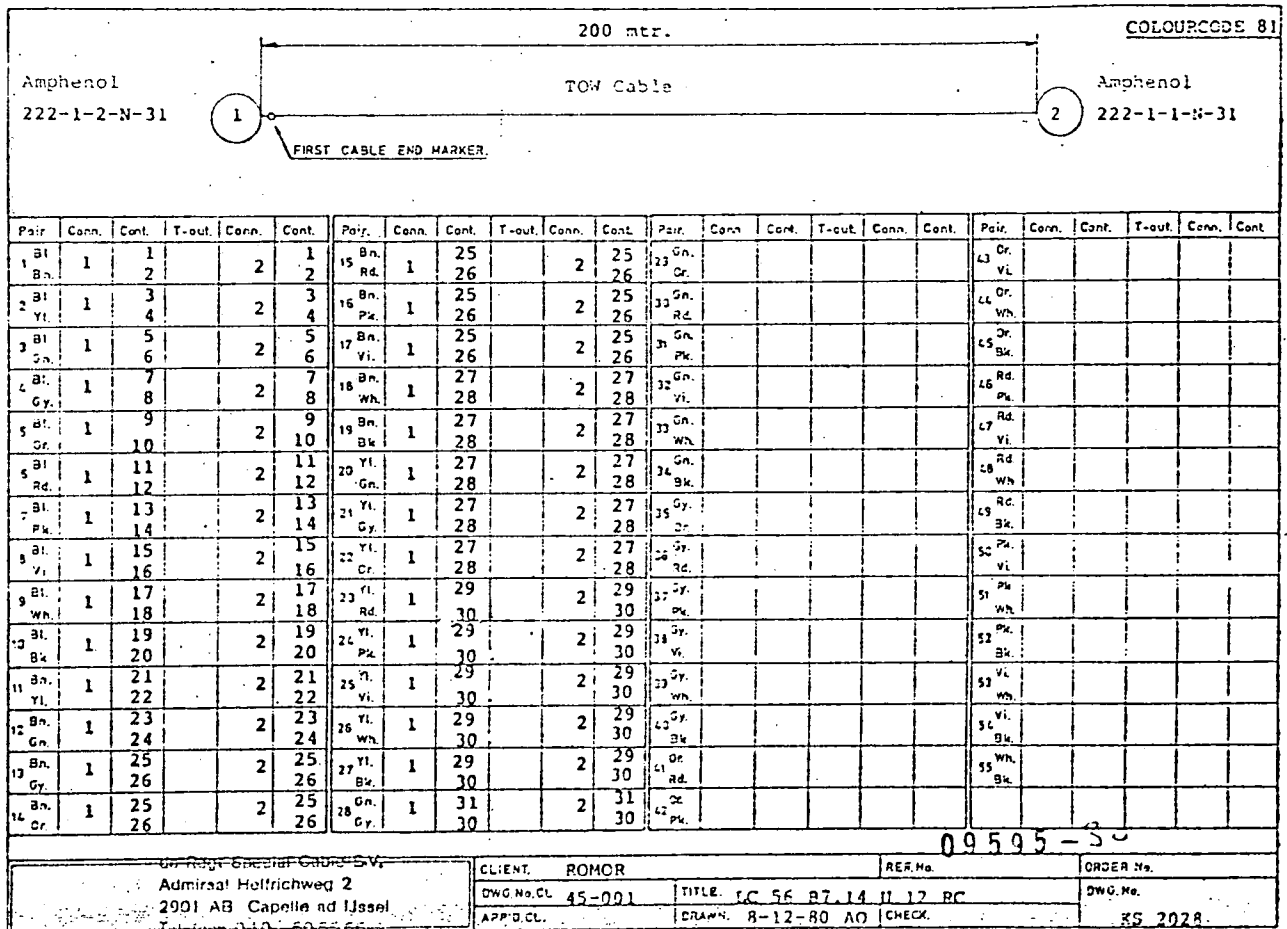


Fig. 2.15

Wiring for main tow cable.

Power supply cable

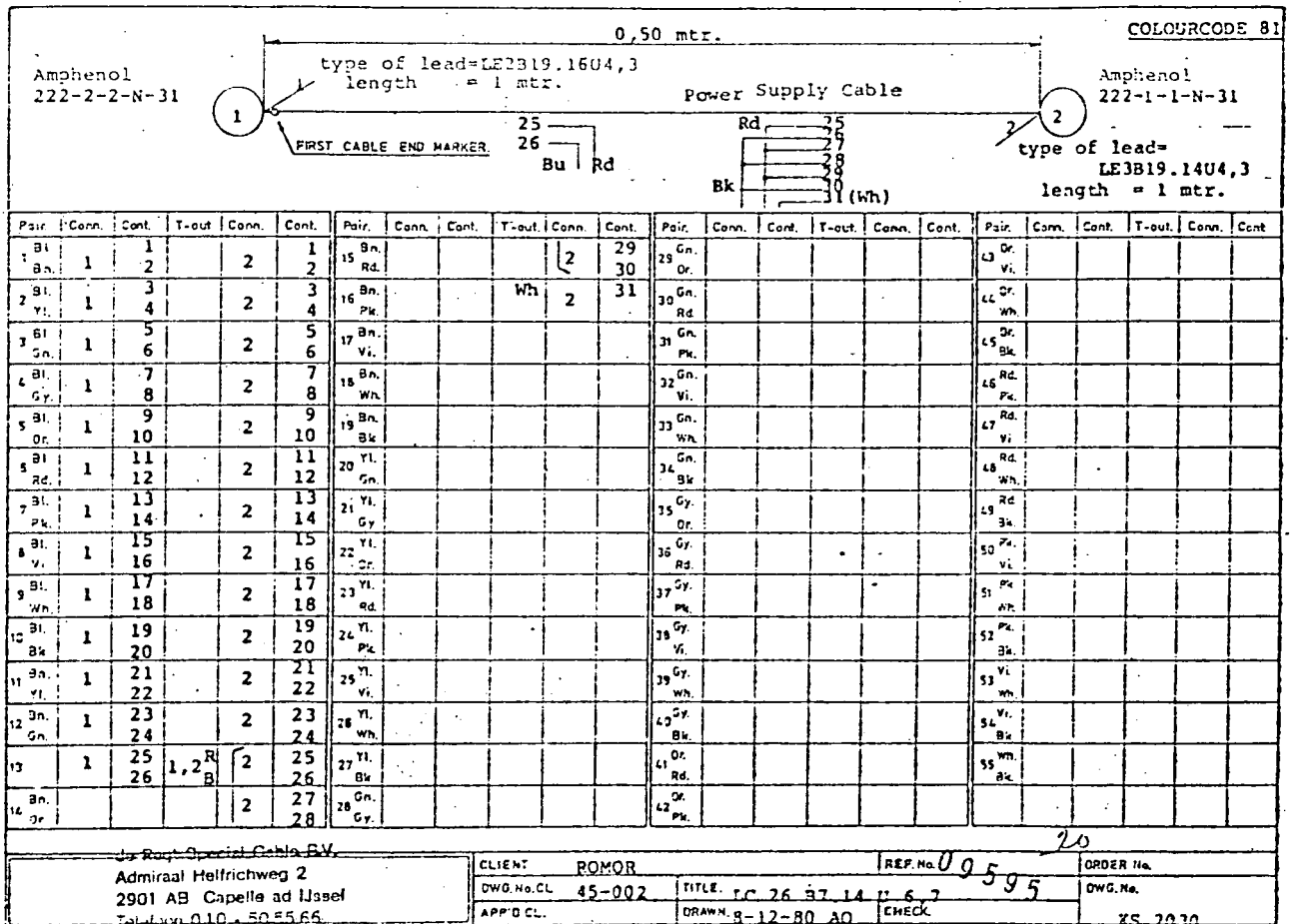


Fig. 2.16

Wiring for short cable used internally in termination box to allow connection of power supply.

Lab. connecting cable to seismic recorder

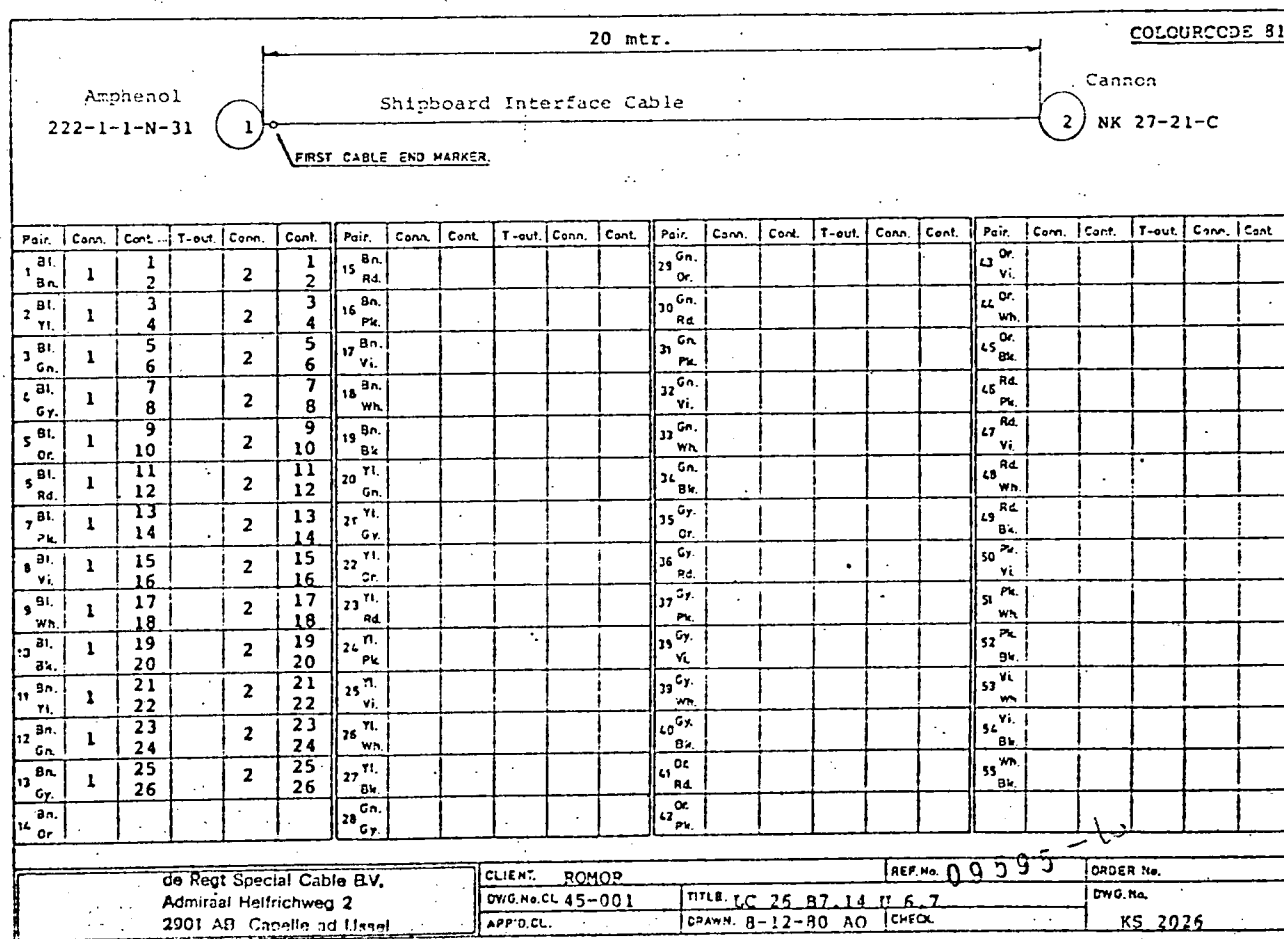


Fig. 2.17

Wiring for cable connecting termination box with seismograph.

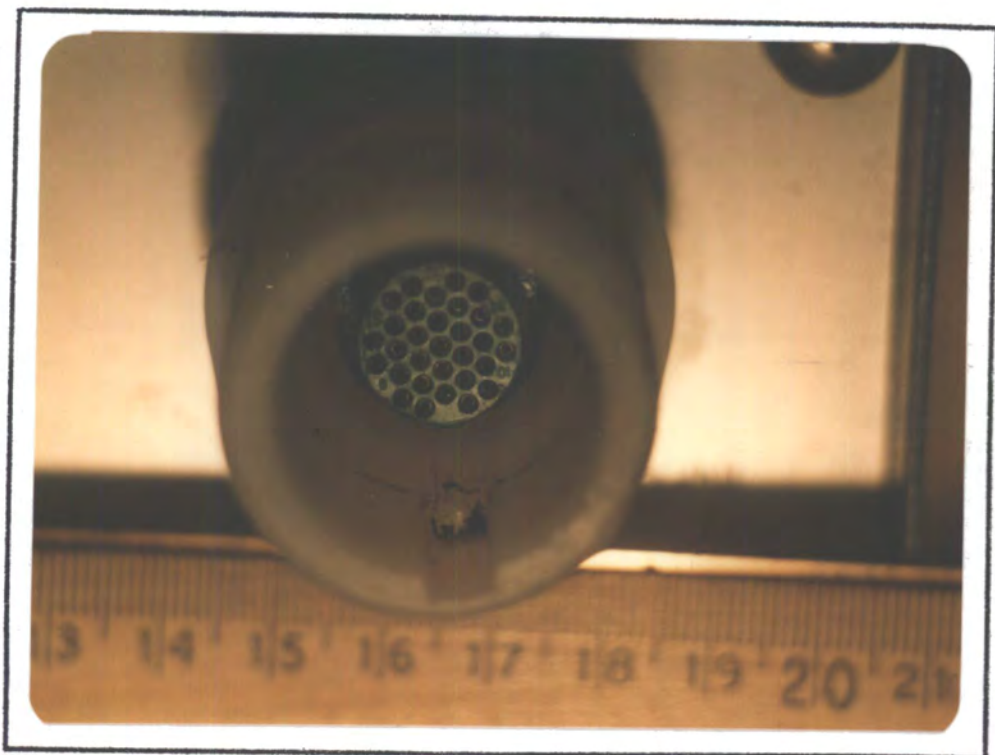


Fig. 2.18

(a) Interior view of the connectors which were made to allow sectional construction of the array.

(b) View of a mated connector pair coupling two array sections. Outside diameter of the white plastic shell nut is ca. 4cm.



Fig. 2.19

(a) The hydrophone elements are supported against the seismic cable by means of a small polyurethane cradle and are tied in place.

(b) A completed section of the array. The black bands seen on the right are wrap-on weights made of thin sheet lead, for buoyancy trimming.

TABLE 2.2

PRE-AMPLIFIER CHARACTERISTICS

No.	Flat Response	Total Gain	-3dB high	-3dB low
1	95 - 14K	46	29K	34
2	90 - 17K	46	30K	33
3	100 - 14K	46	30K	35
4	95 - 14K	46	28K	36
5	88 - 12K	46	25K	34
6	95 - 13K	46	30K	36
7	84 - 13.5K	46	32K	34
8	90 - 12K	46	29K	31
9	85 - 11.5K	46	29K	34
10	95 - 12K	46	31K	38
11	90 - 14.5K	46	29K	35
12	87 - 12K	46	27K	33

This 12-channel array had advantages over the equipment described in section 2.4 in that it offered:

- (i) More channels
- (ii) Neutral buoyancy leading to better towing stability and reduced noise
- (iii) Freedom from crosstalk (better than 40dB separation, compared with c. 12dB).

The principal disadvantage was that it was not incorporated into a near-bottom deployment system complete with depressor and winch (c.f. section 2.1 above). Field results obtained with the 12-channel array are discussed in chapter 9.

3. GASES IN SEDIMENTS.

3.0 INTRODUCTION.

The observation of gases associated with sediments began no later than 1563 when the term "ignis fatuus" was brought into use by Renaissance alchemists to describe the glowing light produced by the spontaneous combustion of marsh gas (Little et al, 1933). The extension of this observation to the lacustrine environment was given by Rossolimo (1935) who reported gases in lake sediments, and by Koyama (1953) who undertook sampling work on sediments in shallow water marine areas. Emery and Hoggan (1958) had undertaken quantitative sampling of gases from deep water marine sediments before the appearance of an extensive literature on the way in which sediment gasification could affect acoustic profiles (Levin, 1962; Jones et al. 1964). The effect of included gas bubbles on the acoustic properties of liquids had been studied more thoroughly by then: by Wood (1931) who gave the equation relating the acoustic compressional wave velocities in a gas and in water to that which would be found in an emulsion of the two fluids; by Minnaert (1933) and by Strasberg (1956) who discussed the production of sound by bubbles in water; by Cartensen and Foldy (1947), MacPherson (1957) and Devin (1959) who considered the effects on velocity of propagation and absorption when resonance occurs in the included bubbles.

This chapter reviews what is known about the origin of gases in sediments, the methods of obtaining samples and quantitative measurements of the amounts of included gases, and the extensions of theories and measurements on gasified water to encompass gasified sediments. The implications for foundation engineering are discussed and field examples are given of gasified sediment in the coastal waters of Eastern and Arctic Canada. This chapter will make clear the gross effect that gasification can have on the velocity at which compressional waves will propagate in a sediment and this is the principal reason for the inclusion of this topic in the overall work.

3.1 METHODS OF DETECTING AND MEASURING GAS IN SEDIMENTS.

Sea water contains dissolved gases. The principal constituents are oxygen, nitrogen, argon and carbon dioxide as would be expected if, as is generally agreed, the main source of these gases is the atmosphere. Emery and Hoggan (1958) extracted gases from sediment cores by displacing them upwards with boiled distilled water perfused through the core while the sediment was fluidised by agitation. They collected the displaced gases in an evacuated bottle and analysed the chemical types by mass spectrometry. In cores taken from marsh and marine environments off California, they found quantities of

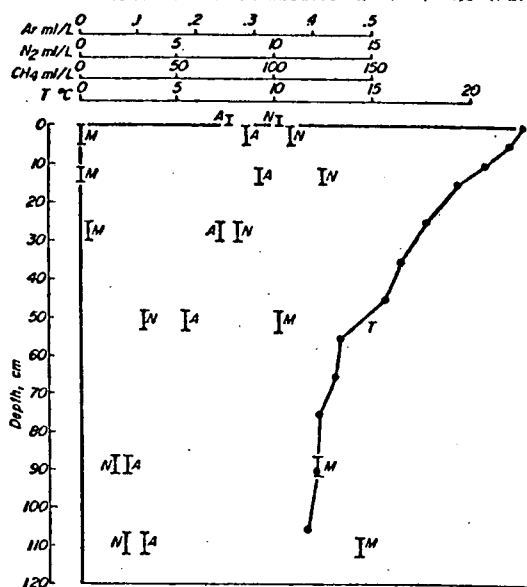
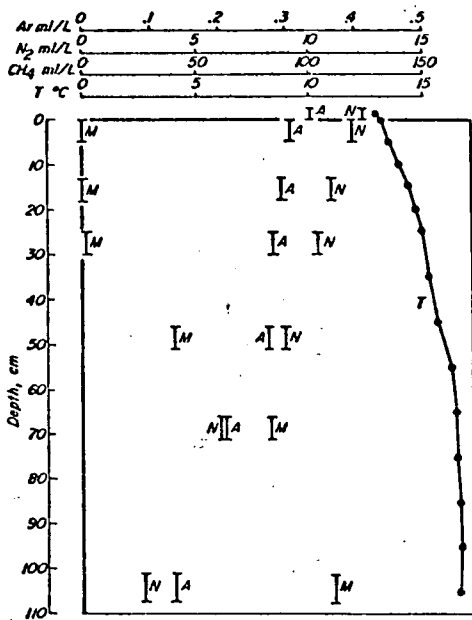
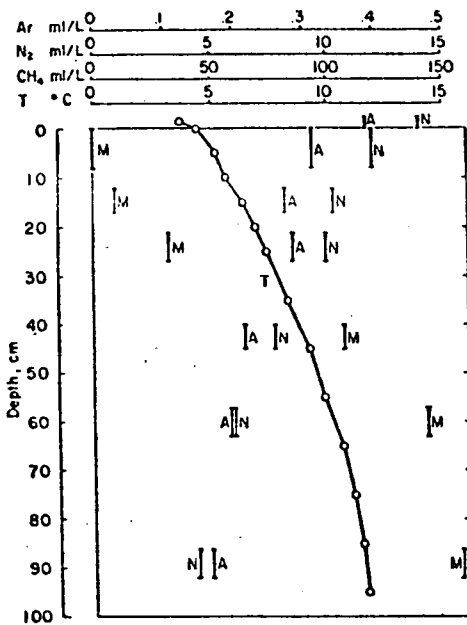
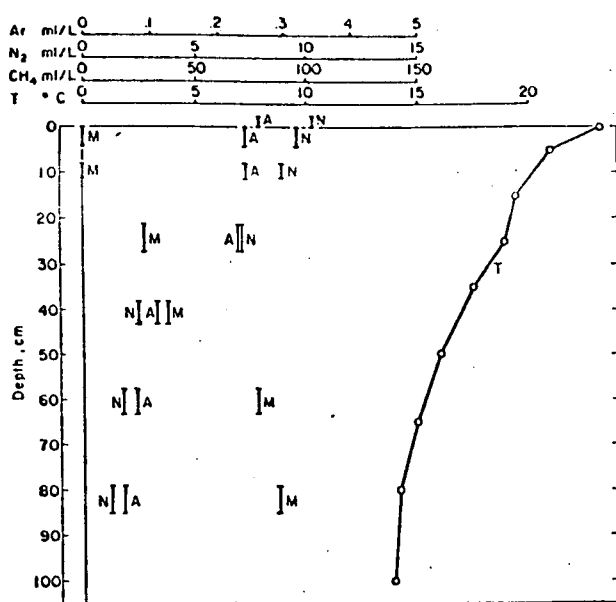


Fig. 3.1

Variation with depths of total Nitrogen, Argon and Methane in sediments of Chesapeake Bay at four different sampling sites (from Reeburgh and Heggie, 1974). Note that Nitrogen and Argon decrease with depth, whereas Methane increases.

methane up to 240ml. per litre of sediment (at atmospheric pressure) together with smaller fractions of higher saturated and unsaturated paraffins, in addition to ammonia, hydrogen sulphide and the atmospheric gases. Methane concentrations were often found to be an order of magnitude greater than those of the atmospheric gases, a finding confirmed by Reeburgh and Heggie (1974) who looked at sediment samples from Chesapeake Bay (fig. 3.1) by a different method. Their technique (Reeburgh, 1968) allowed "the use of smaller samples so that repeated analyses could be made on a single section of core and a better estimate of the experimental scatter be obtained. They claim measurements by gas chromatography of concentrations precise to 5% (fig 3.1). The variations in concentration which they have observed over the first metre of depth below the seabed show nitrogen and argon concentrations decreasing with depth and methane increasing." This in itself suggests an atmospheric origin for the nitrogen and argon and their advection to the water sediment interface by water movements. The increase of methane with depth suggests a deeper origin for that gas. Reeburgh and Heggie note that bubbles of gas are often difficult to release from fine grained sediments. In some cases, vigorous agitation did not speed the process, and the released gases became apparent only after retrieved cores had been stored "for some time". This is an important observation in that it indicates that

included bubbles are normally very small, a point which is significant in the theoretical treatment of the acoustics of gasified sediments (section 3.5).

3.2 ORIGIN OF GASES IN SEDIMENTS.

As mentioned above, there are two possible sources for gases in sediments: from above the seabed by diffusion from the water column into sediment pore waters and from below the seabed by upward migration from underlying strata or else by their formation in-situ by diagenetic processes. Any given gasified sediment may contain contributions from both sources. Upward migration from consolidated strata might be expected to be enhanced along faults or joints so that gas concentrations would have a maximum along these lines of weakness. Some indication that this may occur is given by McKay (1980a) in a report which is included with this work. Such deeper gas which may have migrated into unconsolidated sediments would until recently be presumed to be derived from the petrogenic rocks where it formed by the action of appropriate temperatures and pressures on organic material incorporated in the rocks. There is a possible non-organic origin of gas found at depths up to 10,000 metres in parts of Texas, Oklahoma and Wyoming and predicted to occur extensively in the crust. Gold (1983)

	ΔG° (kcal per mole of glucose equivalent oxidized)
a. Aerobic respiration	
$\text{CH}_2\text{O} + \text{H}_2\text{O} \rightarrow \text{CO}_2 + 2\text{H}_2$	
$2\text{H}_2 + \text{O}_2 \rightarrow 2\text{H}_2\text{O}$	
$\text{CH}_2\text{O} + \text{O}_2 \rightarrow \text{CO}_2 + \text{H}_2\text{O}$	-686
b. Nitrate reduction	
$5\text{CH}_2\text{O} + 5\text{H}_2\text{O} \rightarrow 5\text{CO}_2 + 10\text{H}_2$	
$10\text{H}_2 + 4\text{NO}_3^- + 4\text{H}^+ \rightarrow 2\text{N}_2 + 12\text{H}_2\text{O}$	
$5\text{CH}_2\text{O} + 4\text{NO}_3^- + 4\text{H}^+ \rightarrow 2\text{N}_2 + 5\text{CO}_2 + 7\text{H}_2\text{O}$	-579
c. Sulfate reduction	
$2\text{CH}_2\text{O} + 2\text{H}_2\text{O} \rightarrow 2\text{CO}_2 + 4\text{H}_2$	
$4\text{H}_2 + \text{SO}_4^{2-} \rightarrow \text{S}^{2-} + 4\text{H}_2\text{O}$	
$2\text{CH}_2\text{O} + \text{SO}_4^{2-} \rightarrow \text{S}^{2-} + 2\text{CO}_2 + 2\text{H}_2\text{O}$	-220
d. Carbonate reduction	
$2\text{CH}_2\text{O} + 2\text{H}_2\text{O} \rightarrow 2\text{CO}_2 + 4\text{H}_2$	
$4\text{H}_2 + \text{HCO}_3^- + \text{H}^+ \rightarrow \text{CH}_4 + 3\text{H}_2\text{O}$	
$\text{CO}_2 + \text{H}_2\text{O} \rightarrow \text{HCO}_3^- + \text{H}^+$	
$2\text{CH}_2\text{O} \rightarrow \text{CH}_4 + \text{CO}_2$	-99
e. Nitrogen fixation	
$3\text{CH}_2\text{O} + 3\text{H}_2\text{O} \rightarrow 3\text{CO}_2 + 6\text{H}_2$	
$6\text{H}_2 + 2\text{N}_2 \rightarrow 4\text{NH}_3$	
$3\text{CH}_2\text{O} + 2\text{N}_2 + 3\text{H}_2\text{O} \rightarrow 4\text{NH}_3 + 3\text{CO}_2$	-57
f. Fermentation: heterolactic	
glucose $\rightarrow$ acetaldehyde + CO_2 + lactate + H_2	
H_2 + acetaldehyde $\rightarrow$ ethanol	
glucose $\rightarrow$ lactate + ethanol + CO_2	-49
g. Fermentation: Stickland reaction	
alanine + $2\text{H}_2\text{O} \rightarrow \text{NH}_3$ + acetate + CO_2 + 2H_2	
2H_2 + 2 glycine $\rightarrow 2\text{NH}_3$ + 2 acetate	
alanine + 2 glycine + $2\text{H}_2\text{O} \rightarrow 3$ acetate + CO_2 + 3NH_3	-17

Table 3.1

Metabolic reactions used by micro-organisms (Claypool and Kaplan, 1974).

favours a theory for the formation of the Earth by the assembly of large numbers of carbonaceous chondrites, which would provide sufficient carbon for the formation of these deep-seated gases.

Most authors discuss the in-situ production of gases by chemical and biological reactions in the sediment on the assumption that these processes are the principal source of gasification. Claypool and Kaplan (1974) give the main processes involved in microbial metabolism in marine sediments, (table 3.1, from that source). Where the appropriate reagents are available, it can be expected that the reaction yielding the greatest amount of energy will predominate where organisms are competing for the reagents, but a certain amount of symbiosis is both possible and necessary where one reaction provides the materials for the next. An example of this can be found in the fermentation processes which break down complex carbohydrates and amino acids into the simpler organic compounds which can be used in the aerobic respiration process. Consequently the figure 3.2 is a considerable simplification, but it does indicate the relative stratigraphic horizons at which gases will be produced. An obvious zone which appears to be missing from this figure is that of nitrate reduction, a reaction second only to that of aerobic respiration in energy release. The reason for this is that the reaction is normally

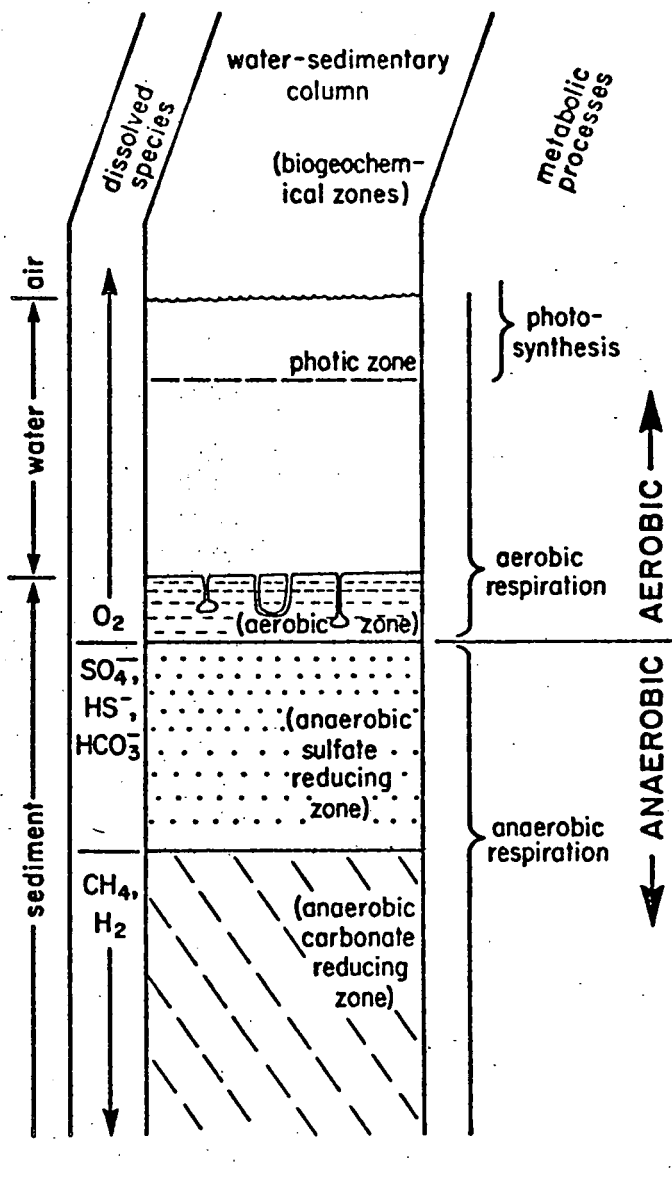


Fig. 3.2

A simplified profile of the various aerobic and anaerobic microbial zones typical of a sediment rich in organic nutrients (from Claypool and Kaplan, 1974).

Temp. °C.	HYDROGEN SULFIDE		METHANE		CARBON DIOXIDE		NITROGEN		OXYGEN		AMMONI
	α	q	α	q	α	q	α	q	α	q	α
0	4.670	0.7066	0.05563	0.003959	1.713	0.3346	0.02354	0.002942	0.04839	0.006945	
1	4.522	0.6839	0.05401	0.003842	1.646	0.3213	0.02297	0.002859	0.04758	0.006756	1176
2	4.379	0.6619	0.05244	0.003728	1.584	0.3091	0.02241	0.002798	0.04633	0.006574	
3	4.241	0.6407	0.05093	0.003619	1.527	0.2978	0.02187	0.002730	0.04512	0.006400	
4	4.107	0.6201	0.04946	0.003513	1.473	0.2871	0.02135	0.002663	0.04397	0.006232	
5	3.977	0.6001	0.04805	0.003410	1.424	0.2774	0.02086	0.002600	0.04287	0.006072	1047
6	3.852	0.5809	0.04669	0.003312	1.377	0.2681	0.02037	0.002537	0.04180	0.005918	
7	3.732	0.5624	0.04539	0.003217	1.331	0.2589	0.01990	0.002477	0.04080	0.005773	
8	3.616	0.5446	0.04413	0.003127	1.282	0.2492	0.01945	0.002419	0.03983	0.005632	
9	3.505	0.5276	0.04292	0.003039	1.237	0.2403	0.01902	0.002365	0.03891	0.005498	947
10	3.399	0.5112	0.04177	0.002955	1.194	0.2318	0.01861	0.002312	0.03802	0.005368	
11	3.300	0.4960	0.04072	0.002879	1.154	0.2239	0.01823	0.002263	0.03718	0.005246	
12	3.206	0.4814	0.03970	0.002805	1.117	0.2165	0.01786	0.002216	0.03637	0.005128	
13	3.115	0.4674	0.03872	0.002733	1.083	0.2098	0.01750	0.002170	0.03559	0.005014	857
14	3.028	0.4540	0.03779	0.002665	1.050	0.2032	0.01717	0.002126	0.03486	0.004906	837
15	2.945	0.4411	0.03690	0.002599	1.019	0.1970	0.01685	0.002085	0.03415	0.004802	
16	2.865	0.4287	0.03606	0.002538	0.985	0.1903	0.01654	0.002045	0.03348	0.004703	
17	2.789	0.4169	0.03525	0.002478	0.956	0.1845	0.01625	0.002006	0.03283	0.004606	775
18	2.717	0.4056	0.03448	0.002422	0.928	0.1789	0.01597	0.001970	0.03220	0.004514	
19	2.647	0.3948	0.03376	0.002369	0.902	0.1737	0.01570	0.001935	0.03161	0.004426	
20	2.582	0.3846	0.03308	0.002319	0.878	0.1688	0.01545	0.001901	0.03102	0.004339	
21	2.517	0.3745	0.03243	0.002270	0.854	0.1640	0.01522	0.001869	0.03044	0.004252	702
22	2.456	0.3648	0.03180	0.002222	0.829	0.1590	0.01498	0.001838	0.02988	0.004169	
23	2.396	0.3554	0.03119	0.002177	0.804	0.1540	0.01475	0.001809	0.02934	0.004087	
24	2.338	0.3463	0.03061	0.002133	0.781	0.1493	0.01454	0.001780	0.02881	0.004007	
25	2.282	0.3375	0.03006	0.002091	0.759	0.1449	0.01434	0.001751	0.02831	0.003931	639
26	2.229	0.3290	0.02952	0.002050	0.738	0.1406	0.01413	0.001724	0.02783	0.003857	
27	2.177	0.3208	0.02901	0.002011	0.718	0.1366	0.01394	0.001698	0.02736	0.003787	
28	2.128	0.3130	0.02852	0.001974	0.699	0.1327	0.01376	0.001672	0.02691	0.003718	
29	2.081	0.3055	0.02806	0.001938	0.682	0.1292	0.01358	0.001647	0.02649	0.003651	586
30	2.037	0.2983	0.02762	0.001904	0.665	0.1257	0.01342	0.001624	0.02608	0.003588	
35	1.831	0.2648	0.02546	0.001733	0.592	0.1105	0.01256	0.001501	0.02440	0.003315	
40	1.660	0.2361	0.02369	0.001586	0.530	0.0973	0.01184	0.001391	0.02306	0.003082	
45	1.516	0.2110	0.02238	0.001466	0.479	0.0860	0.01130	0.001300	0.02187	0.002858	
50	1.392	0.1883	0.02134	0.001359	0.436	0.0761	0.01088	0.001216	0.02090	0.002657	
60	1.190	0.1480	0.01954	0.001144	0.359	0.0576	0.01023	0.001052	0.01946	0.002274	
70	1.022	0.1101	0.01825	0.000926			0.00977	0.000851	0.01833	0.001856	
80	0.917	0.0765	0.01770	0.000695			0.00958	0.000650	0.01761	0.001381	
90	0.84	0.041	0.01735	0.00040			0.0095	0.00038	0.0172	0.00079	
100	0.81	0.000	0.0170	0.00000			0.0095	0.00000	0.0170	0.00000	

Table 3.2

Solubility of gases occurring in sediments (Dean, 1973).

α - volume of gas at S.T.P. which will dissolve in 1 volume of water
 q - weight of gas (gm.) which dissolves in 100 gm. water at a total pressure of 1 atmosphere.

limited by the scarcity of the nitrate radical. This situation may be very different in modern sediments in river estuaries and bays where nitrate leached from fields where it had been spread for agricultural purposes could change the idealised section of fig. 3.2.

3.3 SOLUBILITY OF GASES.

Since the main concern here is with the acoustic effects of gases in sediments, it is important to distinguish between those which are readily soluble in water and those which are not. A dissolved gas does not materially affect the compressibility of a liquid, whereas a gas present in the form of bubbles does do so. The dissolved gas, carbon dioxide may play a part in the production of the bicarbonate radical and the magnesium carbonate molecule which have been discussed by Mellen and Browning (1981) in connexion with the absorption of sound in sea water. This acoustic effect (ca. 1dB/km. at 10kHz in seawater) is small and would not be observable in the presence of the much higher attenuations encountered in sediments. Table 3.2, compiled from a series of tables in Dean (1973) shows the solubility of the common gases detected in sediments. The lowest solubilities are found in nitrogen, oxygen and methane. More soluble by two orders of magnitude are carbon dioxide and hydrogen

sulphide (which arises by the continued reduction of sulphate beyond the stage shown in table 3.1). Ammonia is more soluble again than these, by nearly three orders of magnitude. Consequently, we may expect that any ammonia formed in a sediment even if in large quantities will remain dissolved. Although carbon dioxide and hydrogen sulphide may form bubbles they must be produced in quantities a hundred times greater than methane or nitrogen to have the same effect. From this, it appears probable that the gas which will contribute most to the appearance of bubbles in a sediment will be methane, with some contribution from nitrogen in zones where nitrate reduction can proceed.

3.4 PROPAGATION OF SOUND IN WATER CONTAINING BUBBLES.

The two-phase medium of a liquid with gas bubbles is more amenable to theoretical analysis than is the more complex three-phase medium of a sediment. An early dynamic model of such an emulsion was given by Wood (1931) whose treatment consisted of deriving a mean compressional-wave velocity for the gasified sediment:

$$c = \sqrt{E/\rho} \quad \dots\dots\dots(3.1)$$

on the basis of mean density ρ and elasticity E

where

$$\rho = n\rho_1 + (1-n)\rho_2 \quad \dots\dots\dots(3.2)$$

and

$$1/E = n/E_1 + (1-n)/E_2 \quad \dots\dots\dots(3.3)$$

where n is the fraction of gas by volume and where the subscript "1" refers to the gas and "2" to the gas-free sediment. (See McKay 1980a, included with this work, for a further discussion of the derivation of equations 3.2 and 3.3). Wood's theory predicts the variation of sound speed with gas content shown in fig.3 of McKay 1980b (also included with this work), but does not consider the effects of bubble size and acoustic frequency. The importance of the existence of a resonant frequency for any given size of bubble was noted by Minnaert (1933) who derived this frequency as:

$$\nu = (1/2\pi r_0) \sqrt{(3\delta P_0 / \rho_w)} \quad \dots\dots\dots(3.4)$$

where ν = frequency (Hz)

r_0 = bubble radius

δ = ratio of specific heats of the gas.

P_0 = hydrostatic pressure

ρ_w = water density

It is of interest to note that this formula contains neither the density of the gas nor the elasticity of the water. Most acoustic phenomena associated with gas-liquid emulsions can be treated accurately by assuming a liquid-dominated density and a gas-dominated elasticity. Minnaert was principally concerned with bubbles as acoustic sources, but the later work of Spitzer (1943), Silberman (1957) and Domenico (1982) has led to

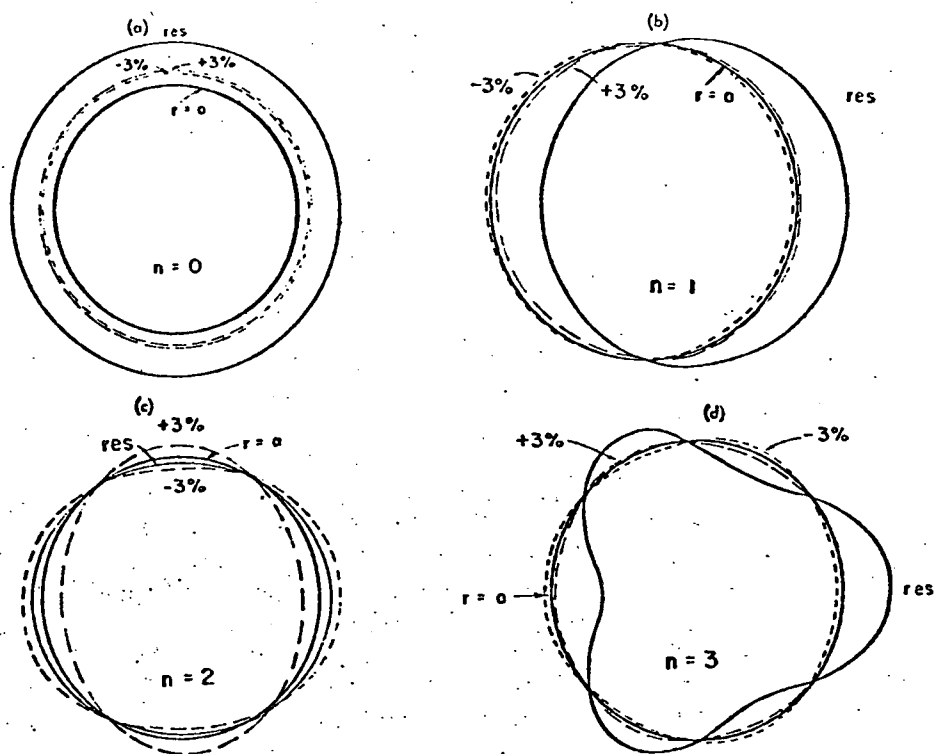


Fig. 3.3

The first few modes of oscillation of a gas bubble. Of these, the "n=0" or "monopole" mode has been used by most authors who have given theoretical values for attenuation. (From Uberall et al. 1979).

expressions for the attenuation and sound speed in gasified liquids. Anderson and Hampton (1980a and b) have summarised elaborations to the frequency expression of Minnaert which have been made by Briggs et al. (1947), Houghton (1963) and Shima (1970) in order to take into account the effects of surface tension and viscosity. Uberall et al. (1979) have discussed different modes of oscillation of an included bubble (fig.3.3). Gaunaurd and Uberall (1981) have calculated the variations in sound speed and attenuation which result from monopole oscillations of bubbles of 3mm. and of 1micron diameters. Van Wijngaarden (1980) has done this for bubbles of 50 micron diameter. Domenico (1982) has summarised the results and concludes that below the bubble resonant frequency, the acoustic velocity is nearly independent of frequency and appreciably lower than the velocity in bubble-free water. Near the resonant frequency the velocity decreases to a sharp minimum, followed by a broad maximum well above the velocity in water. Further increase in frequency leads to a decrease in velocity until that of the bubble-free water is reached. Initially, attenuation increases gradually with frequency and rises rapidly near resonance to a value several orders of magnitude above that at low frequency. Beyond resonance, the attenuation decreases gradually. Figure 3.4 illustrates these findings. The mechanisms of acoustic energy loss according to Domenico (1982) are

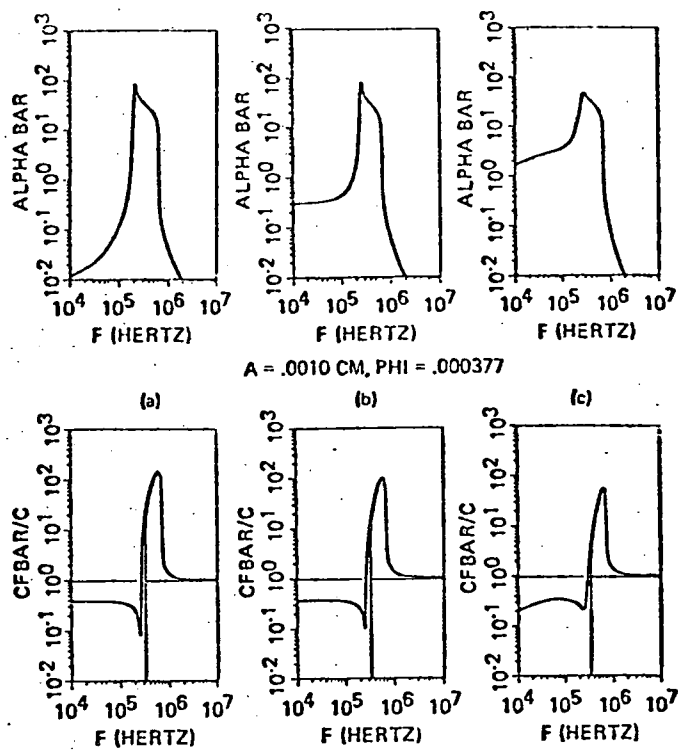


Fig. 3.4

Attenuation and sound speed (relative to the speed in water) for a gaseous volume concentration of 3.77×10^{-4} and a bubble radius of 10^{-3} cm. The case (a) considers energy loss only by re-radiation from pulsating bubbles. Along with re-radiation, case (b) includes viscous damping, and case (c) adds to these the effect of thermal losses (Gaunard and Uberall, 1981).

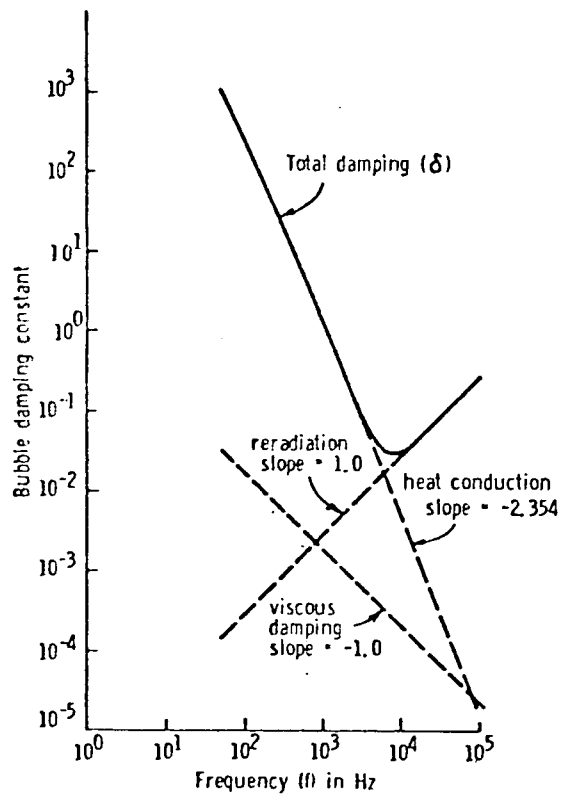


Fig. 3.5

The relative importance of re-radiation, heat conduction and viscous damping as loss mechanisms. The figure (Domenico, 1982) refers to bubbles of radius 0.06 cm.

illustrated in fig. 3.5.

3.5 ACOUSTIC EFFECTS OF GAS IN SEDIMENTS.

As with liquids, the presence of gas in a sediment radically alters its acoustic properties. In the literature, attenuation is often reported as a problem rather than as the subject of a measurement programme. Laboratory studies on sediment samples can become difficult when the sample is not fully water saturated, and in this context, Nyborg et al. (1950) have reported attenuations of 2600dB/m. at 110KHz and Wood and Weston (1964) noted values of 1750dB/m. at 8KHz. These values are at least three orders of magnitude greater than what is typical of a water-saturated fine grained sediment as illustrated in fig.3.6. Similar effects were found to occur at higher frequencies when measurements were made of velocity and attenuation in sand samples which had been re-saturated (McKay, 1967). At 600KHz, attenuations exceeding 2000dB/m. prevented the making of accurate measurements in these samples of either velocity or attenuation when the transmitted acoustic pulse was reduced to the noise level at the receiver after passage through the sediment. A value of about 200dB/m. at that frequency was typical of a sand sample which had been taken underwater and prevented from drying before

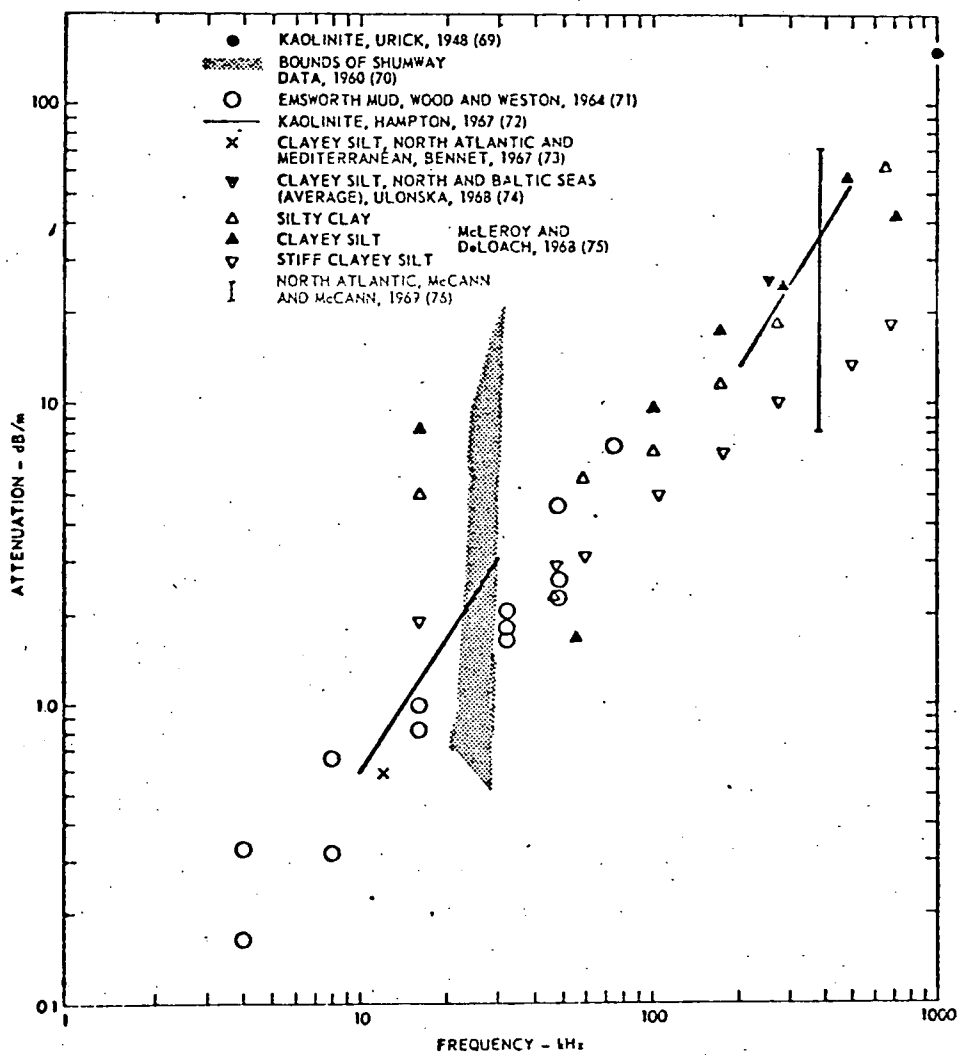


Fig.3.6

This figure from Hampton and Anderson (1974) shows accumulated experimental values for attenuation in water-saturated sediments.

measurements were made.

Theoretical treatment of bubbles in sediments is based on the theory of bubbles in liquids as discussed in section 3.4 above, but when a bubble occurs in a solid, then the restoring force during bubble oscillation will depend in part on the shear modulus of the surrounding solid. Blake (1952) approached the problem by treating an included bubble as a void cavity. He obtained an expression for the ratio of pressure to particle velocity at the cavity surface and equated resonance with the occurrence of a zero in the reactive part of this impedance. The expression he obtained for the resonant frequency, although similar to that of Minnaert (1933) included the effect of rigidity:

$$f_0 = [(\lambda + 2\mu)/\rho \cdot 2(1-2\sigma)/(3\sigma-1)]^{1/2} / 2\pi r \dots (3.5)$$

where f_0 is resonant frequency

r is cavity radius

λ, μ are Lamé parameters for the solid

ρ is density of the solid

σ is Poisson's ratio for the solid

Anderson and Hampton (1980a) point out that this reduces to:

$$f_0 = (4\mu/\rho)^{1/2} / 2\pi r \text{ for } \sigma = 1/2 \dots (3.6)$$

an expression identical with that of Meyer and Skudrzyk (1953) who were considering the motion of the wall of an empty cavity in rubber. The inclusion of the elastic

effect of the gas was given by Andreeva (1964), who was studying the fundamental pulsation frequency of a gas bubble in fish tissue, as:

$$f_0 = [(3\gamma P_0 + 4\mu)/\rho]^{1/2}/2\pi r \quad \dots\dots\dots(3.7)$$

where the parameters are those given along with equations 3.4 and 3.5.

Andreeva's expression reduces to that of a liquid (3.4) when the shear modulus $\mu=0$.

A model for the velocity of propagation in a gasified sediment has been given by Anderson and Hampton (1980b). They adopt the approach of Wood (1931) as discussed above in section 3.4. By considering the additional gas phase along with the pore fluid and particle phases they derive a composite bulk modulus:

$$K = K_g K_w K_m / \{n_g K_w K_m + (1-n) K_w K_g + (n-n_g) K_m K_g\} \quad \dots(3.8)$$

where K = aggregate bulk modulus

K_w = pore water bulk modulus

K_m = mineral particle bulk modulus

K_g = gas bulk modulus ($= \gamma P_0$ for adiabatic compression)

γ = ratio of specific heats of gas

P_0 = ambient hydrostatic pressure.

From this they give the compressional wave velocity as: $c_p = \sqrt{(K/\rho)}$ $\dots\dots\dots(3.9)$

This would be valid for a fluid sediment.

When rigidity is incorporated, the expression is:

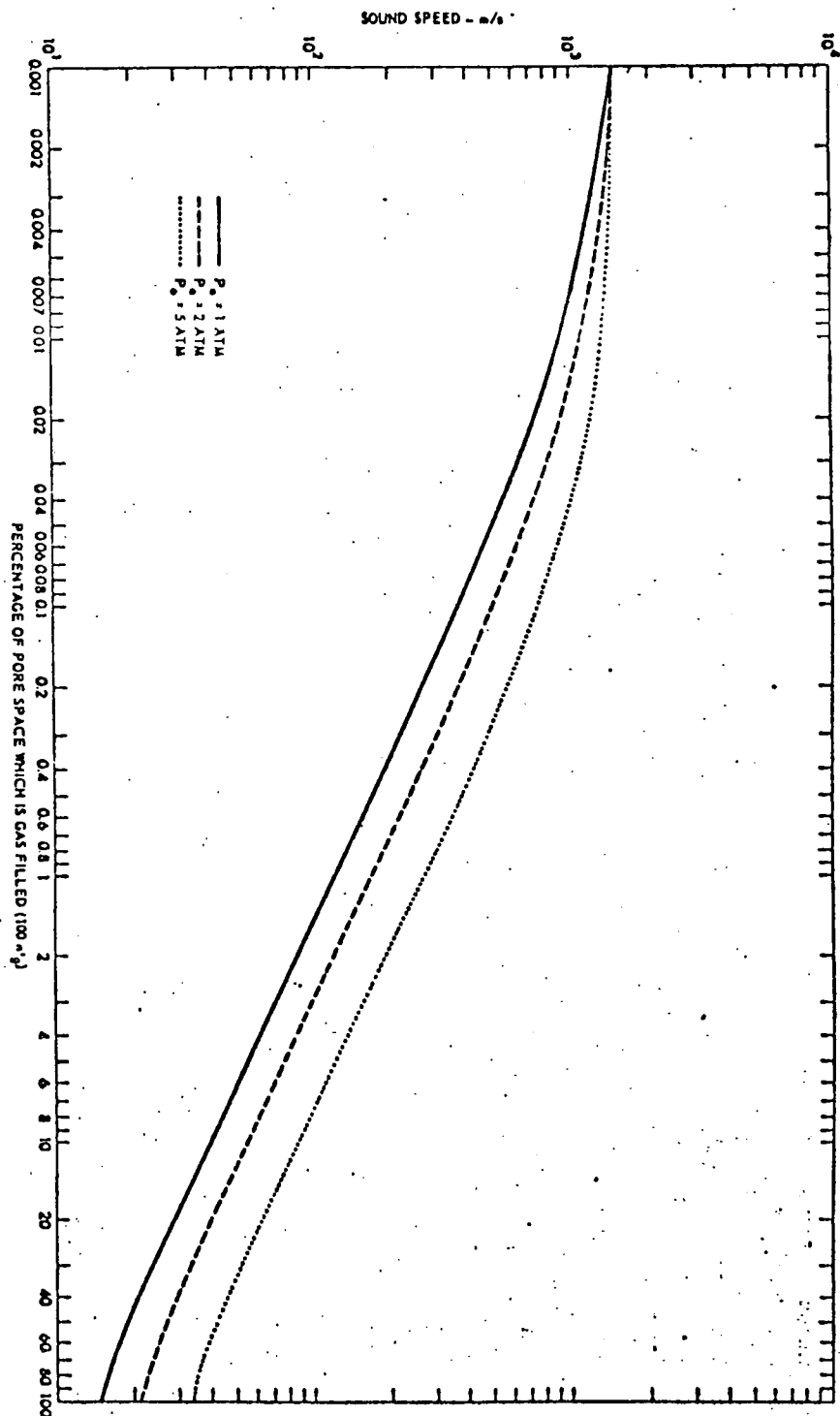


Fig. 3.7

The variation of compressional wave velocity with the fraction of pore space which is gas-filled. Overall porosity is 60%. The curves are for three ambient pressures: 1, 2 and 5 atmospheres (from Anderson and Hampton, 1980 b.).

$$c_p = \sqrt{(K + 4\mu/3)/\rho} \dots\dots\dots(3.10)$$

where μ = the sediment rigidity modulus.

The variation of sound speed with gas content in a silt having 60% porosity is illustrated in fig 3.7 where the curves, for three different ambient pressures are calculated from equation 3.9, and thus valid for frequencies lower than the bubble resonance frequency given by equation 3.7. By considering resonance effects and by including the loss mechanisms of fig 3.5, Anderson and Hampton (1980b) deduce the following relationship for the sound speed:

$$(c_0/c)^2 = \frac{1}{2} \left[1 + \frac{KX_1}{\delta P_0 + 4\mu/3} \right] \left[1 + \left\{ 1 + \left(\frac{KY_1/(\delta P_0 + 4\mu/3)}{1 + KX_1/(\delta P_0 + 4\mu/3)} \right)^2 \right\}^{1/2} \right] \dots\dots\dots(3.11)$$

where

$$K = K_m [(K_f + Q') / (K_m + Q')]$$

$$Q' = K_w (K_m - K_f) / n (K_m - K_w)$$

$$X_1 = n_g (1 - \phi^2) / ((1 - \phi^2)^2 + \delta^2)$$

$$Y_1 = n_g \delta^2 / ((1 - \phi^2)^2 + \delta^2)$$

$$\phi = f/f_0$$

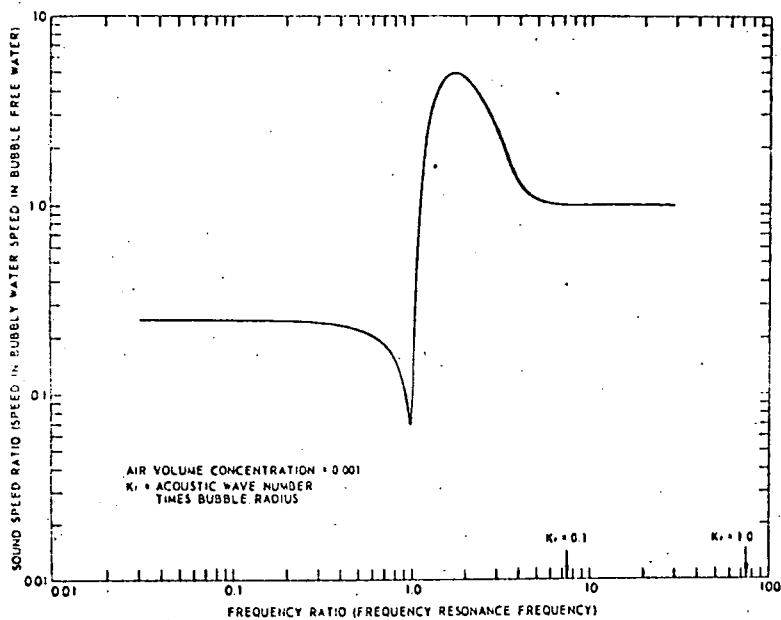
$$\delta = d\phi^2$$

d = damping constant discussed by Anderson
and Hampton, (1980a)

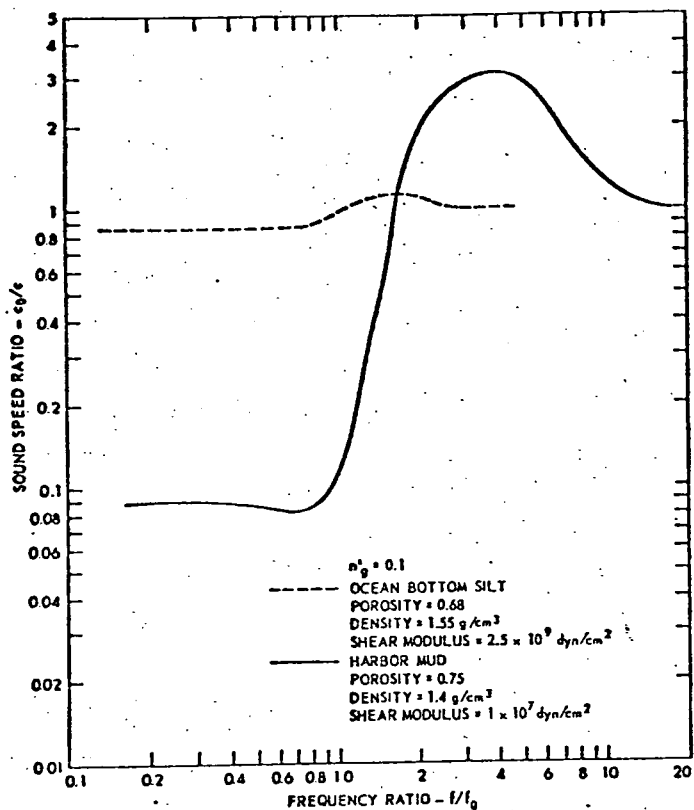
K_f = sediment frame bulk modulus

n = fractional porosity

Figure 3.8 illustrates the difference to be expected on the basis of equation 3.11 between the variation of sound



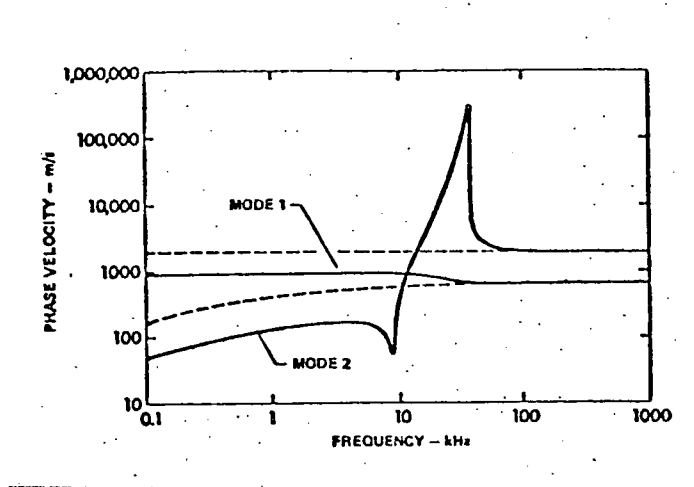
Sound speed ratio versus frequency ratio for water plus a collection of single sized bubbles.



Gassy sediment sound speed ratio versus frequency ratio.

Fig. 3.8

A comparison between the curves for sound speed ratio (speed in bubbly medium to speed in bubble-free medium) versus frequency ratio (frequency to resonant frequency) as given by Hampton et al. (1974) for water (above) and by Anderson et al. (1980) for sediment.



dashed line bubble free
solid line with bubbles

Fig. 3.9

The variation of compressional wave velocity with frequency given by Bedford and Stern (1983) in their treatment of sediment gasification by the Biot-Stoll method.

speed with frequency in water and in sediment. The bubble resonance is less pronounced in the sediment than in the water and the prediction for an ocean bottom silt shows the trough and crest near resonance to be greatly flattened, an effect attributable to the damping by internal friction of the vibration of included bubbles.

Bedford and Stern (1983) have approached the problem of wave propagation in a gassy sediment from the stand-point of a Biot-Stoll sediment model (see section 8.1). Their results are illustrated in figure 3.9. As that theory requires, there are now two compressional waves. The Bedford and Stern result shows that the slow wave (mode 1) has a velocity at low frequency which is considerably higher than for a bubble-free sediment. The normal compressional wave (mode 2) still exhibits a rather sharper peak and trough than for the Anderson and Hampton model (fig. 3.8).

3.6 EXPERIMENTAL RESULTS AND FIELD OBSERVATIONS ON GASIFIED SEDIMENTS.

Apart from the casual observations of very high attenuation values (referred to in section 3.5), laboratory studies on gasified sediment samples are rare. Anderson and Hampton (1980b) have measured compressional

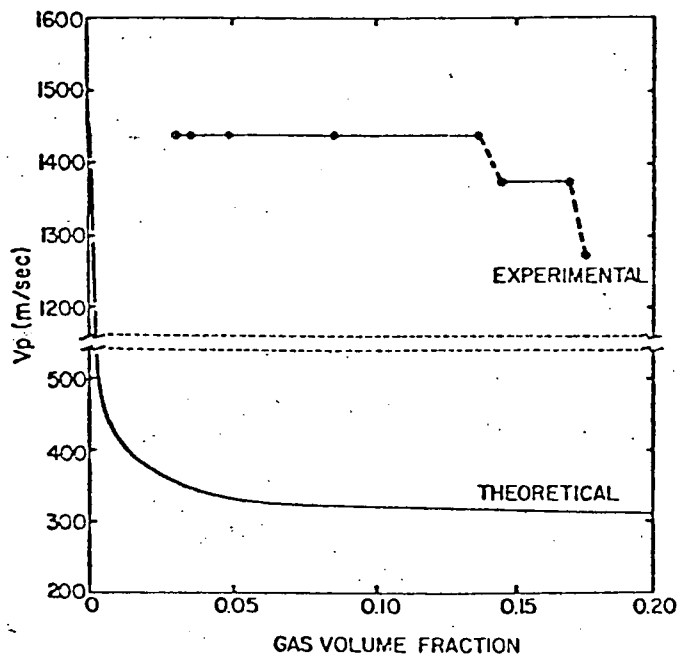
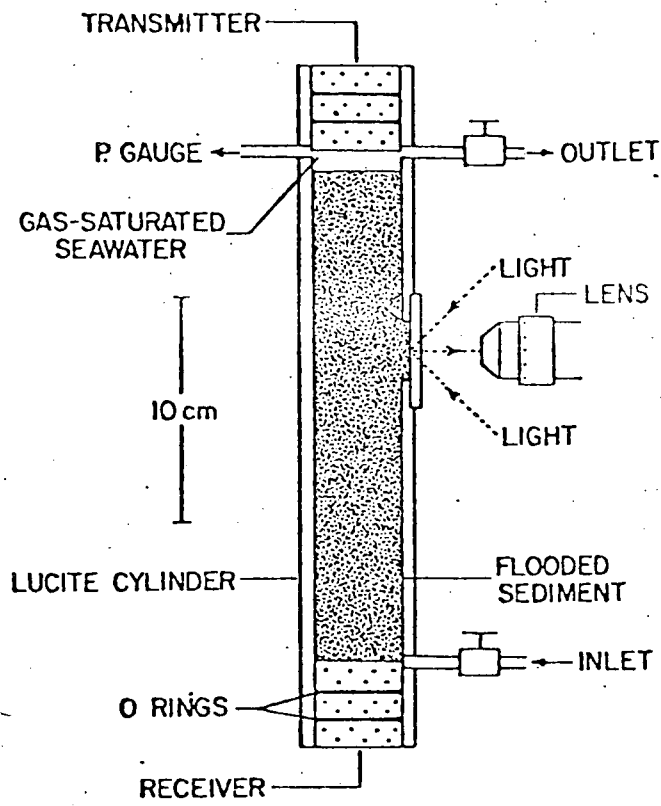


Fig. 3.10

Apparatus of Kepkey and Cooke (1978) together with a comparison between their measured values of compressional velocity and the prediction of Anderson (1974).

wave velocities in kaolinite and sand at 190KHz over a 7.6cm path length, and verified that the velocity was unchanged on the insertion of bubbles in the form of air sacks of resonant frequency 60KHz in sand. This is in agreement with their theory for high frequencies, but they do not appear to have made experimental verification at other frequencies. Kepkay and Cooke (1978) measured velocity at varying concentrations of methane bubbles in a sediment column, and obtained results showing a stepwise variation with gas content (fig. 3.10). This finding seems so peculiar and at variance with all theoretical treatments that it is difficult to accept without stronger verification than they give that propagation through the walls of their sediment tube could not have contributed to spurious values. Another possibility is that they have encountered uneven distributions of bubbles giving rise to complex resonance effects.

Field measurements of acoustic velocity have been made in various locations where gasified sediments occur. Lewis (1966) is reported by Anderson and Hampton (1980b) to have made velocity measurements in Boston Harbour at frequencies below 1KHz. He found values from 2 to 20% lower than in nearby ungasified sediments. Edgerton (1965) had reported these same gasified sediments as strong reflectors at 12KHz and Anderson and Hampton take this as an indication that they may be resonant at about

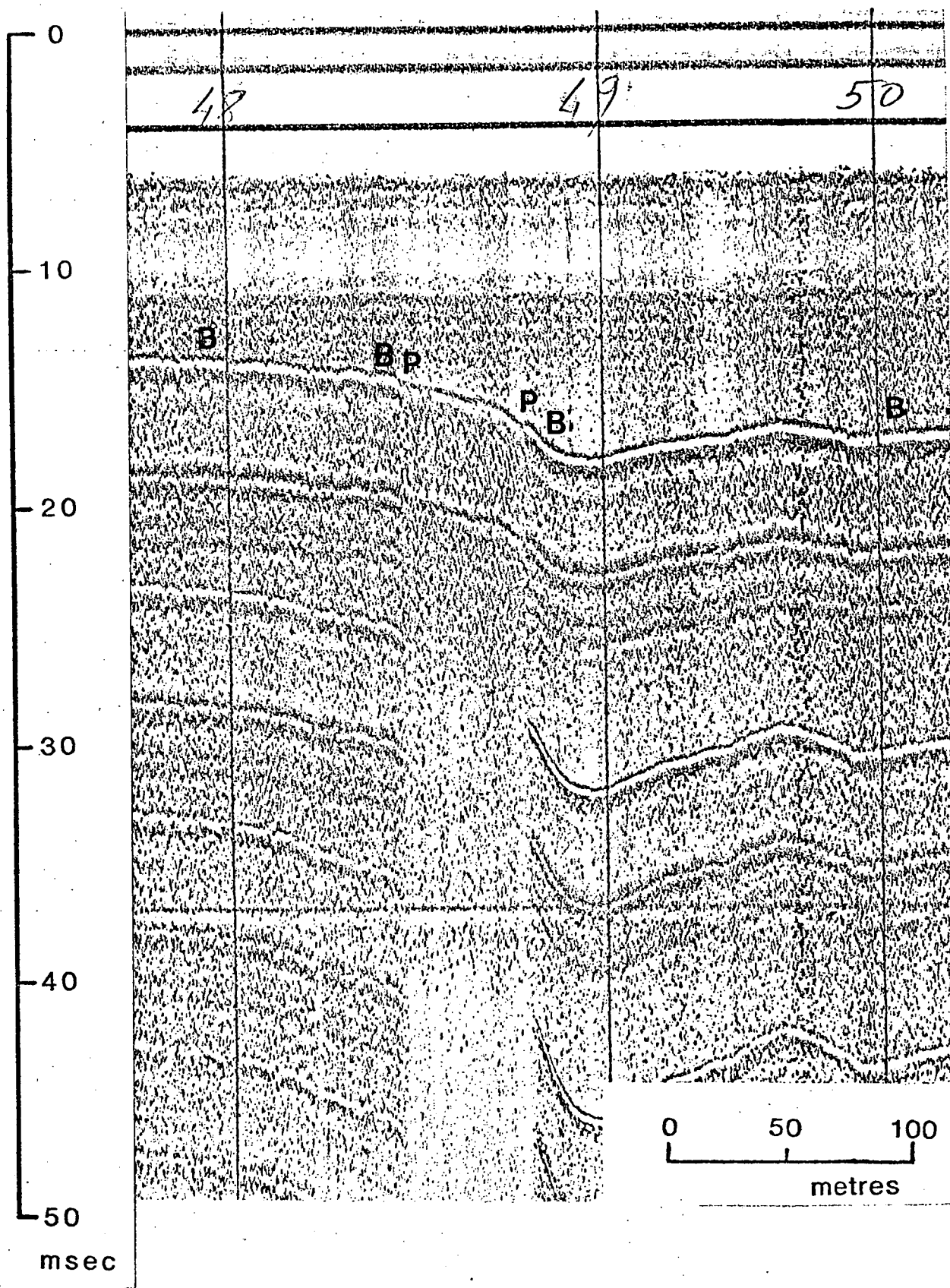


Fig. 3.11

Phase reversal in the bottom-reflected arrival indicating a low-velocity sediment and presumed gasification. Strong multiple reflections emphasize the magnitude of the impedance contrast at the seabed. "Normal" bottom between P-P; gasified bottom at B-B.

this frequency, so that the two sets of observations can be considered consistent with their proposed model. Another field location, Lake Travis in Texas has also been investigated independently at two different frequencies. Jones et al. (1964) observed reversed phase on reflexions from the lakebed at frequencies below 3200KHz, indicating a sound speed in the sediments below that in the water. Subsequent measurements by Muir (1972) at 40Khz gave a velocity the same as that of the water to within 1%. These findings also agree with the Anderson and Hampton (1980b) model if the resonant frequency lies between the frequencies at which the two measurements were made. The findings agree equally well with the model proposed by Bedford and Stern, given all the uncertainties which arise as to bubble size and sediment shear modulus, and considerably more detailed work will be required before either of these two models is to be the preferred one, on an empirical basis.

The commonest way in which gasified sediments are observed acoustically is through their effect on the appearance of a vertical incidence reflexion profile. This may take a number of forms. Figure 3.11 is a section which was obtained in the Miramichi Estuary of New Brunswick (for location see fig 1.8 and Howells and McKay (1977)). The acoustic source was a single-tip sparker firing at 200 Joules of electrical input. The phase

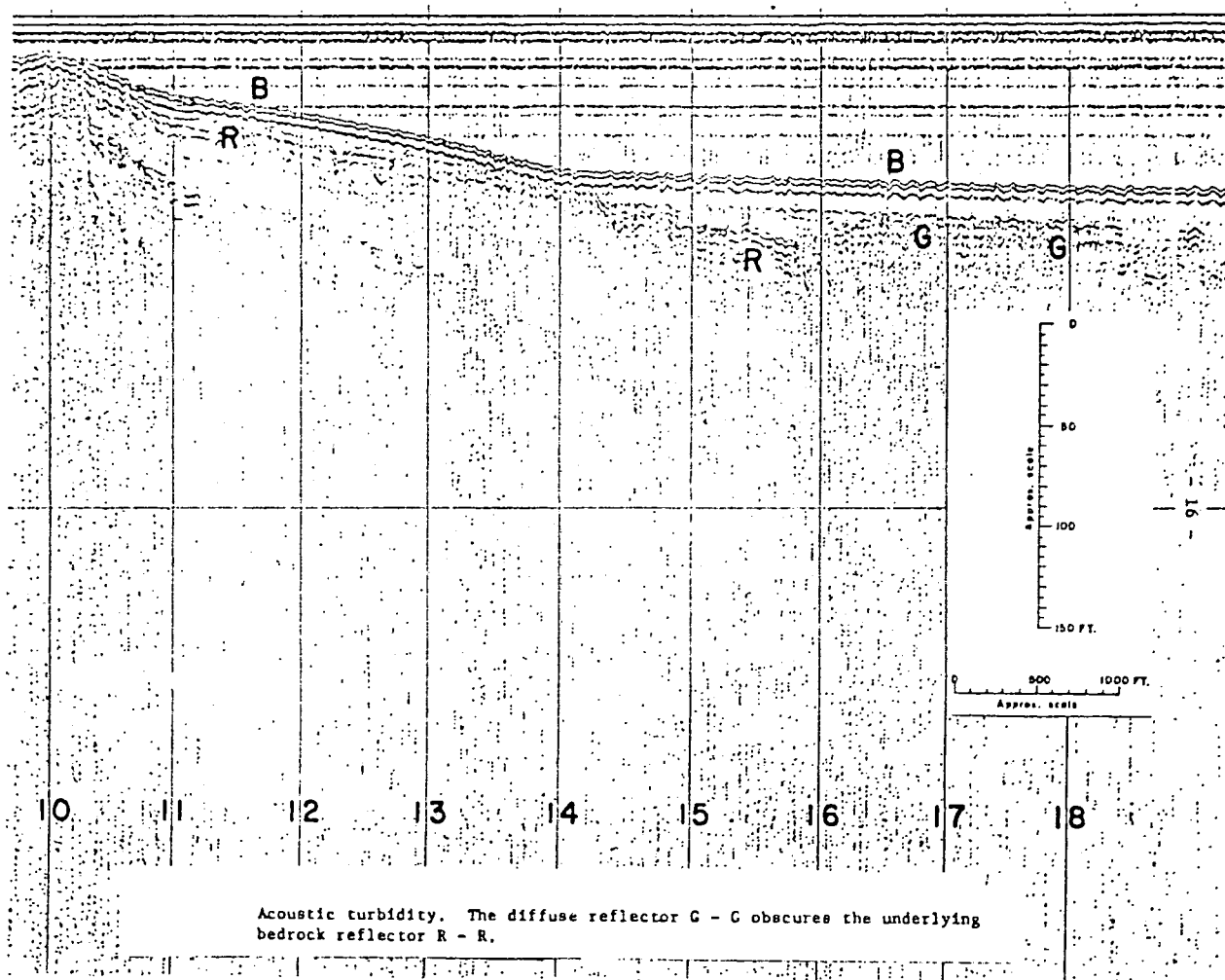


Fig. 3.12

Country Harbour, Nova Scotia (100 miles east of Halifax Harbour - see Fig. 1.8). A diffuse reflector "G" obscures the reflection from rock "R", although the seabed reflector "B" remains unchanged (see also Fig. 3.13).

reversal of the sea-bottom reflexion which takes place at the points 'P' is quite clear. The strong multiple reflexions point to the existence of a high acoustic impedance contrast at the seabed. These observations are consistent with the expected reflexion from a gasified sediment at frequencies below bubble resonance. More common in estuaries and bays in Eastern Canada is the appearance of profiles such as those of figs. 3.12 and 3.13, where the bottom reflexion remains unchanged in character, but there appears a reflector 5 msec. or less below the bottom which obscures all expected deeper reflectors. The examples shown are from bays on the south coast of Nova Scotia. Sometimes such an obscuring reflector may take on a more hummocky appearance (fig 3.14), but usually they remain parallel to the seabed and can be classified as "bottom simulating reflectors" along with the phenomena reported by Hein et al. (1978) in the Bering Sea (fig. 3.15) and by Ewing and Hollister (1970) in the Blake-Bahama Outer Ridge (fig. 3.16) and with the ice-bonding reflectors to be discussed in section 4. The entirely different ways in which such acoustic reflectors can form quite unrelated to stratigraphic horizons makes it important to consider the different possible modes of formation before identifying the reflector unequivocally with any particular one of them. The reflector of fig. 3.15 has been linked speculatively by Scholl and Creager (1973) and by Fullam et al. (1973) with the diagenesis of

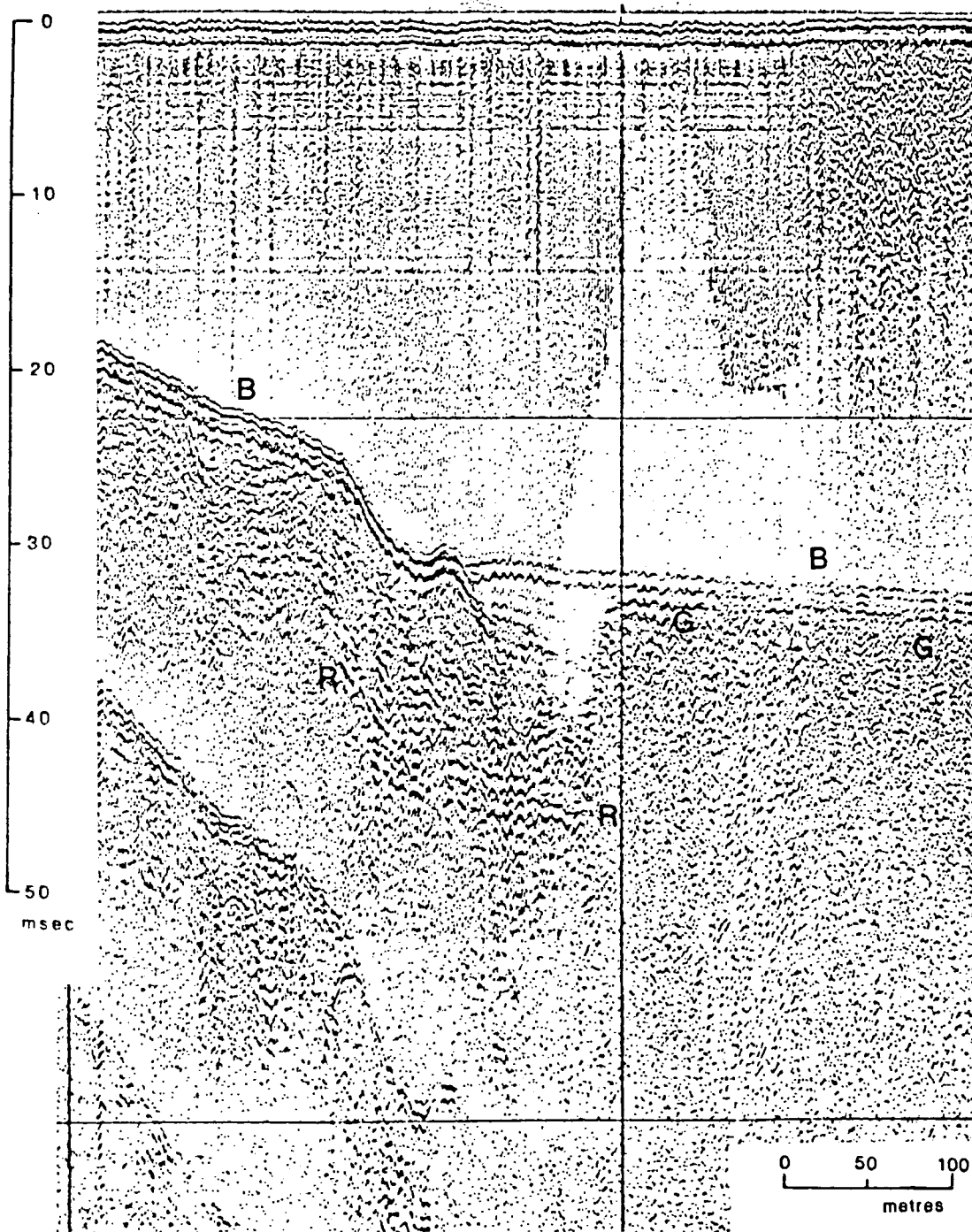


Fig. 3.13

St. Margaret's Bay, Nova Scotia (20 miles west of Halifax Harbour - see Fig. 1.8). Another example of an obscuring reflector "G" (see also Fig. 3.12).

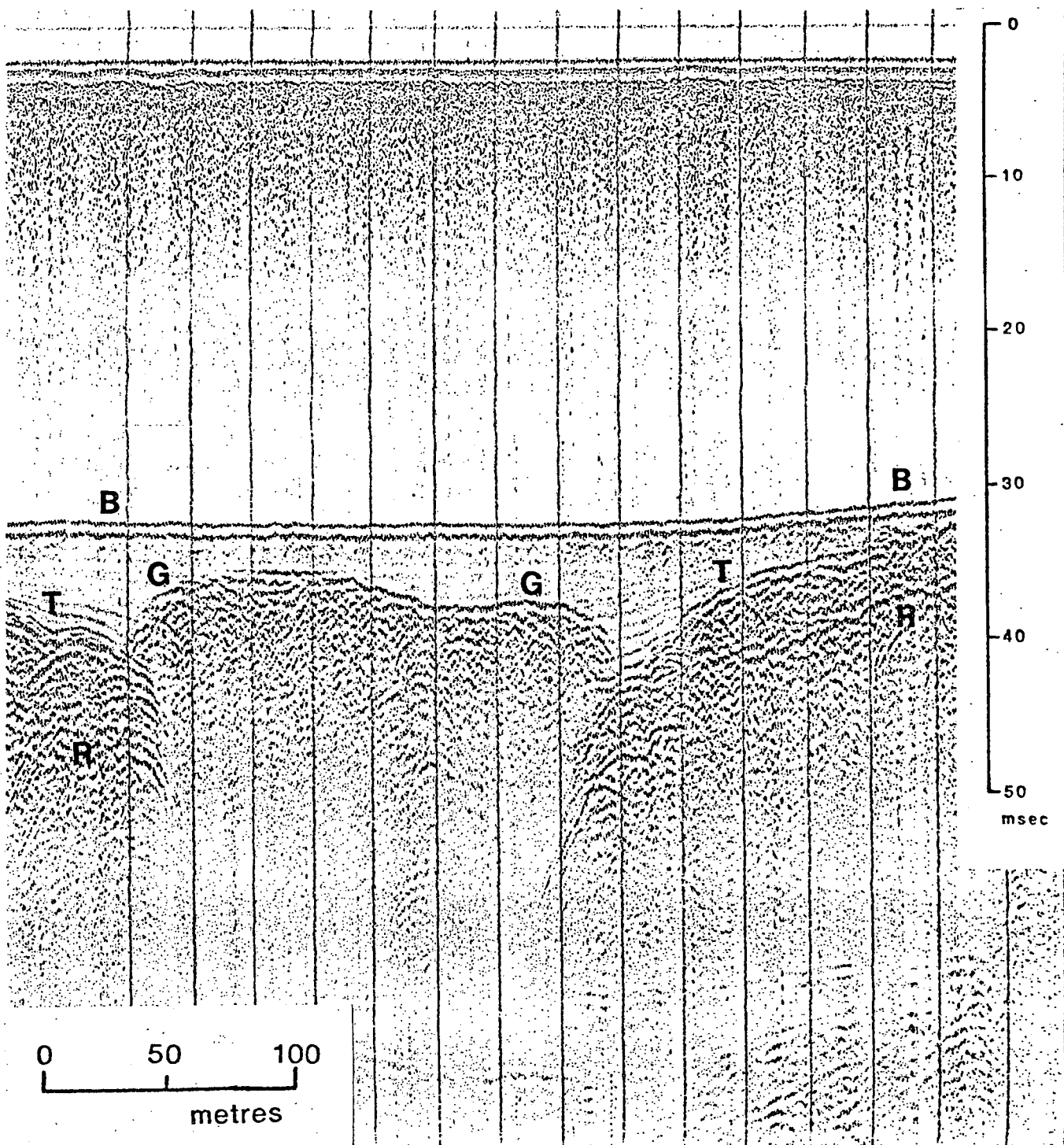


Fig. 3.14

Fairview Cove, Halifax Harbour (see Fig. 1.2 for location). In this section, the reflector (G) which obscures the expected deeper interfaces between fine sediment and till (T) and between till and rock (R) has a more hummocky appearance.

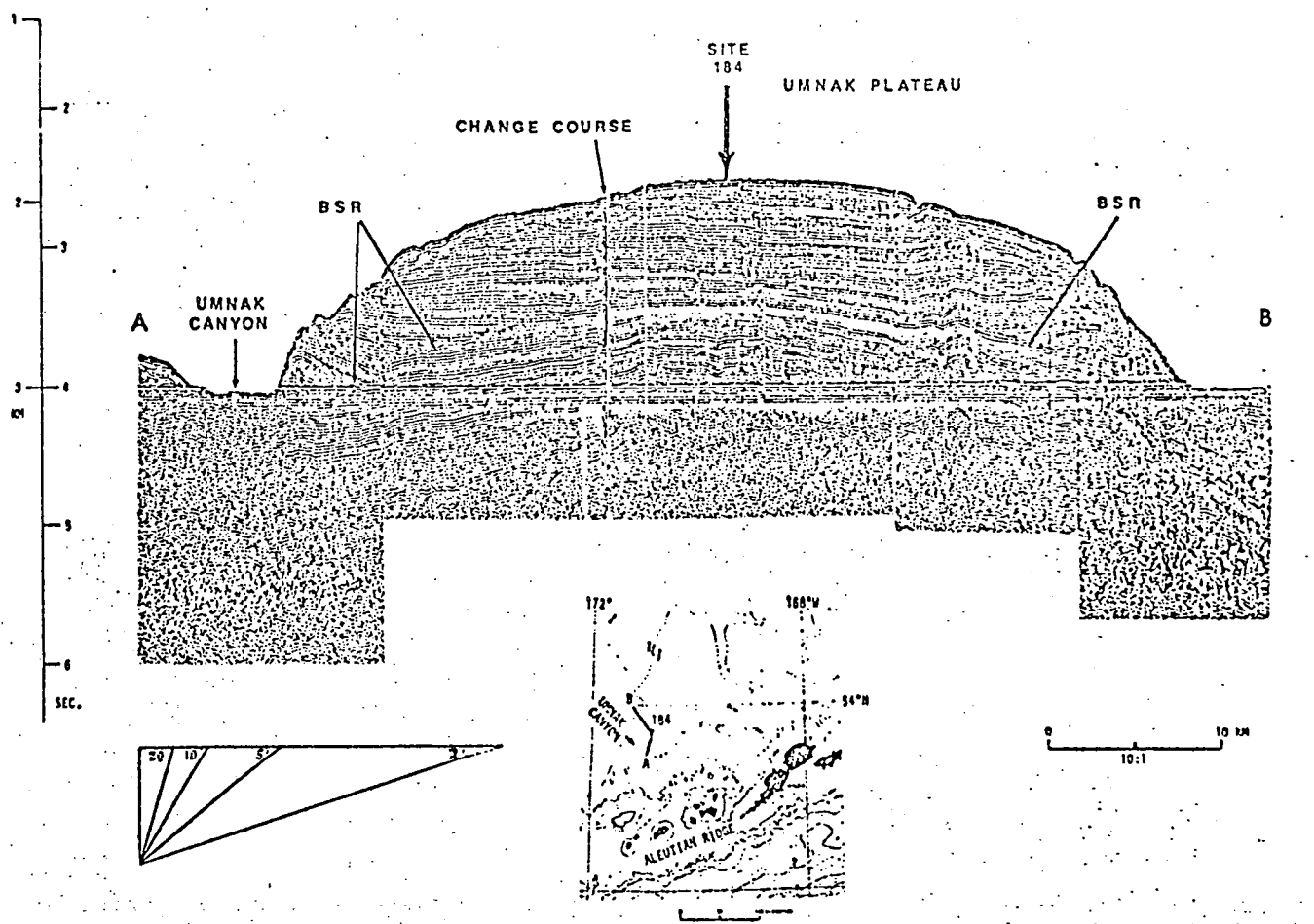
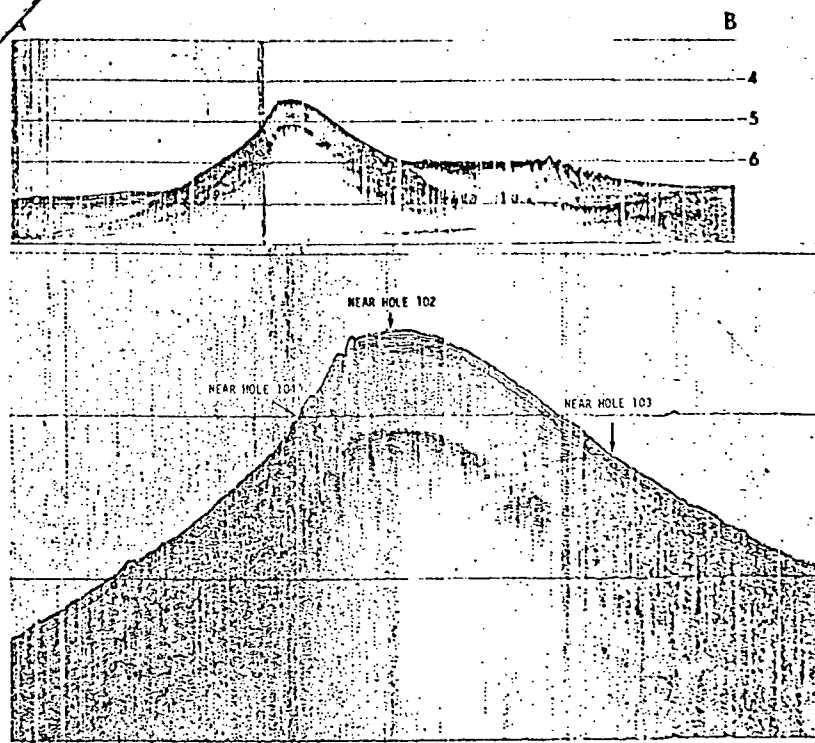
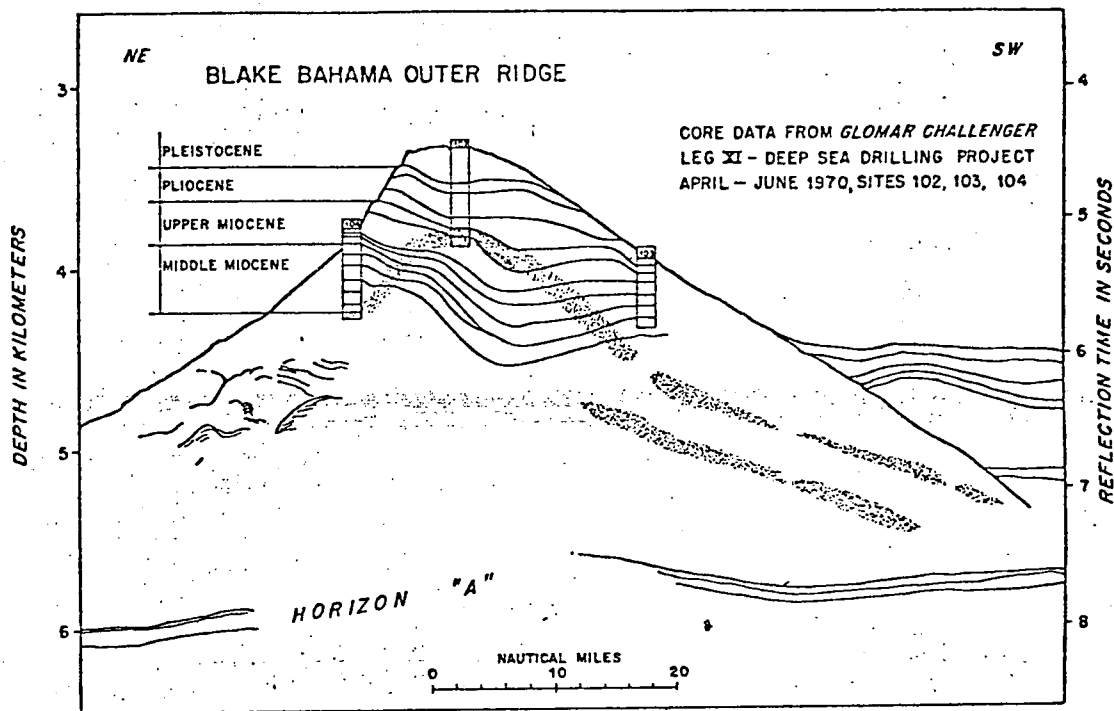


Fig. 3.15

The bottom-simulating reflector (BSR) reported by Hein et al.(1978) in the Bering Sea.



Vema 21 Seismic profiler section across the Blake-Bahama Outer Ridge. (crestal region expanded below).
See Figure 1 for location.



Tracing of the seismic profiler record across the Blake ridge with drill sites indicated. Bedding planes in the crestal region are drawn on the basis of biostratigraphic correlations in the three holes and on behavior of seismic reflectors. Stippling indicates a conformable reflector in the crestal region and diffuse reflectors on the southwest flank. Bedding is disturbed on the northeast flank. See Site Reports for a reproduction of the seismic record.

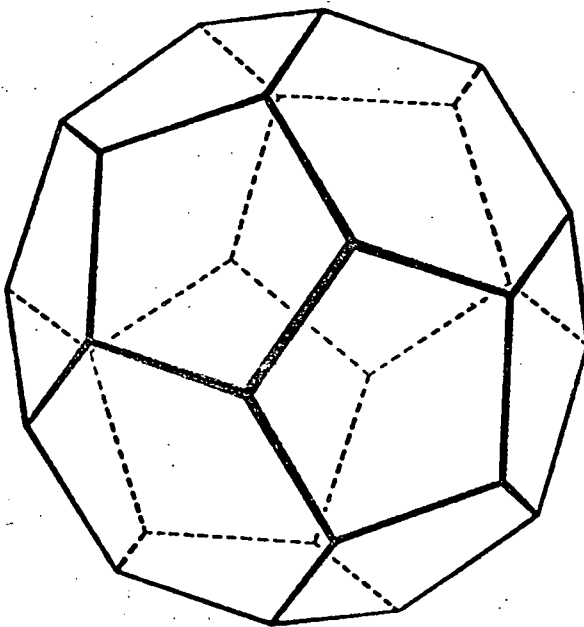
Fig. 3.16

Bottom-simulating reflector in the Blake-Bahama Outer Ridge (Ewing and Hollister, 1970). Upper: seismic profiles. Lower: interpretation showing how the reflector is discordant with the stratigraphy deduced from boreholes.

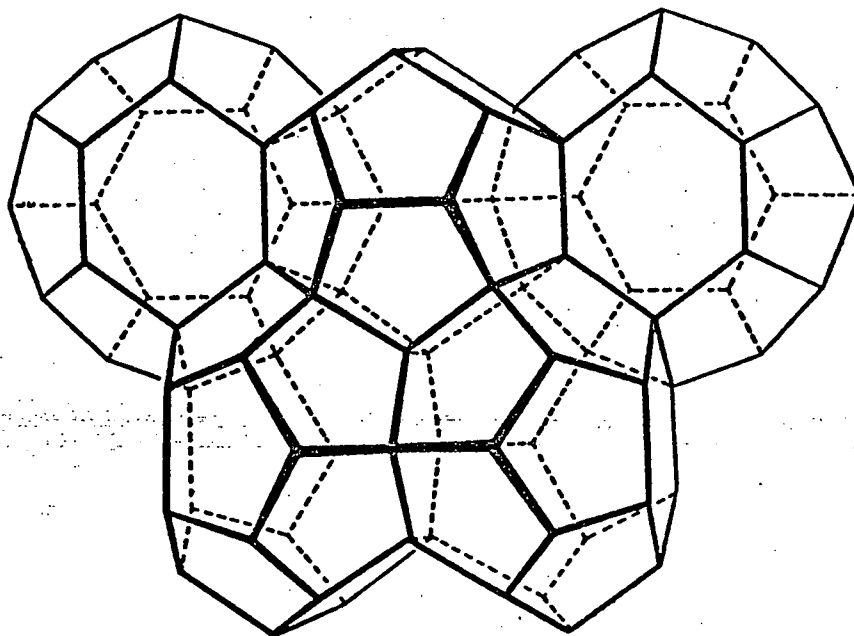
diatomaceous deposits. They suggest that the reflector represents the transition of silica from the opal-A to the opal-CT phase, a transition which is controlled by temperature. Hammond and Gaither (1983) support this hypothesis, having dismissed on temperature-pressure considerations the formation of gas hydrate as a possible agent, as proposed by Kvenvolden and McMenamin (1980) and Shipley et al. (1979), (see section 3.7). The importance of evolving seismic methods of distinguishing the agent responsible for producing non-stratigraphic reflectors will become apparent in the context of engineering operations in the Beaufort Sea (section 4.5).

3.7 GAS HYDRATES.

An important implication of the presence of gas in sediments is the possibility that the gas will, in reaction with the pore water, form a meta-stable compound known as a clathrate hydrate or gas hydrate. Such compounds exist only for certain conditions of ambient pressure and temperature (high pressure and low temperature or both). They are solids which can have a bonding effect on unconsolidated sediments and give them acoustic properties akin to consolidated rocks. This effect is similar to the bonding produced by ice, discussed in chapter 4.



The hexakaidecahedron formed by 28 water molecules in the 17 Å lattice. There are 4 hexagonal faces and 12 pentagonal faces.



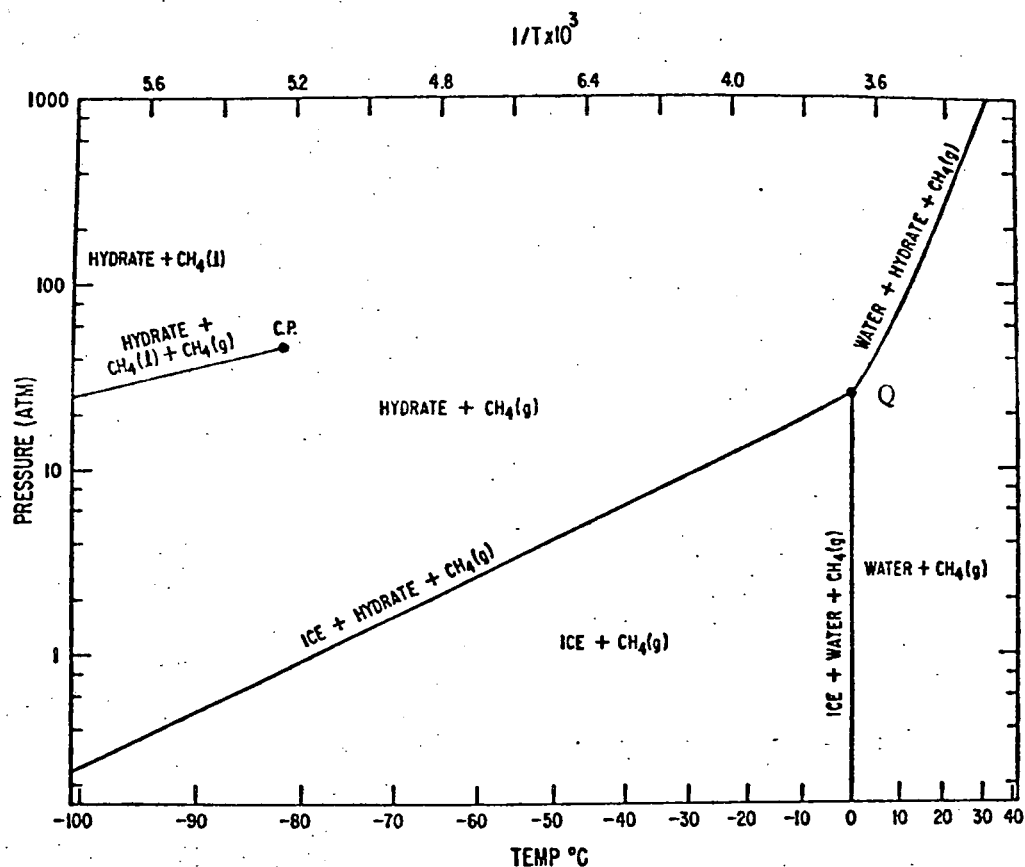
The 12 Å hydrate lattice. The pentagonal dodecahedron (upper center) is formed by 20 water molecules. The tetrakaidecahedra formed by 24 water molecules have 2 opposite hexagonal faces and 12 pentagonal faces.

Fig. 3.17

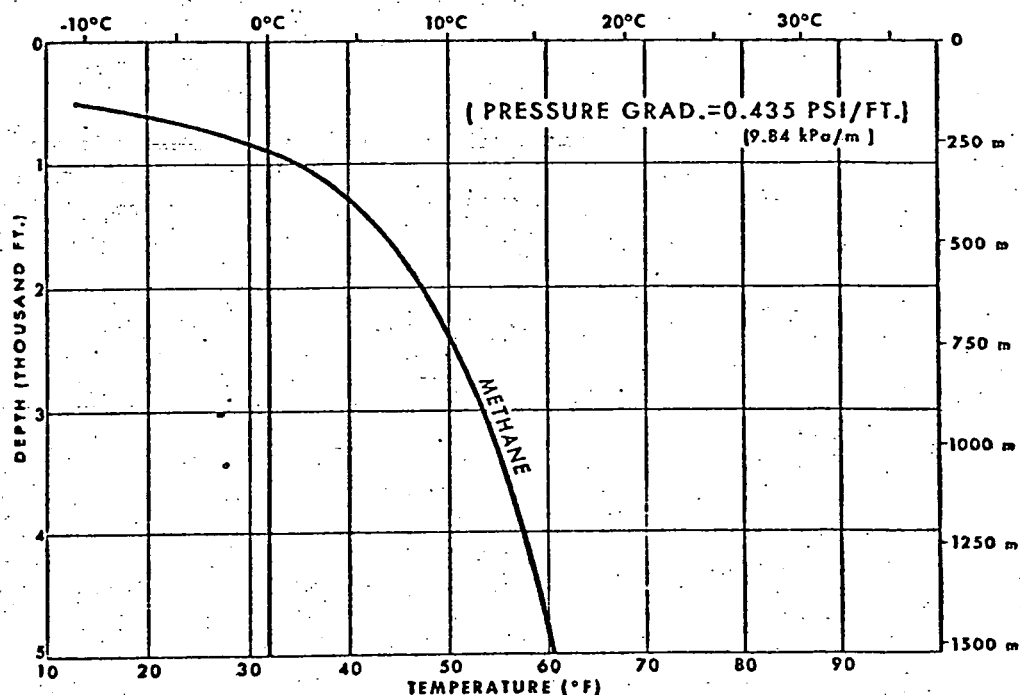
Cage structures formed by water molecules (from Miller, 1974).

A clathrate is a chemical compound having a crystal lattice which contains cages which are large enough to hold molecules of another compound. Such cages occur in several organic compounds, but the principal naturally occurring inorganic clathrates have cages made of either silica or water molecules, the silica giving rise to the class of minerals called zeolites and the water giving rise to gas hydrates. Miller (1974) credits Davy (1811) with the discovery of gas hydrates. Davy noted that the cooling of an aqueous solution of chlorine produced crystals of an ice-like substance which had an approximately constant composition. It has since been shown by van der Waals and Platteeuw (1959), Jeffrey and McMullan (1967) and Davidson (1973) that gas hydrates are non-stoichiometric compounds with cage structures of the type shown in fig. 3.17. The cages are formed of water molecules in a much more open structure than the normal crystalline form. The open structure is stable only when the cage is at least partly filled with gas molecules and when the ambient temperature and pressure satisfy the relationships shown in fig. 3.18 (using methane hydrate as an example).

Outside the laboratory, gas hydrates were first observed not in the natural environment, but in high pressure gas pipelines (Hammerschmidt, 1934; Deaton and Frost, 1946) where they would precipitate and cause



The phase diagram of methane hydrate. The line hydrate + CH₄(l) + CH₄(g) ends very close to the critical point of methane (191°K). Q is the quadruple point (ice + hydrate + water + CH₄(g)).



Depth-temperature curve for methane hydrate.

Fig. 3.18

Upper: Phase diagramme for methane hydrate (from Miller, 1974).
Lower: Depth-temperature stability curve for methane hydrate from Bily and Dick, 1974).

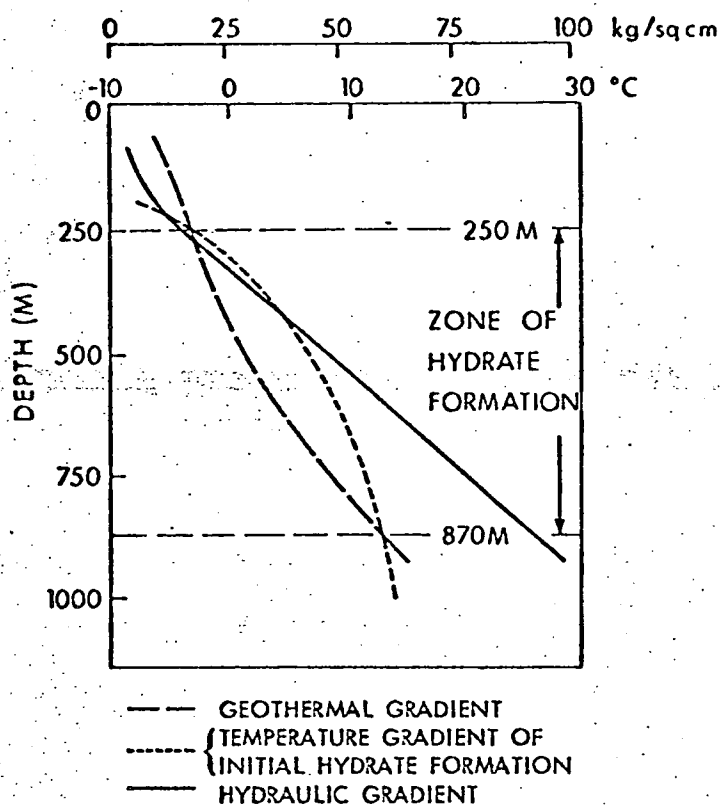
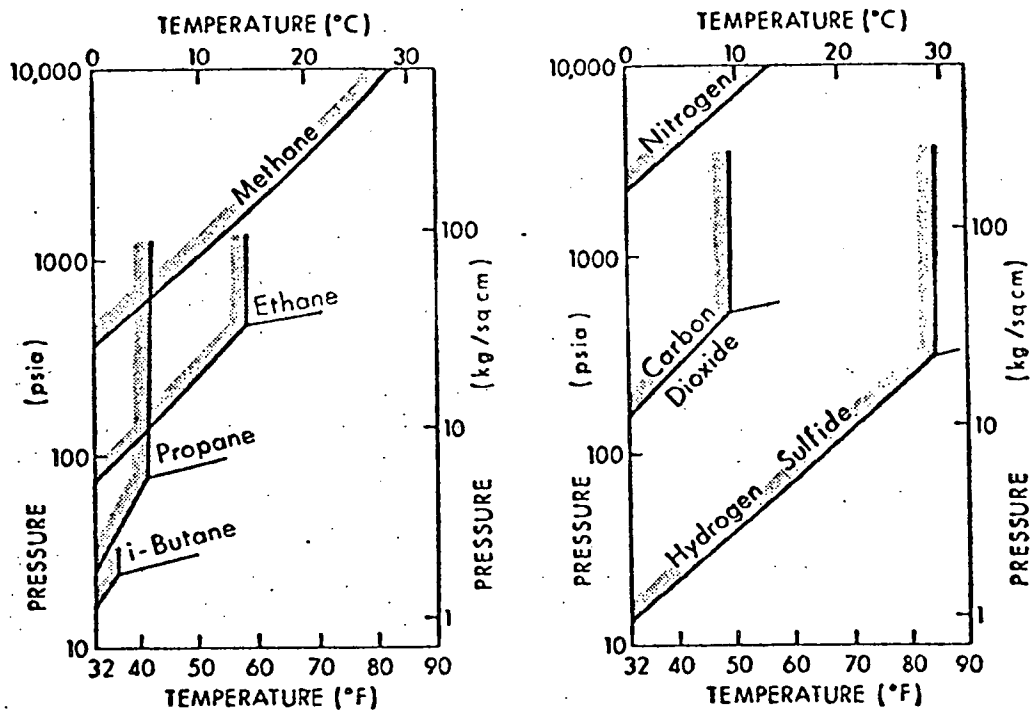
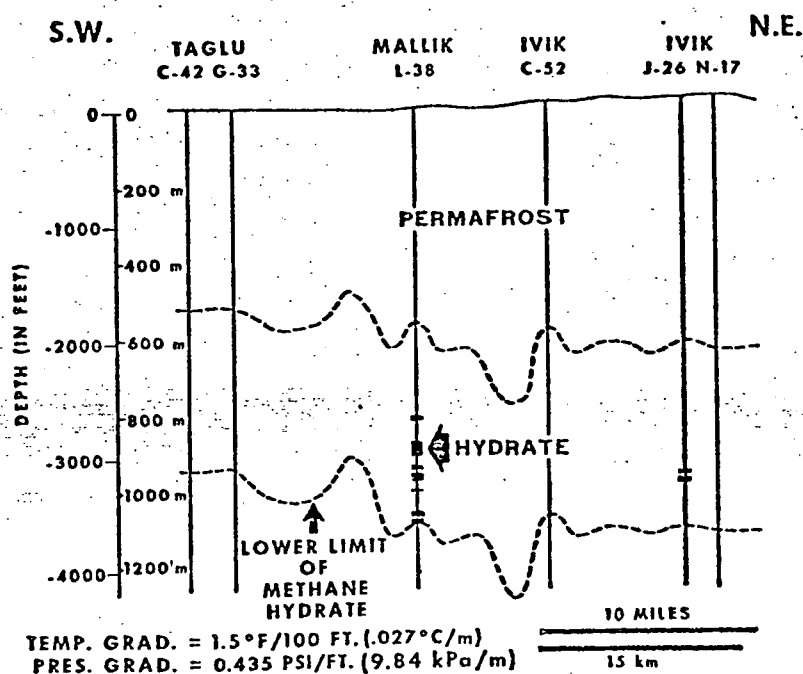
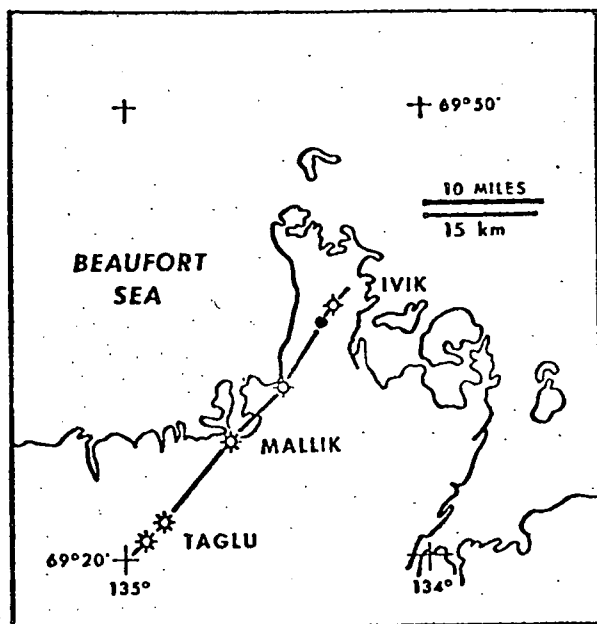


Fig. 3.19

Upper: Phase diagrams for hydrates of saturated paraffins (left) and inorganic gases (right), from Hitchon (1974). Lower: Zone of hydrate formation as predicted by Makogon et al. (1971), for Siberian basins.

blockages, a problem which was solved by the simple expedient of drying the gas before transmission through the pipe. The first sampling of a naturally occurring gas hydrate seems to have been done by Gow et al. (1968) who noted that bubbles of air included in cores taken at Byrd Station ($80^{\circ}01'S$, $119^{\circ}31'W$) on the Antarctic ice cap were not observed at depths greater than 1200 metres below the surface of the ice. Miller (1969) explained this observation in terms of the formation of air hydrates at the temperature ($-28^{\circ}C$) and pressure (70 atmospheres) corresponding to that depth.

Hitchon (1974) has investigated the variation of temperature and pressure with depth in the sedimentary basins of Alberta and the Yukon Territory, and, like Makogon et al. (1971), who have made similar investigations about basins of the U.S.S.R., he concludes that the likely zone of hydrate formation should be between 250 and 870 metres subsurface. In doing this these authors have considered the stability characteristics of different gas hydrates (see fig. 3.19). Hydrates of the paraffins methane, ethane, propane and iso-butane and the inorganic gases hydrogen sulphide and carbon dioxide may occur in this zone. Bily and Dick (1974) devised a pressurised core barrel and sampled gas hydrates from wells in the Mallik area of the McKenzie River delta. Their results (fig. 3.20) are in broad



Mallik - Ivik hydrate zones.

Fig. 3.20

Broad agreement between the actual occurrence of hydrate in wells in the Mallik region of the MacKenzie Delta and the zone of hydrate stability predicted by Bily and Dick (1974) on the basis of assumed gradients of 0.027°C/m. and 9.84 kPa/m.

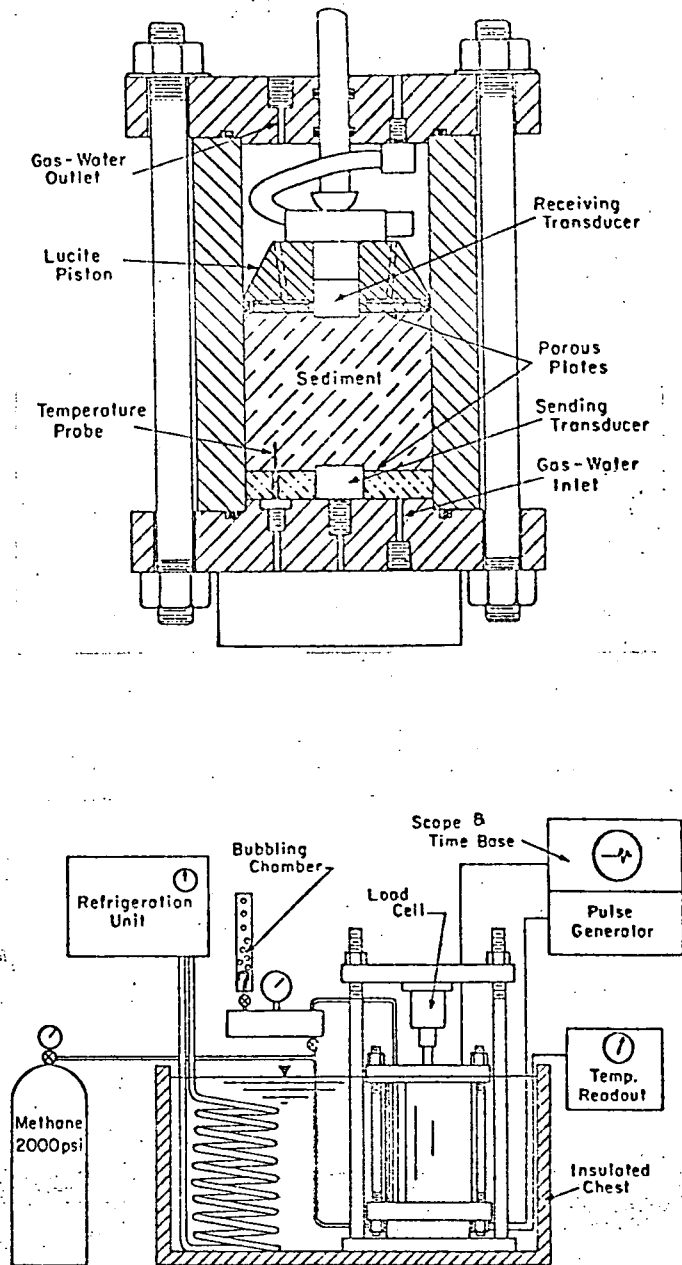


Fig. 3.21

The laboratory apparatus of Stoll et al (1971) for examining the acoustic effects of hydrate deposition.

agreement with the predictions of Hitchon (1974) and Makogon et al. (1971) (fig. 3.19).

Laboratory investigations of the physical properties of sediments containing gas hydrates have not been carried out extensively. Stoll et al. (1971) devised a pressure cell in which they could induce the deposition of gas hydrates in sediment samples by controlling the temperature and pressure while diffusing methane into the sample through porous stones of the type normally used to permit drainage in an oedometer. This apparatus (fig. 3.21) resembles that of McCann (D.M., 1966) or Kepkay and Cooke (1978) (fig. 3.10) in having an ultrasonic transducer and receiver designed to transmit compressional waves axially through the sample. The requirement of gas perfusion did not allow Stoll et al. to adopt McCann's method of using water layers between transducer and sediment sample to obtain good coupling, - but they have retained a sample shape which has the diameter of about the same size as the thickness, thus reducing the risk of having acoustic paths through the container walls such as could lead to arrivals that might be confused with direct transmissions through the sample, an error that Kepkay and Cooke may have made in their measurements of velocities depressed by gasification (section 3.6). Stoll et al. found velocities up to 2.69 km per sec in samples where the pressure was 1200 psi and the temperature 2.0°C. In a

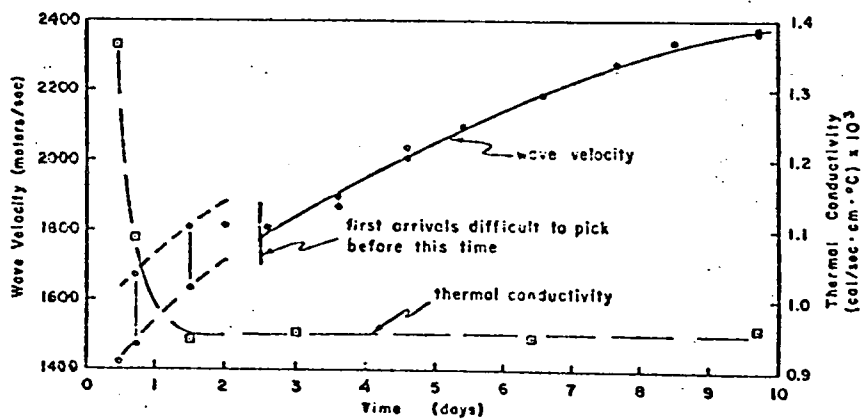


Fig. 3.22

The results of observations by Stoll and Bryan (1979) on the variation of thermal conductivity and compressional wave velocity in a sediment sample with the passage of time after the onset of conditions suitable for the formation of methane hydrate.

later study in which the facility was added to the test cell of measuring the thermal conductivity of the sample, Stoll and Bryan (1979) found that the thermal conductivity of a sediment decreased when hydrate forms in the pore spaces. This is the opposite of what is found when ice is formed there (chapter 4). Stoll and Bryan also investigated changes in acoustic velocity and thermal conductivity which occur with the passage of time after the perfusion with gas of a sediment sample (fig. 3.22). The variation with time of the parameters shown in that figure indicates that some stability has been reached in the value for thermal conductivity, but that the acoustic velocity might, with longer time reach higher values. Stoll (1974b) notes that in experiments of this type, there is a possibility of free gas remaining in the sample even after long periods and that this would have the effect of lowering the value of the acoustic velocity below what might be observed if all such free gas were removed. Consequently, we must not assume that the experimental results obtained for a sediment artificially bonded with gas hydrate would necessarily represent what would be found in a field deposit, especially if one considers the possibility of the formation of a natural gas hydrate from its constitutive elements without the need for the earlier existence of a gaseous phase: higher velocities than have been measured in the laboratory might well occur.

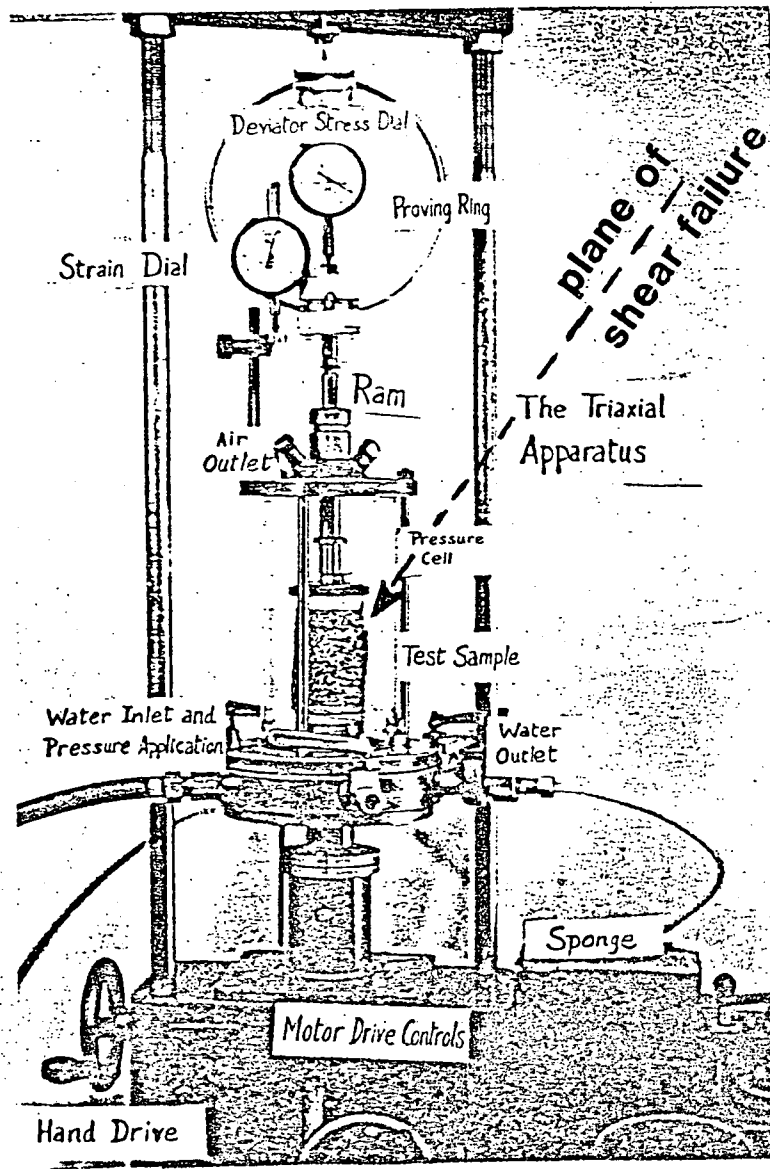


Fig. 3.23

The failure along a shearing plane of a sand sample subjected to a triaxial test (McKay, 1967).

3.8 ENGINEERING IMPLICATIONS OF GAS IN SEDIMENTS.

Monroe (1969) noted that although analysis and experiments had shown that wet clean sand does not flow in response to stress, but rather fails along shearing planes (see fig. 3.23), there could nevertheless be found examples of flow patterns which were preserved in sandstones and showed that clean sands could move as a fluid to form roll-like and dyke-like structures. As further evidence that fluidisation is necessary for the formation of these structure, Monroe cites the transport of fragile fossils in the sand mass without breakage of their shells. He notes that sand structures indicative of fluidisation occur in areas of petroleum accumulation, of recent marine or lacustrine organic sediment deposition, or in sandstone strata within coal measures. These environments have the common factor of being suitable environments for the production of gas. Monroe suggests that gas will reduce inter-grain contact within these sands to the extent that fluidisation can readily occur. In finer grained materials especially in areas of rapid sedimentation such as deltas, instability may arise in underconsolidated sediments merely by a build up of pore pressure. Coleman and Garrison (1977) have identified sedimentation rates of up to 0.5m per year in delta-front water depths of 30m. at the mouth of the Mississippi. Whelan et al. (1977) examined methane concentrations and

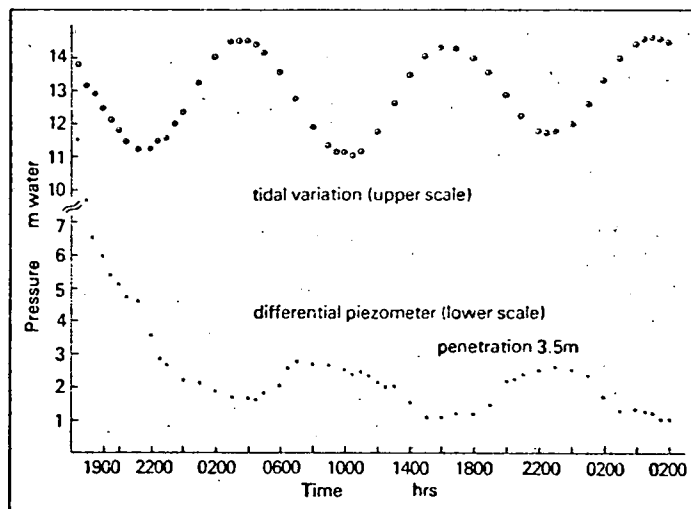


Fig. 3.24

Differential piezometer observations by Sills (1980) compared with tide height in Holyhead Harbour, North Wales.

sediment shear strengths in these sediments. They found that the lowest rate of change of shear strength (as determined by miniature vane tests on recovered samples) coincided with the highest concentrations of methane (as found by purging the samples with helium). The failure of the foundations of several jack-up drilling platforms in near-shore areas of the Gulf of Mexico was noted by Bea (1971), and Garrison and Bea (1977) stressed the importance of sediment gasification for bottom stability in the siting of such platforms.

Direct observations of the effect of gas on pore water pressure have been made in Holyhead harbour by Sills (1980). She found (fig. 3.24) that in the presence of gas bubbles, the seabed pore pressure would have a lower amplitude than the tidal variation, resulting at the seabed in a local hydrostatic gradient which at high tide is inverted in the sense that water would tend to flow from the sea into the sediment. At the Holyhead site, the relaxation time of this gradient appears to have been comparable with the period of the semi-diurnal tide, leading to an inverse phase relationship between tide height and pore pressure. This contrasts strongly with a gas free sediment, where similar investigations of Sills (by means of a differential piezometer) show that pore pressure follows tide height as would be expected. The importance of this observation is that pore water

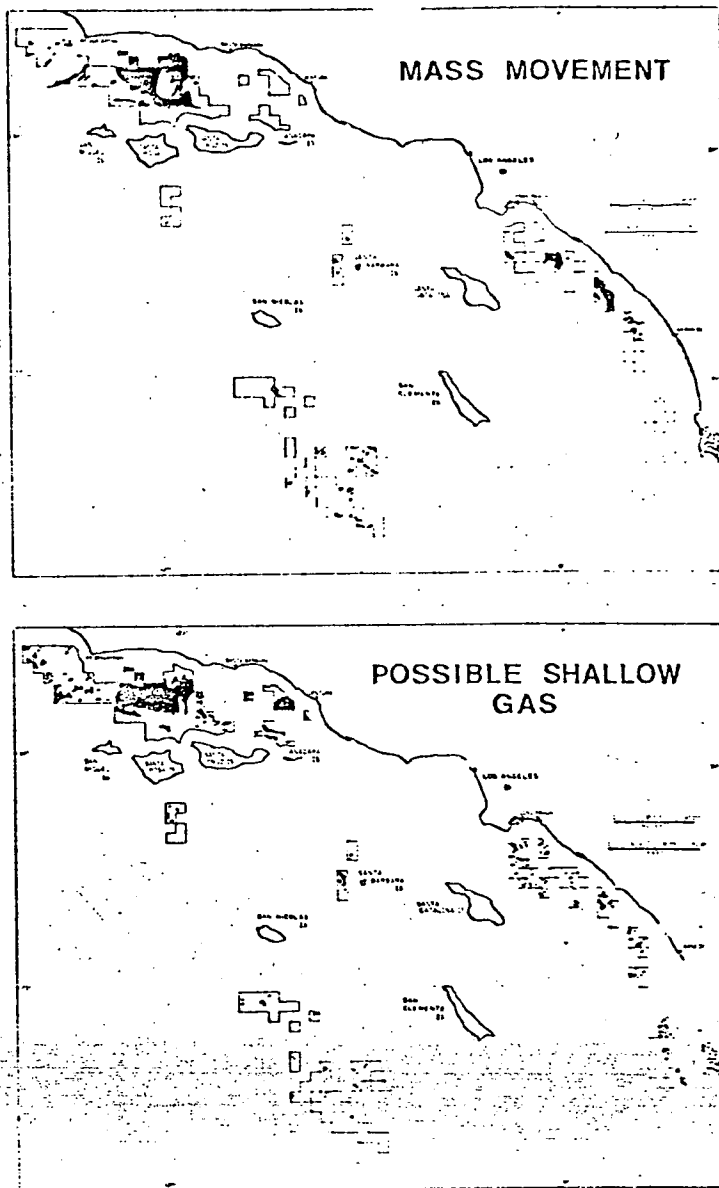


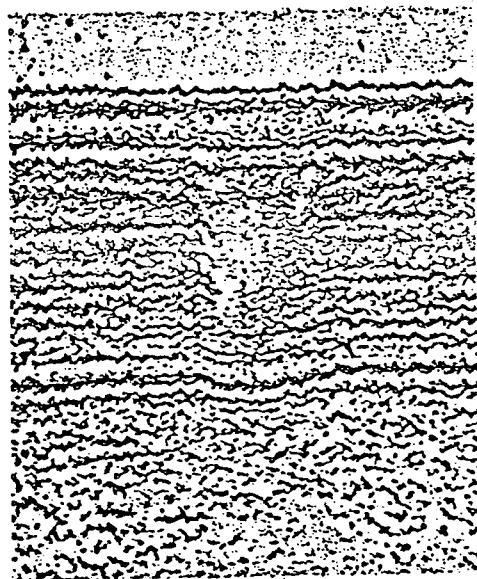
Fig. 3.25

Ploessel et al. (1979) found the above correlation between slumping and gasification off the California coast between Santa Barbara and Los Angeles.

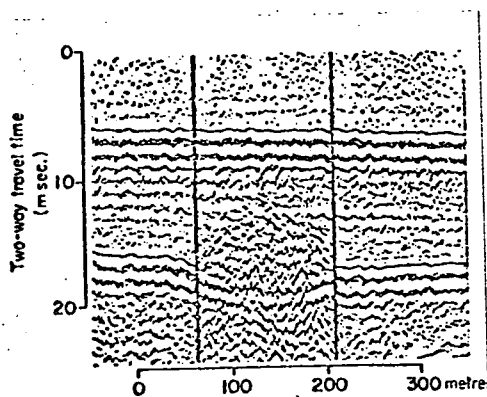
pressures in a gasified sediment do not respond quickly to an imposed load. Consequently any sudden increase in a load must be borne entirely by the framework of solid particles in the sediment. Probably the most significant sudden change in load which might occur on such a sediment would be brought about by the passage of a storm wave. In the event that the sediment fabric could not bear the additional pressure of several atmospheres which could be imposed in this way, failure of structural foundations could occur.

In assessing the potential engineering constraints on the development of proposed continental shelf leases of the U.S. Federal Government off southern California, Ploessel et al. (1979) were able to correlate areas of mass movements of sediment with areas of gasification (see fig. 3.25), both phenomena being identified on the basis of acoustic profiling.

Somewhat interesting, if circumstantial evidence of the effect of gas charged sediments on foundations is given by Stefanon (1980) who has detected on vertical incidence profiles evidence of widespread gasification in the neighbourhood of Venice (see fig. 3.26). McKay (1980b) investigated an area of sediment in Chedabucto Bay (fig. 3.26). Piston cores taken in the area expanded on retrieval and **probably** contained gas. The similarity of



a Northern Adriatic



b Chedabucto Bay (Nova Scotia)

Fig. 3.26

a. Stefanon (1980) describes this feature on a vertical incidence seismic profile made near Venice as a gas pocket.

b. McKay (1980 b) described an almost identical feature from Chedabucto Bay, Nova Scotia.

Stephanon does not give scales on his figure, but his text makes it clear that both horizontal and vertical scales are similar to those given in the lower figure.

Both sections show a similar "pull-down" effect on the deeper reflector.

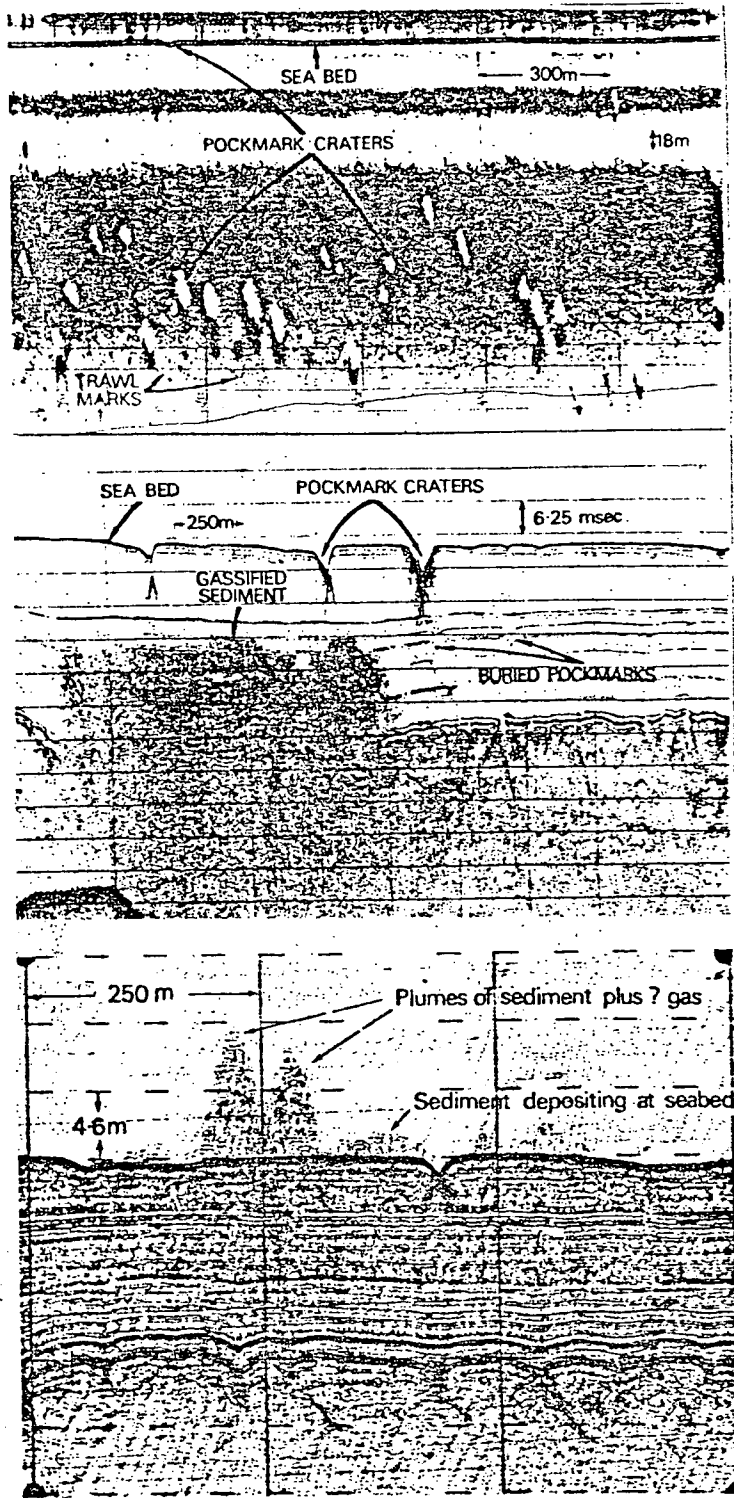


Fig. 3.27

North Sea pockmarks as shown by sidescan (top) and profiler (centre). The bottom frame shows possible sediment plumes rising into the water column. (McQuillan and Fannin, 1979).

the figures a and b of 3.26 lends weight to Stefanon's profiling evidence of gasification in the Northern Adriatic.

A sea bedform normally associated with gas is the "pockmark". King (1970) identified these crater-like depressions on the Scotian Shelf on the basis of echo sounding records and they have been reported in the North Sea by McQuillan and Fannin (1979), and mapped there in detail by Judd (1983). The mapping of pockmark fields in the North Sea is considered an important pre-requisite to the planning of pipe-line routes so that un-supported pipe spans can be avoided. The appearance of pockmarks on profiling and sidescan sonar equipment can be seen in figs. 3.27 and 3.28 . They vary in size, the largest being about 50 to 100 metres wide and 10 metres deep. Their exact mode of formation remains unknown, but they are generally believed to arise from the displacement of sediment by gas bubbling out from deeper levels. McQuillan and Fannin consider that the pattern of pockmarks they have observed in the Witch Ground region (150m depth of water) is inconsistent with a leakage of gas along fissures from underlying hydrocarbon source rocks, and prefer to associate them with gas derived from organic decay in more recent surficial sediments. Although they have been less intensively studied, the pockmarks of the Scotian Shelf, which occur in water

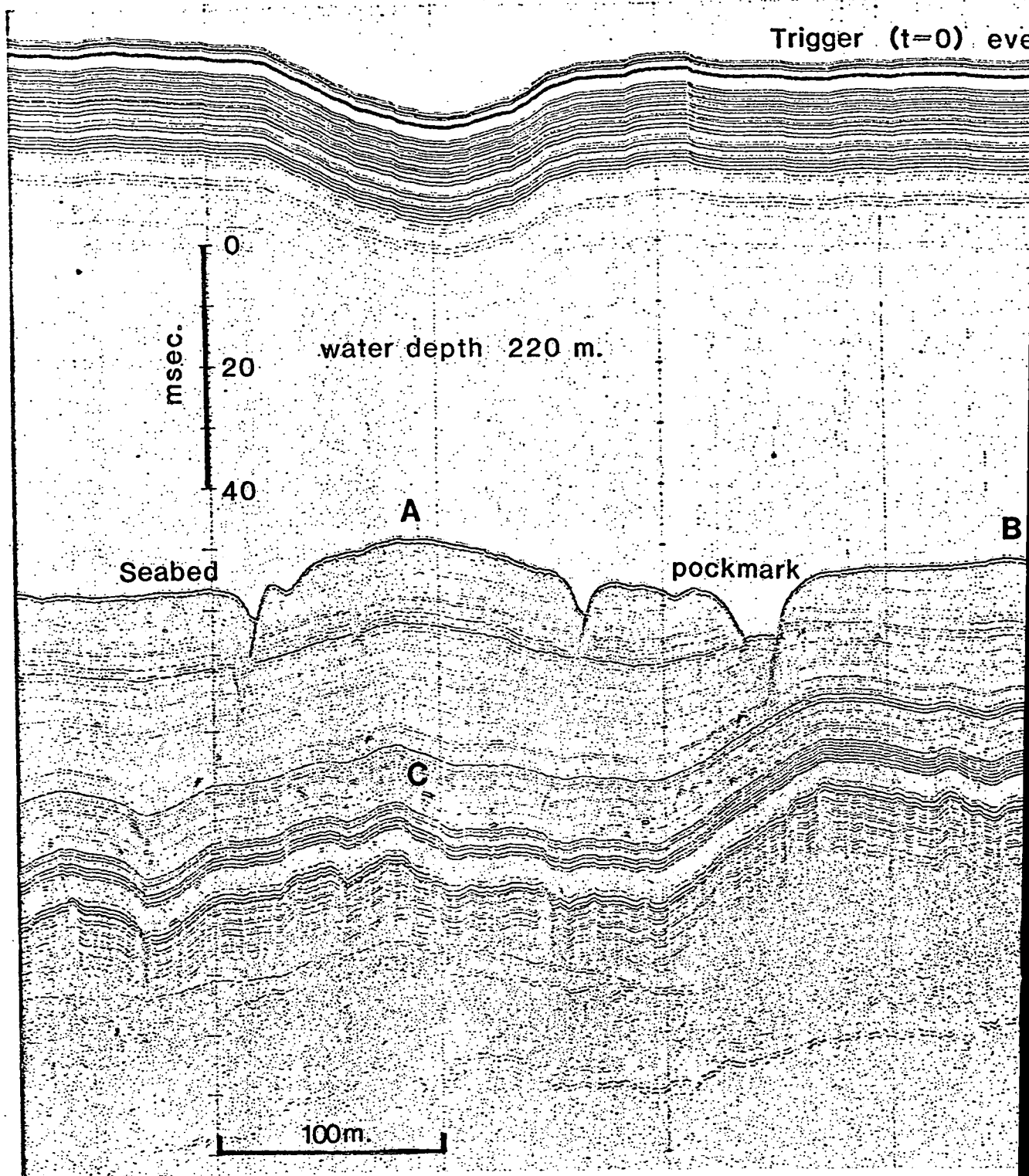


Fig. 3.28

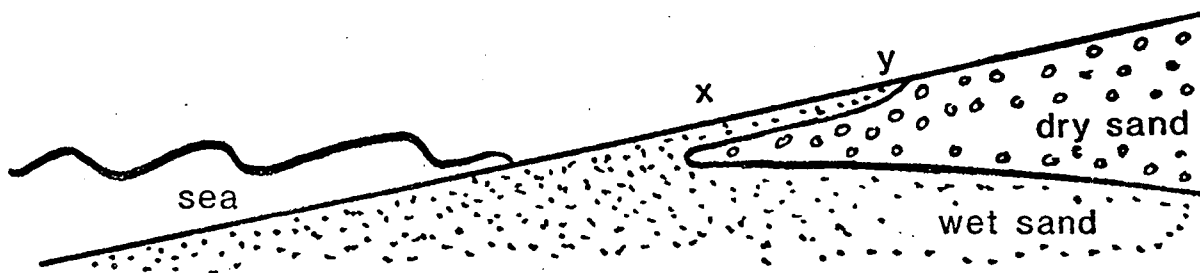
Emerald Basin pockmarks (Scotian Shelf) as observed with a Nova Scotia Research Foundation profiler from M.V. Brandall in February, 1973. The features at A, B, C and D are referred to in the text.

depths mostly over 200 metres in the intra-shelf basins such as the Emerald Basin (fig. 3.28) are generally considered to be scattered in such a way that they would not be associated with the leaking of deeper gas. (The currently known hydrocarbon deposits of the Scotian Shelf occur farther seawards than the pockmarks of the Emerald Basin.) McQuillan and Fannin observed marks on a profiling record (fig. 3.27) which they identify as plumes of sediment with gas rising from the seabed close to an existing pockmark. As there appears to be no disturbance of the seabed immediately under the plumes, nor any evidence of a depressed acoustic velocity in the water column, which should show up as a downwarp in all reflectors beneath the plume if it does contain gas, the claim that fig. 3.27 may show a pockmark in the process of forming has to be regarded as unlikely to be valid.

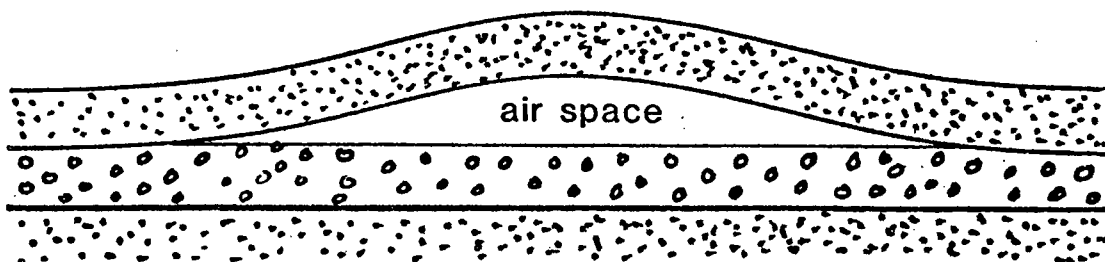
The Emerald Basin example (fig. 3.28) shows "bright spots" in the neighbourhood of "C". These are very localised zones of high acoustic impedance contrast and it seems probable that they are either gas or just possibly gas hydrate concentrations (note water depth of fig. 3.28 and compare with fig 3.18 - the water bottom temperature is ca. 1°C). One might be tempted to try to distinguish between the appearance of small zones of gas and of gas hydrate on the basis of the phase of the reflected signal but Mayer (1979) has pointed out that in regions of finely

stratified sediments, where density or acoustic velocity vary over distances which are small compared to the acoustic wavelength used, clustered reflectors (such as at "D" in fig. 3.28) are unlikely to be identifiable with specific geological interfaces. Thus the nature of the "bright spots" must remain in doubt, but an association with gas in some form seems probable. Whether a pockmark forms catastrophically, or whether it develops over a longer period if gas from organic decay is identified as the erosional agent, then it is necessary to suppose some type of storage mechanism so that sufficient quantities of gas can build up slowly and be released more suddenly, with sufficient vigour to move the seabed sediments. Gas hydrate is a possible storage medium, but only for deeper waters - it is not a possibility in the shallower waters of the North Sea. If gas must be stored otherwise, then it is necessary to suppose that the gas must form in a more permeable region, but be required to traverse a less permeable medium in order to reach the seabed. It may then build up gradually in the more permeable region and, from time to time rupture the less permeable and come out suddenly. A convenient model of such a process may be observed directly on some beaches.

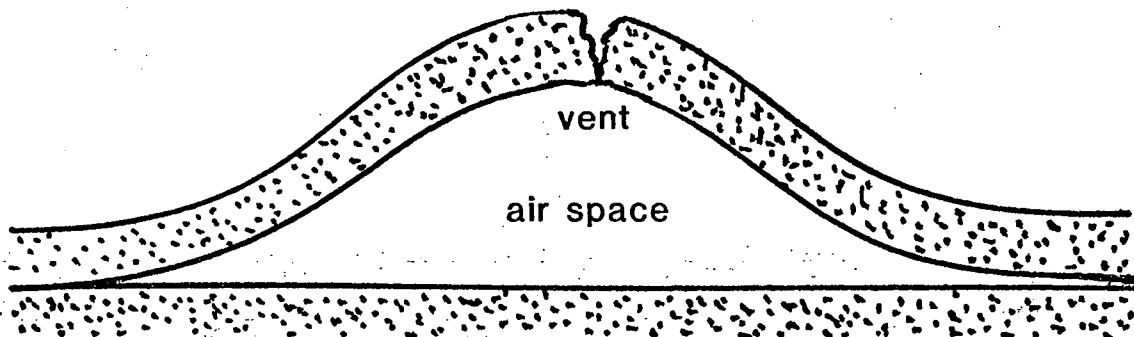
The advancing tide on a sandy beach displaces air in the dry beach sand. In the case of a perfectly calm sea, the water table will simply rise slowly, pushing the air



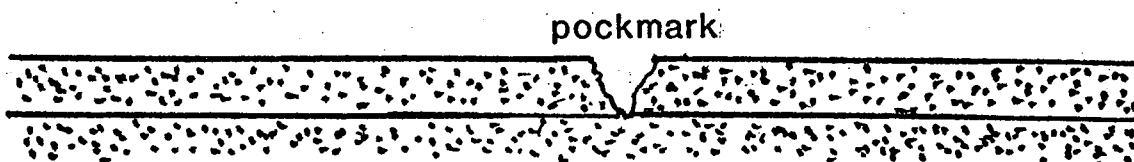
a rising tide waves wash up on beach



b air in sand is displaced upwards



c mature 'beach pingo' with vent forming



d collapsed 'pingo' which forms a pockmark

Fig. 3.29

The evolution of beach pingos and pockmarks

upwards out of the dry sand. Where, as is more usual, there are waves, the upper surface of the sand will be wet as well, for some distance above the current water-level. This thin layer of wet sand can act as a barrier to the venting of air, and gives rise to the following phenomenon. Between the points "x" and "y" on fig. 3.29a, air displaced upwards is compressed until the upper layer of wet sand begins to distort under the stress (fig. 3.29b). This process continues until the rising water table meets the upper layer of wet sand. The small dome (fig. 3.29c) which has formed usually ruptures and splutters out air and wet sand, eventually collapsing to leave a pockmark (fig. 3.29d). The grainsize distribution of the sand in which these features were observed has not been analysed, but it was reminiscent qualitatively of sand from Porth Tre Castell, (McKay, 1967) and, by way of contrast, was quite unlike sand of Porth Oer in its moulding characteristics when damp. These two reference sands are illustrated in fig. 3.30. The pockmark observations were made at Bamburgh Beach (Northumberland) and photographs of the features are shown in figs. 3.31 and 3.32.

In the light of these beach observations, it is interesting to look again at fig 3.28. Two domed features appear on the profile at the seabed - one at "A" and one at "B". The large hump at "A" corresponds exactly with a

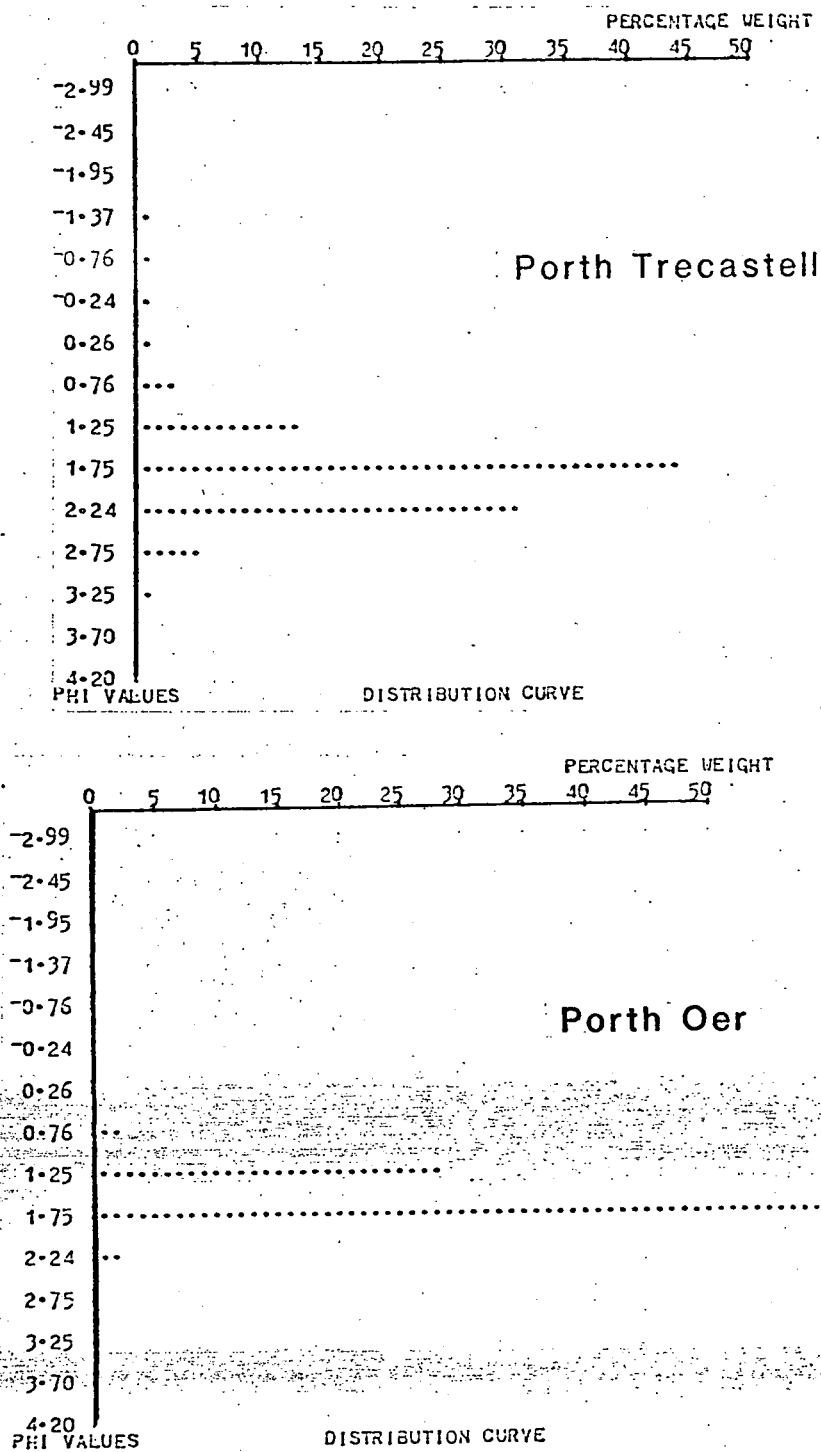
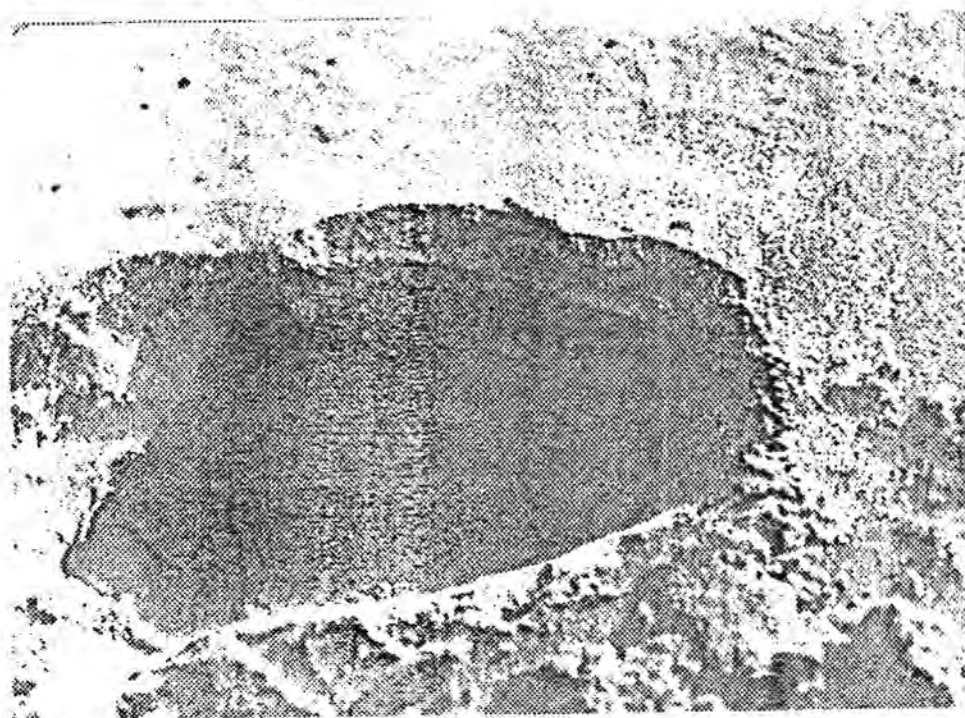


Fig. 3.30

Grain-size distribution curve for two sand types examined by McKay (1967) in North Wales. The Bamburgh beach sand (not analysed quantitatively) resembles that of Porth Treceastell more closely than that of Porth Oer.



a mature pingo



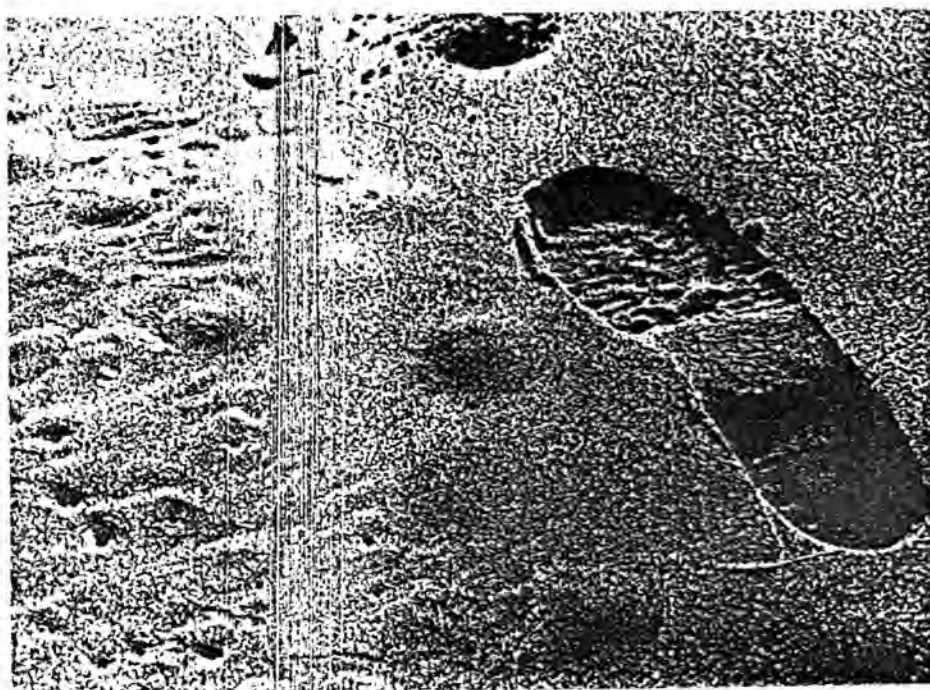
b excavated section of mature pingo

Fig. 3.31

One of the features observed on Bamburgh beach.



a pockmarks



b immature pingos and pockmarks

Fig. 3.32

Pockmarks evolving on Bamburgh beach.

dip in the trigger event, and it must be suspected that the depth compensation mechanism of the deep tow system (see section 2.1) was not operating as it should. The small hump at "B", however, occurs at the seabed when the towfish was in level flight as indicated by the trigger event, so that it is a real hump, and may be an indication of distortion due to gas pressure build-up. The sediments of the Emerald basin have been mapped by King (1970) and it is expected that clays overlie silts and sands.

It is thus quite possible that gas is evolved in the more permeable deeper layers and is trapped by overlying clays. It should be noted that the zone of "bright spots" around "C" and between "C" and "D" does not extend all the way upwards to the seabed.

As well as giving rise to pockmarks and the associated hazards to structures engineered on seabed foundations, I.K.U. (Norway's Continental Shelf Institute) have pointed out (Anon. 1983a) that the sudden release of gas from seabed sediments might produce water of sufficiently decreased density as to allow floating drill rigs or other vessels to lose buoyancy. The effect of aerated water on buoyancy is well known to white-water canoeists, and although there is no evidence that accidents to large sea-going vessels have occurred in this way, the proposed mechanism is possible, but the probability of the

occurrence of such an event is completely unknown.

3.9 SUMMARY.

The existence of gases in marine and lacustrine sediments is well documented, with examples available from diverse environments, but the overall extent of gasified sediments and the quantities of gas or gas hydrate which occur in them is largely unknown. Different sediment models have been proposed to predict the acoustic effects of gasification, and there is insufficient field information to indicate a preference for any particular model, provided it accounts for the gross effects on velocity and attenuation.

Sediment gasification has been implicated as a major factor in bringing about the failure of several marine structures and is regarded as an important hazard to be assessed by ocean engineers who must use the seabed as a foundation.

4. PERMAFROST



Fig. 4.1 N Hemisphere Permafrost

Permafrost distribution in the northern hemisphere. Compiled by Hitchon (1974) after Baranov (1964), Brown (1967 a, b, 1970), Ferrians (1965), MacKay (1972 a, b) and Tricart (1970).

4.0 INTRODUCTION.

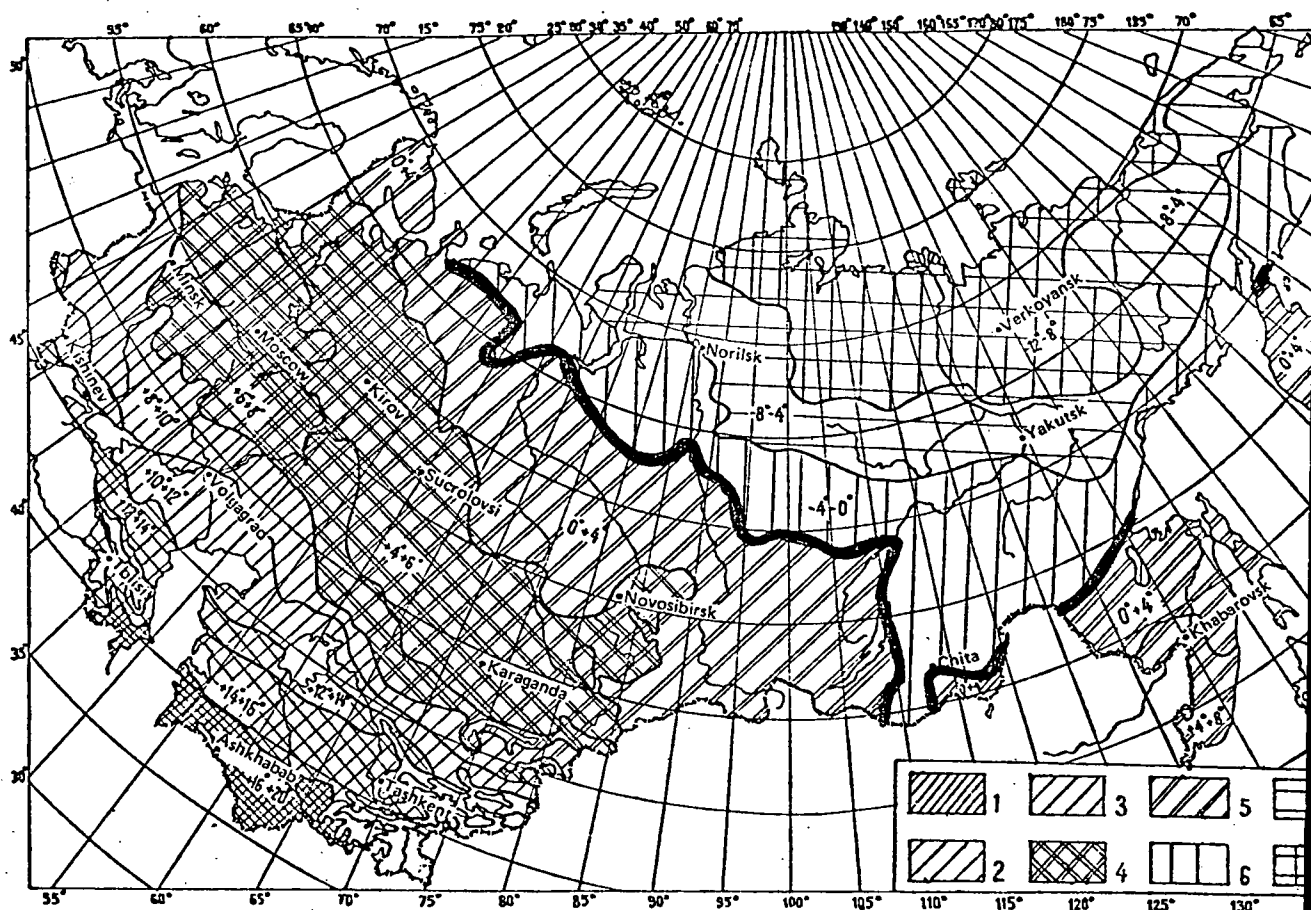
Permafrost, that is, perennially frozen ground, occurs commonly at high latitudes and high altitudes. The distribution of continuous permafrost in the Northern Hemisphere is shown in fig 4.1. Beyond the southern limits shown in this figure, are areas where permafrost occurs intermittently. When areas of continuous and intermittent permafrost are taken together, it is found that the phenomenon affects one half of Canada and the Soviet Union and 85% of Alaska (Ferrians and Hobson (1973)) and is, in consequence of major importance in these lands. Table 4.1 indicates the areas subject to permafrost. Human activity can change the characteristics of soil structures in frozen ground, usually by supplying sources of heat. This causes problems for construction engineering in regions where hydrocarbon or mineral extraction is proposed. According to the Committee on Permafrost of the Polar Research Board of the National Academy of Sciences (U.S.A.), (Sackinger, 1976), "Land permafrost problems are not understood fully, whether from the point of view of basic earth science, engineering geology or environmental protection. Far less is known of offshore permafrost, its distribution, properties and dynamics of formation and destruction." The occurrence of offshore permafrost is indicated in fig. 4.1 and its distribution will be discussed more fully in

Region	Area Occupied (km ² x 10 ⁶)	
Northern Hemisphere:		
U.S.S.R.	11	
Canada	5.7	
Greenland	1.6(?)	
Alaska	1.5	
Mongolian Peoples Republic	0.8	
China (without Tibet)	0.4	
Total Northern Hemisphere		21.0
Southern Hemisphere:		
Antarctica	13.5(?)	13.5
Total for Both Hemispheres		34.5

Table 4.1

Major areas affected by permafrost. (Baranov, 1964).

section 4.4. Its very existence, however, poses an immediate question of definition. Permafrost is commonly agreed to be earth material that is continuously below a temperature of 0°C for two or more years. Normally, on land, this is equivalent to a definition that ice crystals be found in the earth material. However, in conditions of saline porewater, it is quite possible to have temperatures below 0°C without the formation of ice, so that on the temperature definition most of the surface waters of the Arctic and Antarctic oceans might be classed as permafrost, or, if not all such waters (on the argument that 'earth' material is lacking), then at least those with large loads of suspended sediment such as occur in the plumes of the MacKenzie, Lena, Yenisey and Ob. On the other hand, it is conceivable that in some equatorial regions at high altitude where the reduced atmospheric pressure allows freezing of water at a temperature above 0°C , there may be some regions where permanent ice-crystals exist in an environment which does not satisfy the criterion of permanent sub-zero temperature. Usually the factor which is of practical importance is the presence or absence of ice because of its cementing effect, which can completely change the physical properties of a soil, and its potential volume change if artificial sources of heat are applied. As a further complication to the resolution of the definitions of permafrost, Williams (1963) showed that all the water in



Mean annual temperature of soils ($^{\circ}\text{C}$) at a depth of 0.2 m. 1--from +16 to +20, 2--from +12 to +16, 3--from +8 to +12, 4--from +4 to +8, 5--from 0 to +4, 6--from -4 to 0, 7--from -8 to -12, 8--from -12 to -16.

Fig. 4.2

Occurrence of permafrost in the Soviet Union on the basis of soil temperature (Dimo, 1973).

natural soils does not freeze at a single temperature but that the ice content increases as the temperature is lowered. He also found that the amount of unfrozen water depends on whether the soil is freezing or thawing and, if thawing, it also depends on the lowest temperature reached during the freezing.

A general idea of the extent to which conclusions can differ as to whether or not permafrost occurs is given by a comparison of figures 4.2 (mean annual soil temperature) and 4.3 (amounts of visible ice in soils) for the Soviet Union. While there is broad agreement between the two maps, the detailed pattern differs in the vicinity of the 0°C isotherm.

The problem of definition is one which can be of concern when remote geophysical methods (see section 4.3) are used to detect permafrost, since anomalous acoustic velocities, electrical conductivity or other measured physical characteristics associated with ice-bonding will not always correlate with the measured ground temperature.

4.1 PERMAFROST LANDFORMS.

There are a number of geomorphic features which are good indicators of the presence of permafrost. Their

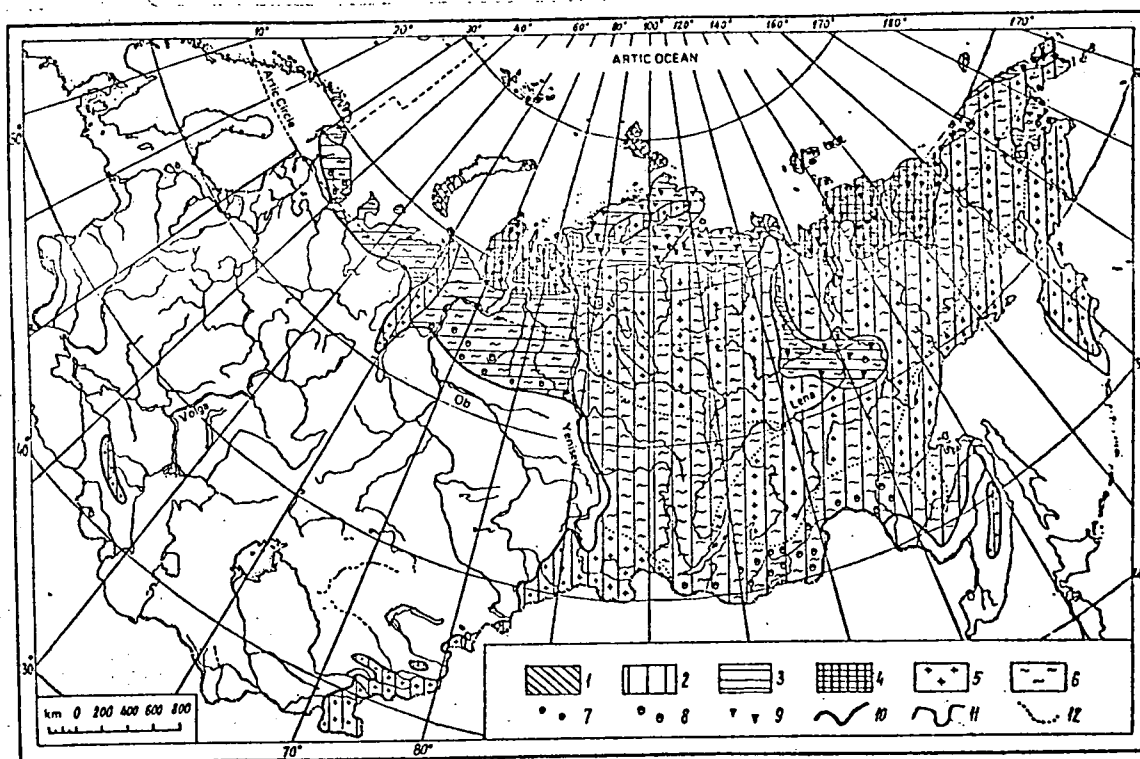
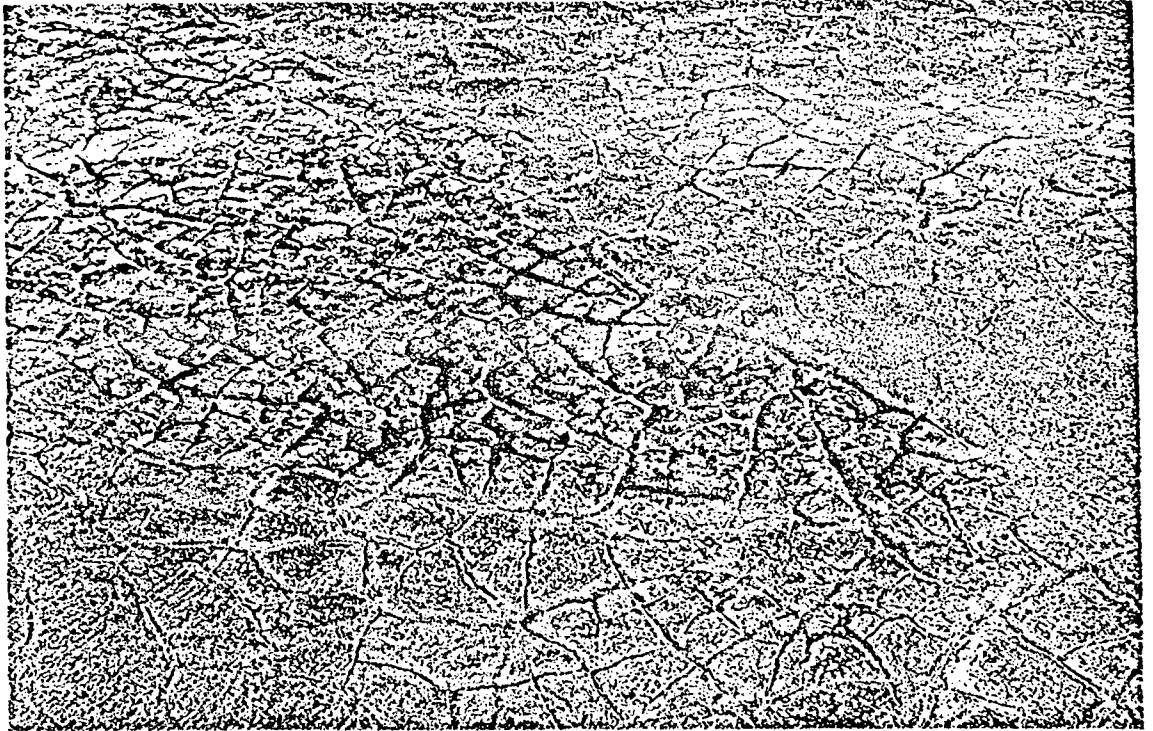


Fig. 4.3

Occurrence of permafrost in the Soviet Union on the basis of ice content (Vtyurin, 1973).



Aerial view of ice-wedge polygons near Tuktoyaktuk, Northwest Territories. The polygons in the center are mainly high centered, those at the bottom and top right are low centered. Each polygon is outlined by a trough beneath which lies an ice wedge. Most of the low centered polygons are primary ones, whereas many of the high centered polygons are secondary, resulting from the subdivision of primary polygons. The polygons range from about 10 to 90 m in diameter.

Fig. 4.4

Ice wedge polygons. (MacKay, 1975).

recognition is an important aid in mapping regions where the cost would be prohibitive of drilling and instrumenting an adequate grid of boreholes to give a detailed delineation of permafrost zones (Harris, 1982). The most commonly identified landforms are ice-wedge polygons, thermokarst, pingos and solifluxion sheets (Ferrians and Hobson, 1973).

Ice-wedge polygons have been described by Leffingwell (1915), Black (1952) in parts of Alaska, Jahn (1972) in the MacKenzie Delta, and MacKay (1975) in the islands off the Northwest Territories. The appearance from the air is shown in fig. 4.4. The sides of the polygons are outlined by troughs, beneath which lie ice wedges. The ice wedges grow as a result of the thermal contraction of hard-frozen ground in the winter and the subsequent infilling of the cracks by snow or melt-water to form ice-veins in the spring (fig.4.5). It is possible for ice-wedges to grow only in permafrost of sufficient horizontal extent to allow the crack-producing stresses to build up, and the presence of polygons can be taken as diagnostic of an area of continuous permafrost (Harris, 1982).

Thaw pits or thermokarst form by the thawing of large underground ice masses which cause severe differential settlement of the ground surface. Thermokarst is more diagnostic of discontinuous permafrost, where frozen and

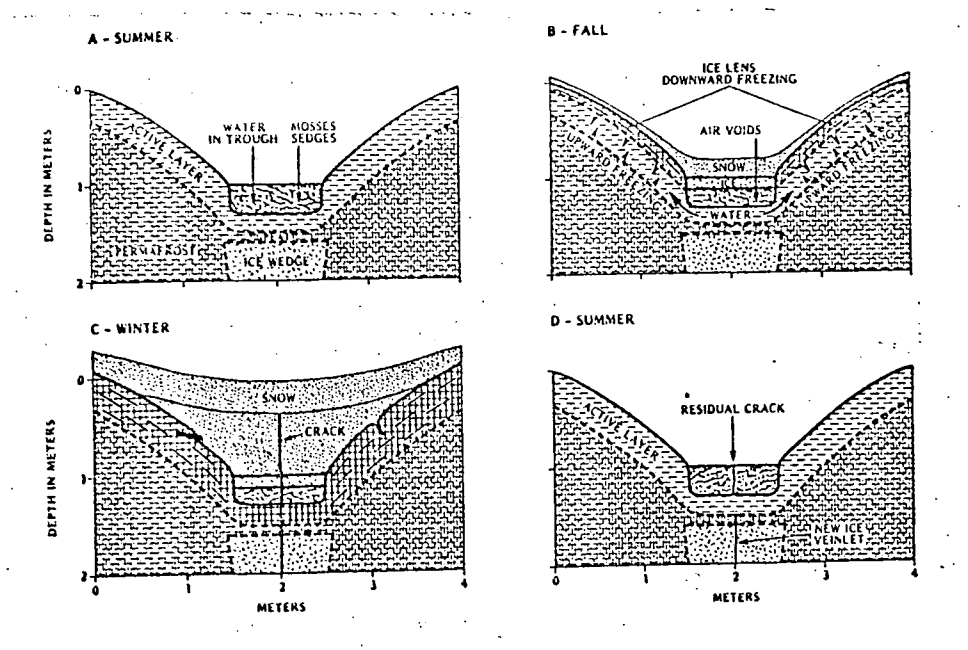


Fig. 4.5

Seasonal variations of freezing and thawing in the surface layer which give rise to the build up of an ice wedge. (MacKay, 1975).



Fig. 4.6

Thermokarst in the MacKenzie Delta.

unfrozen areas are adjacent to one another, whether because of warmer climatic conditions or the proximity to salt water (fig. 4.6).

Pingos, which are small, ice-cored conical hills are common in the zone of continuous permafrost along the coastal plain of Northern Alaska and the Northwest Territories, particularly so in the Tuktoyaktuk peninsula. They vary in size from about 15m. to several km. in diameter and from 3m. to 100m. in height. Often the summit of a pingo is cratered and the base is encircled by a moat. A pingo in the vicinity of Tuktoyaktuk is shown in fig. 4.7. The generally accepted mode of formation of a pingo is shown in fig. 4.8. A lake affords a barrier against the low atmospheric temperatures in winter and allows a large zone beneath its bed to remain unfrozen. The lake may subsequently dry out, either by silting up or, in the case of lakes near the coast, by being breached by the sea during storms. When the water blanket has gone, permafrost will start to invade the unfrozen zone from all sides (fig. 4.8b). Because of expansion on freezing, pressure will build up in the remaining pore water until the ground surface is forced upwards (fig. 4.8c), giving rise to a pingo. The presence of high pressure water under a pingo has been demonstrated by MacKay, (1980) who produced an artesian fountain by drilling into one. The fountain flowed for the summer

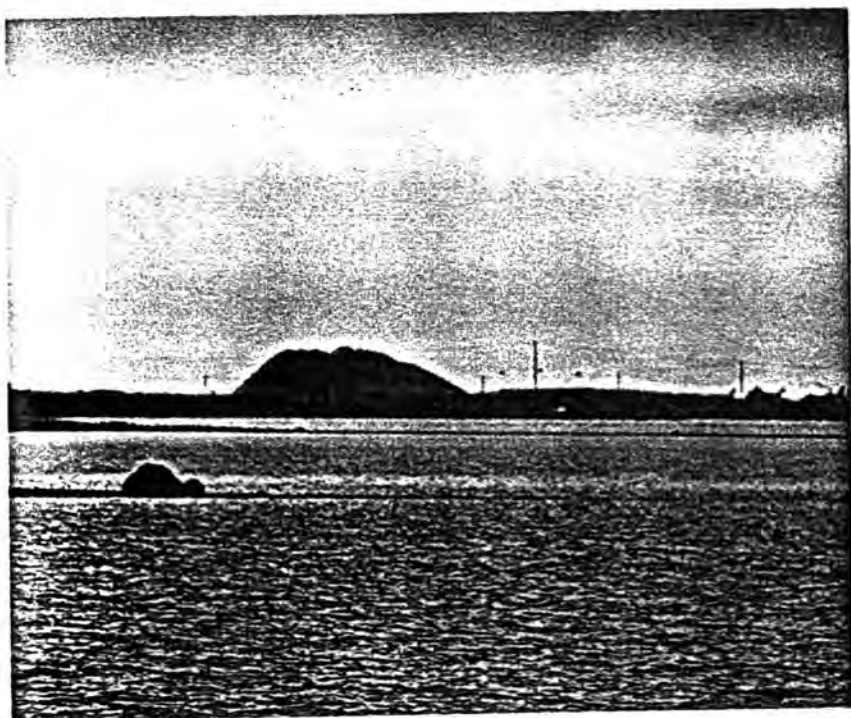


Fig. 4.7

Pingo, Tuktoyaktuk. This pingo is about 80m high and 500m in diameter. The summit crater is clearly visible. The pingo is moated, but the moat is concealed by the low ridge of land in the foreground.

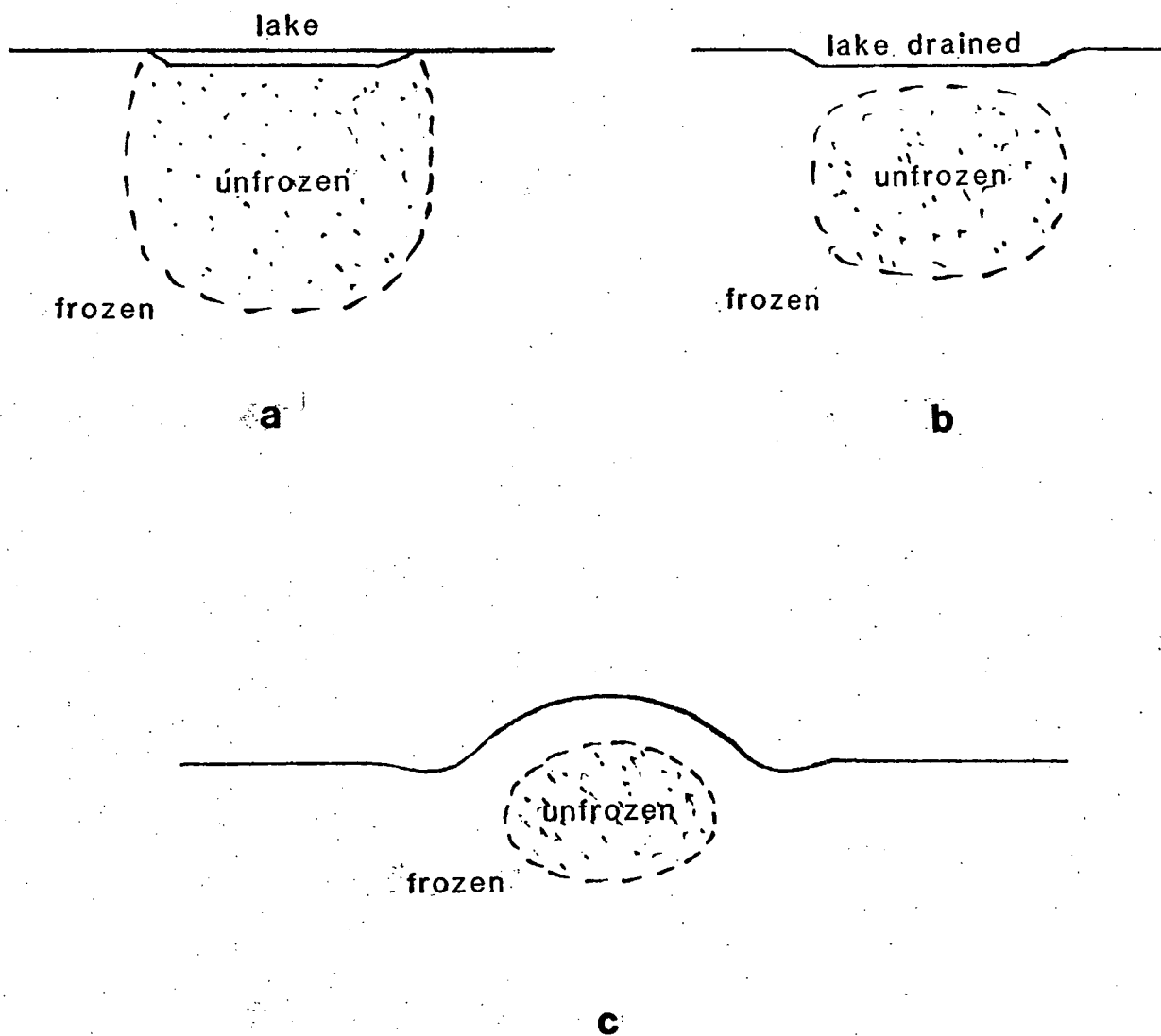


Fig. 4.8 Evolution of a pingo

The drainage of a lake (a) leads to the invasion of permafrost (b) and the build-up of pressure. This gives rise to a small moated hill (c) — a pingo.

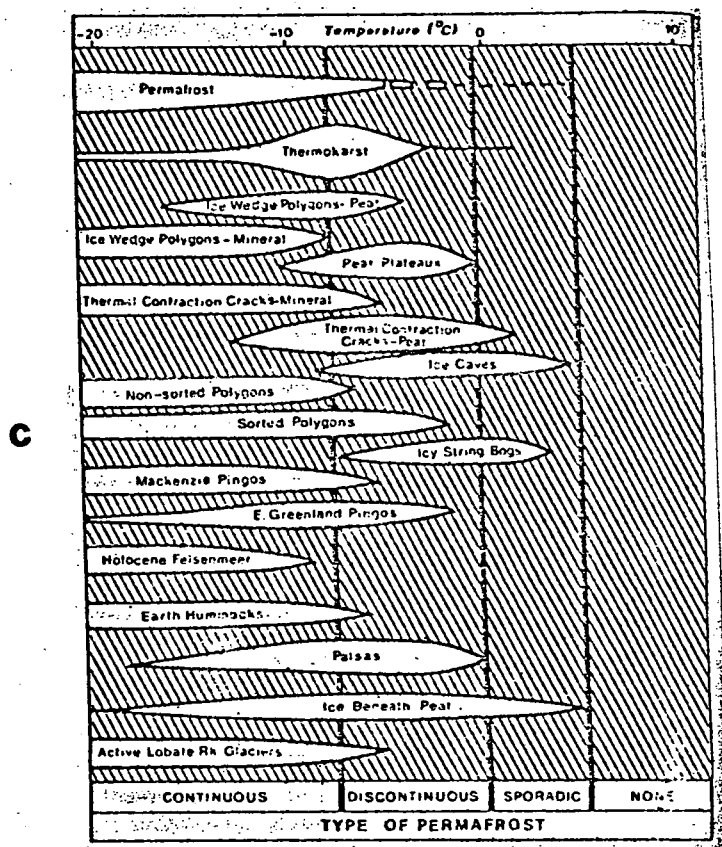
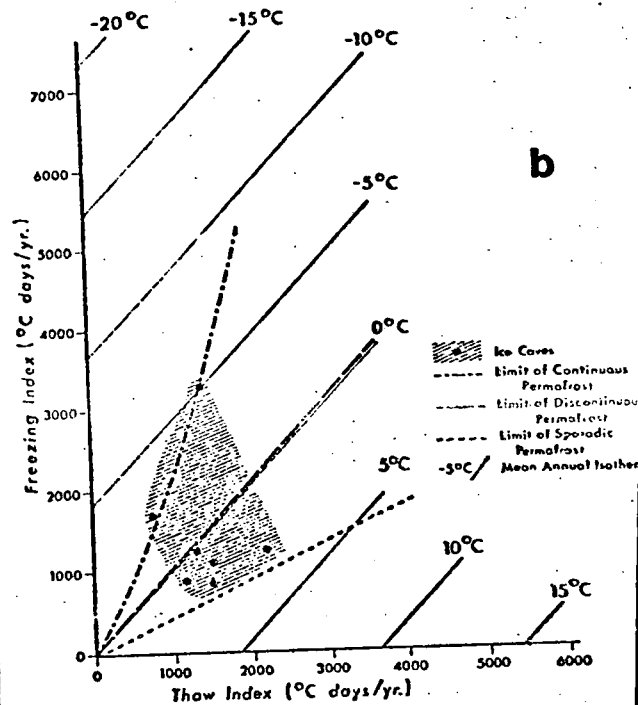
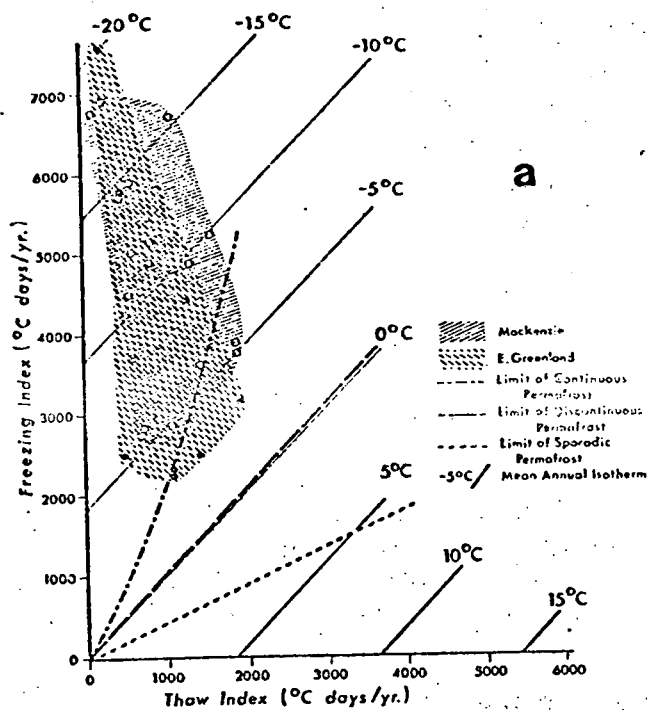


Fig. 4.9

Two landforms, pingos (a) and ice-caves (b) whose occurrence is plotted on the basis of observed temperatures can be correlated with continuous and discontinuous permafrost respectively as shown, along with other landforms, in (c). (Harris, 1982).

season, and the pingo's height was reduced by a few metres. The presence of an ice core above the zone of high pressure water has been observed by Rampton and MacKay (1971) in a horizontal tunnel dug into a pingo by residents of Tuktoyaktuk to serve as a large refrigerated room. Ice cores can also be seen in pingos undergoing erosion at the coast.

Solifluction sheets are masses of unconsolidated sediments formed by the gradual downslope movement of water-saturated sediments. Although not a phenomenon confined to permafrost regions, it is one that is especially abundant there because of the lack of binding vegetation roots and the common occurrence of water-supersaturation in the surficial active layer (Ferrians and Hobson, 1973). This phenomenon can be a good general indicator of possible permafrost.

Harris (1982) has taken large numbers of observations of the above, and other related landforms. He has plotted the occurrences on diagrams of freezing and thawing index. These indices are derived from temperature statistics for the location where each landform was observed. Figures 4.9a and b show Harris's diagrams for two landforms - pingos and ice-caves. It is clear that these two occupy different regions on the plot of thaw index/freezing index. Harris has identified on his diagrams, zones of

continuous, discontinuous and sporadic permafrost on the basis of correlating data from weather stations (maximum and minimum daily screen temperatures) with known permafrost conditions in the locality of the weather station. He derives the fig. 4.9c from his data compilation for landforms described above and for various other features which can occur in frozen ground. In this way he has contributed towards an empirical reconciliation of the potentially conflicting definitions of permafrost discussed in section 4.0.

4.2 PHYSICAL PROPERTIES OF PERMAFROST.

When a soil freezes, changes occur in its physical properties, particularly in electrical and thermal conductivity, density, permeability, and elasticity. The density-effect is the simplest in concept, viz:- the 10% decrease associated with the freezing of water adjusted for porosity and the relative densities of ice and soil particles.

Kersten (1949) measured in both frozen and unfrozen states, the thermal conductivities of samples of soil having various degrees of compaction and water saturation. He found that freezing increased the conductivity of water-saturated samples by a factor between 1.35 and 1.45.

Data for undisturbed permafrost samples

Sample No.	Sp. Gr.	Ice content %	Bulk unit weight (p.c.f.)	Dry unit weight (p.c.f.)	Thermal conductivity (W/m °C)
72-15-1-2(T)	2.48	135.8	78.3	34.7	1.69
72-15-1-2(B)	2.48	135.8	72.2	25.1	1.60
72-15-1-3	2.5	141	76.5	31.7	1.57
72-15-1-4(T)	2.72	95.1	83.8	41.2	1.58
72-15-1-4(B)	2.72	95.1	83.2	40.4	1.71
72-15-2-3	2.75	152	82.6	39.2	1.56
72-15-3-1	2.75	231	70.6	21.4	1.55
72-15-3-5(T)	2.75	112	82.3	38.8	1.63
72-15-3-5(B)	2.75	109	83.0	39.8	1.75
72-15-5-1(T)	2.75	32.9	111.8	73.4	1.49
72-15-5-1(B)	2.75	36.9	111.5	82.2	2.12
72-15-5-2	2.75	38.9	110.0	80.0	2.23
72-15-5-3	2.74	25.4	125.2	93.5	2.16
72-15-6-3(T)	2.75	49.0	100.2	65.4	2.01
72-15-6-3(B)	2.75	49.0	110.6	80.9	2.07
72-15-6-4	2.75	29.6	117.3	90.8	2.19
72-15-6-5	2.75	25.4	120.6	95.4	2.25
72-15-7-1(T)	2.75	47.1	103.0	69.6	2.06
72-15-7-1(B)	2.75	49.8	102.0	68.1	1.92
72-15-7-2(T)	2.75	36.1	95.5	58.4	1.87
72-15-7-2(M)	2.75	36.1	112.0	82.9	2.02
72-15-7-2(B)	2.75	35.9	110.8	81.6	2.24
72-15-8-2(T)	2.75	60.0	97.0	60.6	1.75
72-15-8-2(B)	2.75	80.0	89.7	49.8	1.81

a

WORKING VALUES OF THERMAL PROPERTIES OF SOIL CONSTITUENTS NEAR 0°C

Constituent	Thermal conductivity		Specific heat	
	(W/m °C)	(mcal/cm sec °C)	(kJ/kg °C)	(cal/gm °C)
Air	0.024	0.057	1.0	0.24
Water	0.56	1.34	4.2	1.00
Ice	2.2	5.3	2.1	0.50
Quartz	8.5	20.0	0.8	0.19
Common silicates	2.9	7.0	0.8	0.20
Organic matter	0.3	0.7	1.9	0.45

b

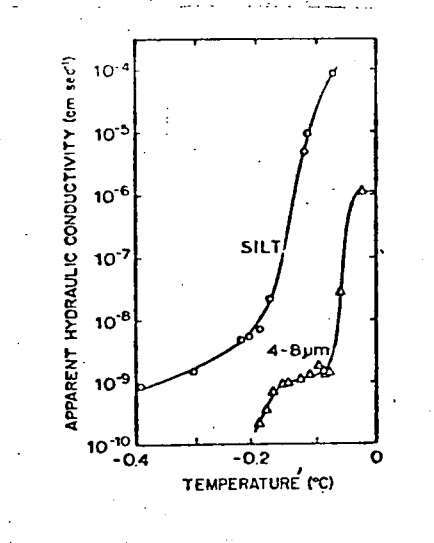
Table 4.2

a. Conductivities of permafrost measured in situ by Slusarchuk and Watson (1975).

b. Conductivities of soil constituents (Hillel, 1980).

When the degree of saturation decreased to about 0.3 or less, there was little change in the conductivity upon freezing. A comparison of thermal conductivities of permafrost and of its constituent materials is given in table 4.2. The permafrost values were measured in situ near Inuvik (Northwest Territories) by Slusarchuk (1975) who used a probe developed by Slusarchuk and Foulger (1973) based on designs of Blackwell (1954) and Jaeger (1956). A comparison of the conductivities given in table 4.2a shows that the conductivities in permafrost are higher the lower the ice content (the samples are from saturated or near-saturated soils). Table 4.2b shows that this might be expected on the basis of the higher conductivities of typical mineral grains.

Permeability is greatly affected by freezing. Figure 4.10 gives results on laboratory samples obtained by Sahin (1973) and by Williams and Burt (1974). The variation of several orders of magnitude which occurs on freezing is of some importance in choosing acoustic models for sediments subject to ice-bonding (section 8.1) The electrical conductivity of a soil sample changes greatly on freezing. Table 4.3a shows that the resistivity of a frozen soil is between 10 and 100 times larger than the resistivity of the same soil when unfrozen. This table is a compilation of data by Barnes (1963) from work of Joesting (1945) in Alaska, Meyer and Roethlisberger (1962) in Greenland and



Temperature dependence of apparent hydraulic conductivity of frozen soils as measured in permeameters by Williams and Burt (1974) for a silt and by Sahin (1973) for a 4-8 μm silt fraction. The flat portion may represent transport by ice, moving by regelation, at rates limited by the thermal conductivity of soil constituents.

Fig. 4.10

Variation of permeability with temperature in the freezing range from the sources cited above as compiled by Hillel (1980).

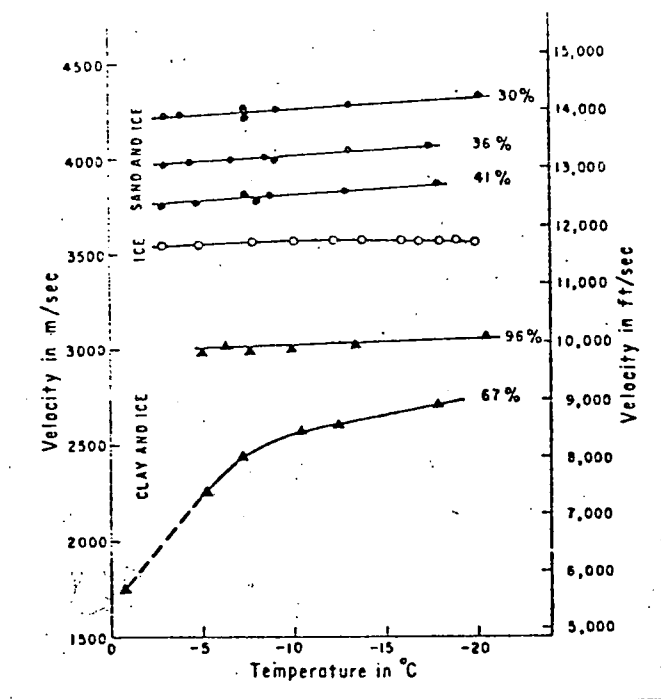


Fig. 4.11

Compressional wave velocity in frozen synthetic cores as a function of temperature below 0°C and of porosity (Mueller, 1961).

Hatherton (1960) in Antarctica. The lowest resistivities shown in table 4.3a for frozen materials are from the Antarctic and were made close to the ocean. Barnes (1963) is of the opinion that they may be influenced by the presence of saline water. Table 4.3b shows laboratory measurements of electrical properties of synthetic permafrost made by Cook (1960). His measurements were made at a frequency of 100MHz and a temperature of -35°C . The values he found are significantly lower than all of the measurements in table 4.3a, with the exception of those from the Antarctic. The difference may either be due to the very different frequency used or to possible differences in the original pore fluid and ice-crystal structure between the laboratory and the field soils.

The change in the elastic behaviour of a soil on freezing and on thawing has been investigated for acoustic and for geotechnical purposes. Mueller (1961) measured the compressional wave velocity in six synthetic sand and clay cores which had different porosities but which were saturated with water before freezing. Figure 4.11 shows his results on the velocity variations with temperature. The flatness of the curve for pure ice and the gentle slope for the four other curves suggest that a small portion of the interstitial water in these cores did not freeze until very low temperatures were reached. The much greater temperature dependence shown by the bottom curve

Rock type	Locality	Resistivities Ohm-cm x 10 ⁻⁴		Approx. Temp.
		Frozen	Unfrozen	
Silt and organic matter	Yukon-Tanana Area, Alaska	20-80	0.36-19	-1
Sand and fine gravel	" "	63-240	2.2-7.1	-1
Gravel	" "	78-410	10.0-18.5	-1
Conglomerate	Circle District, Alaska	120-160	2.2-7.0	-1
Glacial drift	Thule, Greenland	30-400	...	-11
Gneiss	" "	730-1370	...	-11
Basaltic debris	Scott Base, Antarctica	1.5-15	...	-21
Marine gravel	Hallett Base, Antarctica	.32-4.0	...	?

Table 4.3a

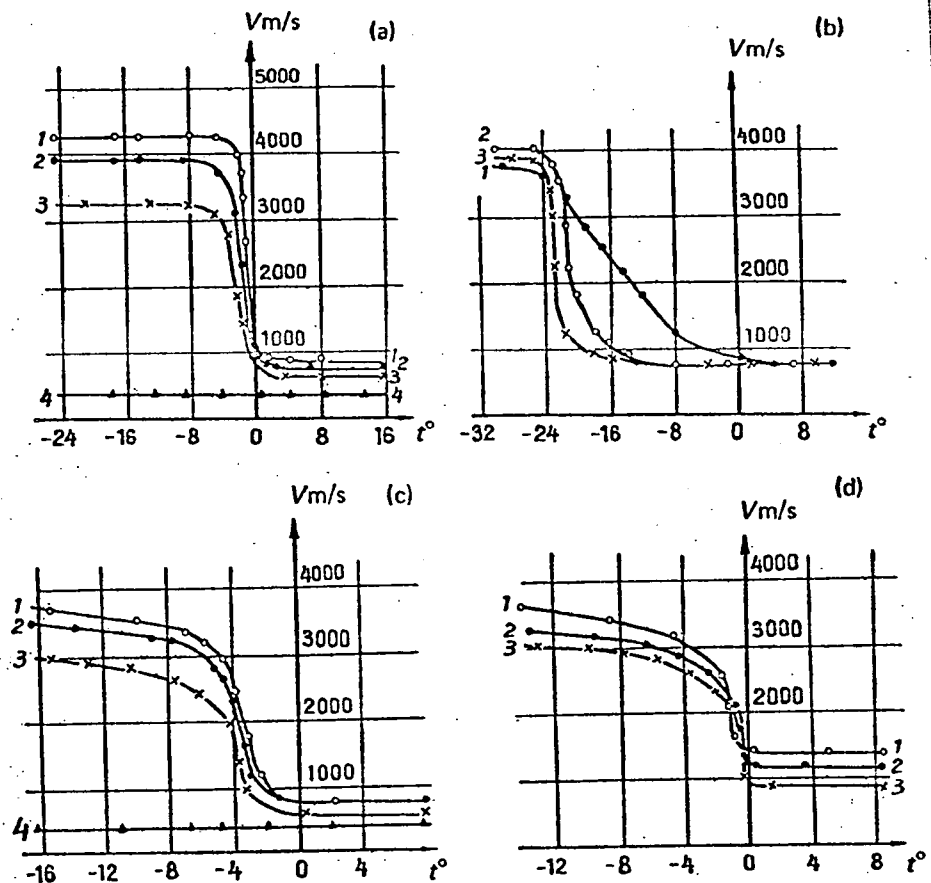
Radio frequency properties of synthetic permafrost
at 100 mc/sec and -35°C

	Resistivity (ohm-cm)	Dielectric constant
Frozen tap water with 0.2% Ca(HCO ₃) ₂	2 500 000	3.5
Frozen tap water in coarse sand	120 000	4.5
Frozen wet sandy clay	13 000	5.0
Brown soil	20 000	4.4
High Fe clay	6 800	7.0
High Al clay	62 000	3.3
Limy mud (lagoonal)	14 000	3.6
Silica mud (diatomaceous)	13 500	1.0

Table 4.3b

a. Field measurements of resistivity compiled by Barnes (1963) from data of Joesting, (1945) in Alaska, Meyer and Roethlisberger (1962) in Greenland and Hatherton (1960) in Antarctica.

b. Resistivity and dielectric constant of synthetic permafrost as measured by Cook (1960). Note that resistivities in Table 4.3a are given in ohm-cm X 10⁻⁴.



Curves showing the variation of the propagation rate of ultrasonic longitudinal waves in frozen and thawing soils as a function of their fineness, moisture content, mineralization, and temperature. (a) Fine-grained sand: 1-- $w = 21$ percent, 2-- $w = 16$ percent, 3-- $w = 10$ percent, 4-- $w = 1$ percent. (b) Fine-grained sand artificially salinized with NaCl at $w = 24$ percent: 1--mineralization 1 percent, 2--3 percent, 3--5 percent. (c) Silty supes: 1-- $w = 30$ percent, 2-- $w = 25$ percent, 3-- $w = 15$ percent, 4-- $w = 1.5$ percent. (d) Silty suglinok: 1-- $w = 28$ percent, 2-- $w = 34$ percent, 3-- $w = 40$ percent.

Fig. 4.12

Variation of compressional wave velocity with temperature above and below $0^\circ C$ for four different soil types (Votyakov, 1973). The Russian term "supes" describes a mud composed principally of silt with a minor proportion of sand and minimal clay. "Suglinok" is also predominantly silt, and has more clay than sand (Sanger, 1973).

(for a clay) suggests that a larger amount of pore-water remained unfrozen at temperatures near 0°C. Votyakov (1973) performed similar experiments to those of Mueller. He used a wider variety of soil types, which were only partially saturated with water. His results are shown in fig. 4.12. It is particularly interesting to note the results of fig. 4.12a and compare them with those of fig. 4.11. Both results are for sands, those of fig 4.11 for water-saturated sands, those of 4.12a for unsaturated. The compressional wave velocity in the frozen state is around 4km/sec in each case, whereas when unfrozen, the unsaturated sand has a velocity about one half of what is to be expected for the saturated sand. That is to say that whereas the acoustic velocity in an unfrozen sediment is very sensitive to gas content (as discussed in section 3.5), the same is not true of the acoustic velocity in a frozen sediment. In addition to further determinations of compressional wave velocities in frozen sediments, Antsyferov et al. (1964) have found values for shear wave velocities and attenuation factors (for compressional waves) by observing travel time and amplitude decay across laboratory samples of 20cm diameter and 10cm or 15cm thick. They made measurements at 28, 45, 65, 75, and 170KHz, and found little or no velocity dispersion. For their frozen sand material (not described in detail) they found $v_p = 4650 \pm 50$ m/sec and $v_s = 2760 \pm 40$ m/sec. This yielded a Poisson's ratio of 0.23 ± 0.02 , a value even lower than

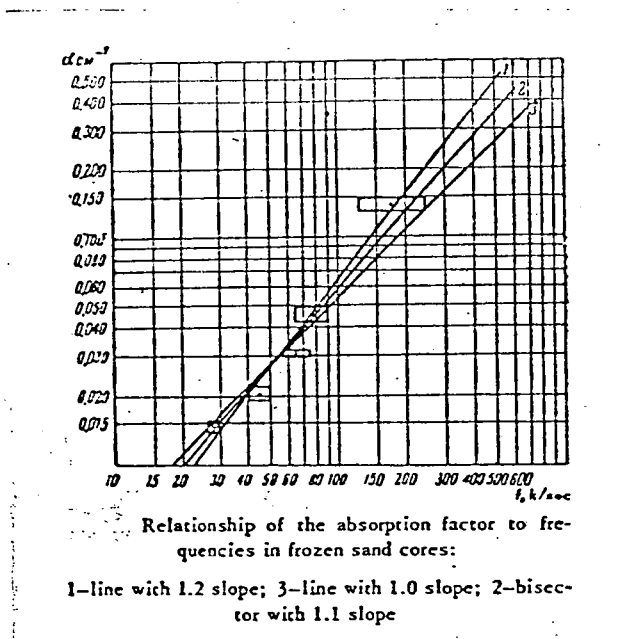


Fig. 4.13

Variation of the absorption factor ($\alpha_f \text{ cm}^{-1}$) with frequency (KHz), in frozen sand cores. (Antsyferov et al, 1964).

the 0.25 which Yoshimura et al. (1982) have found by well-logging observations in soft rocks such as sandstones, tuffs, and mudstones. In contrast to this, the values which they found for attenuation (shown in fig. 4.13) do seem to be typical of those found for water-saturated sediments. For example, from fig. 4.13, they found at 380KHz (extrapolated) attenuation $\alpha = 0.25 \text{ cm}^{-1}$ which can be expressed in dB/cm as $(10/2.303)\alpha \text{ dB/cm}$, i.e. ca. 1.0dB/cm. This is about the average value found at the same frequency by McKay (1967) in laboratory measurements on water-saturated beach sands. That Antsyferov et al. have found a material having a Poisson's ratio typical of a cemented rock and an attenuation typical of an unconsolidated water-saturated sediment is surprising. One explanation could be that they experienced difficulty in coupling the transducers to the frozen sample, although their method of freezing the transducers in place does seem to be a good one, and the regression lines which they have fitted over the different frequencies used do appear valid. Another possibility is the inclusion of air in the sample. It has been noted above that this does not seem to have a large effect on the compressional wave velocity, but it may still have an effect on attenuation.

4.3 GEOPHYSICAL METHODS OF DETECTING PERMAFROST.

The identification of permafrost by geophysical methods depends upon the recognition of the physical properties discussed in section 4.2, or, in the case of remote sensing methods, on the recognition of landforms as discussed in section 4.1.

Ferrians (1972) and Haugen et al. (1972) have used remote sensing data acquired by a satellite (the Earth Resources Technology Satellite) to monitor changing conditions of large scale features which can be discerned on 1:1,000,000 - images of parts of Alaska. Aerial photographs (visible light or infra-red) have been used by Greene (1971,1972) in evaluating test areas along the line of the trans-Alaska pipeline, where the technique proved useful in route selection. The scanning infra-red technique has been used by LeSchack and Morse (1972), again in Alaska for the successful detection of ice-wedge polygons which were not clearly seen from the air by other methods.

The variation in density across ice-cored topography has been detected by Rampton and Walcott (1974) who have run gravity profiles in the neighbourhood of the MacKenzie delta. They modelled the shape and thickness of subsurface ice from these and obtained geologically

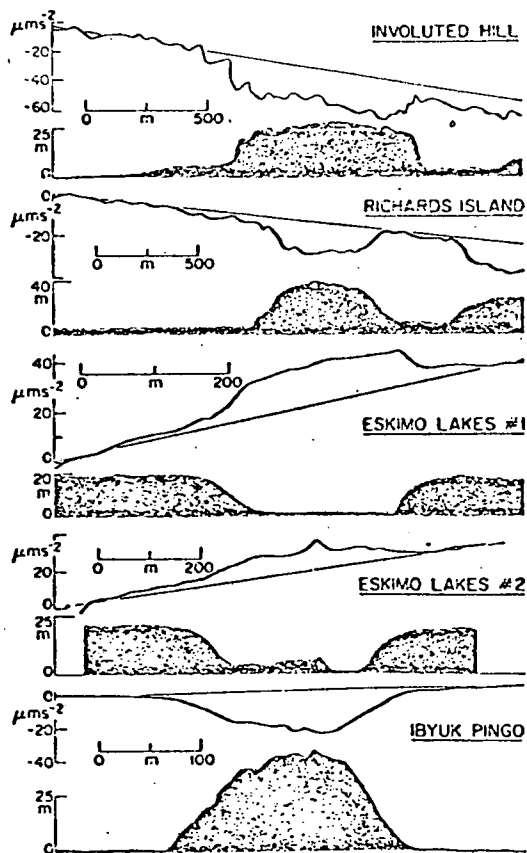


Fig. 4.14 Bouguer anomaly (density of reduction 2.0 Mg m^{-3}) and topographic profiles.

Fig. 4.14

Gravity profiles of Rampton and Walcott (1974), from various locations in the neighbourhood of the MacKenzie delta.

acceptable results, but were unable to verify the models by drilling. They note that an error in the assumed density of the frozen sediments of 5% will give rise to a 10% error in the calculated thickness. The data is interpreted with any local linear trend removed and consequently it is not possible to detect changes in ice-content which also have a linear trend, nor to detect uniform slabs extending the full length of the profile.

The resistivity method is probably used more than any other for the identification of permafrost which may lie beneath proposed structures such as buildings, roads or pipelines. The profile of Barnes and MacCarthy (1964) (fig. 4.15) is a good example of the detection of change in level of a permafrost surface by this method. An expanding electrode spread is required to determine the depth of an unknown permafrost surface. MacKay (1972b) concludes that resistivity methods may be used to identify construction materials such as sand and gravel in permafrost, on the basis of a higher resistivity due to well-frozen pore water in these materials. Difficulties with the method have been pointed out by Hessler and Franzke (1958) and by Gross (1962) who note that electrode resistance increases markedly at temperatures below -20°C . A further potential difficulty with the method has been pointed out by Ostrem (1967) who noted that laboratory specimens of ice had different resistivities in different

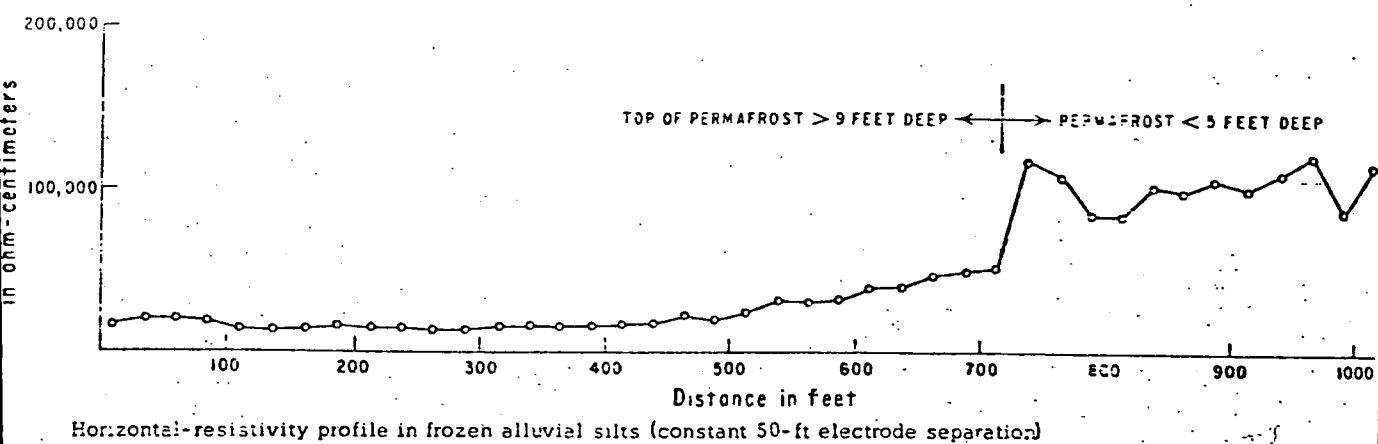


Fig. 4.15

A resistivity profile showing the ability of this method to detect variations in the level of the top of permafrost. Wider electrode spacings can be used for surfaces at greater depth. (Barnes and MacCarthy, 1964).

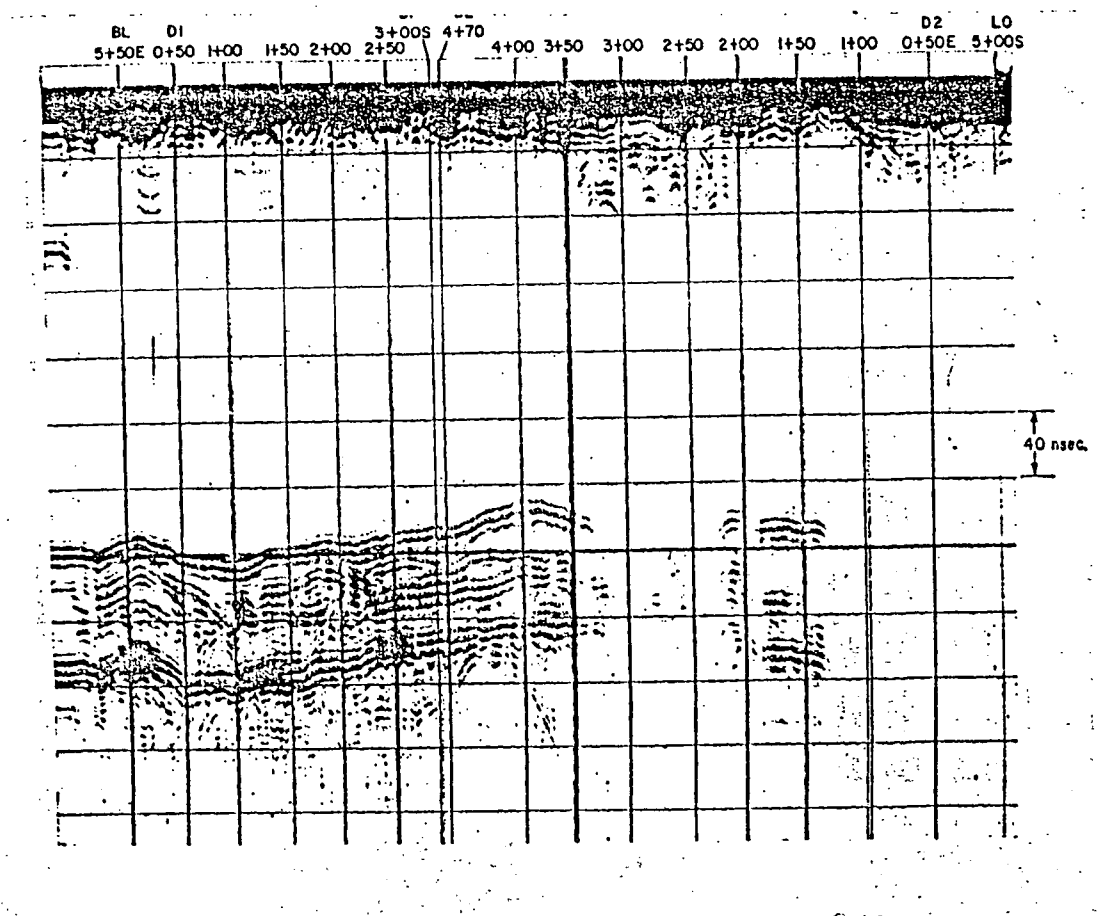


Fig. 4.16

Impulse radar profile of an ice-cored hill near Tuktoyaktuk. The hill had previously been mapped by MacKay (1963, 1966, 1971) and Rampton and MacKay (1971) and was known to have an interior structure consisting of a 20m slab of ice overlain by 5 to 10m of "ice-clay-till" mixed with peat. The above section of Annan et al. (1975) clearly shows the base of the ice (at 280 nsec) and shows spasmodic reflections from within the overlying layer.

directions.

Electromagnetic induction methods have been treated theoretically by Sinha (1976) who considered frequencies of between 10KHz and 1MHz with coils separated by 3 metres, and 1 metre above the ground. He concludes that the technique may be viable, and Sartorelli and French (1982) found it a fast way of doing initial exploratory surveys over sections of the corridors proposed for pipelines in Northwestern Canada. They found that the electromagnetic method was better suited to finding unfrozen anomalies in fine-grained soils than in coarse grained. They did not consider that it was possible to estimate ice content in this way.

An even higher frequency electromagnetic method is that of pulsed radar profiling. This relies on vertical incidence reflexions of radar pulses from discontinuities in subsurface strata. The method yields sections similar in appearance to marine seismic reflexion profiles (fig. 4.16). Since the velocity of propagation of the electromagnetic wave is at least 5 orders of magnitude greater than that of an acoustic wave, the vertical travel time scale is in much smaller units - nanoseconds in the case of fig. 4.16. The technique seems first to have been applied in cold regions by Robin et al. (1969) who used it to investigate the thickness of ice sheets. Annan et

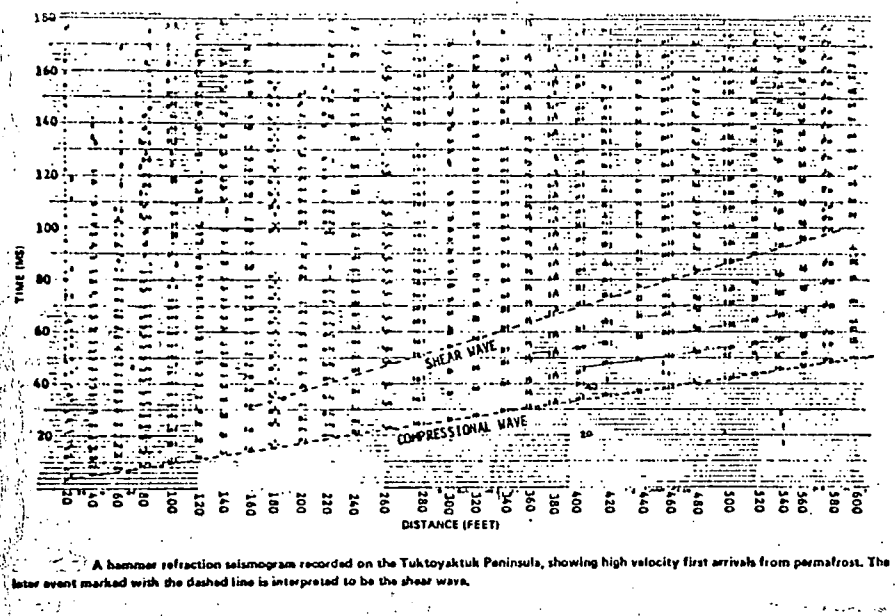
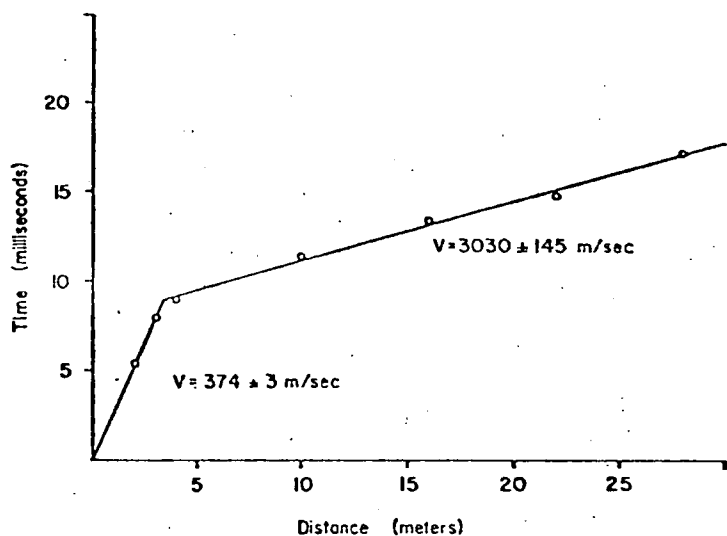


Fig. 4.17

A hammer refraction seismogram recorded by Hunter (1974) on the Tuktoyaktuk Peninsula. The first arrivals show a compressional-wave velocity of 3650m/sec. Later arrivals are interpreted as a shear-wave of velocity 2000m/sec.

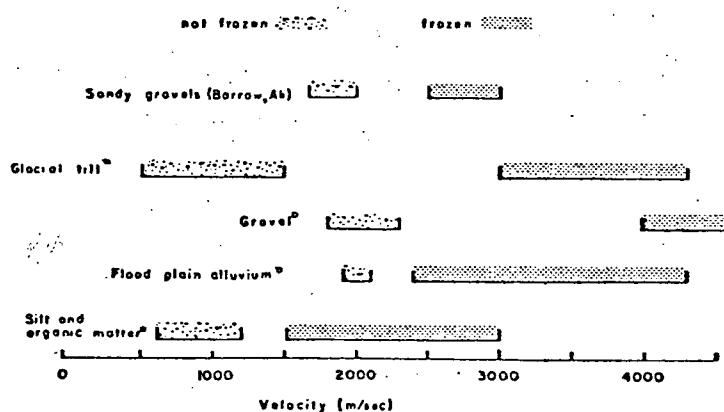
al. (1975) tried out the method over frozen soils in the Tuktoyaktuk Peninsula and found that, in places a near-surface clay layer attenuated the signals appreciably and limited the usefulness of the method in some areas. They attribute the fading of the reflector at 280nsec on the right hand portion of fig. 4.16 to this cause. Davis et al. (1976), who have worked with radar in the same region, conclude that the method is good for detecting interfaces between ice and sand at depths up to about 30 metres. They have experimented with the wide angle reflexion method of measuring propagation velocities and found that this was a good alternative to deriving velocities on the basis of measured dielectric constants of soil samples, but they consider that drillhole information is essential for use in conjunction with radar in order to correlate material changes with observed reflexions.

The seismic method of mapping permafrost in surficial materials is considered by Hunter (1973) to be one of the most promising geophysical techniques. Early field experiments were done by Roethlisberger (1961) in Greenland. His surveys showed that high velocities were associated with permafrost occurrences. Seismic refraction is the method (e.g. Dobrin, 1960) preferred by most workers. A sledgehammer and iron plate is usually an adequate acoustic source and fig. 4.17 is an example of a



a

Time-Distance plot of data taken on Cross Island. The uncertainties in the velocity values are determined from the standard deviations of the straight line fits.



b

Compressional wave velocities in frozen and non-frozen materials. The Barrow data were gathered by the authors. Data marked by an asterisk are from Roethlisberger (1972). The Barrow data were compiled from 28 velocity measurements, 12 of which were from reversed refraction lines. Neither the number of records nor whether the data were from reversed lines was reported by Roethlisberger.

Fig. 4.18

a. Time-distance plot of a refraction line of Morack and Rogers (1981) near Point Barrow (Alaska).

b. Compiled data on velocities and sediment types from Morack and Rogers (1981) and Roethlisberger (1972).

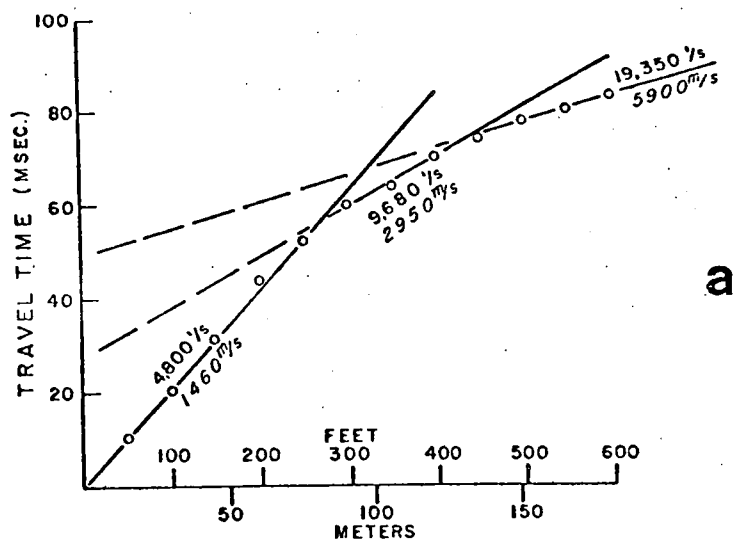
hammer seismic refraction profile made in this way by Hunter (1973), using a Huntect FS-3 seismograph. This is a single-channel device which records in variable density format all waveform peaks above some threshold level. As such, it does not preserve the actual waveforms received. Hunter has measured compressional wave velocities from reversed lines made in this way and has sometimes been able to pick convincing shear-wave arrivals (as in fig. 4.17) but these generally are obscured by the general clutter of arrivals occurring after the headwave. A table of velocity values determined in a field location where the soil materials were well-exposed to view (viz. in a tunnel near Fairbanks, Alaska) is shown in table 4.4 (Hunter, 1974). Morack and Rogers (1982) worked in a similar way farther to the north, on islands along the Alaska coast of the Beaufort Sea. Fig. 4.18a shows one of their time-distance profiles, which gives a very clear velocity break between a loosely compacted overburden and an underlying ice-bonded material. Morack and Rodgers compiled their own data from reversed refraction lines near Point Barrow with that of Roethlisberger (1972) to produce fig. 4.18b, and they note that this compilation shows that it is not always possible to separate frozen and unfrozen materials by using velocities alone. It is also necessary to have some idea of the type of material being investigated - the principal characteristic which should be known is the approximate grain size.

SITE	SEDIMENT TYPE	COMPRESSIONAL VELOCITY		SEISMIC RECORD QUALITY	ICE CONTENT BY VOLUME	
		Measurement	Average		Measurement	Average
1	Silt	3.20 km/sec	3.14 km/sec	Good	68% 76	72%
		3.05				
		3.13				
		3.29				
		3.02				
2	Ice wedge	3.45	3.00	Fair	100	100
		2.97				
		2.53				
		3.05				
3	Ice wedge (clear ice)	3.40	3.22	Fair	100	100
		2.45				
		3.81				
4	Silt	2.98	2.83	Fair	57	62.5
		2.67			68	
5	Silt	3.20	3.17	Good	71	71.5
		3.13			72	
6	Silt	2.96	3.01	Good	59	59
		3.05				
7	Silt	3.02	3.02	Fair	63	63
	Gravel	--	--	No record	8.9 - 10.0	
Bedrock	Schist	3.20	3.20	Poor	6.5 - 19.9	

Note: Temperature of tunnel walls at time of experiment -4.2°C.

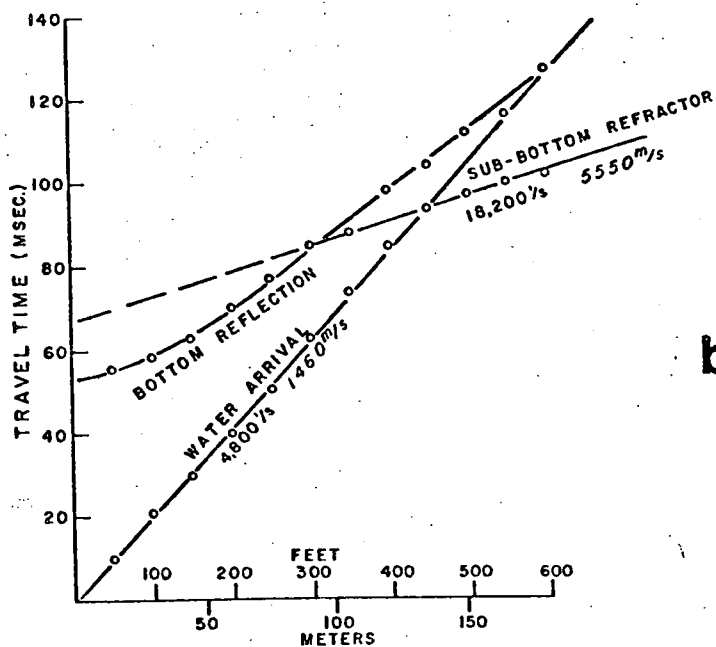
Table 4.4

Compressional wave velocity measurements made by Hunter (1974) in a tunnel near Fairbanks, Alaska.



SOURCE-RECEIVER DISTANCE

A first arrival time-distance plot of a marine refraction record showing a three layer interpretation.



SOURCE-RECEIVER DISTANCE

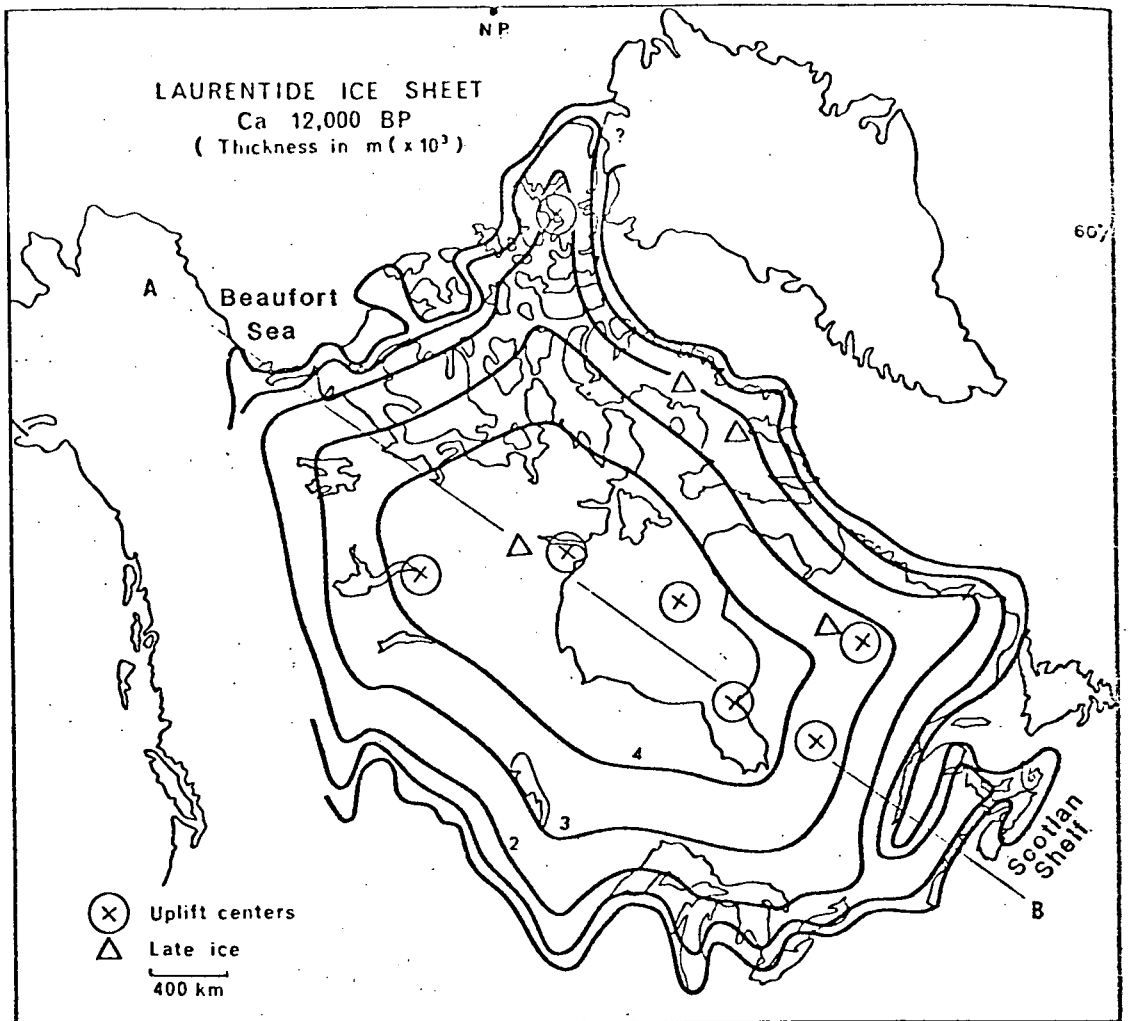
A time-distance plot showing a "hidden layer" case where the intermediate refractor velocity is not seen as a first arrival.

Fig. 4.19

Time-distance plots of refraction lines of Hunter and Godfrey (1975).

- Line showing 3 distinct velocities.
- Line with intermediate layer hidden.

The conventional land seismic refraction technique described above indicated that permafrost extended to the sea-coast. The first attempts at carrying this investigation beyond the shoreline were made in various shallow bays along the Beaufort coast (Hunter, 1973; Hunter and Godfrey, 1975; Carson et al., 1975). Beyond inter-tidal depths, where bay cables were used, with geophones, it became necessary to convert to hydrophones. Hammer acoustic sources were replaced with detonators or small explosive charges. Hunter and Godfrey were able in this way to observe (fig. 4.19) good velocity breaks from subsea layers in Cunningham Inlet, Somerset Island. Their results show a difficulty, which is known on land, but can be a greater problem at sea for refraction studies because of the increased velocity and depth of a water layer compared with that of the uppermost soil layer on land. This difficulty is the "hidden layer" problem. It can be seen that in fig. 4.19a, three distinct velocities can be derived from the first arrivals: 1460m/sec, identified with the water layer; 2950m/sec, identified with ice-bonded sediment and 5900m/sec, identified with bedrock. In fig. 4.19b, the intermediate velocity does not appear. Because of the greater water depth, the headwave associated with the interface between water and seabottom permafrost never appears as a first arrival. Such observations, however, suggested that the mapping of subsea permafrost might be carried out by seismic methods.



Ice margin about 12,000 BP (from Bryson *et al.*, 1969); tentative contours on the ice sheet (no allowance made for isostatic depression) and locations where strandline tilts intersected (from Andrews and Barnett, 1972).

Fig. 4.20

The Laurentide Ice Sheet (near maximum) as outlined by Andrews (1973).

Before this is discussed further, the way in which permafrost forms beneath the sea is recounted in the next section.

4.4 SUBSEA PERMAFROST IN THE BEAUFORT SEA.

In order to understand the origin of subsea permafrost in the Beaufort Sea, it is necessary to consider the recent history of sea level changes in that area. Such changes have been dominated not only by the general world wide effects of locking up water in large ice sheets during the Wisconsin epoch (30,000 - 7000 yr. B.P.) but also by local tectonic effects associated with the proximity of the Beaufort region to the edge of the Laurentide Ice sheet, Andrews (1973). Sea level is normally determined by the recognition of strand-line geomorphologic features, particularly raised beaches, in the case of a falling sea level or else by the study of such micro-fossils as can be used to distinguish between fresh and salt-water environments (Scott and Medioli, 1978, 1979). A detailed discussion of the species of estuarine foraminifera useful for making this distinction is given by Scott et al. (1980).

Figure 4.20 shows Andrews' (1973) interpretation of the shape and thickness of the Laurentide ice sheet. The

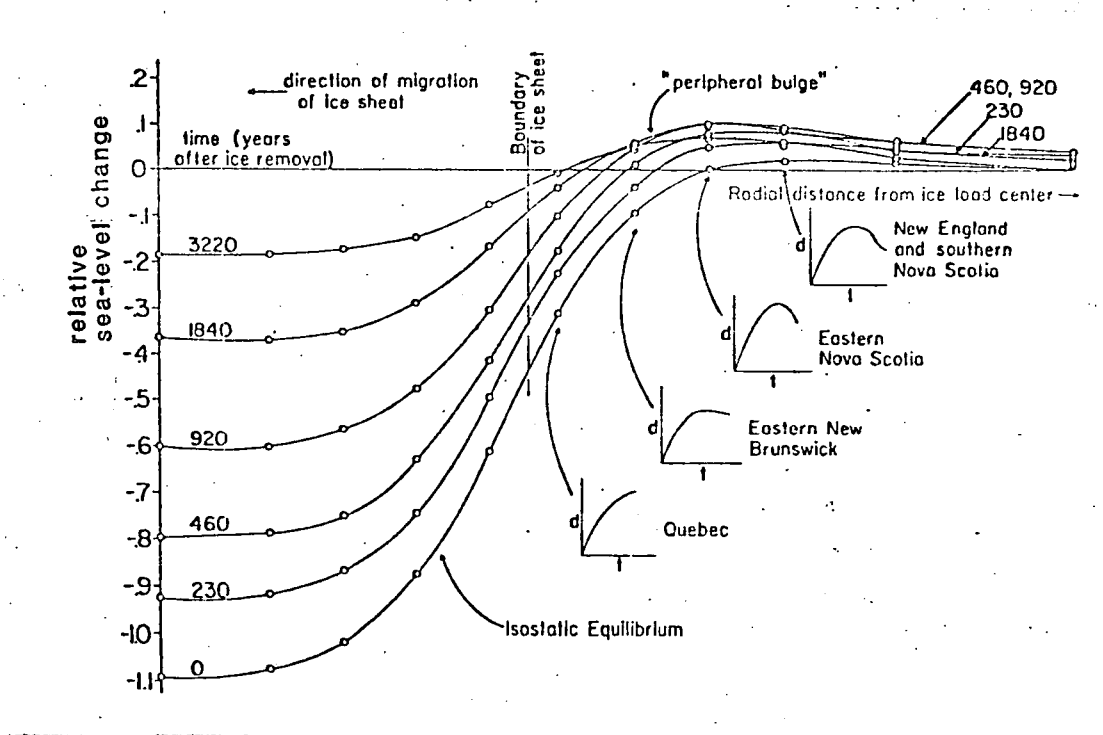
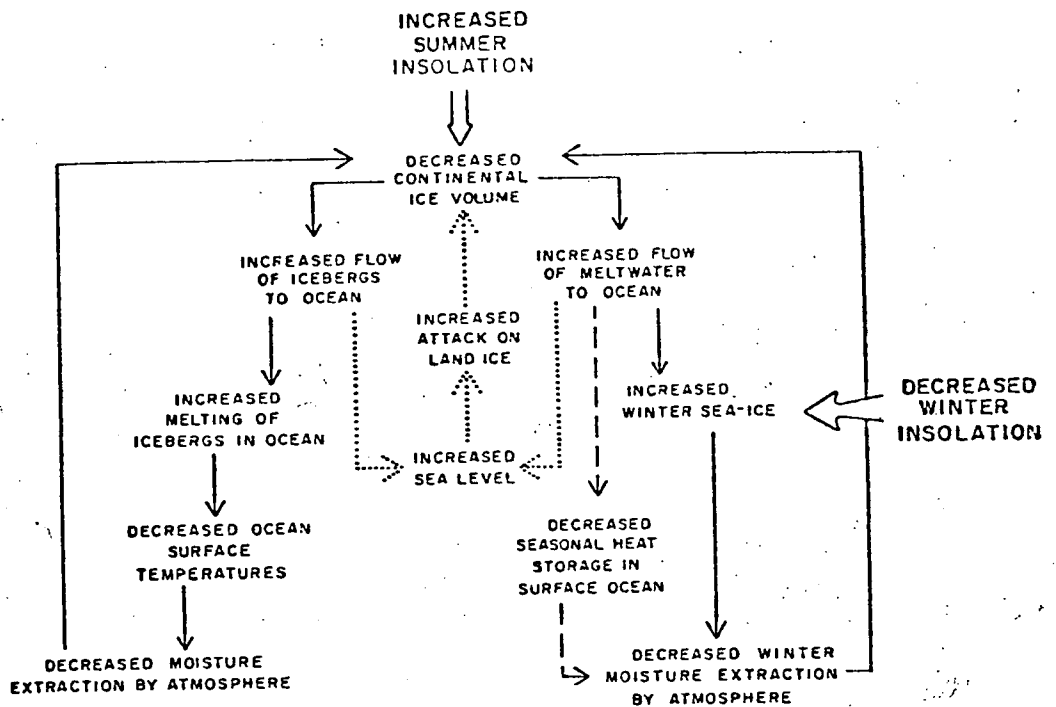


Fig. 4.21

Predicted change in sea level since Wisconsin maximum on the basis of expected rebound of a depressed elastic lithospheric plate overlying a viscous substrate. The small insets show the observed sea level changes at different localities in Eastern Canada and New England. (Scott and Medioli, 1978).

Beaufort Sea lies diametrically opposite the Scotian Shelf on the periphery of the sheet, which is centred on Hudson's Bay. It is reasonable to expect that the relative vertical movements between land and sea would be similar in these two peripheral areas. Accordingly, it is useful to look at studies of sea level changes in Eastern Canada. A comparison between theoretical and observed sea level changes during the melting of the ice sheet is given in fig. 4.21 (Scott and Mediolli, 1978). The "peripheral bulge" shown in this figure is expected on the basis of the lateral displacement of viscous mantle material from beneath the ice sheet. It is clear from the figure that in the vicinity of this bulge, an initial uplift followed later by subsidence is both predicted and observed. Clark et al. (1978) have generalised comparisons between different parts of the world on the basis of the effects of changes in post-glacial sea-level, and they classify the Beaufort Sea and the Scotian Shelf as lying within the same zone in their scheme. The formation and breakup of large ice-sheets has been discussed by Ruddiman and McIntyre (1981) who have studied microfossil abundances and calcium carbonate content of sediment cores taken from the floor of the N. Atlantic Ocean. They identify cyclic periods of 23,000 years, both for temperature (based on calcium carbonate solubility) and for surface-water productivity (based on foraminifers and coccoliths), but they note that sea-surface temperature lags in phase by



Positive feedback loops of potentially critical importance in explaining rapid deglaciations ("terminations"). All four mechanisms rely on outflow of icebergs and water from the disintegrating ice sheets to accelerate the deglaciation.

Fig. 4.22

Positive feedback mechanism for breakup of ice sheets as proposed by Ruddiman and McIntyre (1981).

one quarter cycle. This suggests to them that sea-surface temperature varies in phase with the negative first derivative of ice volume, that is to say , with the meltwater or iceberg outflow. A positive feedback mechanism is indicated by this observation, wherein both increasing sealevel and decreased precipitation combine to bring about rapid break-up of the ice sheet (fig. 4.22). The suppression of biotic productivity during a period of ice breakup and global warming has been attributed by Adam (1973) to the formation of a thick layer of low-salinity water which inhibits the upwelling of nutrients into the photic zone.

A curve (fig. 4.23) showing change of sea-level in the Beaufort Sea is quoted by Hunter et al. (1976) from MacKay (1972b) who has taken it from Mueller-Beck (1966). This curve appears to be really of a generalised nature, for world-wide sea level changes, after Zeuner (1959), and became applied specifically to the Beaufort Sea area when Mueller-Beck used it in discussing the migration of palaeolithic hunters from Asia into America across the Bering Strait. Undoubtedly the curve gives a good general idea of sea level variations in this region, but the indicated relative rise in sea level over the past 4000 years seems unrealistically small, when coastal erosion is known by a comparison of aerial photographs of 1935 and 1970 (MacKay, 1972b) to be proceeding at a rate of up to

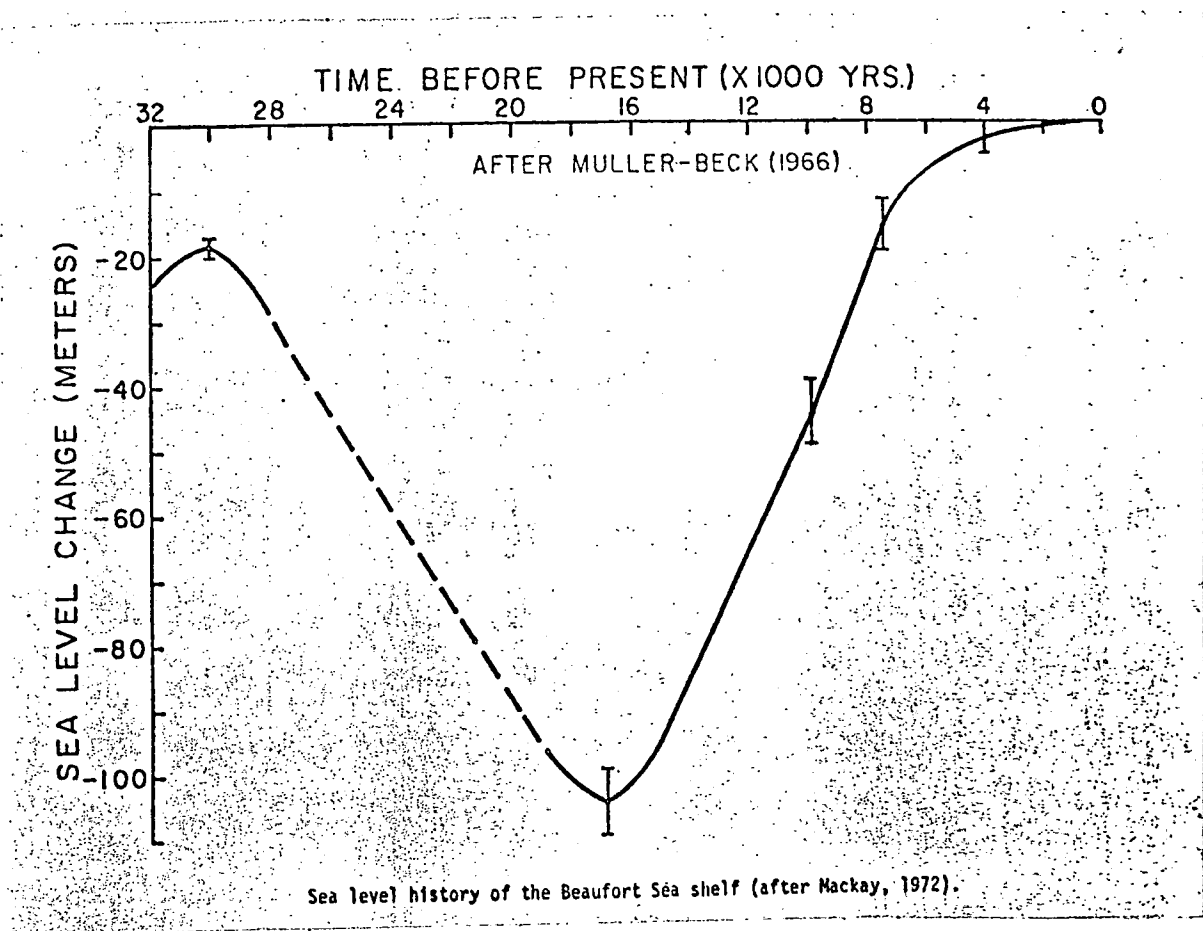
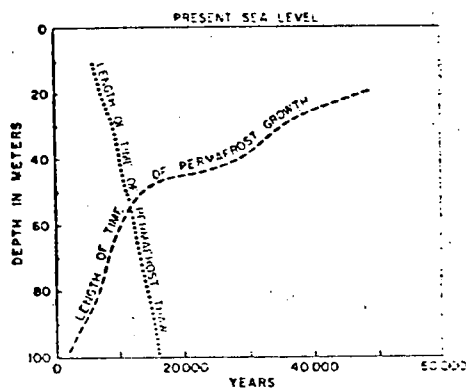
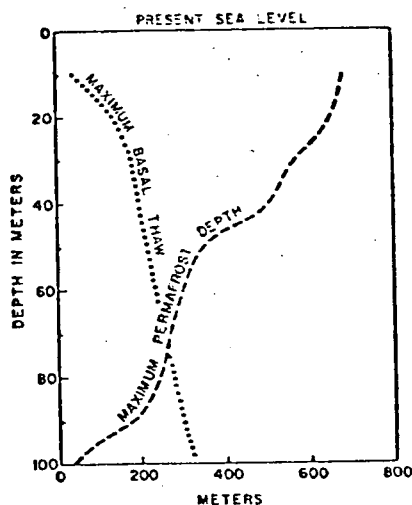


Fig. 4.23

Sea-level changes in the Beaufort Sea as given by Hunter et al. (1976). The significance of the error bars and the dotted line are not made clear in that text, and they do not appear in the quoted sources (MacKay, 1972 and Mueller-Beck, 1966).



a



b

Fig. 4.24

A comparison between the available thawing and freezing times (a) with the maximum depth at which permafrost is predicted to occur (b). (MacKay, 1972). The lengths of time available for permafrost growth and thaw are taken from the sea-level variations. The curve of maximum permafrost depth has been found for a value $b=3$ (intermediate between that of equations 4.3 and 4.4). The curve of maximum basal thaw assumes a rate of 2 cm/yr.

200 metres in 35 years. The principal use made by MacKay (1972b) of the curve in fig. 4.23 was to estimate the depth to which permafrost might exist beneath the recently inundated shelf areas of the Beaufort Sea (fig. 4.24). MacKay (1972b) notes that it is impossible to compute accurately the expected depth of permafrost, even if the exact variation of sealevel were known. The invasion of permafrost into a former sea-bed at an emergent coastline will depend, in addition to the length of time of subaerial exposure, upon the climate during emergence, on the thermal conductivity of the frozen and unfrozen sediment and the salinity and rate of flow of ground-water. MacKay's estimates of fig. 4.24b are based on the relationship:

$$h = b\sqrt{t} \quad \dots\dots\dots(4.1)$$

where h is the depth of freezing, t is time, and b is the solution of:

$$KTe^{-\frac{b^2}{4\alpha}/\sqrt{(\alpha)}} \operatorname{erf}\{b/2\sqrt{\alpha}\} = -\sqrt{\pi}/2Lnb \dots(4.2)$$

where K = thermal conductivity of the frozen soil

T = mean ground surface temperature (C)

L = latent heat of fusion

n = porosity of the frozen soil

α = diffusivity of the frozen soil

MacKay reduces this relationship by the substitution of appropriate values for the parameters to:

$$h=4\sqrt{t} \text{ for a low porosity sand } \dots\dots\dots(4.3)$$

$$\text{and } h=1.5\sqrt{t} \text{ for an icy clay } \dots\dots\dots(4.4)$$

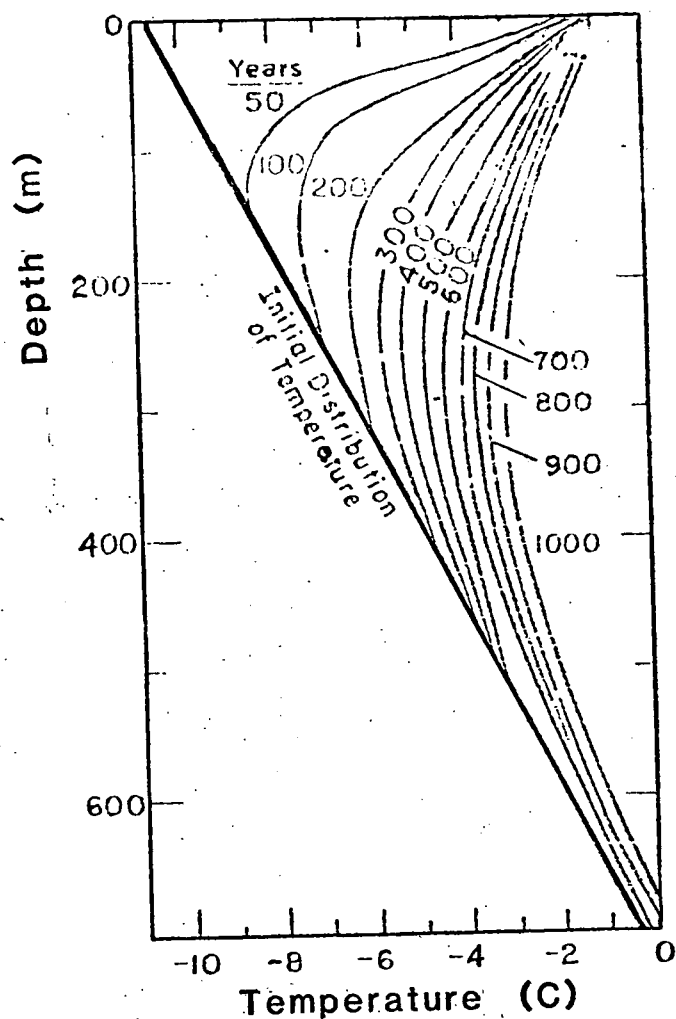


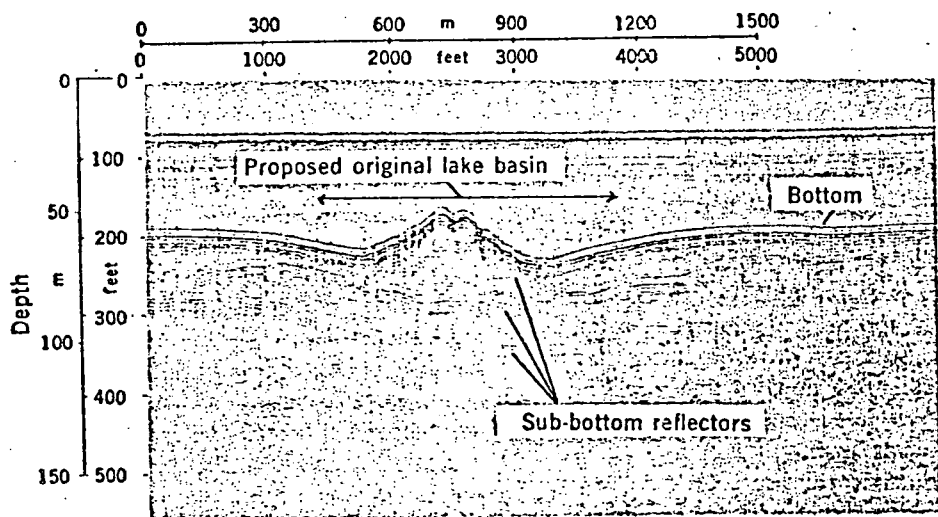
Fig. 4.25

The time response to a step change of 10°C of the geothermal profile as predicted for the conditions expected in the Laptev Sea by Molochuskin (1973).

where the depth is in metres and the time in years.

The submergence of a frozen land area beneath sea water having a temperature below 0°C will bring about a change in the geothermal gradient, so that the permafrost will degrade from the bottom, even if it does not degrade from the top. This point was noted by Terzaghi (1952) in connexion with subaerial permafrost and he showed that the maximum rate of upward thaw in a frozen sand of porosity 30% is about 2cm/yr. when the geothermal heat flux is about 1 micro-calorie/cm²/sec. Relaxation times of the geothermal gradient following on a step change in temperature at the surface (fig. 4.25) have been given by Molochuskin (1973). Hunter et al. (1976) give a similar diagram and note that the time to achieve a new equilibrium state after a step change of about 10°C (as occurs on inundation in the neighbourhood of Tuktoyaktuk) is "many tens of thousands of years". They note that after 300 years, the melting of frozen ground is expected to penetrate only 12 metres beneath the seabed.

Conflicting predictions to that of MacKay (1972b) have been made concerning possible subsea permafrost. Black (1950) considered that the existence of ice beneath the seabed would not be possible because of the high salinity he expected in porewater. Lachenbruch (1957) expected offshore permafrost to exist only within a few thousand metres of shorelines, unless active shoreline recessions



Seismic reflection trace over one of the pingo-like features.

Fig. 4.26

Subsea pingo as found by Shearer et al. (1971) seen on a high resolution profiling section.

were occurring. Werenskiold (1953) suggested that many narrow fjords might be completely underlain by permafrost, but did not expect it far offshore. By the time of MacKay's (1972b) paper, several field observations had been made which no doubt gave him confidence in publishing what was considered a startling prediction about the thickness of permafrost which might be encountered on the Beaufort shelf, a prediction which was not generally accepted by the planners of exploratory hydrocarbon drilling until they had later encountered difficulties from this very cause (J. Hunter, personal communication). These were the observations by Baranov (1959) of perennially frozen ground beneath water in recently submerged sections of the Lapteev and East Siberian Seas, the appearance of frozen sediments beneath Deception Bay in Northern Quebec (Samson and Tordon, 1969), the recovery of a core containing freshwater ice from the seabed off Cape Bathurst (Northwest Territories) and the hypothesis by Shearer et al. (1971) who explained the formation of prominent mounds on the bed of the Beaufort Shelf as underwater pingos (figs. 4.26 and 4.27). Some years after the initial discovery of these mounds, I observed a few on subbottom profiling sections, and these are shown in fig. 4.27. The similarity in appearance of fig. 4.27a and the photograph of the subaerial pingo near Tuktoyaktuk (fig. 4.7) is remarkable. The stratification apparent in the sediments to the left and right of fig. 4.27a is clearly

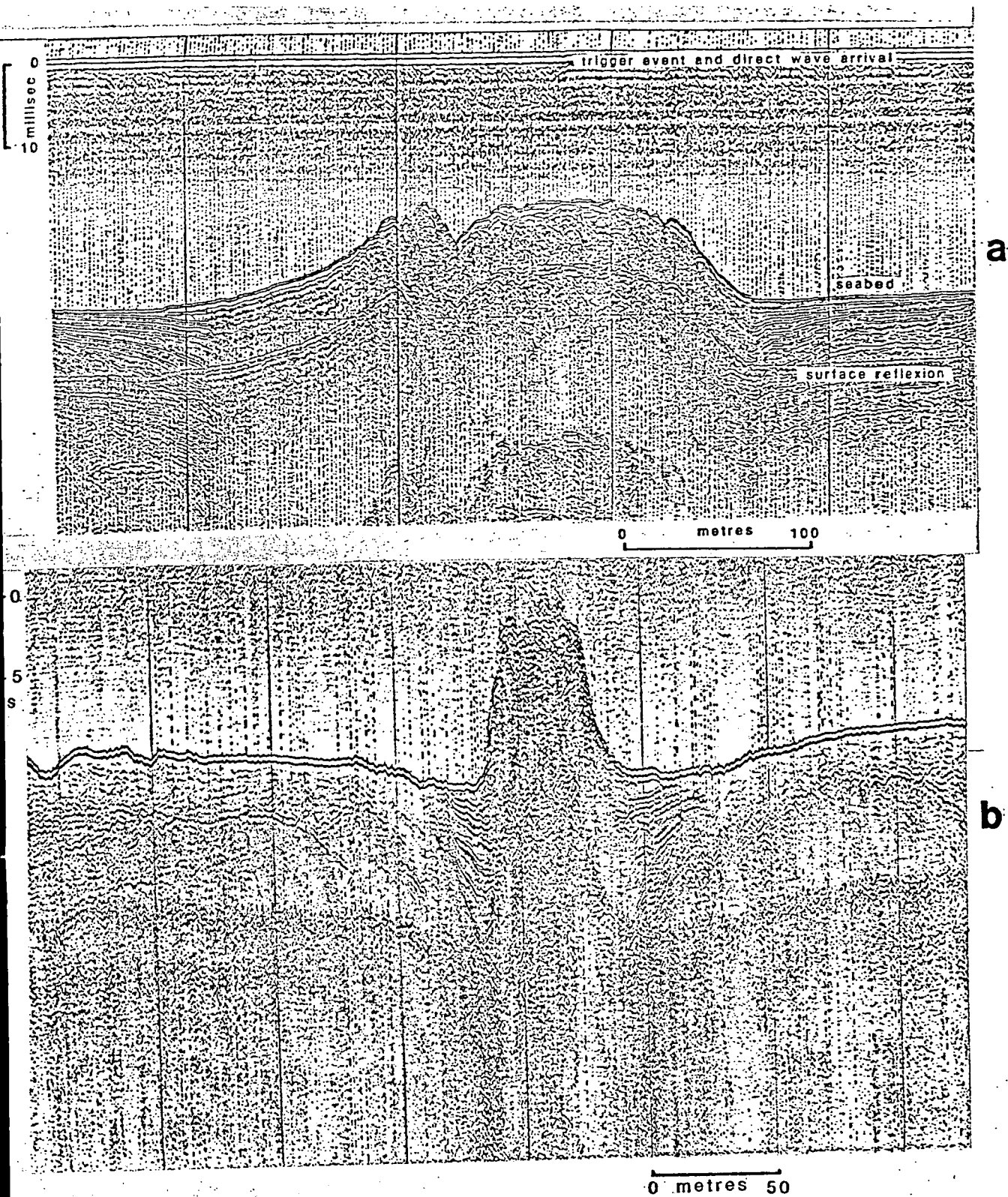


Fig. 4.27

Features similar to that shown in Fig. 4.26. These profiles were obtained during 1981 with a boomer as acoustic source and 3m long hydrophone array as receiver.

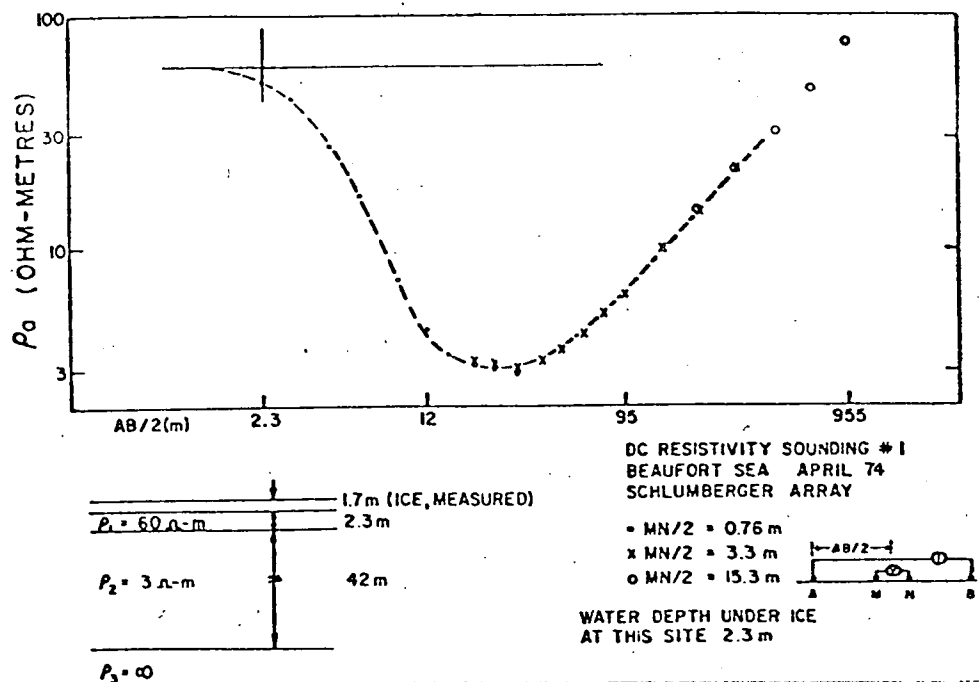
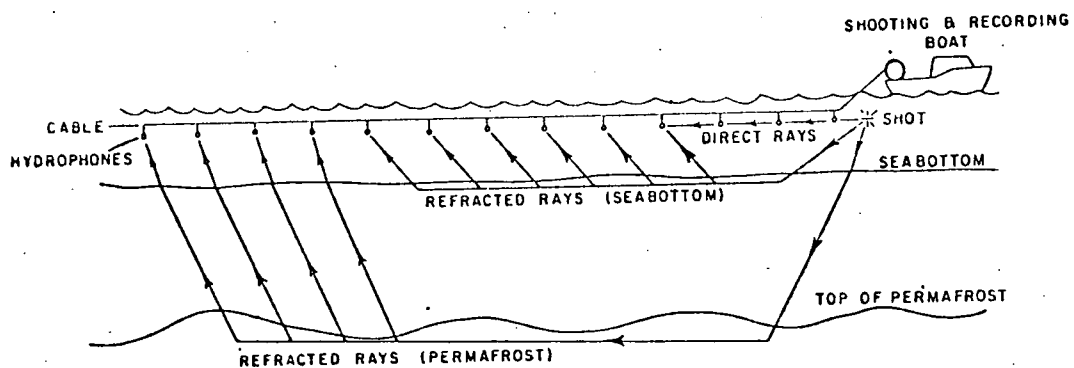


Fig. 4.28

Resistivity sounding through ice in Kugmallit Bay (Scott and Hunter, 1977). The interpreted thickness and resistivity of layers are given for the simple model shown on the right.

disrupted by the mound. The lower example (fig. 4.27b) appears to have the "moat" found around subaerial pingos.

Soon after the presence of frozen sediments beneath the seabed became established, considerably more effort was made to explore its extent, both by drilling (Lewellen, 1975; Judge et al., 1976) and by geophysical techniques. Both resistivity and seismic methods proved useful. Scott (1975) carried out seabed resistivity work in Kugmallit Bay, by deploying electrodes on the seabed. In the winter he did this by lowering electrodes into the water through holes cut in the sea ice. In summer, he deployed them from small boats. One of his resultant resistivity profiles, for expanding spacing between electrodes is shown in fig. 4.28. The interpretation given in that figure by Scott and Hunter (1977) is of a 42m. layer of material of 3 ohm-m. resistivity, probably composed of clays and silts overlain by a measured 2.3m of water and resting on a substrate of frozen material. They consider that the high value of resistivity (60 ohm-m.) for the water layer is the result of dilution of the seawater by fresh water flowing from the MacKenzie River. They also observe that even although the current electrode half-spacing was increased to 5km., no change was found in slope of the curve as seen towards the right hand of fig. 4.28. This appears to indicate that the penetration of a thick layer with high resistivity would require



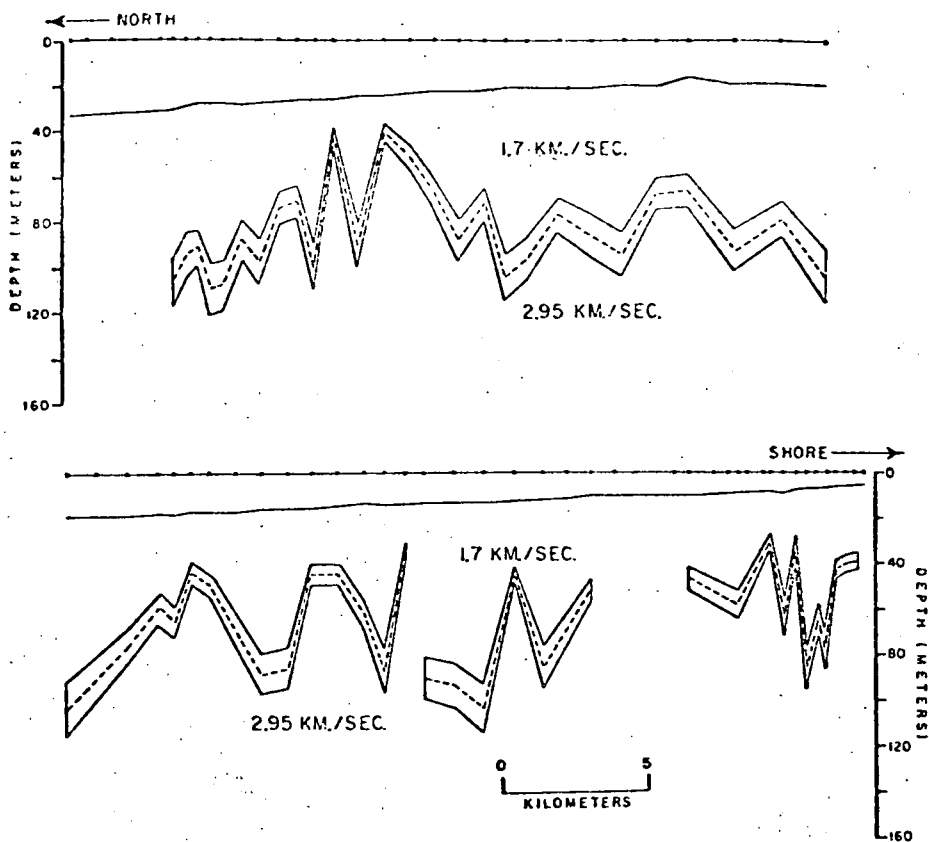
The marine seismic refraction technique.

Fig. 4.29

The seismic refraction technique of Hunter and Hobson (1974). They fabricated a marine seismic cable by suspending hydrophones on two-metre long pigtails from takeouts on a land seismic cable designed for use in swamps.

inconveniently large electrode spacings, and they conclude that the thickness of sub-seabottom permafrost cannot be reliably found from resistivity soundings. The method could be used to define the top of the frozen layer, although this would become more difficult as the depth of water increases.

Hunter and Hobson (1974) attempted to extend further the seismic methods which had been used on land and in shallow water to detect permafrost (see section 4.3). They suspended hydrophones by 2 metre long 'pigtails' from the takeouts of a land seismic cable intended for use in wet conditions (fig. 4.29). As a seismic source, they used small charges fired 2 metres below the surface. In this way they were able to map an abrupt velocity transition in an area off Atkinson Point on the ~~Tuktoyaktuk~~ peninsula (fig. 4.30). ~~This~~ very irregular surface proved to be fairly typical of the top of the permafrost layer so that this discontinuity, although of a non-stratigraphic nature can scarcely be said to be "bottom simulating" (section 3.6). The increasing amount of exploration work for hydrocarbon deposits on the Beaufort Shelf made available a large quantity of deep reflexion seismic data. Neave et al. (1978) arranged to have much of this data reprocessed to enhance the early arrivals, and were in this way able to obtain velocity breaks indicative of permafrost. This technique was



Seismic section, north from Atkinson Point into Beaufort Sea.

Fig. 4.30

The irregular top surface of a permafrost zone as mapped by Hunter and Hobson (1974) using the seismic refraction technique (Fig.4.29).

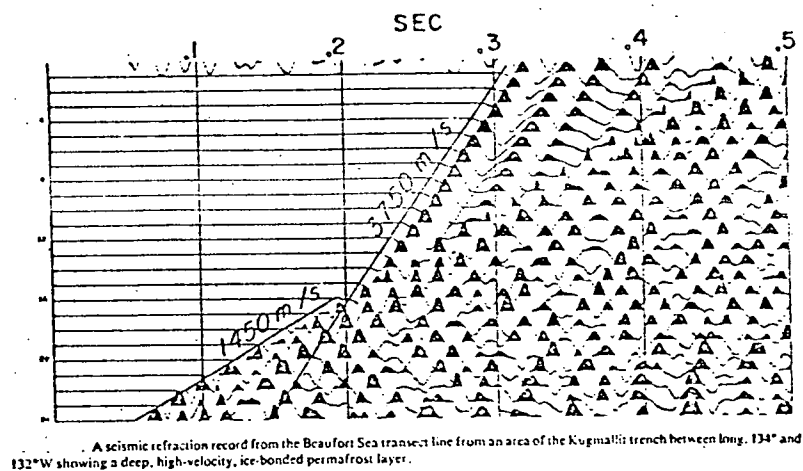
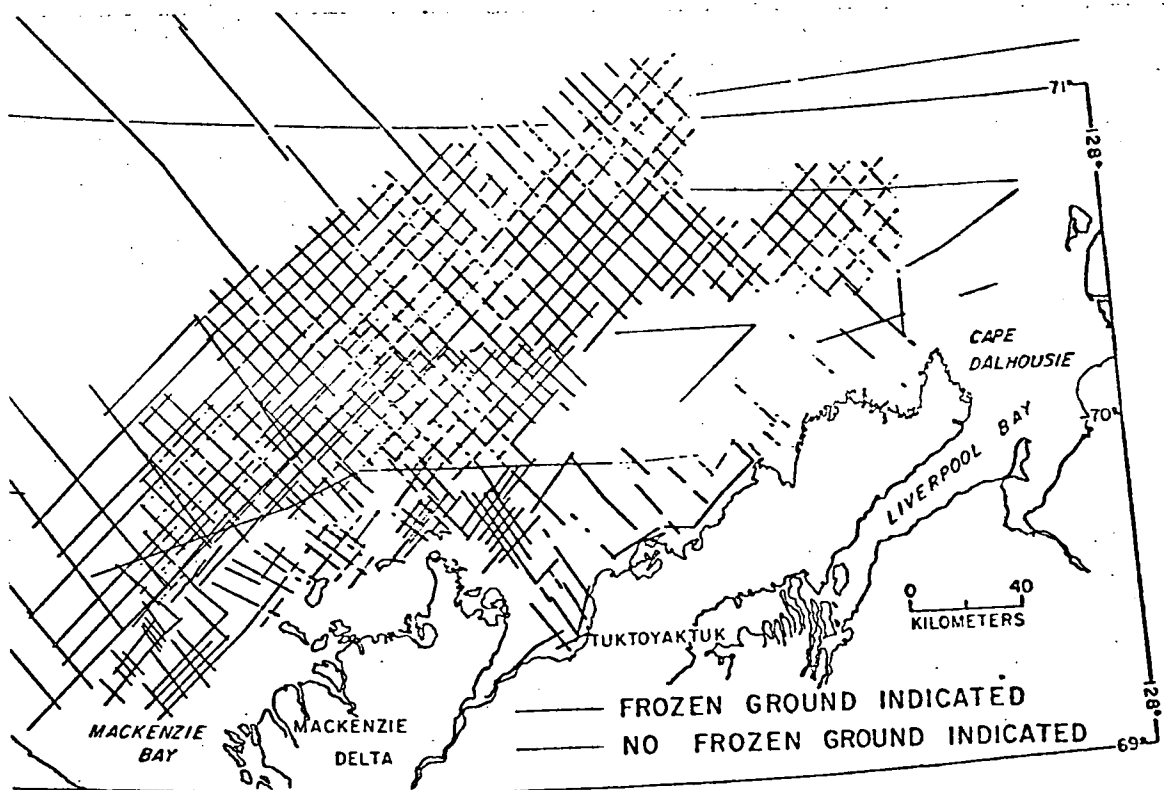


Fig. 24.31

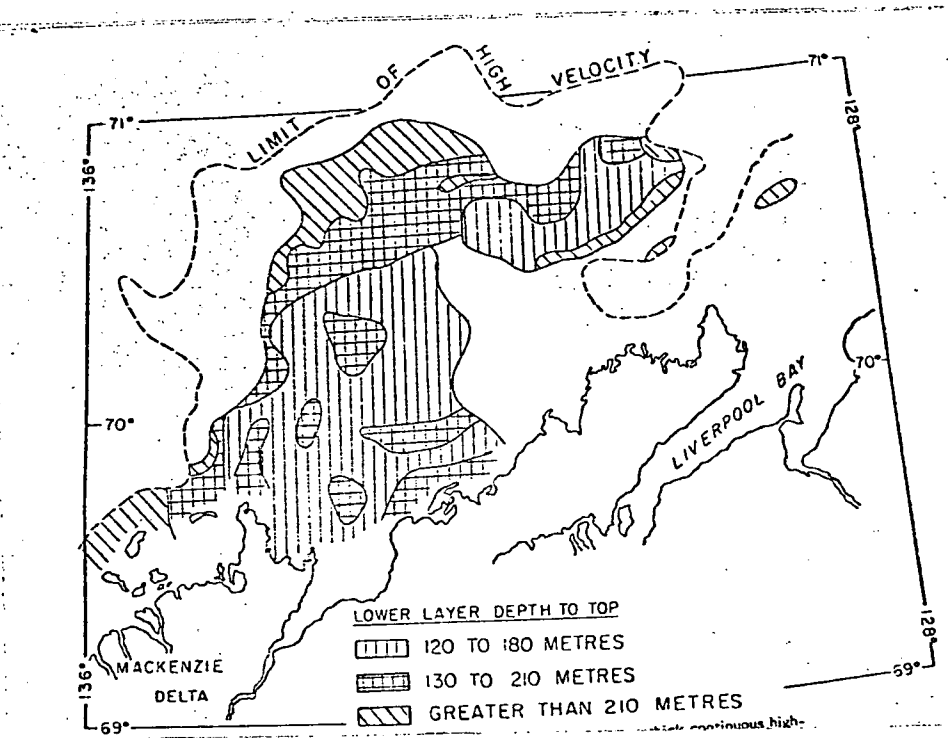
Clear headwave arrivals obtained by MacAulay and Hunter (1982) by reprocessing oil-industry 24-channel high-resolution seismic data.

further refined by MacAulay and Hunter (1982) who reprocessed multichannel high resolution data in a similar way (fig. 4.31). Neave et al. mapped the permafrost as shown in fig. 4.32, and their result is seen more clearly in its regional setting in fig. 4.33. It has also been found possible to observe this upper surface by single-channel vertical-incidence profiling with acoustic sources of between 150 and 1500 Hz. An example of such a profile is shown in fig. 4.34, where the acoustic source used was a small airgun of 16ml volume, and the receiver was the leading channel of the 12 channel array described in section 2.5.

Neither this reflexion method nor the refraction method of Neave et al. (1978) is useful for finding the basal surface of the permafrost zone. A way of doing this seems to be the electromagnetic sounding method. Some details of this method are given in an undated anonymous application note of Geo-physi-con (fig. 4.35a). Some results obtained on a transect from land to sea across an unspecified region of the Beaufort Sea Coast are given in fig. 4.35b which shows resistivity values plotted against the square root of the period of the signal transmitted, and an interpretation in terms of the top and bottom of a permafrost layer is given in fig. 4.35c. No details are given as to the logistical problems associated with deploying the 500 metre loop in the sea nor of the



a



b

Fig. 4.32

a. Track plot of reprocessed deep-seismic reflection data (Neave et al, 1978).

b. The interpretation for depth to the top of the frozen layer as obtained from the data in (a).

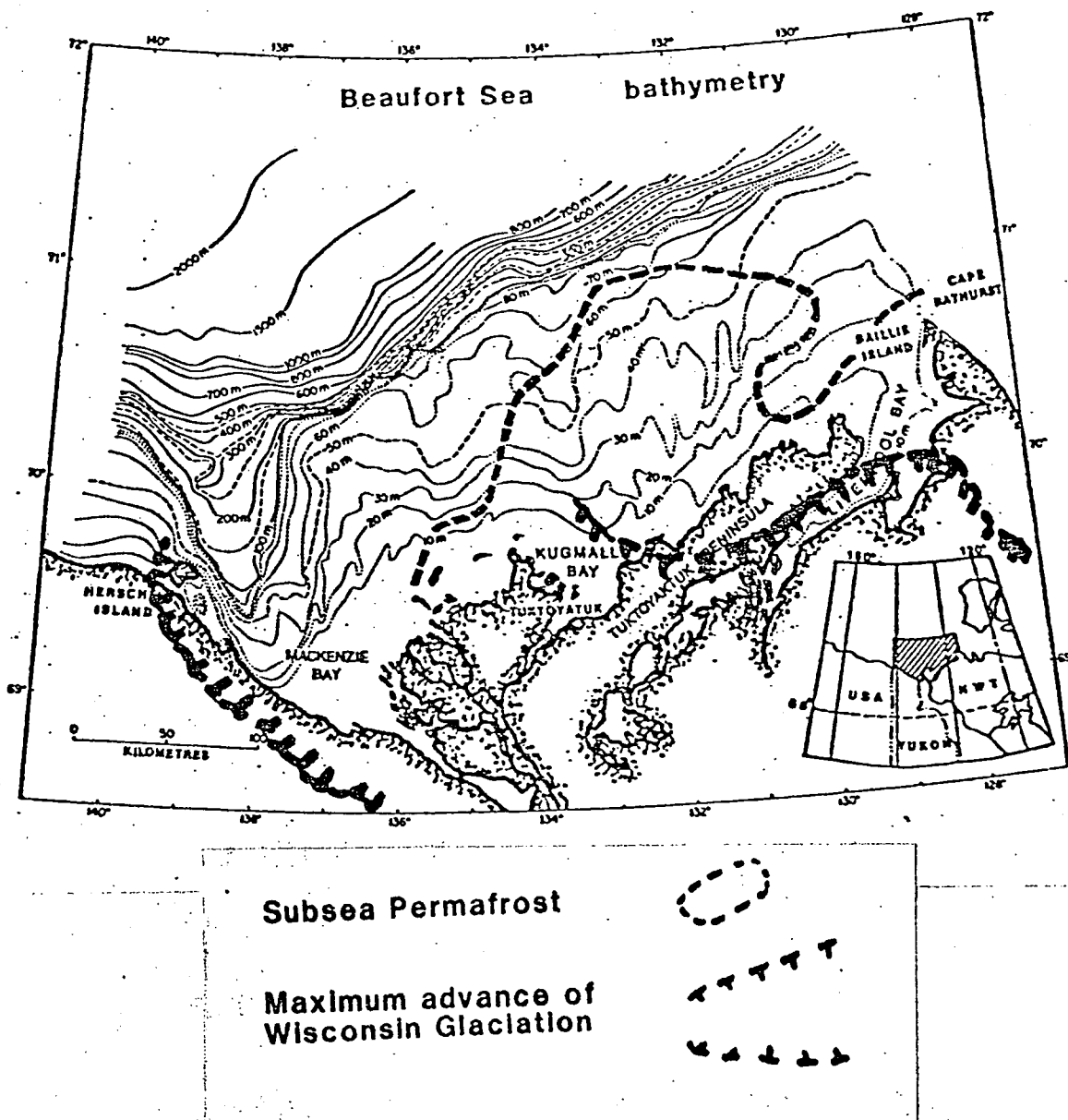


Fig. 4.33

A compilation by McKay (1983) of the distribution of subsea permafrost (Neave et al, 1978) together with bathymetry (Pelletier, 1975) and the maximum advance of the Laurentide Ice Sheet during the Wisconsin glaciation (MacKay et al, 1972).

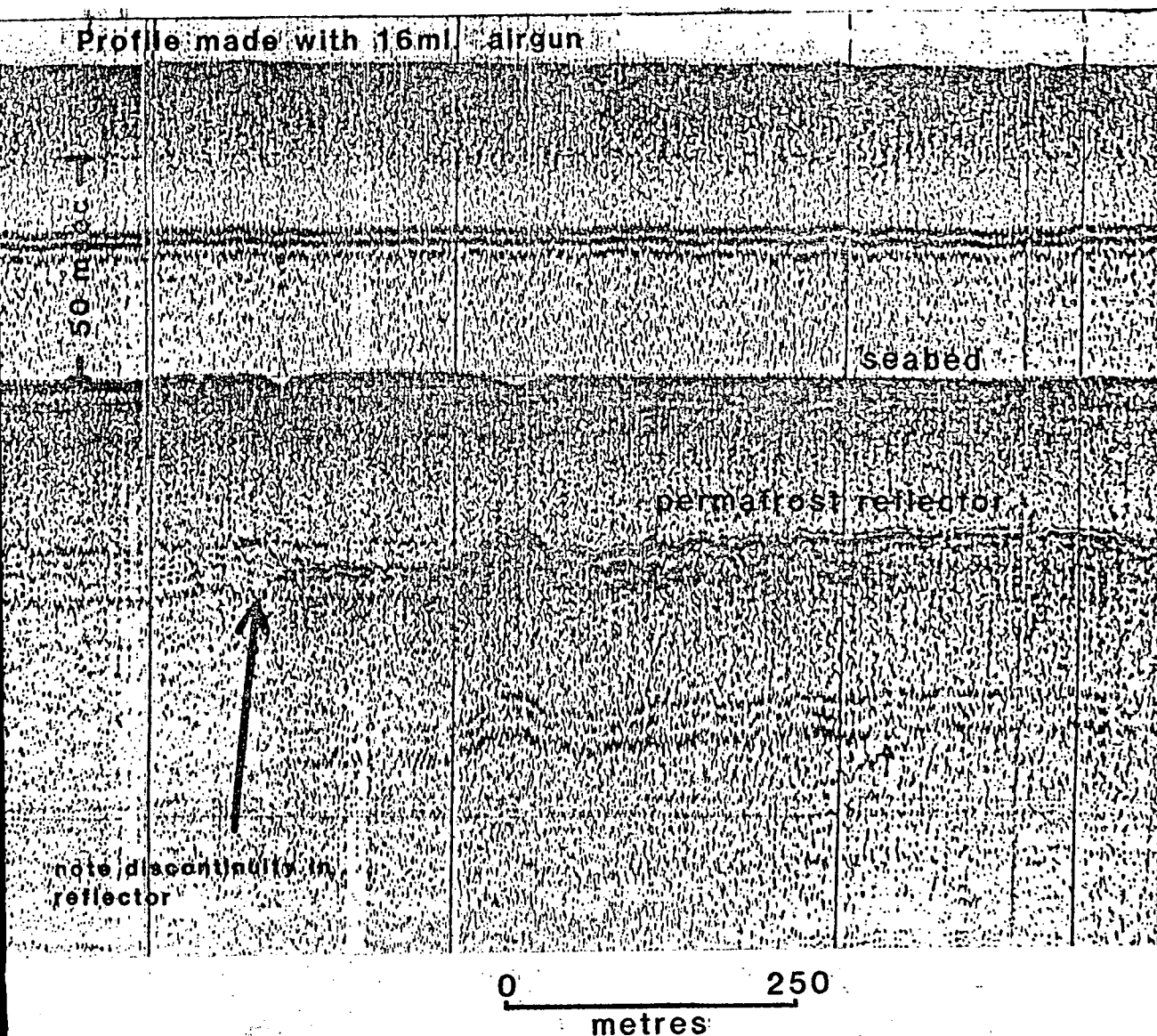
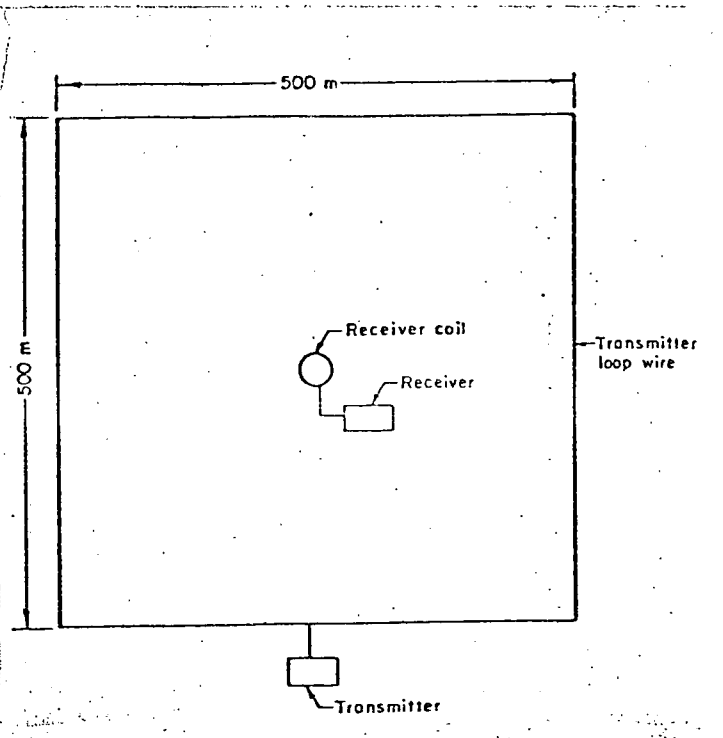
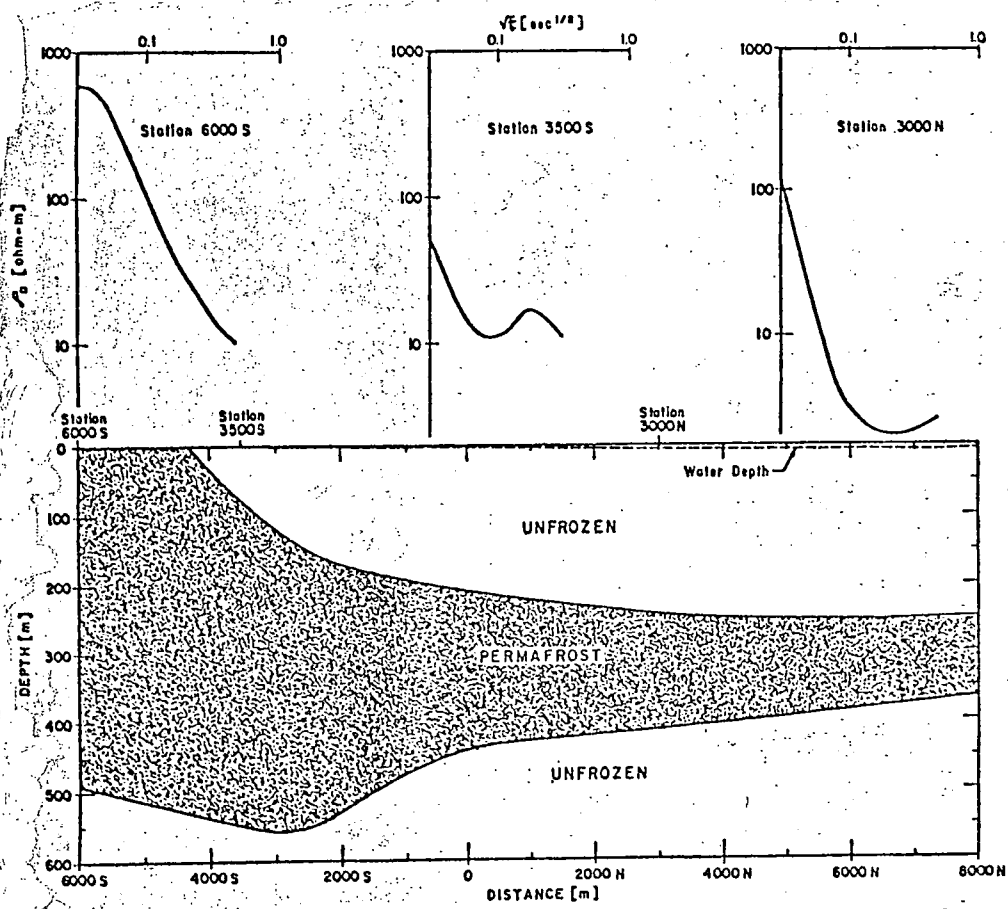


Fig. 4.34

The variable-density raster display of a single channel from the 12-channel array described in section 2.5. The profile was made in water about 50m deep and shows a strong reflector about 30m below the seabed (assuming velocities of about 1500m/sec, which may not be correct for the sediment layer, but give a general idea of the scale of the figure).



a



b

c

Permafrost profiles, on a line perpendicular to the coast of the Beaufort Sea, derived from transient electromagnetic soundings. Typical transient apparent resistivity curves are also shown.

Fig. 4.35

(a) equipment layout; (b) resistivity curves and (c) interpretation for an electromagnetic method of permafrost thickness measurement.

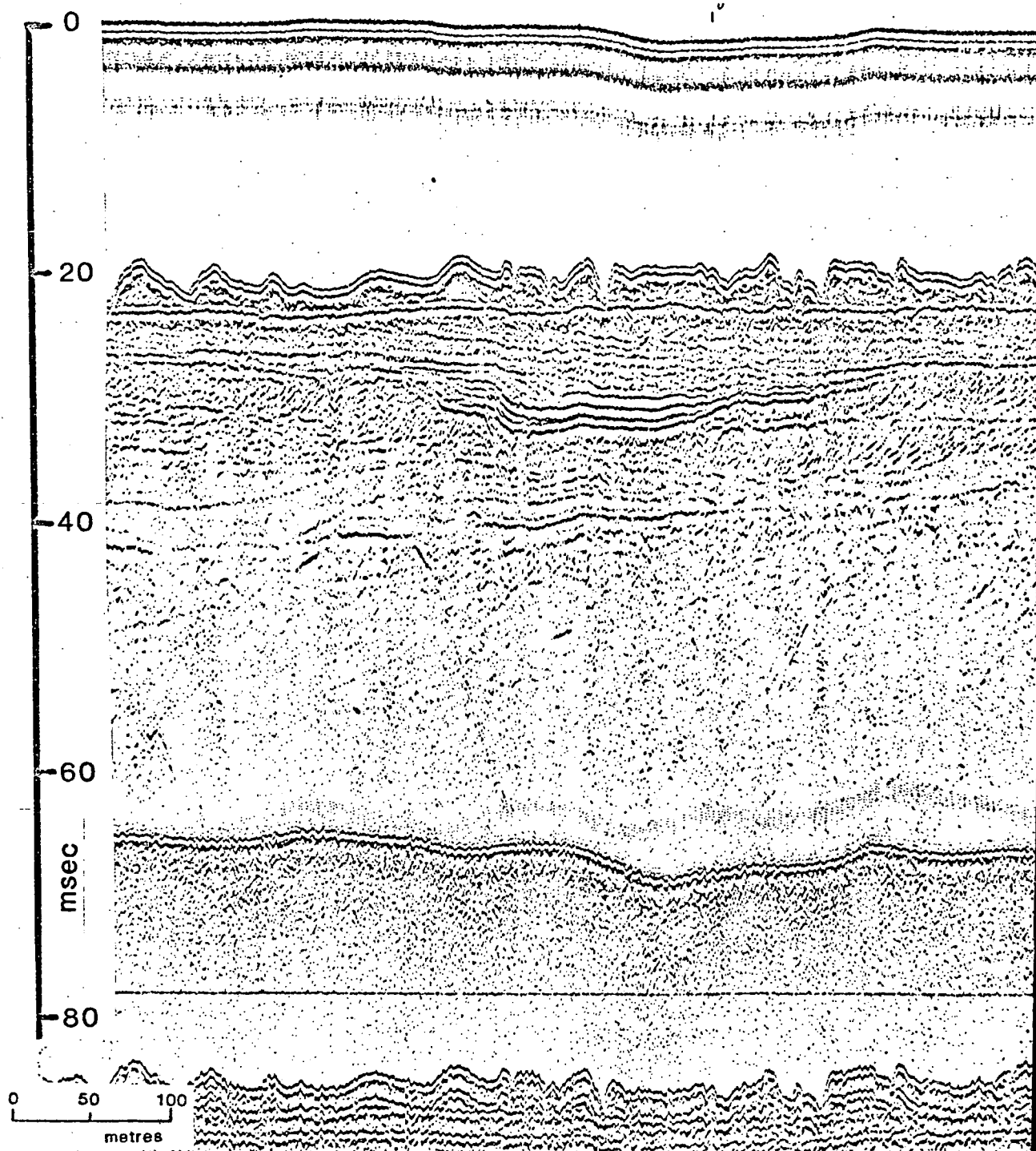
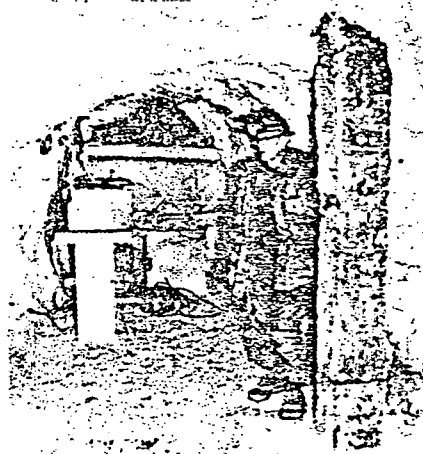
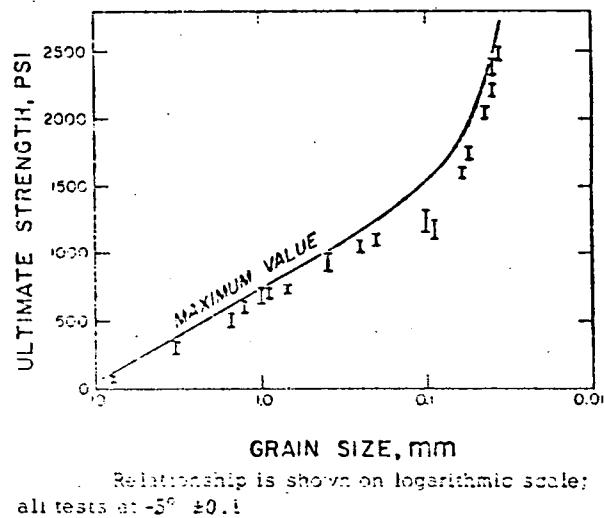
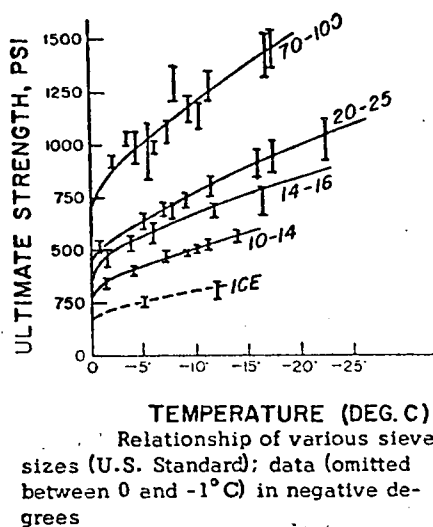


Fig. 4.36

Complex features as seen on a deep-tow seismic profile. There is no way of finding out from a record of this kind whether or not permafrost is present.

exciting currents required. Scott and Hunter (1977) consider that the electromagnetic sounding method is "one of the most promising" for the determination of permafrost thicknesses.

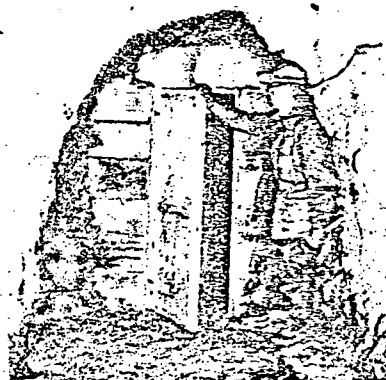
A problem which remains is the detection of ice-bonded or partially ice-bonded zones within the upper sediment layer. Small near-surface features which appear on vertical incidence profiles made at frequencies of 1-2KHz (fig. 4.36) are beyond the resolution of the techniques of Neave et al. (1978) or MacAulay and Hunter (1982), but it is important to identify possible permafrost in such features because they could lie immediately under the foundations of structures or pipelines. That permafrost can occur at this level is clear from the obvious bathymetric curiosities of the subsea pingos (figs. 4.26 and 4.27). Less clear indications on subbottom profiles may go unnoticed or be confused with changing lithologies (particularly in this deltaic environment). For this reason it is of value to have for measuring seismic velocities, a method capable of higher spatial resolution than those described above. This was the principal reason for the construction of the 12-channel array described in section 2.5 and for the treatment of results obtained from it, to be discussed in chapter 9.



a.



b.



c.



d.

Application of artificial frozen soil mixtures: a. Column supporting a loose roof slab; b. tunnel section protected by a combination of poured permacrete and precast bricks; c. bulkhead made of large precast bricks; d. diesel fuel storage pit poured in place and lined with fine grain frozen material. Finish: Clay slurry applied with paint brush

Fig. 4.37

Variation of bearing strength of "permacrete" with temperature and grain size of the soil used (Swinzow, 1963). The lower part of the figure illustrates the use of permacrete as a column, arch, bulkhead and tank.

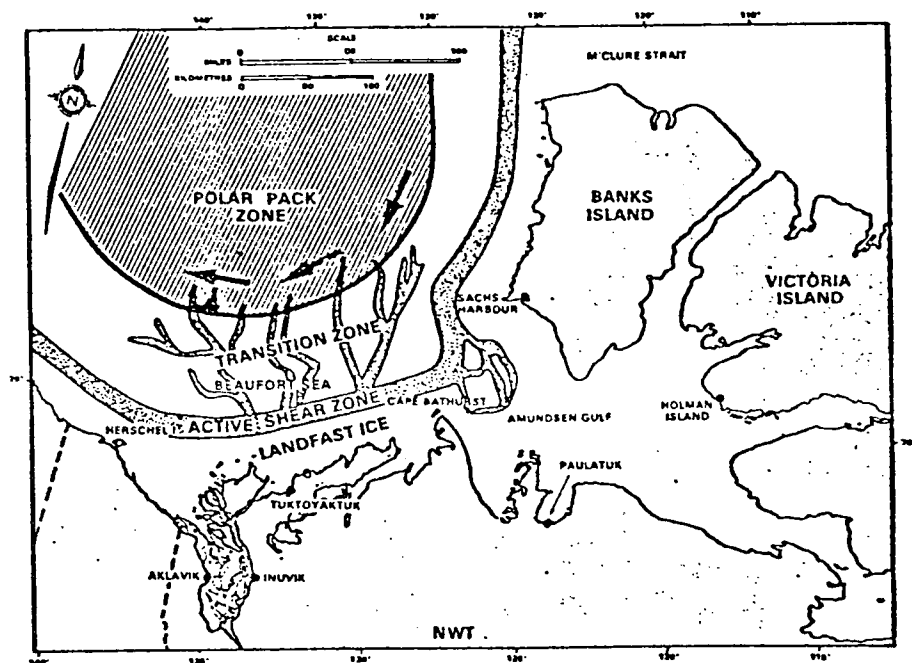
4.5 ENGINEERING IN THE BEAUFORT SEA.

Frozen soil provides an excellent foundation provided it remains frozen. It has even been used as a structural material (Swinzow, 1963) under the name of "permacrete". Swinzow initiated a study of the material after noting its usefulness as a "plaster" for repairing fractured tunnel walls near Thule in Greenland. With standard concrete testing equipment, he found the relationships between temperature, grain size and direct flexural strength as shown in fig. 4.37. He considers that the material could be used as if it were concrete, and has the benefits that it required no curing time after it was solidified, and it could, if necessary be rewarmed and reused. The behaviour of frozen soil as a foundation material is a more usual concern. When it remains frozen, it provides an excellent foundation, but if it thaws, it will tend to subside differentially, the amount of subsidence depending on the localised ice-content and thermal input. Building methods on land have been evolved to preserve the frozen integrity of permafrost foundations. This is done by raising buildings on piles a metre or so above the ground to allow cold air to circulate under them (fig. 4.38). Water, sewage and other services are also carried above the ground, in heated conduits. A similar concept was tried by Rowley et al. (1973), who built an experimental pipeline loop. This pipe, 600 metres long and 1.2 metres



Fig. 4.38

Photograph of arctic construction methods. This prefabricated sectional building is the shore base for operations by the Dome Petroleum Company in the Beaufort Sea. Note the wide air space beneath the building to avoid thawing the foundations.



The Beaufort Sea area is characterized by three main types of ice. In the nearshore area the ice becomes landfast and is therefore relatively stable. Far offshore is located the permanent polar pack which is constantly moving in a generally clockwise direction. In between is the transition zone, where moving ice constantly grinds against the stationary landfast ice. During the summer months, the landfast ice melts away and the transition zone clears, to permit drilling in this area.

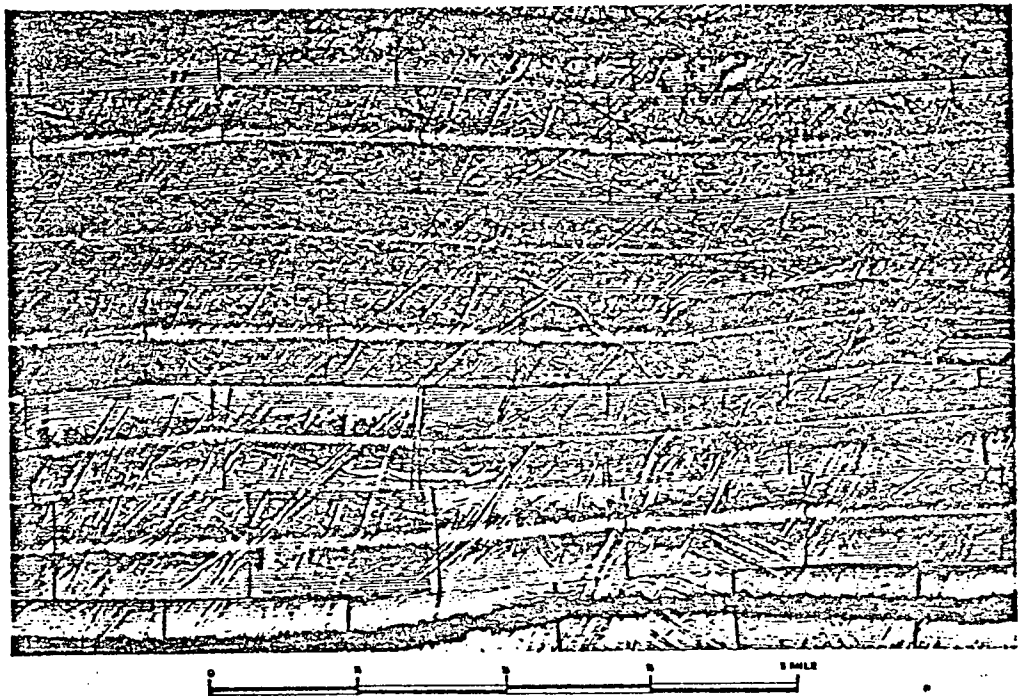


Fig. 8—Typical sonar mosaic.

Fig. 4.39

- a. Distribution of winter ice in the Beaufort Sea. (Anon., 1982).
- b. A mosaic of sidescan sonographs made with a 100KHz Klein device. (Hnatiuk and Wright, 1983).

in diameter, was supported about 1 metre above various frozen soil types. It was instrumented with temperature sensing devices, and elevations along its length were surveyed before and after a 14 month test period when warm oil was circulated through the pipe. The investigators found no significant movement of the pile supports nor any evidence of permafrost degradation other than an increase of about 0.3m in the depth of the active layer within a radius of 1m around each pile. By contrast, a section of the same pipe which had been laid within a gravel berm, although thermally insulated to cut down on heat loss, showed settlement of up to 1.5m between the start and end of the experimental period. These findings are useful in planning large pipeline projects which may be needed in Arctic lands, but only serve to emphasize the difficulty of laying pipelines on the floor of a sea where ice-bonded sediments may occur. The supporting of a pipeline above the seafloor is quite impractical because of the hazards from pressure ridges of ice which build up particularly in the active shear zone between the polar pack and the winter land-fast ice (fig. 4.39a). These ice-ridges give rise to scour marks over most of the Beaufort Shelf to depths of up to 80 metres. It is possible that those in the deeper water may not be recent, but marks with great detail of "en-echelon" drag features along the edges seem to indicate active scouring to depths of at least 50 metres. A mosaic of side-scan sonographs (fig. 4.39b)

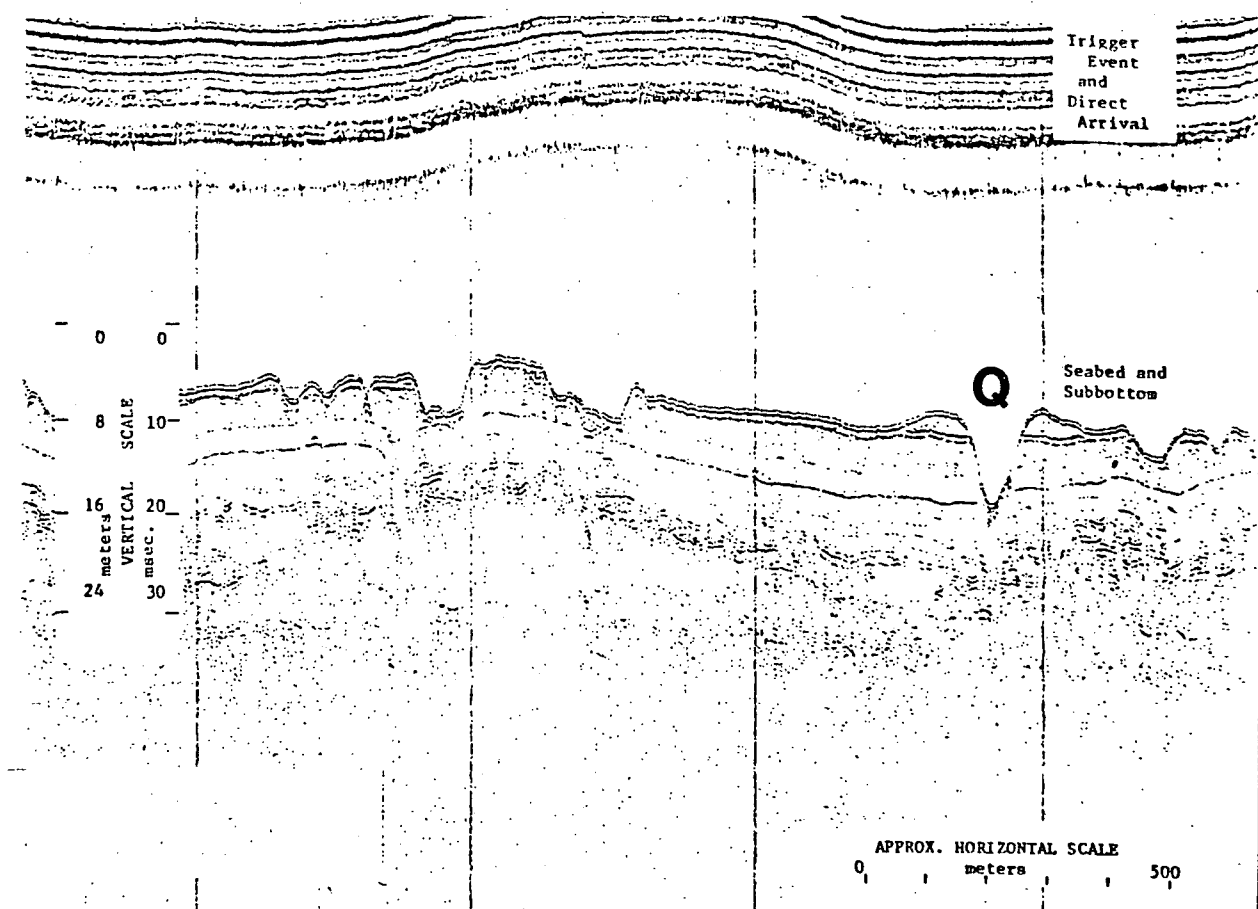


Fig. 4.40

Ice-berg scour marks as seen in section by subbottom profiler .
Karlsefni area of the Labrador Shelf (McKay, 1976):
Note the levées on both sides of the scour at "Q"

gives some idea of the density and size of these scour marks (Hnatiuk and Wright, 1983). They are very similar to scour marks made by ice-bergs on the Labrador Shelf (fig. 4.40; McKay, 1976) where active scouring is in progress. Although no icebergs occur in the Beaufort Sea, and the scouring is due entirely to sea ice, scours at least as deep as 5 metres can occur. During investigations in 1980 into the feasibility of laying a power cable across Lake Erie, pressure-ridge scours up to 3 metres deep were observed in up to 20 metres of water. Large masses of ice carried by wind and water currents clearly put major constraints on the methods which can be used for works of ocean engineering in waters to the north, south, and east of Canada.

Sackinger (1976) points out that the offshore activities most likely to be influenced by permafrost are the drilling of wells, the siting of pile-founded and gravity structures, and the laying of pipelines. There is consequently a requirement both to evolve engineering techniques so that greater immunity to permafrost problems is provided and also to improve field survey methods for the detection of ice-bonded sediments so that, whenever possible, they can be avoided if a proposed structure would be likely to upset their thermal balance. The development of survey techniques making use of the characteristic acoustic properties discussed in section 4.2

BEAUFORT SEA AND MACKENZIE DELTA

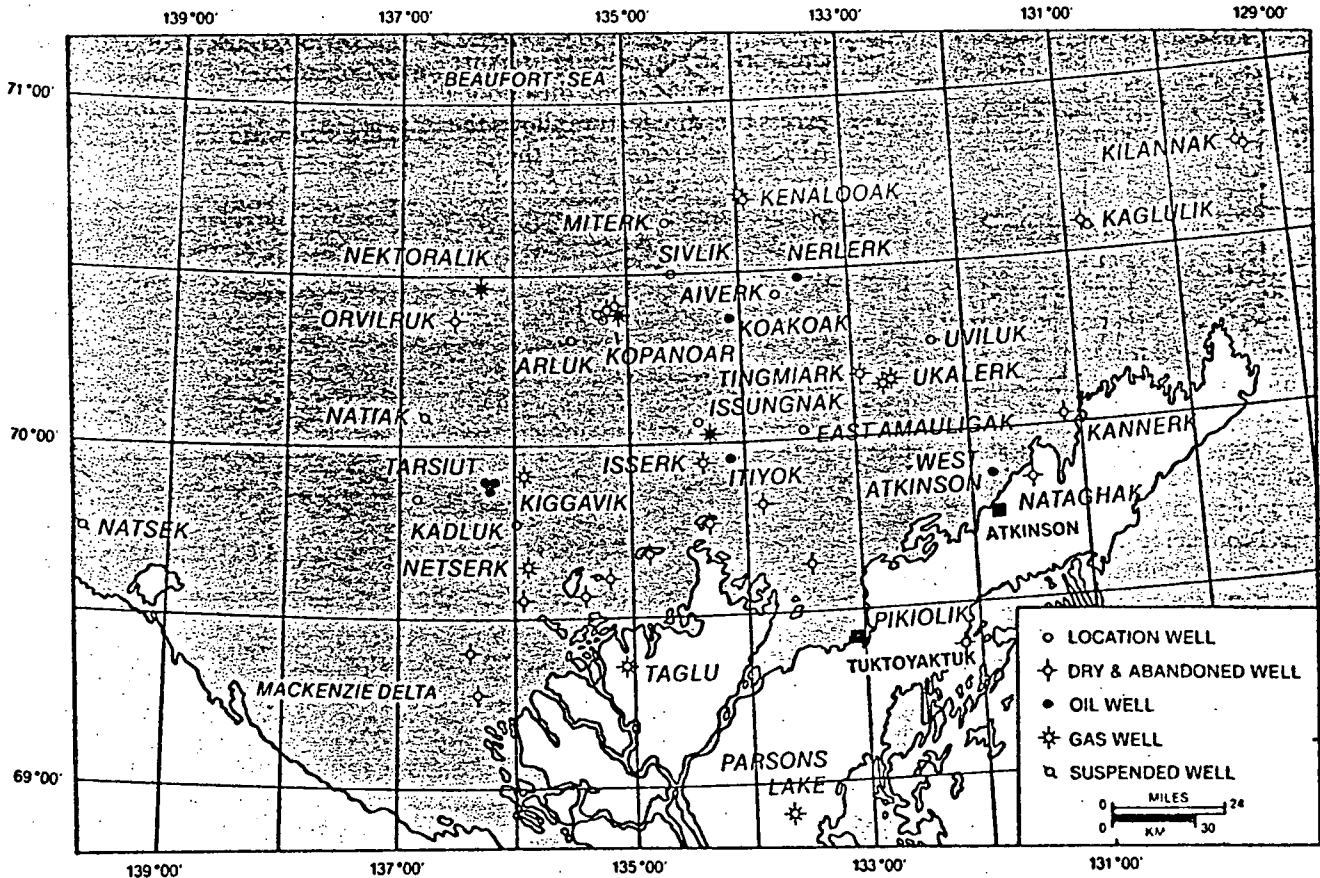
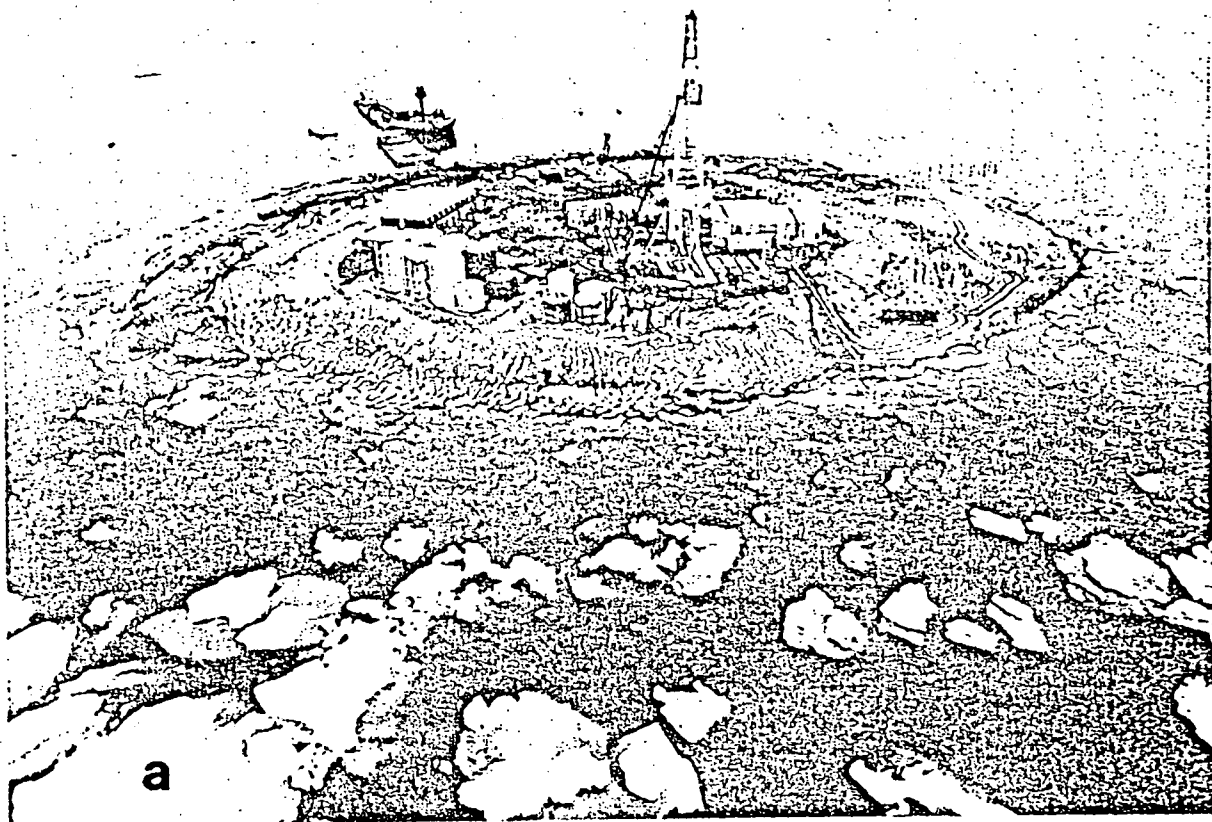


Fig. 4.41

Exploratory activity in the Beaufort Sea area. Most of the names for the drill sites are taken from Eskimo (Inuit) words. (Anon.,1983).

is treated in later sections.

One of the most interesting engineering methods to have been evolved in the Beaufort area to meet the problems both of sub-seabed permafrost and of sea ice is that of building artificial islands. These were first used in very shallow water of only a few metres, close to the coastline such as at Kannerk (fig. 4.41) but were soon scaled up to enable year-round drilling work at locations farther off shore such as Issugnak (19m. of water), shown in fig. 4.42a. Even at this depth, 4.9 million cu metres of sand was required to build the island. For deeper water still, such as at Uviluk (31 metres) (fig. 4.42b) or Tarsuit (22 metres), (figs. 4.43a,b, and c, 4.44a and b, and 4.45), it became necessary to use a caisson to avoid the requirement for prohibitively large quantities of sand. The Tarsuit caisson is similar to the design (fig.4.43) discussed by McCreath et al. (1982). A sand berm is first constructed on the seabed. The caisson sits on this and has its interior filled with sand and gravel. The drill rig can then be set up on the platform formed in this way. The height of the platform above high tides was seven and a half metres. Myers et al. (1983) have described their observations in monitoring Tarsuit Island during the open-water season of 1982, which, they point out, was the first open-water season when such observations could be made on a caisson-retained island.



Issungnak Island in the Canadian section of the Beaufort Sea in September. This is a "sacrificial beach" island; the outer rim breaks storm waves in summer and causes ice sheets to break up and form a protective rubble field around the island. Sandbags and filter cloth reduce erosion. The island is made of 4.9 million cubic meters of sand, dredged from the seabed and delivered to the site through a floating pipeline. It is 26 kilometers offshore in water 19 meters deep. From November through June, there is no open water around the island. In the winter, the sun does not appear on the horizon. (Photo courtesy of Esso Resources Canada, Ltd.)

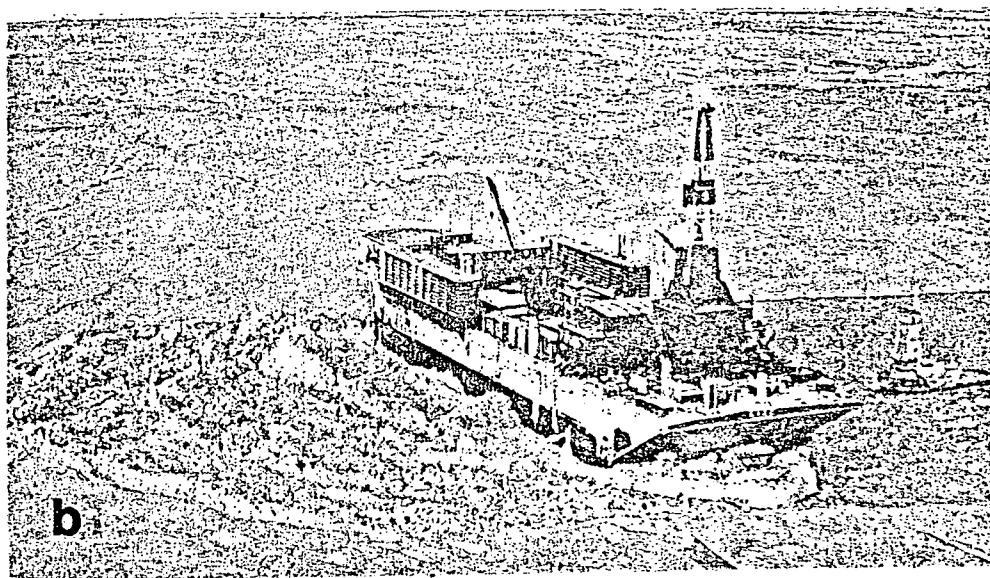
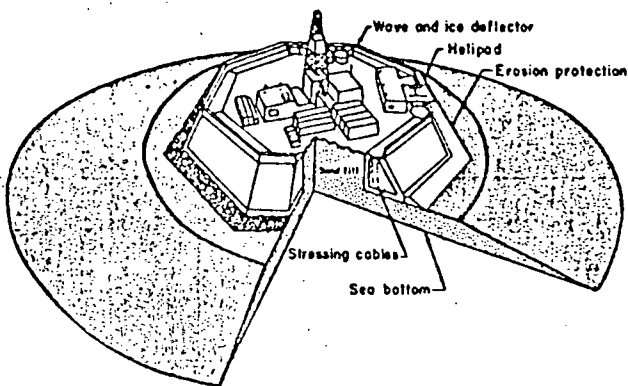


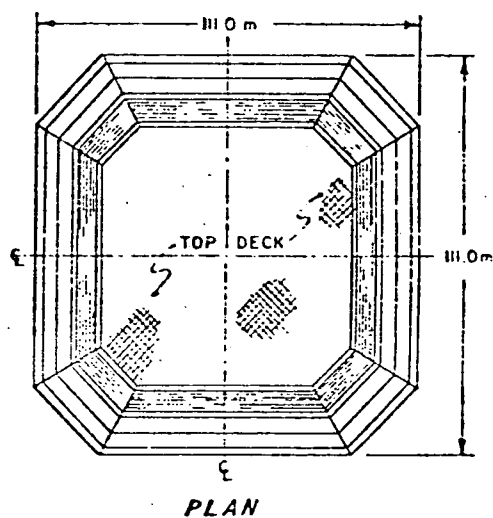
Fig. 4.42

a. The artificial island at Issungnak.

b. The converted hull of a very large crude-oil carrying tanker in use at Uvilik as a platform. The tanker is filled with sand. (photograph of Dome Petroleum).



a CAISSON RETAINED ISLAND



b

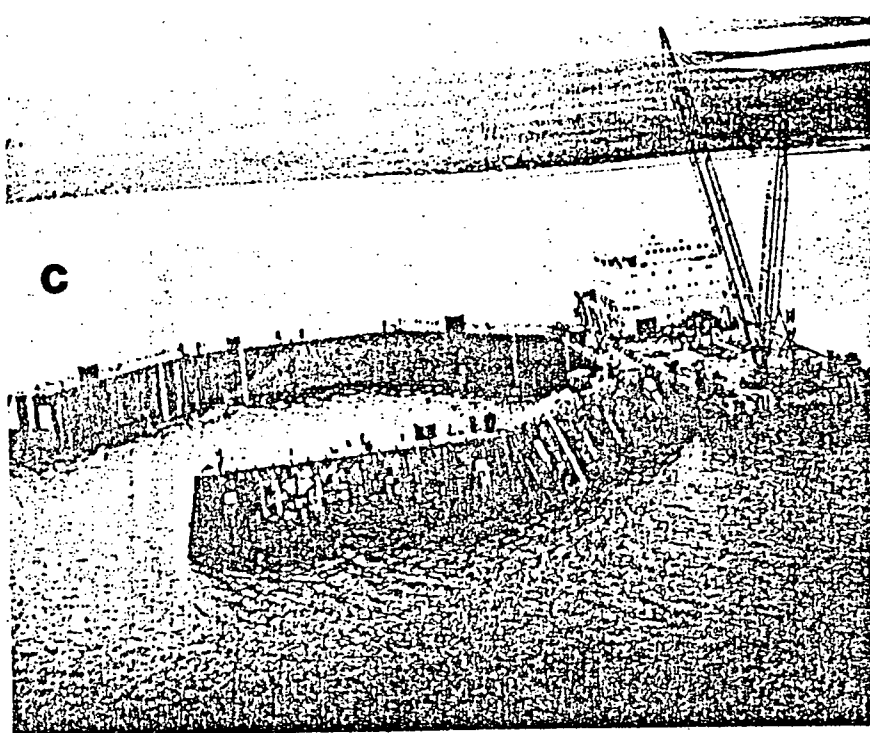


Fig. 4.43

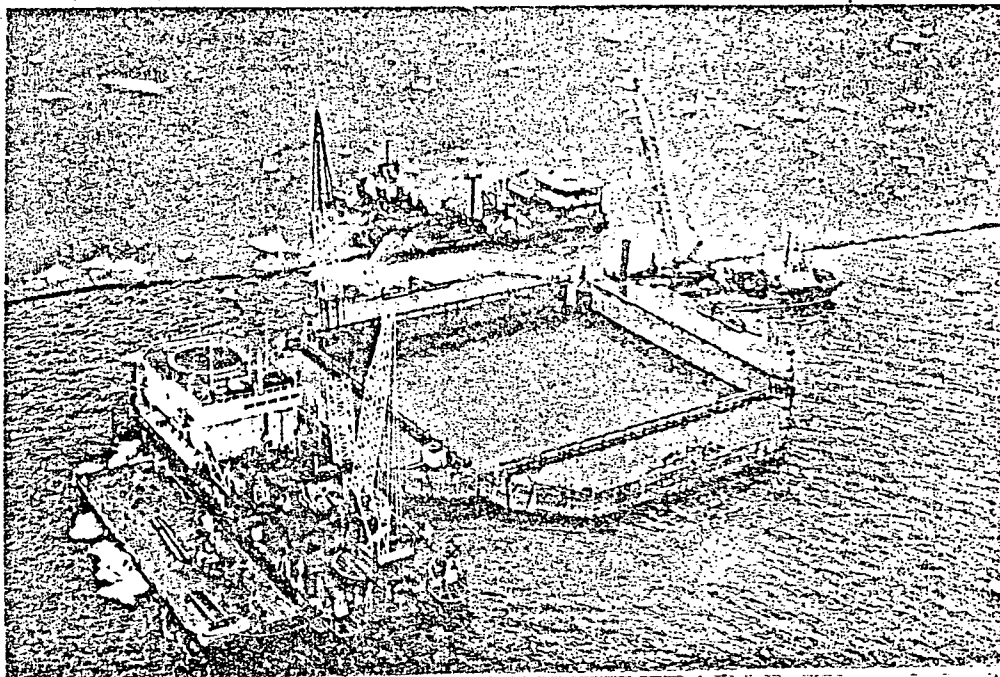
a. General view of a planned caisson-retained island.

b. Plan and elevation.

c. View of caissons afloat. This type of caisson has evolved from what has actually been built at Tarsuit (Figs. 4.45 and 4.46) and is designed to be re-usable. (Mancini et al, 1983).

With the four concrete caissons in position, Tarsuit takes form in the Beaufort Sea. After being filled with sand and topped off with a gravel surface, drilling equipment was put in place using the giant cranes shown around the circumference. Each caisson weighs 5,300 tons and measures 80 metres (262 feet) long. They were built at a cost of \$27 million dollars in Vancouver, B.C. and towed by barge to the Beaufort in July of this year. When empty the caissons float; they are sunk in position by flooding with water, then filled right to the top with sand to form the perimeter reinforcement for the island.

a



The four independent concrete caissons of Tarsuit, viewed from the air, form a hollow square which is filled with sand and gravel fill. Each hollow caisson weighs 5,300 tons and is connected at the corners by steel doors to contain the fill.

b

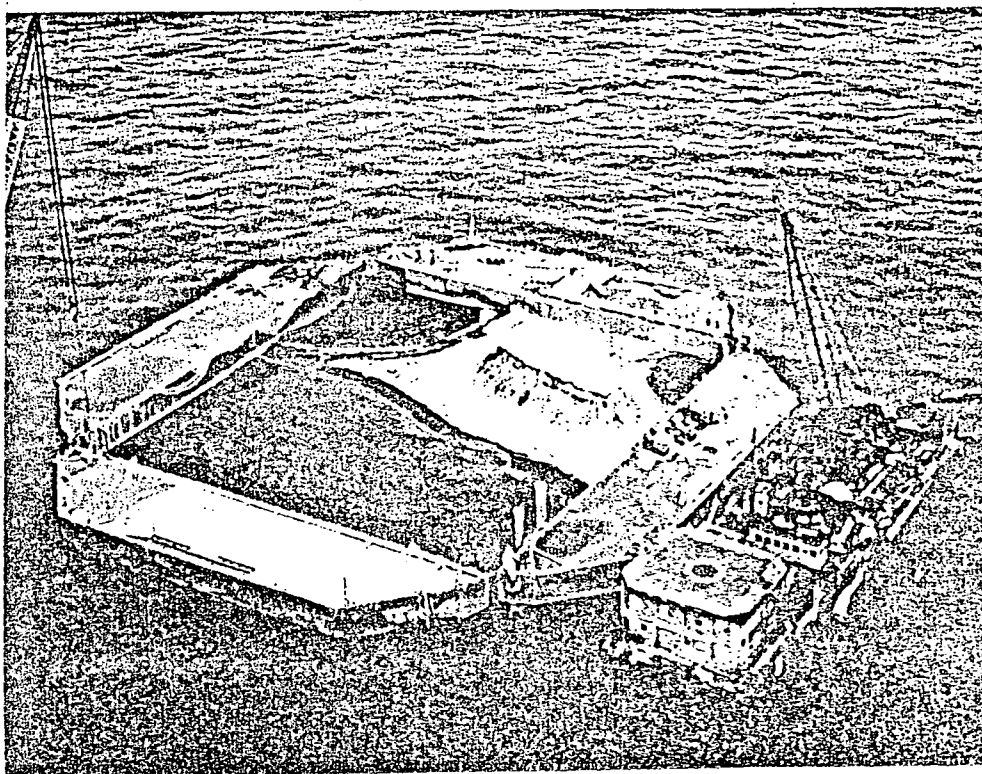


Fig. 4.44

a. Caissons being set in place at Tarsuit.

b. Sand being fed into the caisson-enclosed space. (Van de Wal, 1981).

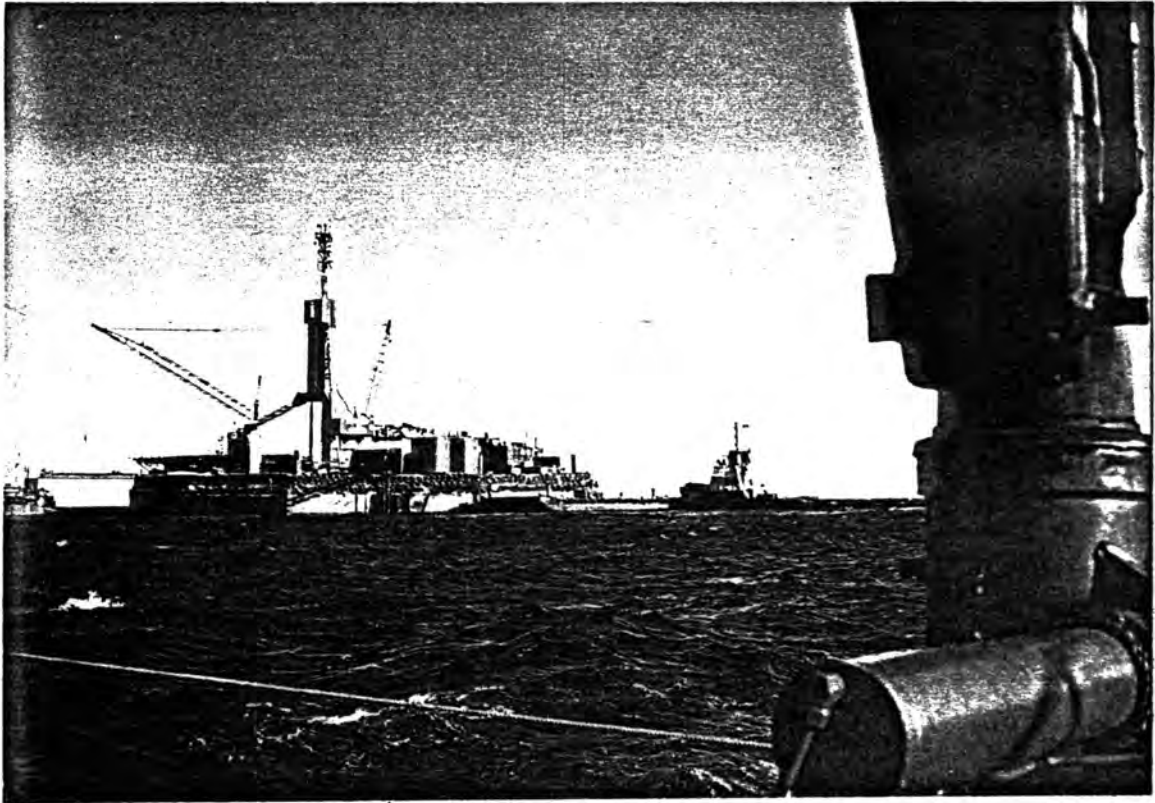
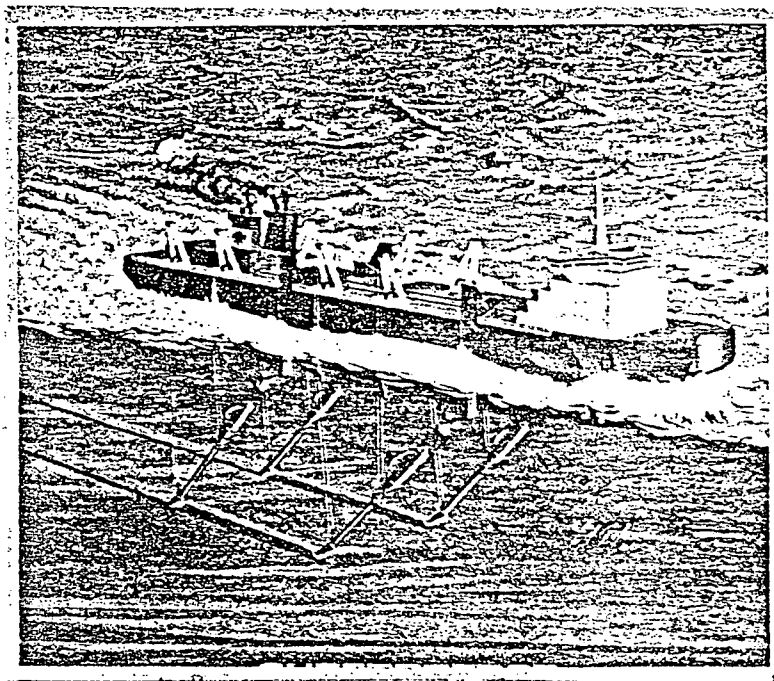
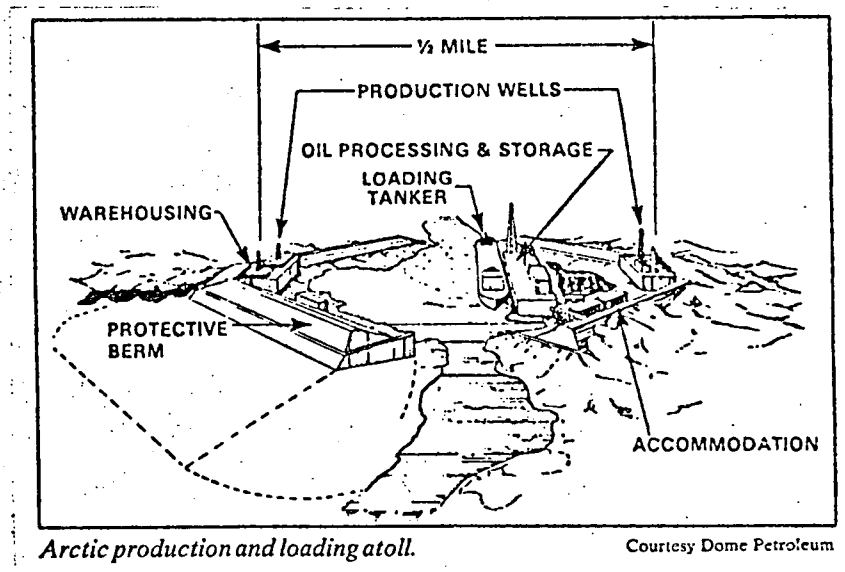


Fig. 4.45

A view of the completed island at Tarsuit.



a



b

Fig. 4.46

(a) Artist's view of a large suction dredge used to gather the sand required for building the islands.

(b) If production and shipment of oil or gas is to be accomplished without the extensive use of seabed pipelines, then atoll-like islands will have to be made to allow berthing and loading of tankers. The sketch shown above is of a production atoll which would require almost 100 times as much sand as the exploratory drilling island at Tarsuit (Fig. 4.43). (Dome Petroleum).

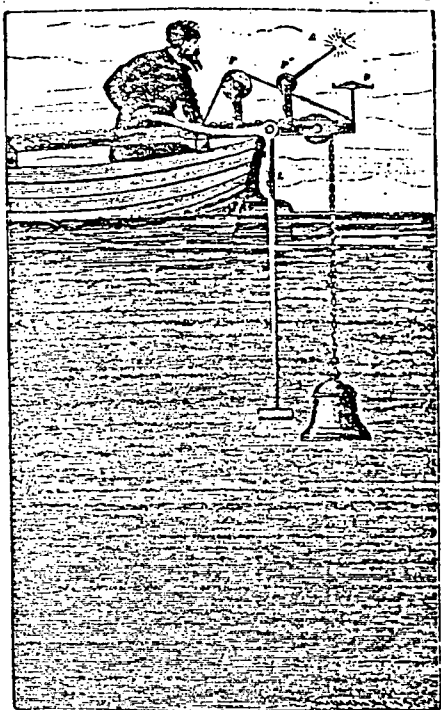
They found that one of the main problems of wave action on the island was the erosion of the sand berm on which the caissons rest. This they intended to control by placing a rock apron at the toe of the caissons and a layer of gravel 3 metres thick over the upper part of the berm. Difficulties with exactly this manner of erosion by wave action in the vicinity of the toe of such structures had been predicted by Kobayashi et al. (1981).

The technology of dredging on the scale required for island building is being advanced by the Canadian Marine Drilling Company who have designed a suction dredge (fig. 4.46a) to operate in winter ice conditions and to have a dredging and carrying capacity of 25,000 cu. metres. The suction dredges currently in use operate only in summer and carry about half of that quantity of material. The need to find suitable material for the building of islands has been one of the main reasons for conducting subbottom profiling and sidescan sonar surveys. These can help to define the extent of sand layers, but identification of the material is done by sampling. In addition to its use in detecting permafrost which has interfered with dredging (Sackinger, 1976), the making of detailed acoustic measurements including the determination where possible of the elastic properties of the seafloor material could aid in the identification of suitable fill and cut down on the need for sampling.

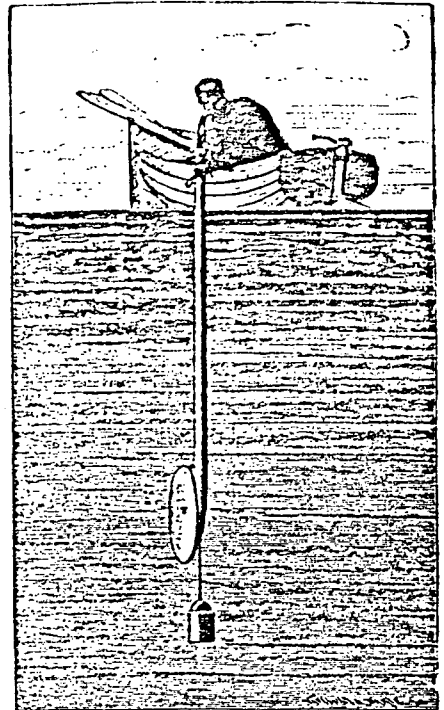
The island-building technique is one that appears to work quite well. Some temperature measurements have been made in holes drilled through existing islands. Brown and Barrie (1976) measured temperatures of 2°C to 4°C at a depth of 4 metres below the original seabed. At Pelly Island they found temperatures of +2°C to +4°C; at Unark they found -3°C. Temperature differences before and after building are not available, but the observations do seem to indicate that, at Unark, the permafrost was not rapidly degraded by the presence of the island. Difficulty with differential thaw subsidence on other islands has not been reported.

Sackinger (1976) pointed out the "critical" importance of recognising offshore permafrost as a major problem of which far greater understanding was required if offshore oil exploration is to proceed in safety. It is probably true to say that since Sackinger's publication, the problems have become widely recognised, but that much of the understanding is still lacking.

5. ACOUSTIC PROPAGATION IN THE OCEAN.



Bateau expéditeur du son.



Bateau receveur du son.

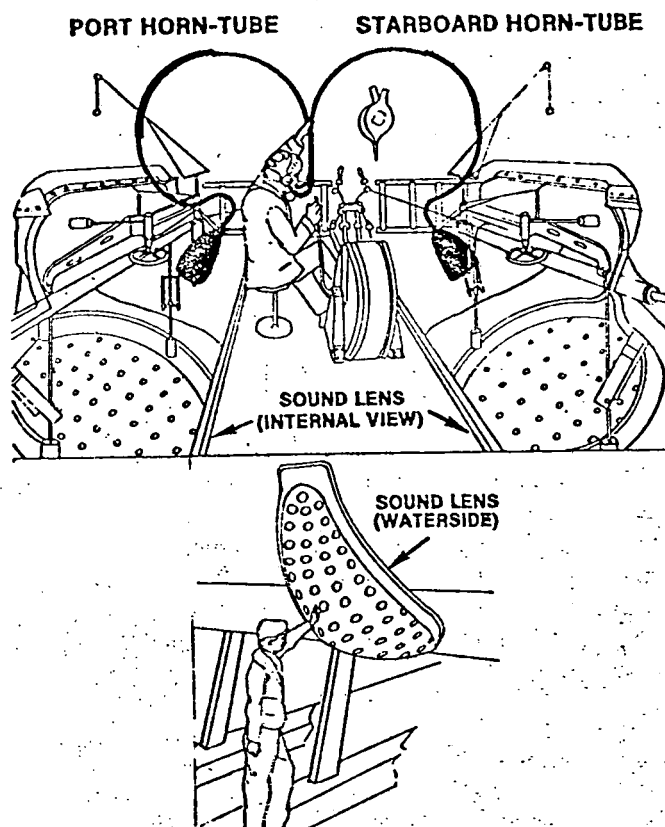
Fig. 5.1

Colladon's experiment to determine the speed of sound in water. In the transmitting boat, the bell was struck and a flash of light was produced simultaneously. The observer in the other boat set running a stop watch when he saw the flash of light and stopped it when he heard the sound. The determined value of 1435 m/sec. at 8°C is within 0.2% of the presently accepted figure. (Lasky, 1977).

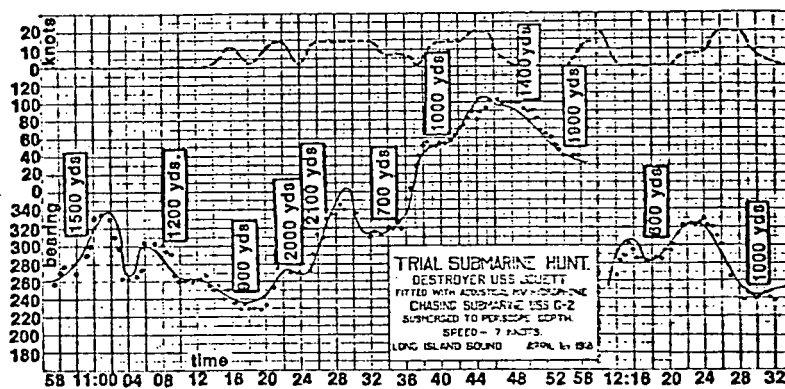
5.0 INTRODUCTION.

In addition to their possible use in predicting conditions of seabed sediments which are of importance to ocean engineers, as discussed in chapters 3 and 4, the physical properties of the sea floor are of interest to underwater acousticians who are concerned with the propagation of sound in the water column. Usually what is of importance is the distance at which sounds can be detected by passive listening devices or the range at which active sonar can locate certain objects, as well as the general problem of using the water as a medium for communication.

The first measurement of the velocity of sound in water was made by Colladon and Sturm (Lindsay, 1973) in Lake Geneva in 1826. The experimental procedure is illustrated in fig. 5.1 and comprises a bell as sound source and a human ear assisted by an underwater trumpet as the receiver. Colladon performed this experiment with a distance of 10 miles between the transmitting and receiving boat and found an acoustic velocity of 1435m/sec at 8°C. What was at least as significant as the determination of the velocity was the observation that a relatively small signal could be detected over a long distance in water with very crude receiving equipment. Little use was made of this observation until the



Shipboard binaural listening system (circa 1915).

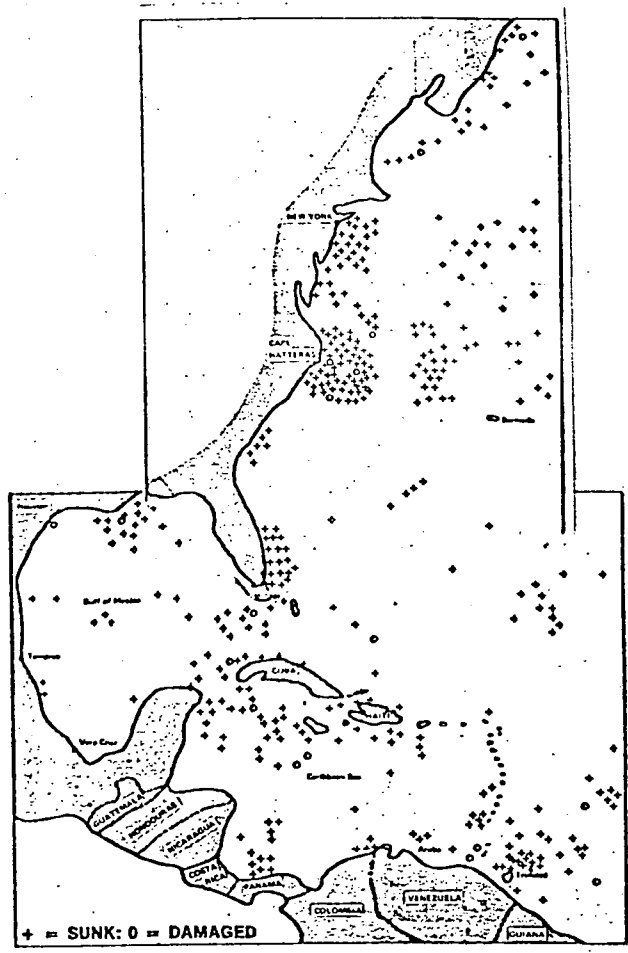


Acoustic-optical ranging trials on periscope-depth submarine, 1915. Dashed line: speed of U.S.S. JOVETT. Solid line: optical bearing on periscope. Filled circle: hydrophone bearing. Open rectangle: acoustical range from acoustic sensor cross fix.

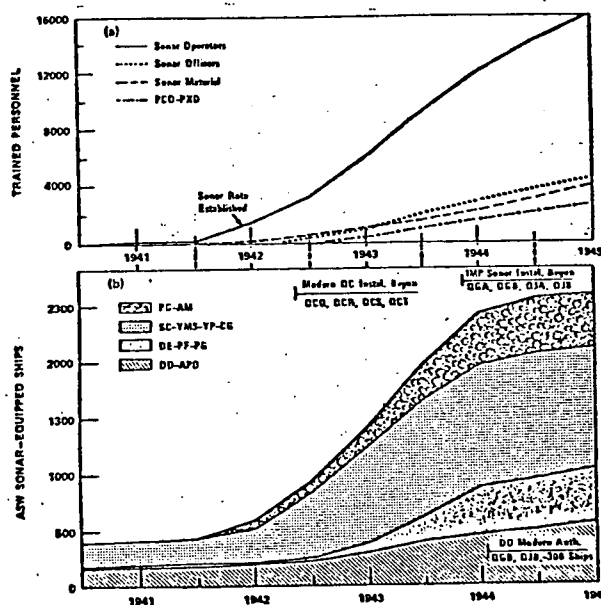
Fig. 5.2

- Directional shipboard listening device.
- Comparison between line-of-sight bearing and acoustic bearing obtained on a trial submarine hunt. (Lasky, 1977).

establishment of the Submarine Signal Company in 1901 by Gray and Mundy (Lasky, 1977). They proposed to make use of the low attenuation of sound in water to improve on the airborne signals produced by foghorns to give warning of hazards to navigation. Their method was to use an underwater bell which was driven pneumatically by compressed air stored in a buoy. Waterproofed versions of Edison's carbon microphone would act as receiving transducers. By 1912, many trans-Atlantic ships were fitted with this apparatus and submarine bells were maintained in dangerous locations. After that time, interest in the technique declined because of the increasing use of radio direction finding equipment, but another development was instigated by a need for vessels underway to be able to locate unknown objects, particularly icebergs. This was first accomplished by Fessenden, as reported by Quinan (1914), who detected an iceberg by underwater echo-ranging at a distance of two miles, by using a moving coil device as transmitter and receiver. At this time, anti-submarine warfare became important (Hayes, 1920) and naval ships were fitted with directional listening devices (fig. 5.2a), which by 1918 had become sufficiently refined that during trials by the destroyer U.S.S. Jouett, which tracked a submarine at periscope depth, good agreement between acoustic and visual bearings could be obtained (fig. 5.2b). During the second world war, this work became even more important, as



a



b

Fig. 5.3

a. Shipping sunk or damaged in the western North Atlantic by submarines during the first six months of 1942.

b. Rise in numbers of sonar operators and sonar-equipped ships between 1940 and 1945. (Lasky, 1977).

can be seen from the map of ships damaged and sunk by submarines in the western North Atlantic (fig. 5.3a) and from the rapid increase in the number of ships equipped with sonar equipment and trained personnel (fig. 5.3b). By this time, piezoelectric transducer development by Boyle (1928) and others and the increasing refinement of equipment by the application of electronics had brought sonar to the point where limitations imposed by the natural environment became important, and became the subject of investigation.

5.1 ACOUSTIC PROPAGATION IN SEAWATER.

Assumptions which are usually made in deriving the wave equation used in underwater acoustics are:

- (i) The medium is a fluid and does not support shear stress.
- (ii) The only fluid motion which is occurring is that due to the acoustic disturbance.
- (iii) No heat transfer takes place (adiabatic conditions).
- (iv) The acoustic disturbances are of small amplitude.
- (v) Gravitational forces can be ignored.

With these assumptions, the wave equation can be derived from the principles of conservation of mass and momentum and the equation of state relating density to

pressure changes.

In a volume element of fluid of density ρ , where particle velocity is given by $\underline{v}$, mass conservation is expressed as:

$$\dot{\rho} + \underline{v} \cdot (\rho \underline{v}) = 0 \quad \dots\dots\dots (5.1)$$

Momentum conservation requires:

$$\rho \dot{\underline{v}} + (\underline{v} \cdot \underline{v}) (\rho \underline{v}) = -\underline{v} p + \rho \underline{g} \quad \dots\dots\dots (5.2)$$

where p is the ambient pressure and $\underline{g}$ is the gravitational acceleration.

Bulk modulus can be defined for an adiabatic process as:

$$\kappa = (1/\rho) \{ (dp/dt) / (d\rho/dt) \} \quad \dots\dots\dots (5.3)$$

Allowing for spatial as well as time variation by writing:

$$\begin{aligned} d/dt &= (\partial x/\partial t) (\partial/\partial x) + (\partial y/\partial t) (\partial/\partial y) + (\partial z/\partial t) (\partial/\partial z) + \partial/\partial t \\ &= \underline{v} \cdot \underline{v} + \partial/\partial t \end{aligned}$$

We can obtain:

$$\dot{p} + \underline{v} \cdot \underline{v} p = (\kappa/\rho) (\dot{\rho} + \underline{v} \cdot \underline{v} \rho) \quad \dots\dots\dots (5.4)$$

For small perturbations in density ($\delta\rho$ from ambient ρ_0), pressure (δp from ambient p_0) and velocity ($\underline{ov}$ from rest), equations 5.1, 5.2 and 5.4 become respectively:

$$(\partial/\partial t) (\delta\rho) + \underline{v} \cdot (\rho_0 \underline{\delta v}) = 0 \quad \dots\dots\dots (5.5)$$

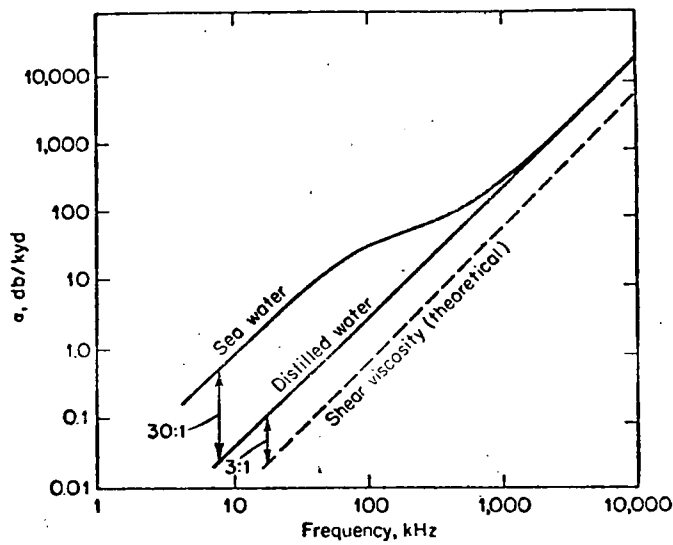
$$\rho_0 (\partial/\partial t) (\underline{\delta v}) = -\underline{v} (\delta p) + (\delta\rho) \underline{g} \quad \dots\dots\dots (5.6)$$

$$(\partial/\partial t) (\delta p) = (\kappa_0/\rho_0) (\partial/\partial t) (\delta\rho) \quad \dots\dots\dots (5.7)$$

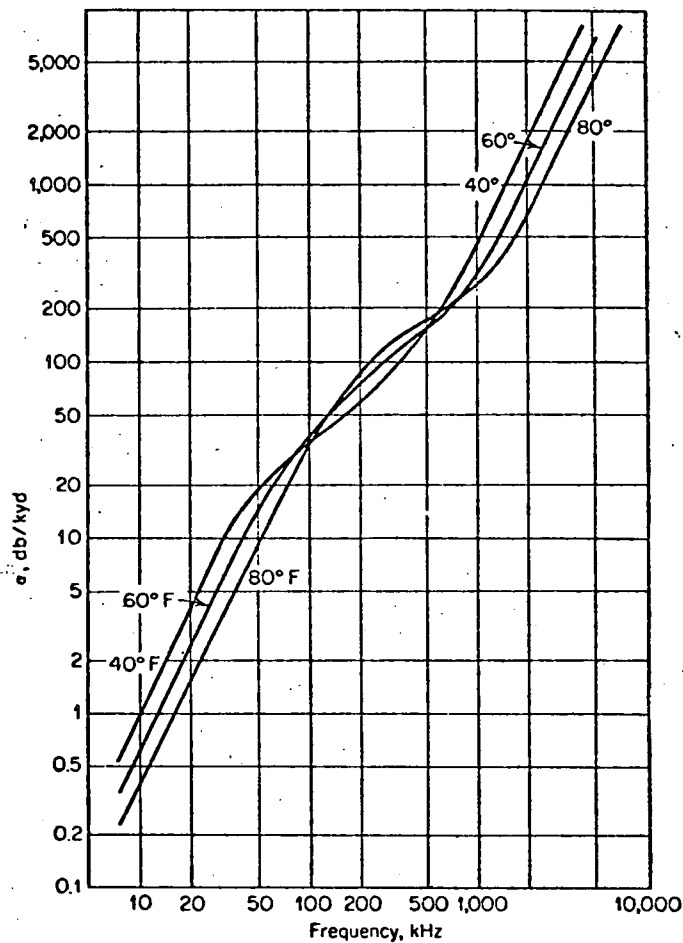
Substituting (5.7) in (5.5):

$$(\rho_0/\kappa_0) (\partial/\partial t) (\delta p) + \underline{v} \cdot (\rho_0 \underline{\delta v}) = 0 \quad \dots\dots\dots (5.8)$$

Taking the time derivative:



Absorption coefficients in sea-water and in distilled water and theoretical absorption due to shear viscosity alone.



Absorption coefficient in seawater of salinity 35 ppt as a function of frequency at three temperatures.

Fig. 5.4

Attenuation of sound in water. Observed values are higher than predicted from quasi-static measurements of viscosity because of resonance effects in dissolved ions. Figures from Urick (1967, p.88).

$$(\rho_0/K_0) (\partial^2/\partial t^2) \delta p + \rho_0 \nabla \cdot (\partial/\partial t) \delta \underline{v} = 0 \quad \dots (5.9)$$

Substituting from (5.6) for the second term of (5.9) on the assumption that $g(\delta\rho)$ is negligible:

$$\nabla^2(\delta p) - (1/c^2) (\partial^2/\partial t^2) \delta p = 0 \quad \dots (5.10)$$

where $c^2 = K_0/\rho_0$

This differential equation has a solution of the form:

$$\bar{p} = \bar{p}_1 \exp i(\underline{k} \cdot \underline{r} - \omega t) \quad \dots (5.11)$$

where $\bar{p}$ is written for δp and $\bar{p}_1$ for the maximum value taken by δp during the wave cycle.

To find the conditions under which the neglect of the gravitational term (arising in the momentum equation 5.2)

is justified, compare the magnitude of ρg with that of ∇p .

Since $p = p_0 + \bar{p}$, $\nabla p = \nabla \bar{p}$ (p_0 constant)

Then:

$$|\rho g|/|\nabla \bar{p}| = g/2\pi f c$$

where f is the frequency of the acoustic wave. For values of $g=10\text{m/sec}^2$ and $c=1500\text{m/sec}$ it can be seen that the effect of gravity can be ignored provided $f \gg 10^{-3}\text{Hz}$, a condition which is satisfied for many practical purposes (Bergmann, 1946).

A further approximation in the above treatment is the neglect of absorption of sound. The magnitude of this effect is illustrated in fig. 5.4. Rayleigh (1955) derived, on the basis of ordinary ("shear") viscosity, the relationship:

$$\alpha = 16\pi^2 \mu_s f^2 / 3\rho c^3 \quad \dots (5.12)$$

where α = intensity absorption coefficient (cm^{-1})

μ_s = shear viscosity (poise)

ρ = density (gm/ml)

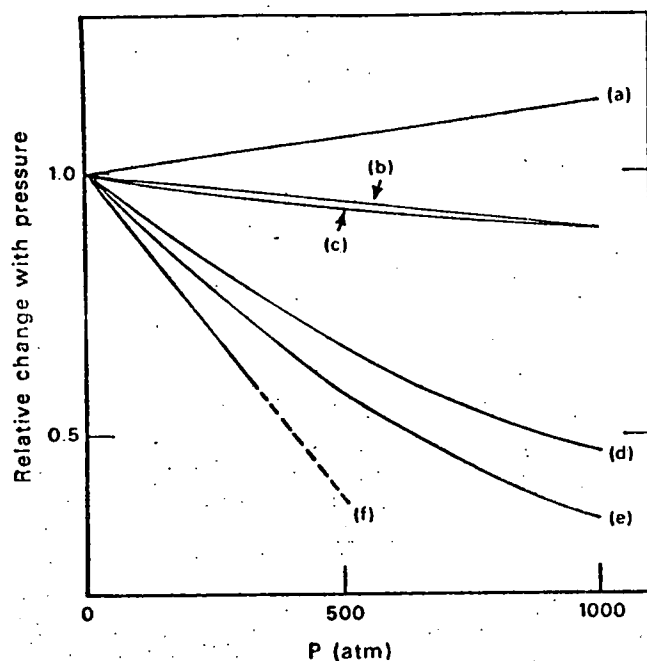
c = sound velocity (cm/sec)

f = frequency (Hz)

This predicts a value which is low by a factor of 3 from what is observed in de-ionised water. The difference is attributed to relaxation processes within the polymeric molecular structure of water, and is empirically taken into account by the concept of a "volume" viscosity (Urick, 1967,p.87). In seawater, the discrepancy at frequencies below 100KHz between observed and calculated (eqn. 5.12) values of absorption is much greater. This effect is attributed to absorption by dissolved ions, whether by the provision of energy for dissociation of molecules or for other relaxation processes (Leonard et al. 1949). Magnesium sulphate is the most important salt in this respect (Francois and Garrison, (1982). At very high frequencies, say over 10^8 Hz, high values for absorption may dictate a modified approach to the derivation of the wave equation. Most underwater acoustics is pursued at frequencies no greater than ca. 1MHz, so that for practical purposes, it is sufficient to incorporate absorption into the equation (5.11) by writing the wavenumber as the complex number:

$$k = K + i\bar{K}$$

where $\bar{K}$ describes the absorption, (e.g. Brekhovskikh



Relative change with pressure of (a) MgSO_4 ion-pair concentration according to Kester and Pytkowicz, (b) MgSO_4 ion-pair concentration according to Millero, (c) MgSO_4 ion-pair concentration according to Fisher, (d) sound absorption according to one Eigen-Tamm multistate model for MgSO_4 , (e) sound absorption in a 0.5M MgSO_4 solution measured by Fisher, (f) sound absorption in the ocean measured by Bezdek.

Results for $(\alpha\lambda)_m$ and f_r at 25 °C.

$P(\text{atm})$	$(\alpha\lambda)_m \times 10^6$	$f_r(\text{kHz})$	Solution
1	61.3 ± 0.5	140 ± 11	Lyman and Fleming seawater
71	57.8 ± 0.5	140 ± 11	
188	52.9 ± 0.3	133 ± 8	
307	45.2 ± 0.3	142 ± 7	0.02M MgSO_4
1	80.8 ± 0.6	149 ± 10	
307	65.0 ± 0.4	145 ± 8	
1	20.2 ± 0.3	162 ± 20	0.02M MgSO_4 + 0.6M NaCl
307	16.2 ± 0.2	150 ± 14	

Fig 5.5

a. Relative change with pressure in the absorption of sound in seawater (compiled by Hsu and Fisher, 1983). There is clearly considerable disagreement in the manner of this variation as found or predicted by different workers. The three lower curves (d,e,f) show some degree of consonance between the theoretical model (d) of Eigen and Tamm (1962) and the experimental data of Hsu and Fisher (1983) in the laboratory (e) and of Bezdek (1973a) in the sea (f).

b. A table of results of Hsu and Fisher (1983) on laboratory samples of MgSO_4 solutions and on synthetic seawater (Lyman and Fleming, 1940).

1980). This same approach to loss is taken in the model of a solid material used by Fuchs and Mueller (1971) as described in section 8.2.

Variation of attenuation with pressure may be significant in the deep ocean. Hsu and Fisher (1983) have made laboratory measurements of sound absorption at frequencies between 30 and 300KHz and pressures between 1 and 300 atmospheres in 0.02 molar aqueous solution of $MgSO_4$ in an attempt to effect some reconciliation between laboratory measurements of Kester and Pytkowicz (1970) with theoretical treatments of Millero (1971) and Eigen and Tamm (1962) and field observations in seawater by Bezdek (1973a,b). Their findings (fig. 5.5) favour the model of Eigen and Tamm and are in approximate agreement with the field observations of Bezdek, but it is clear that further effort is required before the manner in which absorption varies with depth in the ocean can be said to be understood.

5.2 RAY TRACING.

The general concept of ray tracing to define the path along which acoustic energy will travel through a medium where the sound velocity is varying is one which arises intuitively from a consideration of Snell's law, usually

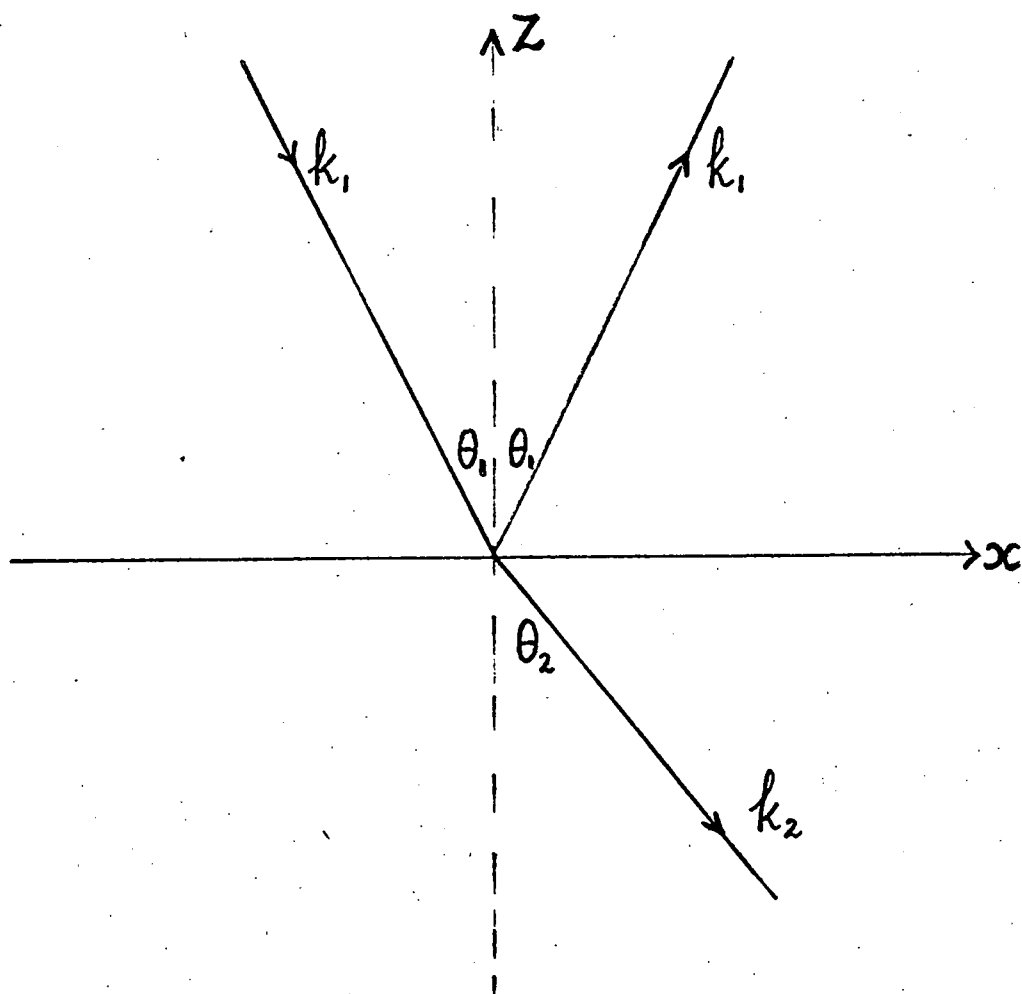


Fig. 5.6

Reflection and refraction of a plane wave at the boundary between two homogeneous fluid half spaces. A consideration of continuity of pressure and particle velocity across the boundary lead to equation (5.2) (Brekhovskikh, 1980, p.7).

derived formally by looking at the reflection and refraction of a wave at the plane interface between two fluid media (fig. 5.6). For a wave of unit amplitude, considerations of continuity of pressure and particle velocity at the interface require:

$$1 + V = W \exp(i(k_2 \sin \theta_2 - k_1 \sin \theta_1)x) \quad \dots\dots(5.13)$$

where V = reflexion coefficient

W = transmission coefficient

k_1 = wavenumber in the upper medium

k_2 = wavenumber in the lower medium

θ_1 = angle of incidence

θ_2 = angle of refraction

Since the left hand side of equation (5.13) does not depend on x , neither should the right hand side (Brekhovskikh, 1980,p.7). This condition is met if:

$$k_1 \sin \theta_1 = k_2 \sin \theta_2 \quad \dots\dots\dots(5.14)$$

which is Snell's law. A "ray parameter"

$$a = \sin \theta / v \quad \dots\dots\dots(5.15)$$

can be associated with each raypath between source and receiver, a parameter which remains constant for all points on the ray, and whose value depends on the direction in which the ray left the acoustic source.

A somewhat more formal derivation of the ray concept makes clearer the nature of the approximations used to simplify the mathematics and so provides the conditions under which the theory remains valid:

Consider an oscillatory pressure field $p(\underline{x}, t)$ having a slowly varying amplitude $A(\underline{x})$ and phase $k_0 W(\underline{x})$:

$$p(\underline{x}, t) = A(\underline{x}) \exp\{i(k_0 W(\underline{x}) - \omega t)\} \dots\dots(5.16)$$

where $k_0 = \omega/c_0$, where c_0 is a reference sound speed.

The index of refraction is defined as:

$$n(\underline{x}) = c_0/c(\underline{x})$$

The wave equation (5.10) then gives, by substitution:

$$\nabla^2 A + k_0^2 A (n^2 - |\underline{\nabla} W|^2) + ik_0 (2\underline{\nabla} A \cdot \underline{\nabla} W + A \nabla^2 W) = 0 \dots\dots\dots(5.17)$$

Since the real and imaginary parts of this must each be equal to zero:

$$|\underline{\nabla} W|^2 - n^2 = (1/k_0^2 A) \nabla^2 A \quad (\text{real part}) \dots\dots(5.18)$$

and

$$2\underline{\nabla} A \cdot \underline{\nabla} W + A \nabla^2 W = 0 \quad (\text{imaginary part})$$

$$\text{or } \underline{\nabla} \cdot (A^2 \underline{\nabla} W) = 0 \dots\dots\dots(5.19)$$

The instantaneous particle velocity $\underline{y}$ is given by:

$$\underline{y} = - (1/\rho) \underline{\nabla} p \dots\dots\dots(5.20)$$

and the energy flux by $p \underline{y}$ (Clay and Medwin, 1977, p81.)

The time-averaged energy flux for the wave specified in eqn. 5.16 is:

$$\langle p \underline{y} \rangle_t = (1/\rho_0 c_0) A^2 \underline{\nabla} W \dots\dots\dots(5.21)$$

That is, $\underline{\nabla} W$ shows the direction of energy transport and equation (5.19) shows that its divergence is zero, which is to say that it is confined to a 'ray'. The method of tracing these rays having direction $\underline{\nabla} W$ is obtained from equation (5.18) with the simplifying assumption that:

$$(1/k_0^2 A) \nabla^2 A \ll 1 \dots\dots\dots(5.22)$$

This is the same approximation which is made in geometric optics. The quantity $(A^{-1} \nabla^2 A)^{1/2}$ defines a length over which A varies significantly. Provided that the acoustic wavelength is much smaller than this length, the right hand side of equation (5.18) may be set equal to zero. In practical terms, this is a requirement that the wavelength be much smaller than the water depth, the radius of curvature of the rays, or any length over which the sound speed or density changes significantly. The equation (5.18) then simplifies to:

$$|\underline{\nabla} W| = n \quad \dots\dots\dots (5.23)$$

Rays can be expressed as functions $\underline{x}(s)$ where s is the arc length along the ray. The customary unit tangent vector of differential geometry is then:

$$\hat{\underline{t}}(s) = (d/ds) \underline{x}(s) \quad \dots\dots\dots (5.24)$$

This vector must be tangent to the path of energy transport to define the ray, that is, from equation (5.23):

$$\hat{\underline{t}} = (1/n) \underline{\nabla} W$$

Differentiation of the vector $n \hat{\underline{t}}$ along the ray gives:

$$(d/ds) n \hat{\underline{t}} = (d/ds) \underline{\nabla} W = \underline{\nabla} (dW/ds) = \underline{\nabla} (\underline{\nabla} W \cdot \hat{\underline{t}}) = \underline{\nabla} n$$

i.e. $d/ds \{n(d/ds) \underline{x}\} = \underline{\nabla} n \quad \dots\dots\dots (5.25)$

Equation (5.25) relates the ray function $\underline{x}$ to the index of refraction, and the function can be obtained by integration (numerically if necessary). Phase along the ray can then be derived from:

$$dW = n ds \quad \dots\dots\dots (5.26)$$

Usually in the real ocean variations in the index of refraction occur mainly with depth, and at any given depth, n is constant in horizontal directions. In these circumstances, the vector equation (5.25) simplifies to:

$$(d/ds)(n \, dx/ds) = 0 \quad \dots\dots\dots(5.27a)$$

$$(d/ds)(n \, dy/ds) = 0 \quad \dots\dots\dots(5.27b)$$

$$(d/ds)(n \, dz/ds) = dn/ds \quad \dots\dots\dots(5.27c)$$

The equations (5.27a) and (5.27b) imply that

$$dy/dx = \text{constant}$$

so that any given ray is confined to a plane perpendicular to the x - y plane, reducing the problem to a two dimensional one. For a ray propagating in the plane $y=0$, the ray equations become:

$$ndx/ds = a \quad \text{where } a \text{ is a constant} \dots\dots\dots(5.28a)$$

$$d/ds(ndz/ds) = dn/dz \quad \dots\dots\dots(5.28b)$$

$$\text{where } (dx/ds)^2 + (dz/ds)^2 = 1 \quad \dots\dots\dots(5.28c)$$

Equations (5.28a) and (5.28b) can be re-written as:

$$dx/ds = a/n$$

$$\text{and } dz/ds = \{1 - (a^2/n^2)\}^{1/2}$$

which in turn give:

$$dx/dz = a/\sqrt{(n^2 - a^2)} \quad \dots\dots\dots(5.29)$$

Raypaths can be obtained from this equation for any particular velocity profile. The constant a is the same "ray parameter" which is discussed and used in geophysical methods (see section 7.4) for the measurement of sound velocity in seabed sediments (e.g. Bryan, 1974). Sound velocity profiles have normally been derived from

$$\begin{aligned}
 V &= 1449.14 + V_T + V_P + V_S + V_{STP} \\
 V_T &= 4.5721T - 4.4532 \times 10^{-4}T^2 - 2.6045 \times 10^{-6}T^3 + 7.9851 \times 10^{-9}T^4 \\
 V_P &= 1.60272 \times 10^{-3}P + 1.0268 \times 10^{-5}P^2 + 3.5216 \times 10^{-8}P^3 - 3.3603 \times 10^{-10}P^4 \\
 V_S &= 1.39799(S-35) + 1.69202 \times 10^{-4}(S-35)^2 \\
 V_{STP} &= (S-35)(-1.1244 \times 10^{-4}T + 7.7711 \times 10^{-7}T^2 + 7.7016 \times 10^{-9}P - 1.2943 \times 10^{-11}P^2 + 3.1580 \times 10^{-13}PT \\
 &\quad + 1.5790 \times 10^{-15}PT^2) + P(-1.8607 \times 10^{-4}T + 7.4812 \times 10^{-7}T^2 + 4.5283 \times 10^{-9}T^3) \\
 &\quad + P^2(-2.5294 \times 10^{-7}T + 1.8563 \times 10^{-9}T^2) + P^3(-1.9646 \times 10^{-10}T).
 \end{aligned}$$

In this equation the units of temperature, pressure, salinity, and sound speed are °C, kg/cm², parts per thousand, and meters/sec, respectively. The standard deviation from the mean is 0.30 m/sec for all data obtained in the ranges -4°C < T < 30°C, 1 kg/cm² < P < 1000 kg/cm², and 0‰ < S < 37‰. The new equation may consequently be used to predict sound speeds to within 0.30 m/sec over the indicated temperature, pressure, and salinity ranges.

$$C = C_0 + C_T + C_S + C_P + C_{TSP}$$

$$\begin{aligned}
 C_0 &= 1402.394, C_T = 5.01132T - 5.513036 \times 10^{-3}T^2 \\
 &\quad + 2.221008 \times 10^{-4}T^3, C_S = 1.332947S, C_P = 1.605336 \\
 &\quad \times 10^{-3}P + 2.12448 \times 10^{-5}P^2, C_{TSP} = -1.266383 \times 10^{-4}TS \\
 &\quad + 9.543664 \times 10^{-7}T^2S - 1.052396 \times 10^{-8}TP^2 + 2.183988 \\
 &\quad \times 10^{-11}TP^3 - 2.253828 \times 10^{-13}SP^3 + 2.062107 \times 10^{-15}TS^2P.
 \end{aligned}$$

Check value: at T = 2°C, S = 34.7‰, P = 6000 dbar; C = 1559.499 m/s.

Note again that the above three equations differ from previous sound-speed equations in that they employ decibars (0 at surface) in the pressure terms. For those users without access to computer programs that determine pressure from oceanographic data, we suggest a modified version of Leroy's simplified pressure equation

$$P(Z) = 1.0052405(1 + 5.28 \times 10^{-3} \sin^2 \phi)Z + 2.36 \times 10^{-4}Z^2,$$

where P represents pressure in decibars, Z depth in meters, and φ latitude. Leroy indicates that use of this pressure equation should not introduce errors in sound speed exceeding 0.05 m/s for all ocean-sea areas outside the Black and Baltic seas. However, the relatively small Sula Sea proves another exception, with errors > 0.05 m/s below 2000 m, and ~0.2 m/s at maximum depth.

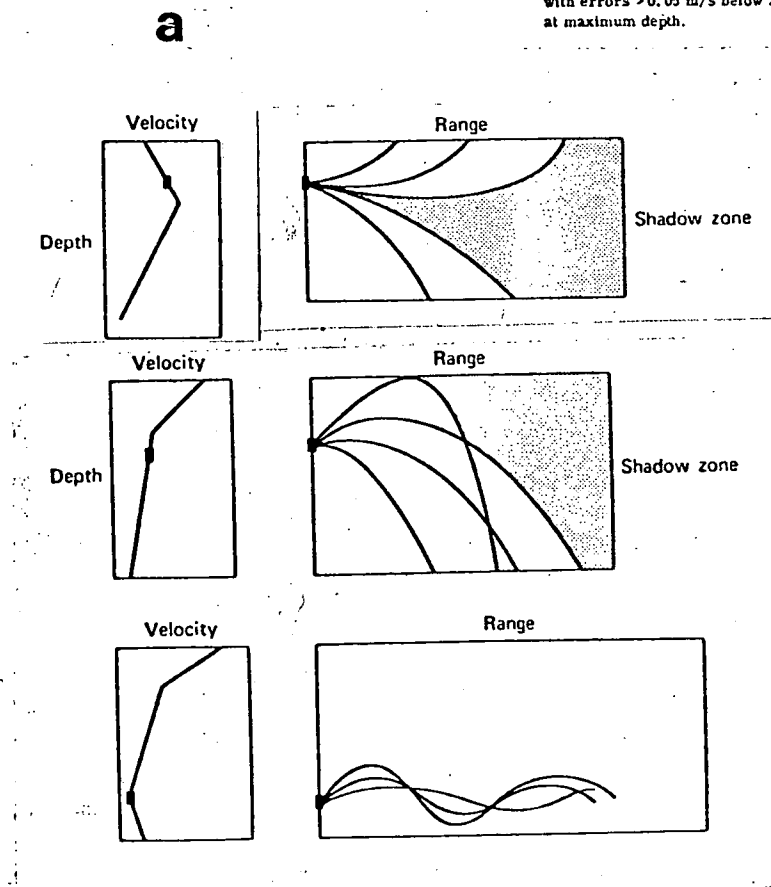


Fig. 5.7

a. Empirical equation for the speed of sound in seawater as a function of temperature, salinity and depth (Wilson, 1960).

b. Lovett (1978) considers that Wilson's equation gives the best pressure dependence, but that an equation given by Del Grosse (1974) on the basis of work by Del Grosse and Mader (1972) provides the best relationship with temperature and salinity. Lovett has merged the two equations to create a third, which is shown in this figure. Note that pressures are derived from depths for this equation by a method of Leroy (1969).

The expected differences in velocity as predicted by the equation in (a) and (b) is no greater than 0.5 m/sec. (about 1 part in 3000). This precision probably represents the limit with which velocity can be predicted in this way.

c. Examples of ray-tracing for three different velocity-depth profiles. (Camp, 1970, p.81).

oceanographic station measurements of temperature and salinity at many depths through the water column. The relationships between these parameters and the acoustic velocity is consequently of importance. A recent discussion about the relationship is given by Lovett (1978), who compared equations of Wilson (1960), given in fig. 5.7a and of Del Grosso (1974). Lovett produced from these yet another equation (fig. 5.7b), which probably represents the extent to which this technique of finding the speed of sound can be refined.

When a sound speed profile is known, a ray trace can be generated for a sound source at the desired depth (fig. 5.7c) by making use of equation (5.29). This figure illustrates the way in which shadow zones can be formed, that is, areas which will not be reached by sound propagating from the source, because of the bending of rays. Also shown is the situation where a sound source is located in a low velocity region, such that the velocity increases both above and below the source. This has the effect of producing a channel or wave guide to which rays are confined and along which they may propagate for a long distance without loss by spreading. Bryan et al. (1963) investigated the possibility of observing acoustic arrivals at very great distances by locating both source and receiver in the low velocity zone which occurs consistently beneath the thermocline in tropical and

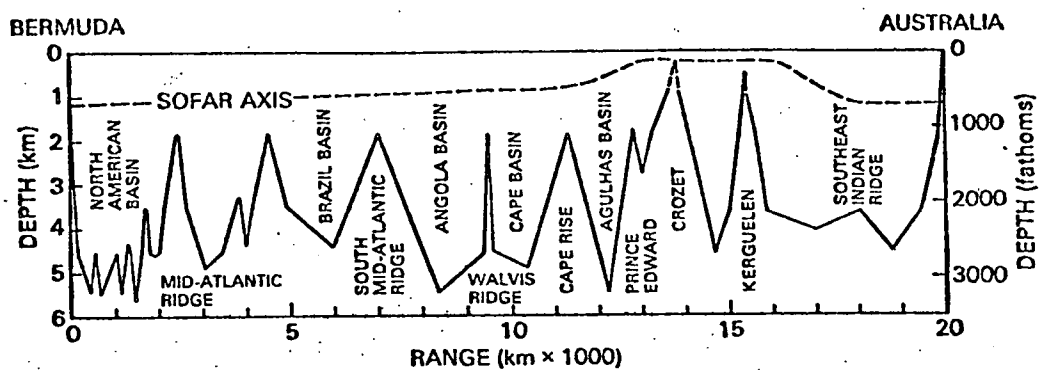
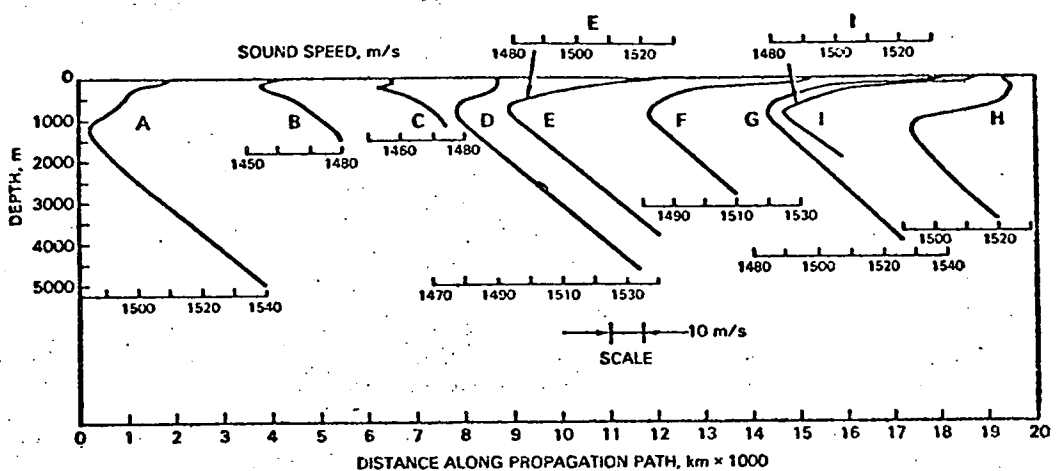
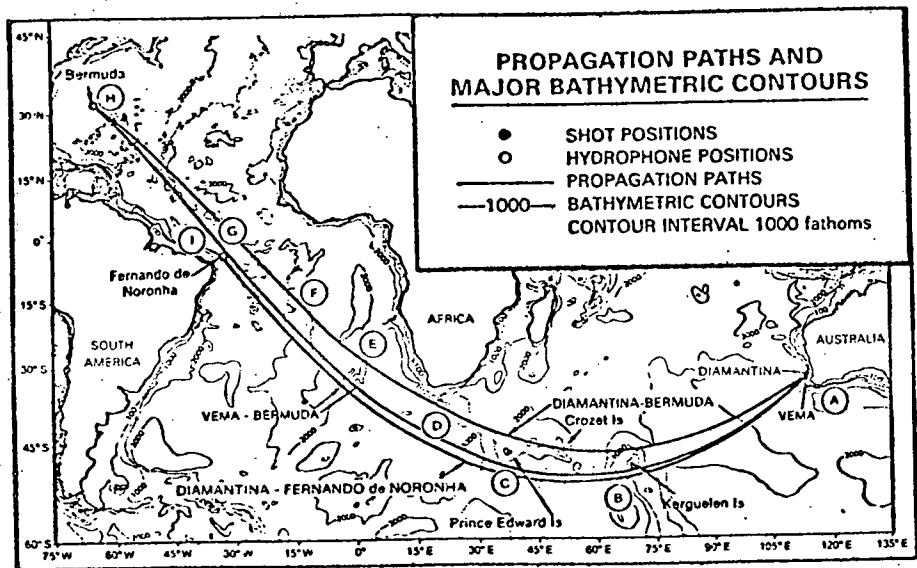


Fig. 5.8

a. Propagation paths for sound generated by dropping depth charges set to fire on the axis of the "SOFAR" channel i.e. the low velocity zone.

b. Sound speed profiles from oceanographic stations.

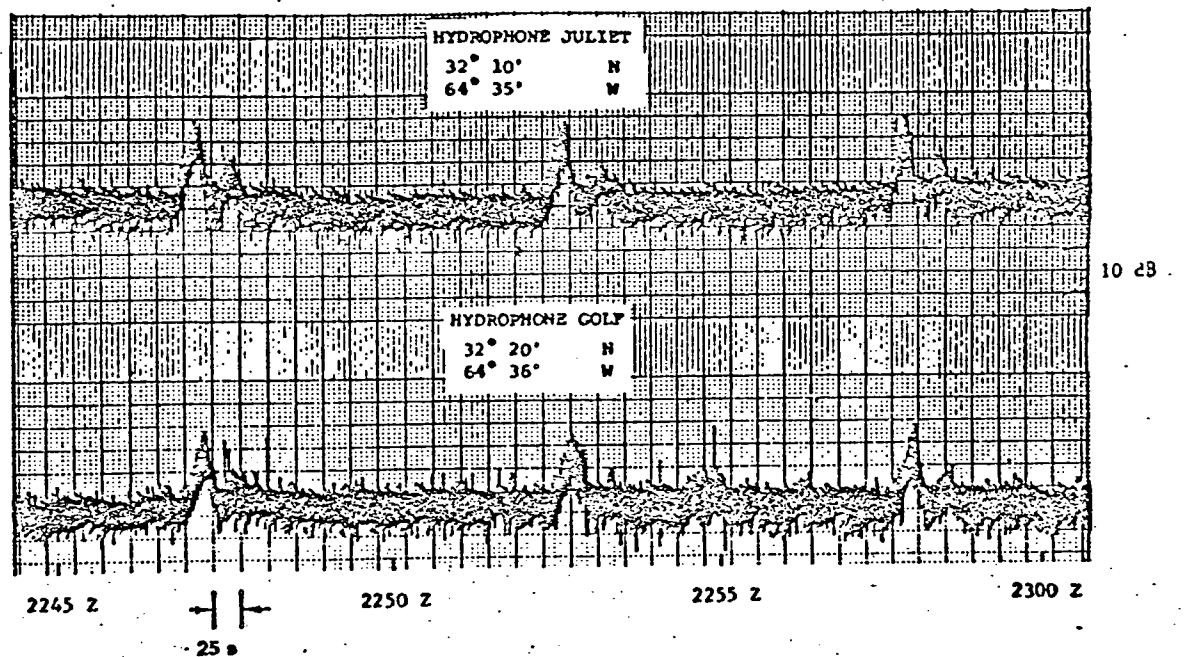
c. Axis of the "SOFAR" channel between Bermuda and Australia.

(figures from Shockley et al., 1982)

subtropical seas (Ewing and Worzel, 1948) and which can extend in suitable weather in the summer season into higher latitudes. At the time of the investigations, geodetic methods making use of satellites did not exist, and the primary interest was to establish the accuracy with which great circle distances between mid-oceanic islands (fig. 5.8a) might be found by acoustic methods (Ewing et al., 1959). Some of the results obtained by firing depth charges on the axis of the low velocity zone (fig. 5.8b,c) are shown in fig. 5.9a. These have been discussed by Shockley et. al. (1982) in relation to a modified adiabatic invariant approximation for predicting the speed of guided waves (fig. 5.9b). It seems doubtful from the results that the geodetic aims of the experiment could have been realised even if the alternative earth-satellite method had not materialised, but the work gives one of the best illustrations available of the ranges possible for acoustic waves in the ocean.

Since the principal purpose of ray tracing is to predict signal levels, it is important to consider the way in which amplitudes are computed.

Consider a tube of adjacent rays capped at each end by surfaces of constant phase. This defines a volume in space in which Gauss's integral identity may be applied to give (from eqn. 5.19):



a

Observed and calculated signal speeds for the VEMA and DIAMANTINA shots.

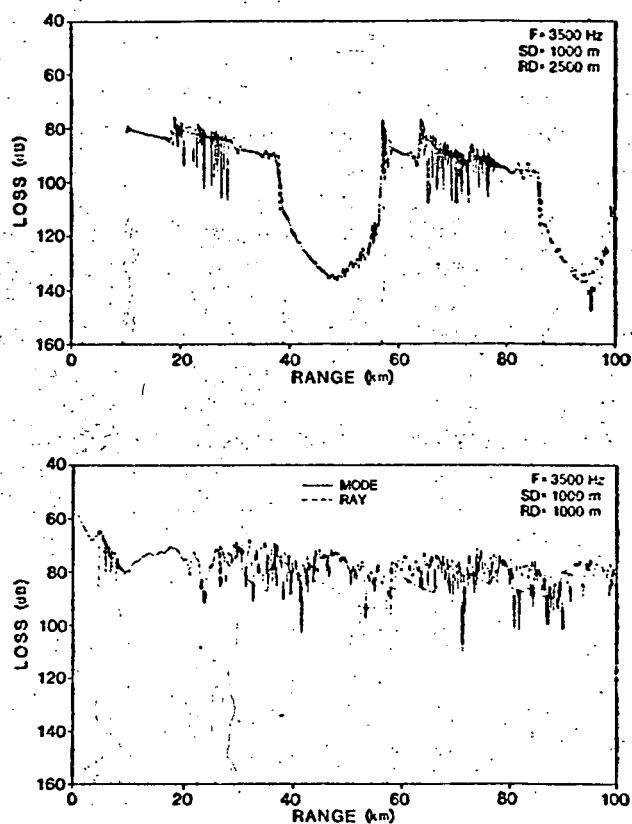
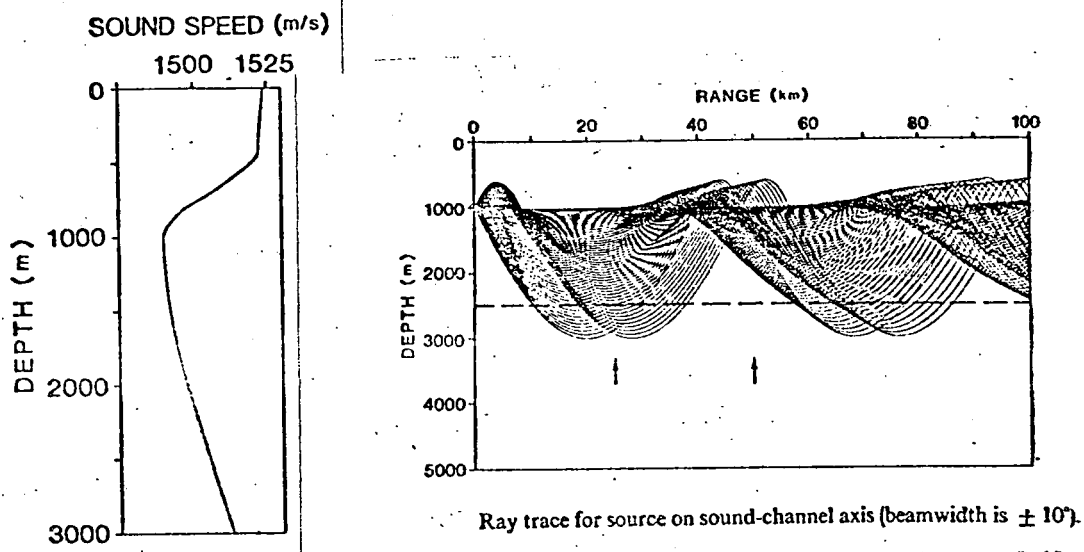
Propagation path	Signal speeds (m/s)			
	Measured	Long-range raytracing	AIA based on raywave	AIA based on bathymetry
VEMA-Bermuda	1485.0 ± 5.2	1477.5 ± 0.4	1477.0	1481.2
DIAMANTINA-Bermuda	1481.6 ± 0.6	1478.3 ± 0.8	1478.3 ± 0.5	1480.4
DIAMANTINA-Fernando de Noronha	1480.4 ± 0.9	1473.6 ± 1.2	1477.5 ± 1.4	1479.5 ± 1.5

b

Fig. 5.9

a. Oscillogrammes produced by Hartdegen (unpubl., noted in Shockley et al., 1982), showing the arrival at Bermuda of shots fired near Perth (Australia). The signals (unprocessed) are quite clear against background noise.

b. Measured speed of guided wave propagation for the propagation paths shown in Fig. 5.8a. These are compared with ray-theoretical predications of the propagation speed (Shockley et al., 1982). The variations among the theoretical values arise from different ways in which horizontal variation in the sound speed profile is treated. (A.I.A. stands for adiabatic invariant approximation, whose basic assumption is that $\oint c^{-1} \sin \theta \, dz$ is a constant, where integration is carried out over one ray loop. This may be compared with the regular ray-trace assumption that $c^{-1} \sin \theta = \text{constant}$.)



Propagation loss versus range for source on sound-channel axis.

Fig. 5.10

a. Sound speed profile and raytrace for an acoustic source on the axis of a low velocity channel.

b. A comparison between the propagation loss expected when the frequency is 3.5KHz and the source is located at 1000m in the cases where the receiver is at 2500 m (upper) and at 1000m (lower).

(Ferla et al.,1982)

$$\int_V \nabla \cdot (A^2 \nabla W) dV = \int_S A^2 \nabla W \cdot \hat{m} dS = 0 \quad \dots\dots(5.30)$$

where $\int_V dV$ and $\int_S dS$ imply integration throughout the volume and over the surface of the tube, respectively, and where $\hat{m}$ is the unit vector normal to the surface S . Because the sides of the tube are formed by rays, then $\nabla W \cdot \hat{m} = 0$ on this surface, leaving only the end caps, which are phase fronts where $\nabla W \cdot \hat{m} = n$. If the tube is taken to be narrow enough that A is constant over each end cap then:

$$A_1^2 n_1 dS_1 = A_2^2 n_2 dS_2 \quad \dots\dots(5.31)$$

where the areas of the endcaps are dS_1 and dS_2 .

From this, it follows that:

$$A_2 = A_1 (n_1 dS_1 / n_2 dS_2)^{1/2} \quad \dots\dots(5.32)$$

This gives the method for finding the amplitude of the acoustic field from the source strength as defined at some reference distance r_1 (often 1 yard in the American literature). When two or more ray tubes pass through a given receiver point, the resultant amplitude can be calculated by summing the individual ray amplitudes in the form:

$$A \exp(ik_0 W) = \sum_j A_j \exp(ik_0 W_j)$$

in order to take account of possible phase differences.

From calculations based on the above theory, it is possible to predict transmission losses. An example can be found in Ferla et al. (1982). This is shown in fig. 5.10. It is clear from a comparison of the two diagrams showing propagation loss (5.10b) that the position of the receiver in the water column can have a profound effect on



the expected loss. When sound is confined to a guided channel within the water column, the properties of the seabed are irrelevant to calculations of propagation loss, but when the sound speed profile does not give rise to a waveguide, interaction with the bottom becomes significant and a knowledge of the physical properties of the seabed materials is required before valid predictions can be made about propagation. This is usually of greatest importance in shallow water, where a different analysis of acoustic propagation is often used. This is discussed briefly in the next section before the question of bottom interaction is pursued further.

5.3 NORMAL MODE THEORY.

Consider a pressure wave of the form of eqn. (5.11) propagating in a space which is of infinite extent in x and y , but is bounded in z by an upper pressure release surface and a lower rigid boundary. The wave equation (5.10) then becomes:

$$\nabla^2 p + k_0^2 p = 0 \quad \dots\dots\dots (5.33)$$

$$\text{where } k_0 = \omega/c$$

Since the problem is assumed to be unchanged by translation in x and y , the solution will take the form:

$$p(x,y,z) = R(x,y)U(z)$$

where R and U satisfy the equation:

$$d^2R/dx^2 + d^2R/dy^2 + k^2R = 0 \quad \dots\dots\dots(5.34a)$$

$$d^2U/dz^2 + (k_0^2 - k^2)U = 0 \quad \dots\dots\dots(5.34b)$$

where the separation constant, k , corresponds to the horizontal wave number, i.e.

$$R(x,y) = \exp\{i(k_x x + k_y y)\}$$

$$\text{with } k_x^2 + k_y^2 = k^2$$

The equation (5.34b) has solutions $U_n(z)$ only for certain values of k , given by:

$$k_n^2 = k_0^2 - (n-1/2)^2 \pi^2 / H^2 \quad \dots\dots\dots(5.35)$$

where n is a positive integer.

For k_n defined by eqn. (5.35) the solutions for $U(z)$ which satisfy the boundary conditions $U(0) = 0$ (corresponding to $p=0$) at the upper ($z=0$) pressure-release boundary, and $dU/dz=0$ for $z=H$ at the lower rigid boundary, the solutions are:

$$U_n(z) = (z/H)^{1/2} \sin \sqrt{k_0^2 - k_n^2} z \quad \dots\dots\dots(5.36)$$

It follows that the general solution of the wave equation for waves propagating in the x direction is:

$$p(x,y,t) = \sum_n a_n U_n(z) \exp\{i(k_n x - \omega t + \phi_n)\} \quad \dots\dots(5.37)$$

where the amplitudes a_n and phases ϕ_n depend on how the modes are excited by the source.

From eqn. (5.35) it follows that k_n^2 is positive for only a certain number of modes. That number is:

$$N = (\text{integral part of}) \{2Hf/c + 1/2\} \quad \dots\dots(5.38)$$

The velocity of phase fronts for mode " n " is given by:

$$c_n(f) = \omega/k_n = c \{1 - (n-1/2)^2 (c/2Hf)^2\}^{1/2} \quad \dots\dots(5.39)$$

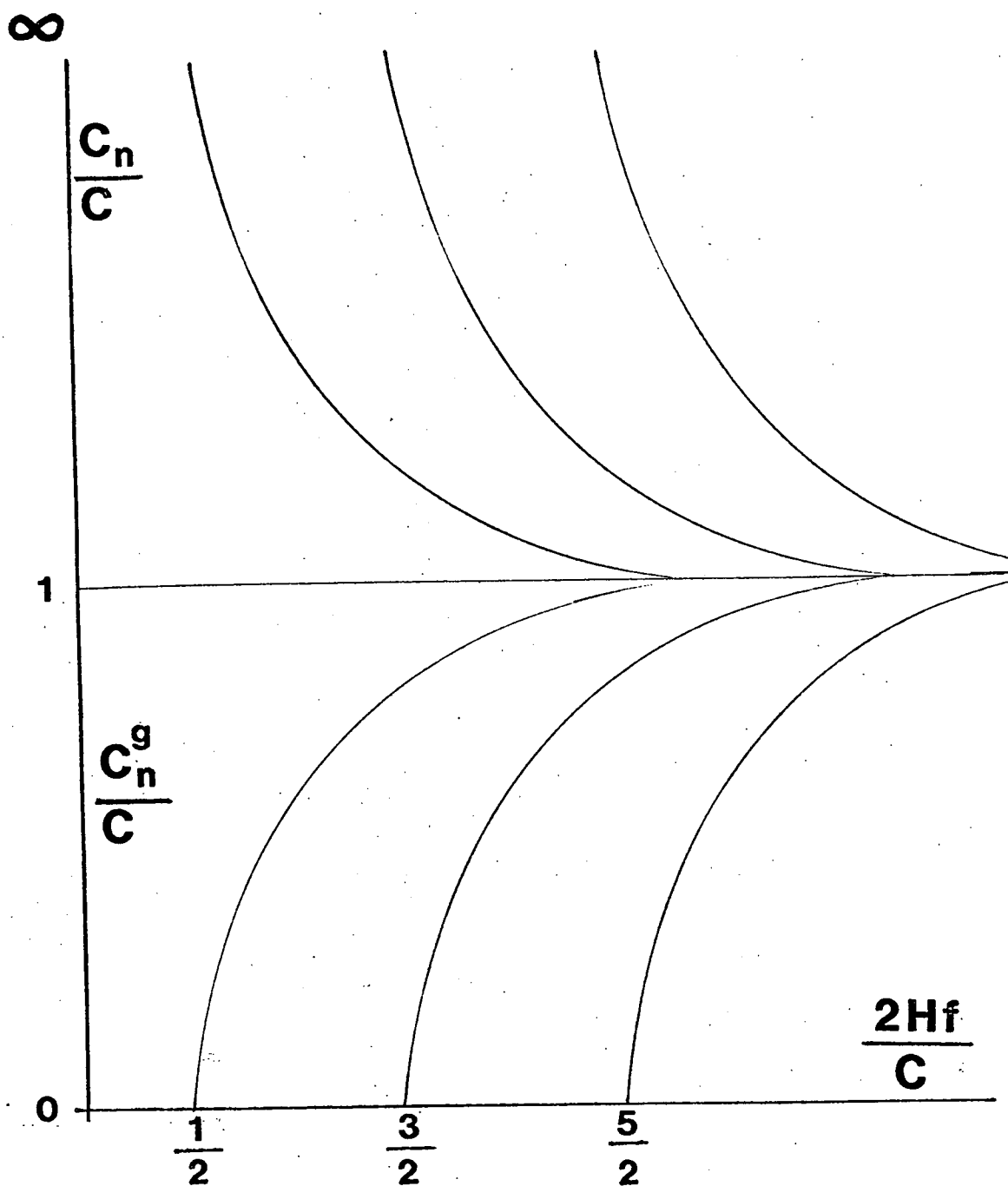


Fig. 5.11

Variation of the phase and group velocities of the first three modes relative to the compressional wave velocity as a function of frequency.

This expression depends on frequency, which implies that a wavelet consisting of a combination of several frequencies disperses as it propagates.

The velocity of energy propagation is:

$$c_n^g(f) = d\omega / dk_n = c \{ 1 - (n-1/2)^2 (c/2Hf)^2 \}^{1/2} \dots (5.40)$$

The variation of $c_n(f)/c$ and $c_n^g(f)/c$ is shown in fig.

5.11. Each mode has a cutoff frequency below which no energy can propagate by that mode. This frequency is:

$$f_n = (n - 1/2) (c/2H) \dots \dots \dots (5.41)$$

The main features of the normal mode treatment are:

- (i) the solution of the wave equation is separable
- (ii) there exist a finite number of discrete, orthonormal propagating modes
- (iii) the modes are frequency dispersive
- (iv) there is a cutoff frequency.

These considerations carry through to the case where the bottom of the fluid layer is not rigid, but consists of a lossy elastic medium. In this instance, the mode functions $U(z)$ penetrate into the bottom material to varying depths (fig. 5.12). The way in which the acoustic field is influenced depends on the physical properties of the bottom material. This has led to a considerable pre-occupation in the literature of underwater acoustics with estimating or measuring these properties.

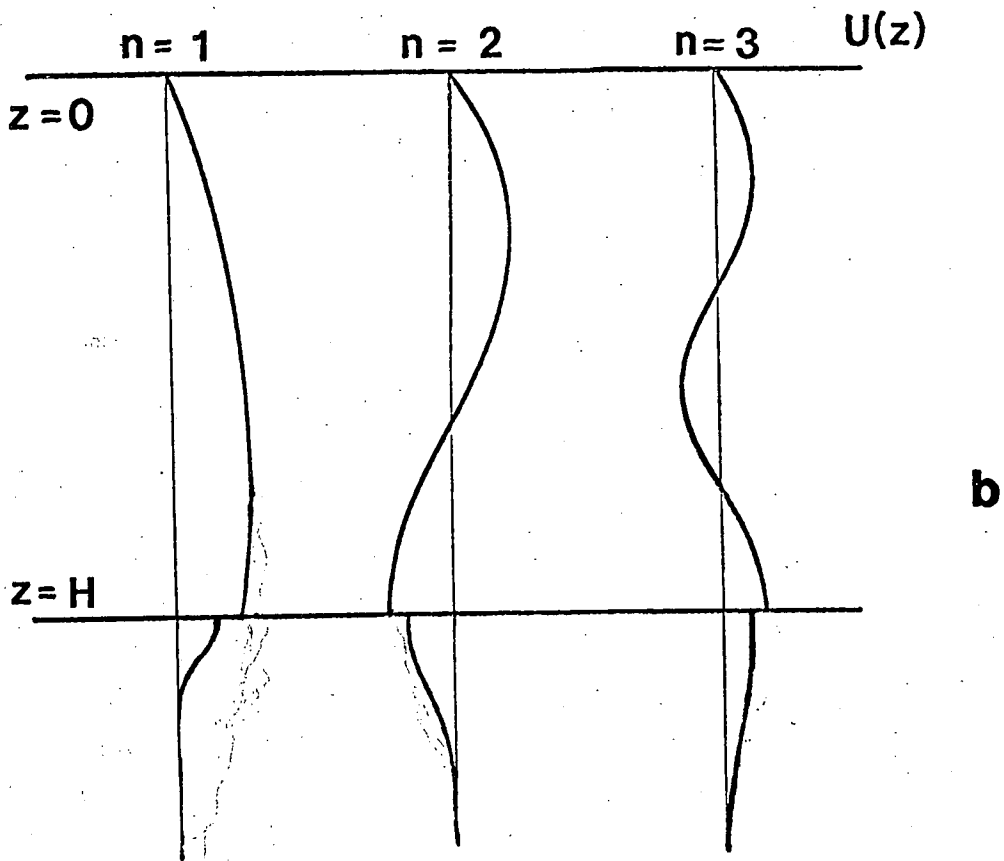
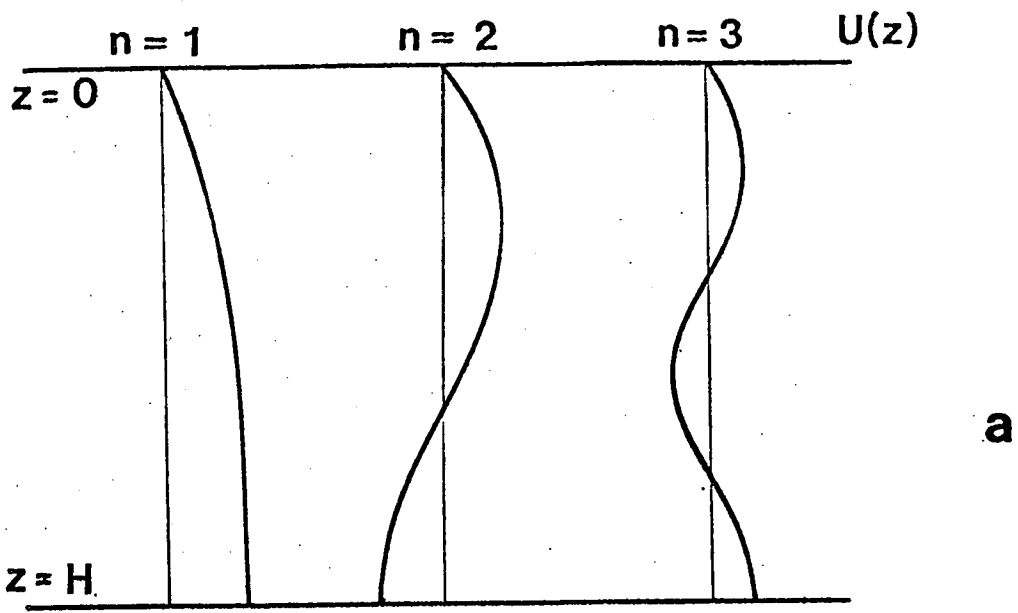


Fig. 5.12

a. The wave function $U(z)$ plotted against depth (z) for the first three normal modes in a fluid bounded at $Z = 0$ by a pressure release surface and at $Z = H$ by a perfectly rigid surface.

a. As for (a), except that now the lower surface represents the top of a viscoelastic medium.

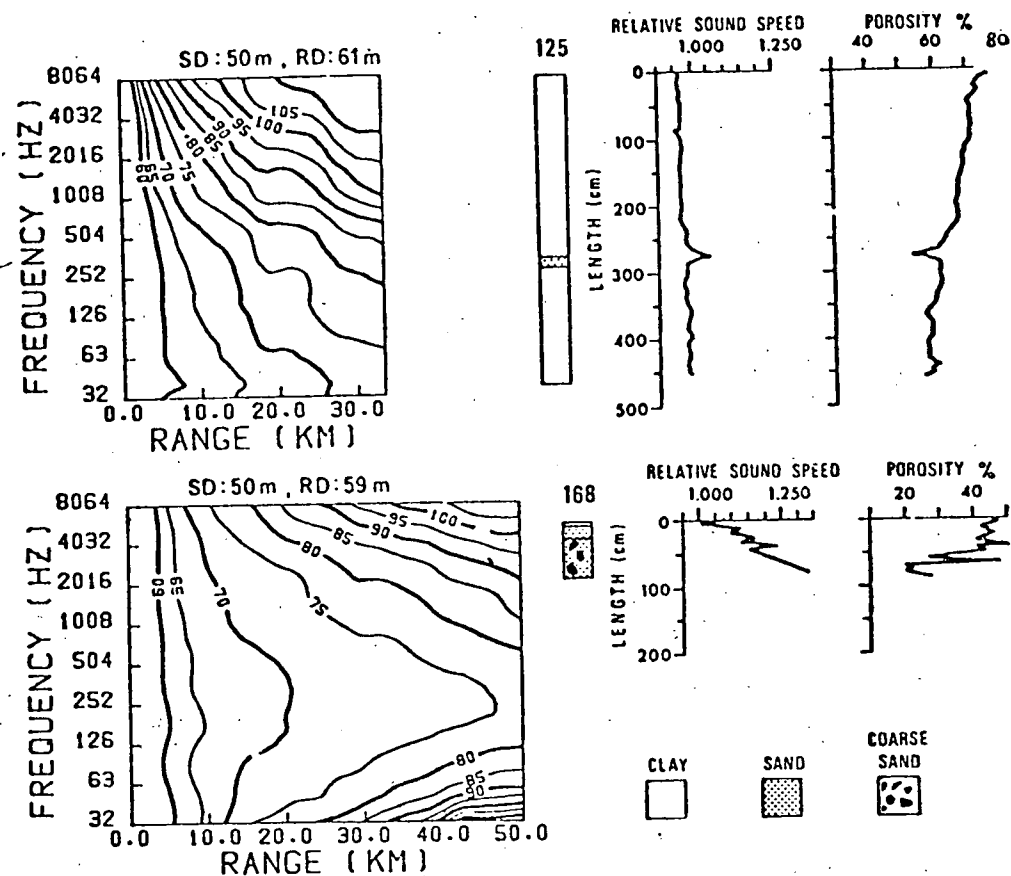
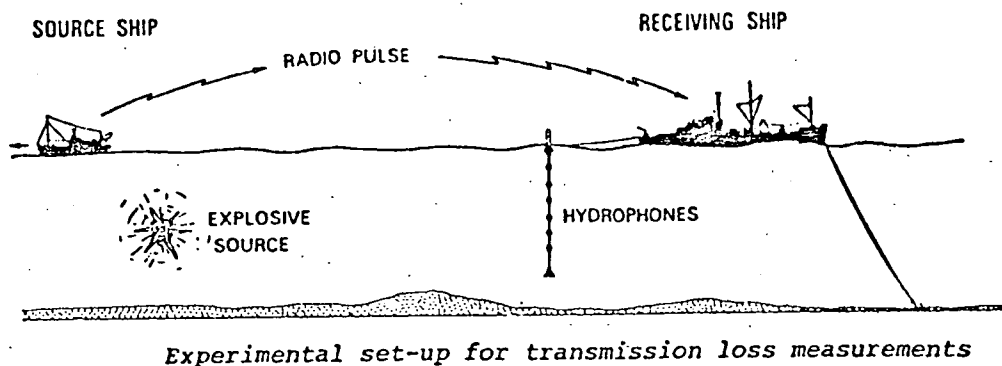


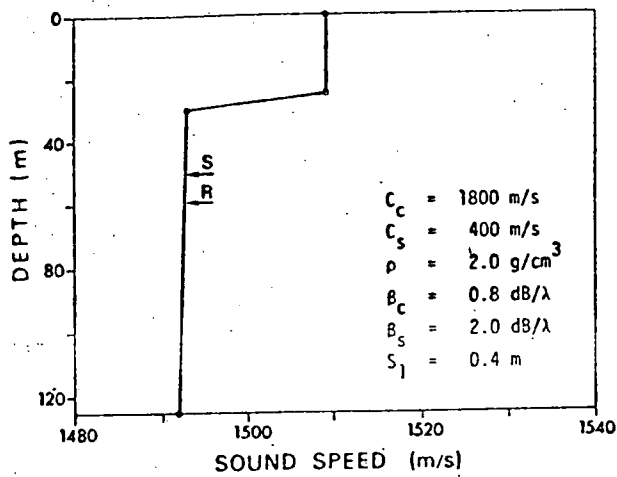
Fig. 5.13

a. Equipment deployment for transmission loss experiments.

b. Measurements by Akal (1981) of the variation of transmission loss (contours in dB) with frequency and range for two different bottom types whose sound speed and porosity have been obtained from the cores illustrated in section.

5.4 INFLUENCE OF SEABED PROPERTIES ON SOUND PROPAGATION.

In contrast with sound propagation in the 'SOFAR' channel discussed in section 5.2, the transmission characteristics in shallow water depend on such environmental parameters as the roughness of the sea surface and the seabed and on the physical properties of the seafloor. Propagation paths which interact with the bottom many times between source and receiver become very sensitive to the boundary conditions at that interface. Akal (1981) used a typical experimental configuration (fig. 5.13a) to investigate the effect of different bottom types on propagation. His results are shown in fig. 5.13b which illustrates the difference in losses for the two bottom types, whose sound speed and porosity were found from core samples. These results, and those of many other workers suggest that there is an optimum frequency at which sound should be transmitted to achieve maximum range. This has been investigated by Jensen and Kuperman (1983) who obtained data from a "coastal area of the North Atlantic" by the use of two ships in the manner shown in fig. 5.13a. The sound speed profile in their area appears in fig. 5.14a and their experimental results in fig. 5.14b. Jensen (1981) points out that it is to be expected that propagation loss will increase with range and with frequency due to attenuation in the water as discussed in section 5.1. At sufficiently low frequencies, however,



Sound-speed profile and bottom properties for shallow-water area in the eastern North Atlantic. C = sound speed, ρ = density, β = attenuation, S_1 = rms bottom roughness, subscripts c and s refer to compressional and shear waves, respectively. Source and receiver depths are indicated on the sound-speed profile.

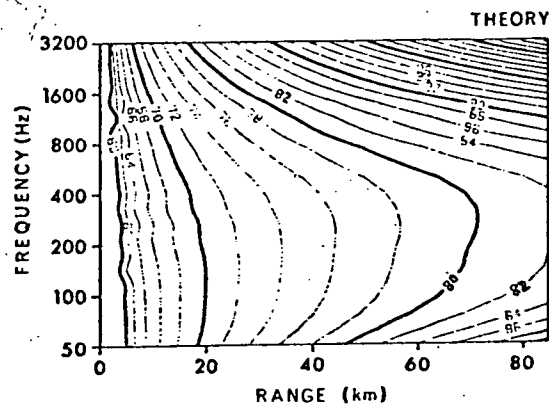
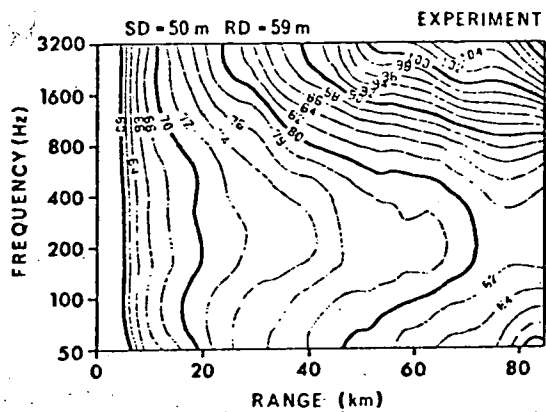


Fig. 5.14

a. Sound-speed profile in the region of the experiment of Jensen and Kuperman (1983).

b. Broadband experimental data from test area.

c. Theoretical propagation loss.

	POROSITY	RELATIVE DENSITY	RELATIVE SPEED	COMPRESSIONAL SPEED	SHEAR SPEED	COMPRESSIONAL ATTENUATION	SHEAR ATTENUATION
BOTTOM TYPE	P(%)	ρ_B/ρ_W	C_B/C_W	C_B (m/s) *	C_S (m/s)	β_C (dB/ λ)	β_S (dB/ λ)
CLAY-SILT	60	1.6	1.01	1515	100	0.5	1.0
SAND-SILT-CLAY	55	1.7	1.02	1530	150	0.8	1.5
SILT	50	1.8	1.05	1575	200	1.0	2.0
SAND-SILT	40	1.9	1.1	1650	400	0.8	1.5
COARSE SAND	35	2.0	1.2	1800	600	0.7	1.5
CHALK-LIMESTONE	-	2.2	1.5	2250	1000	0.4	1.0
LIMESTONE	-	2.4	2.0	3000	1500	0.3	0.5
BASALT	-	2.6	3.5	5250	2500	0.2	0.5

* $C_W = 1500$ m/s

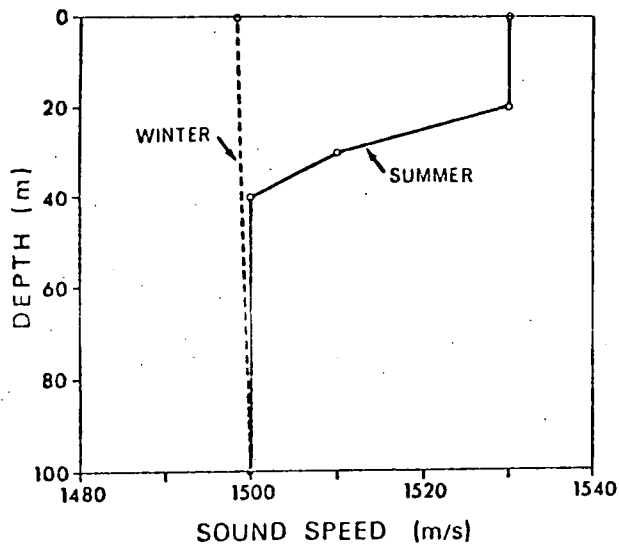
Table 5.1

Geo-acoustic parameters for different bottom types as compiled by Jensen and Kuperman (1983).

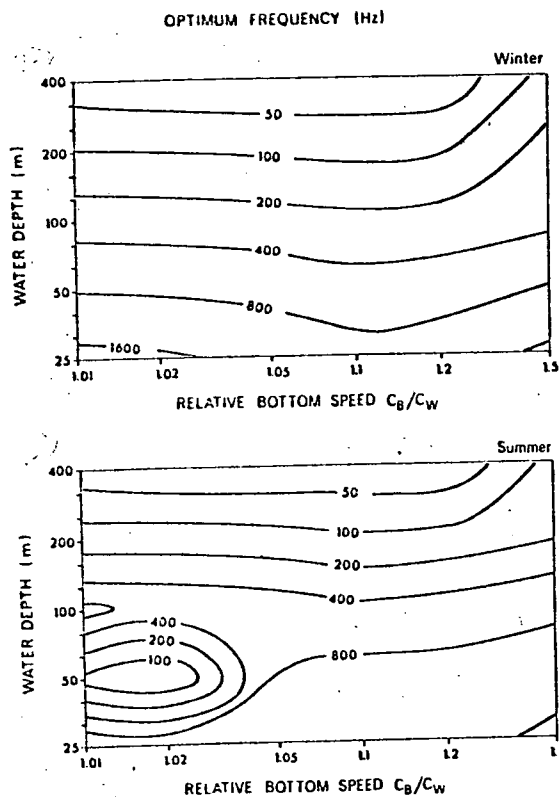
the effect of mode cutoff frequency (section 5.3) also leads to high losses (Tindle et al., 1978). In general, in addition to environmental parameters, the optimum frequency depends on both range and on the depth of source and receiver. The theoretical losses (fig. 5.14c) are based on a smoothed summation of normal modes (Jensen, 1981), which have been calculated from compressional velocity and density measured from core samples. Shear wave velocity and attenuation in the bottom material have been adjusted to obtain an optimum fit between figs. 5.14b and c so that fig. 5.14c can be considered to be only a partially theoretical prediction of propagation loss, although approximate values for such parameters can be obtained from Akal (1974) or Hamilton (1980) as summarised in Table 5.1. Having obtained a reasonably good fit between experiment and theory (figs. 5.14b&c), Jensen and Kuperman were able to make predictions for a different sound speed profile in the same area (fig. 5.15).

Akal (1981) has plotted changes in the optimum propagation frequency on the basis of his observations (fig. 5.16) and notes a correlation with change in bottom type.

Jensen and Kuperman (1982) in discussing the merits of different techniques for modelling sound propagation in the water column (including still further ways of solving



a

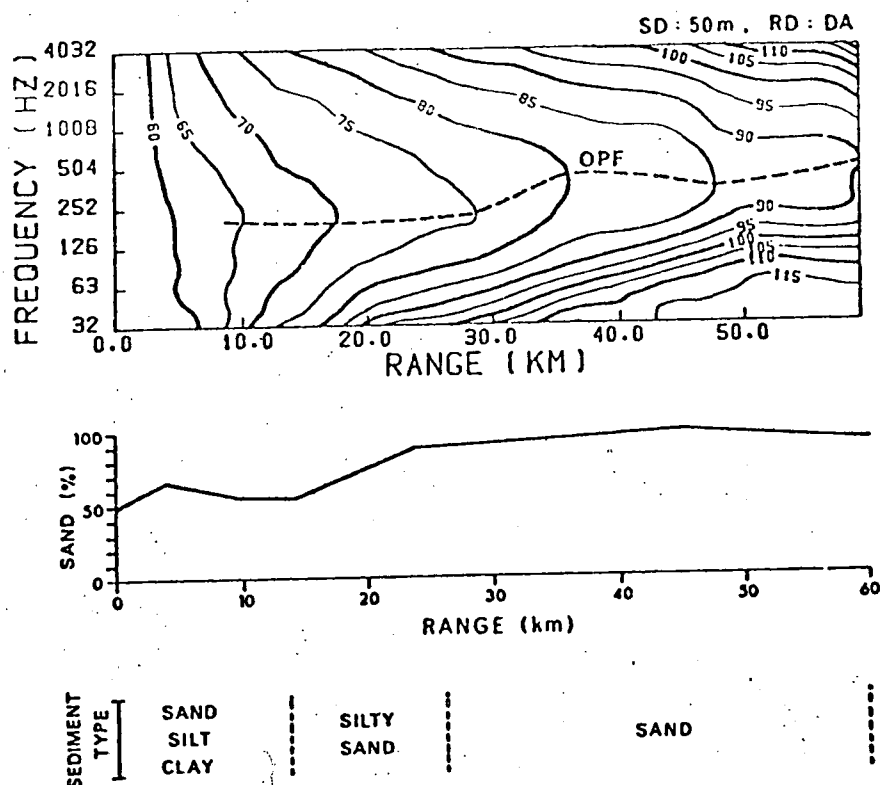


b

c

Fig. 5.15

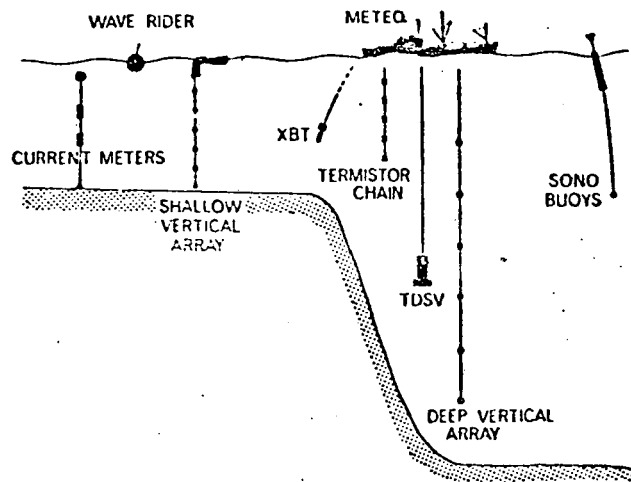
- a. Winter and summer sound speed profiles.
 - b. Prediction of optimum frequency for winter conditions.
 - c. Optimum frequency from summer observations.
- (Jensen and Kuperman, 1983).



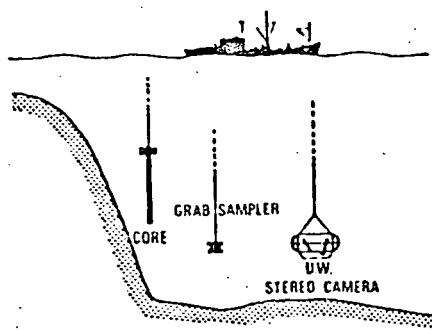
*Effects of spatial variation of sediments
on acoustic propagation*

Fig. 5.16

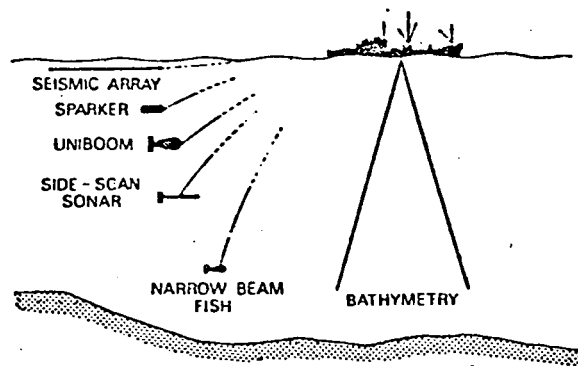
Change in the optimum propagation frequency apparently resulting from change in bottom sediment type as noted by Akal (1981).



a



b



c

Summary of environmental data collected during acoustic measurements

Fig. 5.17

- a. Methods of measuring parameters of the water column.
- b. Direct methods of investigating the subbottom.
- c. Indirect methods of investigating the subbottom.

the wave equation in addition to ray and normal mode theory) note that "considerable experience" with any given modelling technique is often necessary in order to obtain meaningful results. This may be a result of analytical singularities and related problems in some instances, but often the difficulty is simply that of having insufficient environmental information to use in the model. Jensen and Kuperman (1979) observe that the principal problem is how to obtain sufficiently detailed environmental information to be able to predict propagation in any given coastal area. Akal (1981) describes the effort which is put into gathering such information, and this is summarized in fig. 5.17. It is relatively easy to obtain sound speed profiles in the water column. Bottom properties are much more difficult to measure. Although in principle, one might expect to be able to obtain a fit between experiment and theory (e.g. fig. 5.14 b and c) and so derive some or all of the bottom properties, Jensen and Kuperman (1979) in discussing those same results point out that around the optimal propagation frequency of 200Hz, the important propagation paths have grazing angles of less than 5° , so that, over a range of 80km, each path interacts with the bottom between 50 and 100 times. If a 3dB fit between prediction and experiment were regarded as satisfactory, then the reflectivity of the bottom would have to be known to an accuracy of 0.03dB at all points along the 80km track. Since this is not feasible, the approach

recommended by Jensen and Kuperman is as follows:

- (a) obtain sound speed profiles in the water column
- (b) take core samples for determining the stratification in the immediate subbottom
- (c) delineate deeper layering by conventional seismic reflexion profiling
- (d) conduct a two-ship propagation experiment
- (e) attempt to fit with the best propagation model available.

This approach has allowed confidence in modelling propagation conditions for different seasons to be built up gradually on the basis of experience in known areas. The way in which near-bottom multichannel seismic surveying could help would be in adding considerably to the amount of information gathered in step (c), possibly to the extent of allowing less coring (b) to be required. Equipment designed for this purpose was described in chapter 2 and results are discussed in chapters 7 and 9.

6. VELOCITY FROM TRAVEL-TIME DATA.

6.0 INTRODUCTION.

An understanding of the response of fluid and solid earth materials to an applied acoustic field is the aim of seismology in the various forms in which that science is pursued, and a full discussion of the topic is beyond the scope of this present (or indeed any one) work. This chapter discusses briefly some of the approaches which have been taken to analyse observations of acoustic waves which have travelled through seabed materials. The association of anomalous values for compressional wave velocity with the seabed conditions discussed in chapters 3 and 4 and the importance of that velocity in predicting sound propagation characteristics in the water column (Ellis, 1981; and chapter 5) have already been discussed. Simple kinematic methods for its evaluation are discussed here.

6.1 TRAVEL TIME METHODS.

Cervený and Ravindra (1971) note that Merten, in 1927 first applied the Huyghens principle to obtain a description of the propagation of an acoustic wave across a plane interface dividing two media of differing propagation velocity and found that he could, under the appropriate relationships between refractive index and

incidence angle, obtain a description of the interface wave, variously called a lateral, conical or head-wave. This graphical approach has been used by Dix (1939) and Jardetsky (1952), but it remained for Brekhovskikh (1960 p.241) to give a more formal treatment of the behaviour of a spherical wavefront at such a plane interface and demonstrate convincingly that the processes occurring at that interface to produce the travel-time characteristics of the wave fields observed in exploration geophysics are a function of the geometrical disposition of source and receiver with respect to the interface and of the velocity contrast, but are quite independent of the mechanical characteristics of the media in question. All of the travel time methods start with this premise, and use various contrivances to invert the problem, that is, to derive the velocity structure of a layered space from the arrival time of acoustic signals.

The methods may be broadly classified as "travel time" or "differential travel time", according to whether measurement is made of an arrival time or of its gradient, a distinction which has more basis in practice than in theory. Field records whose time and distance origins are not well defined may still give useful results if receivers or shotpoints occur sufficiently densely to allow accurate measurement of travel-time gradients - the basis of the layer parameter method for dealing with wide

angle reflexion profiles (Bryan, 1974). On the other hand, if observations are available at only a very few different separations between source and receiver, then careful attention to the zero points of time and distance will still allow results to be obtained. An illustration of this latter approach to what is more normally dealt with by means of the travel time gradient method of a standard "refraction line" is given by McKay and McKay (1982), McKay (1983), and discussed further in section 7.3. Another illustration of this approach comes from the wide angle reflexion work of Bennell et al. (1982).

Most effort of refining travel time methods has been put into dealing with wide angle reflexion data. This approach to velocity determination was first described by Green (1938) who ignored the difference between least-time and least-distance propagation paths to give the " T^2-X^2 " method. Where velocity differences are small (of the order of 20%) Green's method remains a good one because errors which arise by the neglect of the 'least-time' principle are often smaller than the experimental error in measuring travel times. This is particularly important when the material studied is a surficial seafloor sediment, whose velocity usually lies within 20% of that of the bottom water, and where velocities beyond that range indicate unusual properties which may merit further investigation by the modelling procedures discussed later

in section 8.2 and in chapter 9. A useful criterion for deciding on whether Green's method is adequate in a given situation is to compare the experimental limits of the interval velocity V_2 derived from the velocity V_1 of an upper layer and V_a , the average in the two layers when determined from each of the two relationships:

$$V_2^2 = \{V_a^2 (T_1 + T_2) - V_1^2 T_1\} / T_2 \quad \dots\dots\dots (6.1)$$

$$V_2 = \{V_a (T_1 + T_2) - V_1 T_1\} / T_2 \quad \dots\dots\dots (6.2)$$

where T_1 and T_2 are the vertical-incidence travel times in the layers corresponding to V_1 and V_2 . Equation 6.1 derives from a method of Dix (1955), whereas 6.2 arises from the simple $T^2 - X^2$ approach. If there is no significant difference in the result obtained, then there is probably little to be gained by adopting other approaches, although a number of these are available.

The use of the wide angle reflexion method at sea was originally discussed by Katz and Ewing (1956). Houtz and Ewing (1963,1964), Knox (1965) and Clay and Rona (1965) also observed wide angle reflexions in the deep ocean before Le Pichon et al. (1968) published the method for the routine gathering of this kind of information by use of disposable sonobuoys. Houtz et al. (1968) discussed the application of this technique to the sediments overlying the basalts of the N. Atlantic sea floor, and later papers have given results of this kind from many parts of the oceans e.g. Ludwig et al. (1975) in the

Columbia and Venezuela basins of the Caribbean, Houtz (1980) from the Barents Sea, Bachman and Hamilton (1980) in the Arabian Sea. Houtz (1977) extended the depth capability of the technique to observe mantle reflexions in the East Pacific. Although this and other deep sea wide angle reflexion work, particularly that carried out in conjunction with the Deep Sea Drilling Project has over the past 15 years given a great deal of information on the evolution of the ocean basins, much less direct information has accrued about the first few tens of metres of the sediment column immediately beneath the seabed. This is the zone which has the greatest influence on the propagation of sound in the water column (discussed in section 5.4), and is also of most significance in ocean engineering for the assessment of foundation properties (chapter 1). Hamilton (1979a) has derived velocities in the near-seabed sediments by the upward continuation of gradients recognised by measurements in deeper layers, but notes that such estimates are only extrapolations. One approach of extending the travel-time techniques to the problem of the velocity structure of the immediate subbottom has been taken by Bryan (1980) who allowed a pinger to drag along the seabed away from a hydrophone, also on the seabed, the whole assembly being dangled from a standard hydrographic STD cable. This method removes the major problem of the large error in estimating the interval velocity in a thin layer when that layer lies

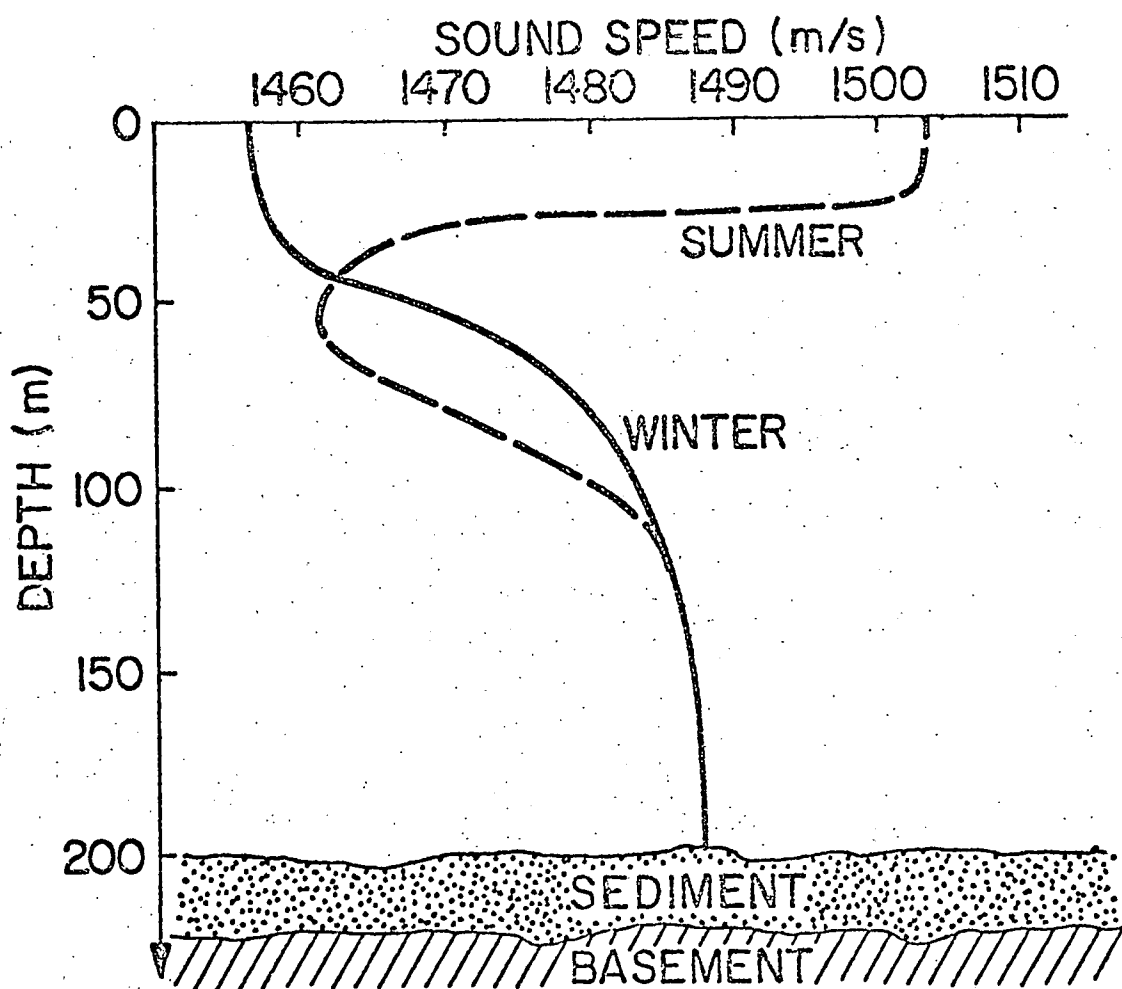


Fig. 6.1

Profile of the changes in the compressional velocity of sound in water with depth for a location on the Scotian Shelf. (Data of the Defence Research Establishment, Canada). The stable iso-velocity zone near the seabed can be contrasted with a more complex structure in the surface layers.

beneath a deep water column (as is evident from eqn. 6.1 or 6.2), but introduces the serious logistical problem of having to hold station in order to make the necessary observations. McKay (1975) showed that hydrophones could be towed conveniently at different offset distances from existing deep-towed profiling equipment, thus opening up the possibility of maintaining almost as favourable a ratio between water column and sediment thickness as that achieved by Bryan (1980) while retaining the logistical advantage of a continuous profiling method, previously used in shallow water by Porter et al. (1974) and refined by Bennell (1977,1978) in that same environment. The use of equipment near the seabed yields data suitable for reduction by the T^2-X^2 method (with a simple but important modification discussed in section 7.2) for two reasons:

- (i) the effective water column is small
- (ii) sea bottom water is usually free of the more complex velocity structure often found in near-surface layers (fig. 6.1).

Limitations result mainly from the uncertainty in phase shifts which occur on reflexion at an interface. The magnitude of these is discussed in McKay and McKay (1982) and are such that the precision of ca. 0.3% sought by Bennell (1978) in connexion with attempts to define velocity gradients is in principle unattainable from travel-time methods, without consideration of a mechanical model on which to base estimates of phase shift.

More precise travel time methods have been evolved for use in crustal problems by means of ray tracing (e.g. Cerveny et al. 1977). These avoid approximations inherent in the techniques discussed above and can be used to treat problems involving more complex earth structure. Field data of a quality which justifies this approach is beginning to be gathered (chapter 9), but it may prove more rewarding to treat it by means of constructing synthetic seismogrammes based on some appropriate mechanical model for the sediments concerned (chapter 8), rather than by deducing the kinetic properties alone. Although these more precise travel-time methods must be borne in mind, the earlier simple methods discussed above can be used successfully to add substantially to such knowledge of sediments as is currently deduced from the type of acoustic observation (a vertical incidence profile) regarded as acceptable in present-day research studies of the seabed (e.g. Piper and Normark, 1983), particularly in areas where marked changes in compressional wave velocity can occur, as in the environments discussed in chapters 3 and 4.

Such simple methods are used in dealing with the results described in chapter 7. Until such time as equipment (section 2.5) becomes more readily available for gathering information suitable for more elaborate treatment (chapter 8), these simple methods will remain of importance.

7. FIELD RESULTS.

7.0 INTRODUCTION.

Field results from various locations around coastal regions of the Canadian Maritime Provinces from Lake Erie and from the Scotian Shelf illustrate the ways in which information can be gathered in a form suitable for treatment by kinematic methods for the determination of acoustic velocities in the subbottom materials. Although current methods of handling seismic data (see section 8.2) suggest a more unified approach, there is still some practical convenience in classifying kinematic methods as 'refraction' and 'wide angle reflexion' depending on whether equipment is deployed with the geometrical relationships between acoustic source, receivers and layer interfaces such as to allow the observation of head waves as first arrivals or not. The acoustic sources used have been boomers and sparkers.

The coalfield work at Mabou (section 7.1) gives a good example of the way of measuring velocity by the variable-length hydrophone method using equipment described in section 2.3. An extension of this approach in the deeper water encountered in the Sydney Basin (section 7.2) required that equipment be deployed near the seabed if headwaves were to be observed as first arrivals. The differences in geometry and consequent appearance of field records and data reduction technique which results

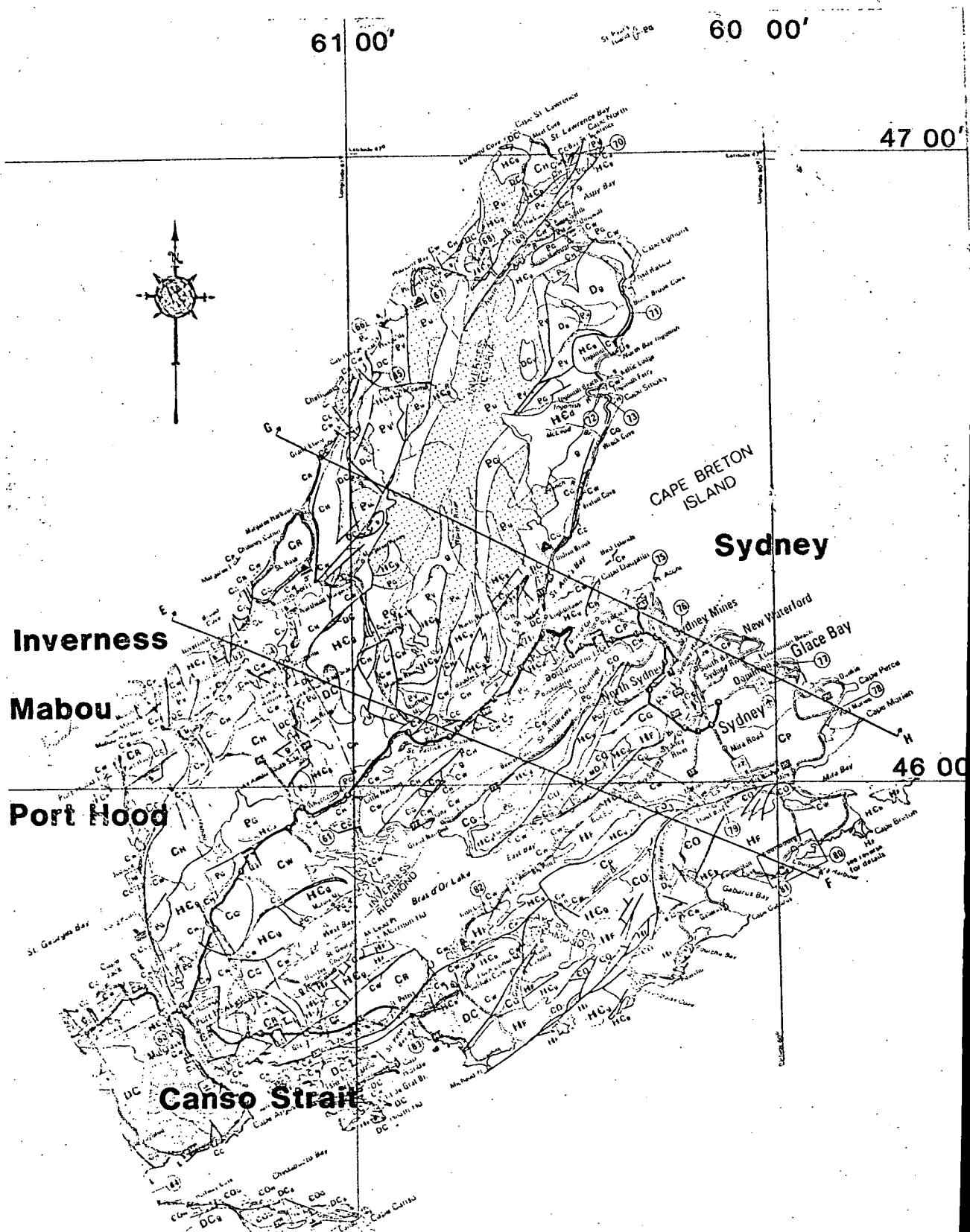


Fig. 7.1

Geology map of Cape Breton Island showing the location of sedimentary basins of Carboniferous age. (Bujak and Donohoe, 1980).

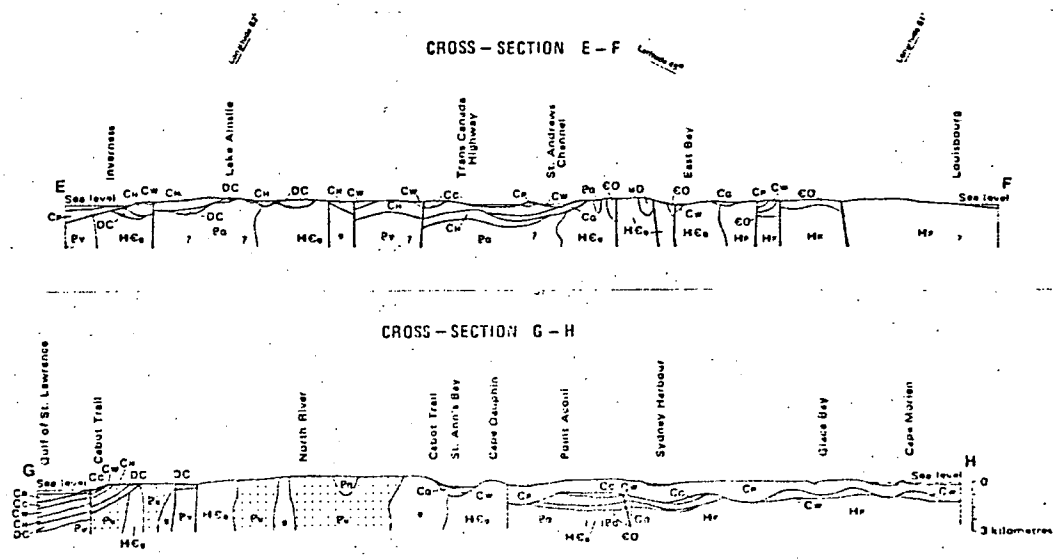
from this approach is described in that section. Success with the technique employed at Sydney suggested that a similar but modified approach could be used on fieldwork in the Beaufort Sea (section 7.3) to allow the observation of headwaves as first arrivals without disturbing the basic operation of a system for gathering vertical-incidence reflexion profiles.

The wide-angle reflexion technique is illustrated by results from Lake Erie where a hydrophone of variable tow-length was deployed at the surface (section 7.4). The results of attempts to extend this method to greater depths by the adaptation of a deep tow seismic profiler (section 2.1) are described in section 7.5, with reference to areas of the Scotian Shelf.

7.1 MABOU COALFIELD - SEA SURFACE TECHNIQUE.

A resurgence in interest of the coal deposits of Nova Scotia in the mid 1970's led to the surveying by seismic methods of a number of offshore coalfields. The general location of basins in Cape Breton Island is shown in fig.7.1 and the Sydney and Inverness basins are seen in section in fig. 7.1b. At Mabou, the coal measures are exposed at Findlay Point and Coalmine Point (fig. 7.2) but

Geological Map and Table of Formations in the St. John's Region						
AGE/ERA	PERIOD	EPOCH/OROGENY	SOUTHERN MAINLAND	NORTHERN MAINLAND	CAPE BRETON ISLAND	INTRUSIVE ROCKS
MESOZOIC	CRETACEOUS	Early		K Kensington Group (sandstone, shale, lignite)		
	JURASSIC	Early		J ST. JOHN'S BAY & MUDY BAY FORMATIONS (sandstone, shale, limestone, chert)		
	TRIASSIC	Late		TJ ST. JOHN'S BAY & MUDY BAY FORMATIONS (sandstone, shale, limestone, chert)		
	PERMIAN	HEACYNIAN		P REDMOND & HOLLYHILL FORMATIONS (sandstone, shale, limestone, chert)		
PALEOZOIC	CARBONIFEROUS	Early		CP PULLEY, STELLANTON & MURPHY GROUPS (sandstone, shale, limestone, chert)		
		Mid		CC CANNON GROUP (sandstone, shale, limestone, chert)		
		Mid		CC CANNON GROUP (sandstone, shale, limestone, chert)		
		Mid		CW WINDWARD GROUP (sandstone, shale, limestone, chert)		
		Early		CC CANNON GROUP (sandstone, shale, limestone, chert)		
	DEVONIAN	Early		CC CANNON GROUP (sandstone, shale, limestone, chert)		
		Mid	ACADIAN	DC DORSET GROUP (sandstone, shale, limestone, chert)		
		Mid		DC DORSET GROUP (sandstone, shale, limestone, chert)		
		Mid		DC DORSET GROUP (sandstone, shale, limestone, chert)		
		Early		DC DORSET GROUP (sandstone, shale, limestone, chert)		
PROTEROZOIC	CAMBRIAN	Early	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Early	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
	HADRINIAN OR OLDER	Early	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Mid	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		
		Early	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)	OS TERRAPINE FORMATION (sandstone, shale, limestone, chert)		



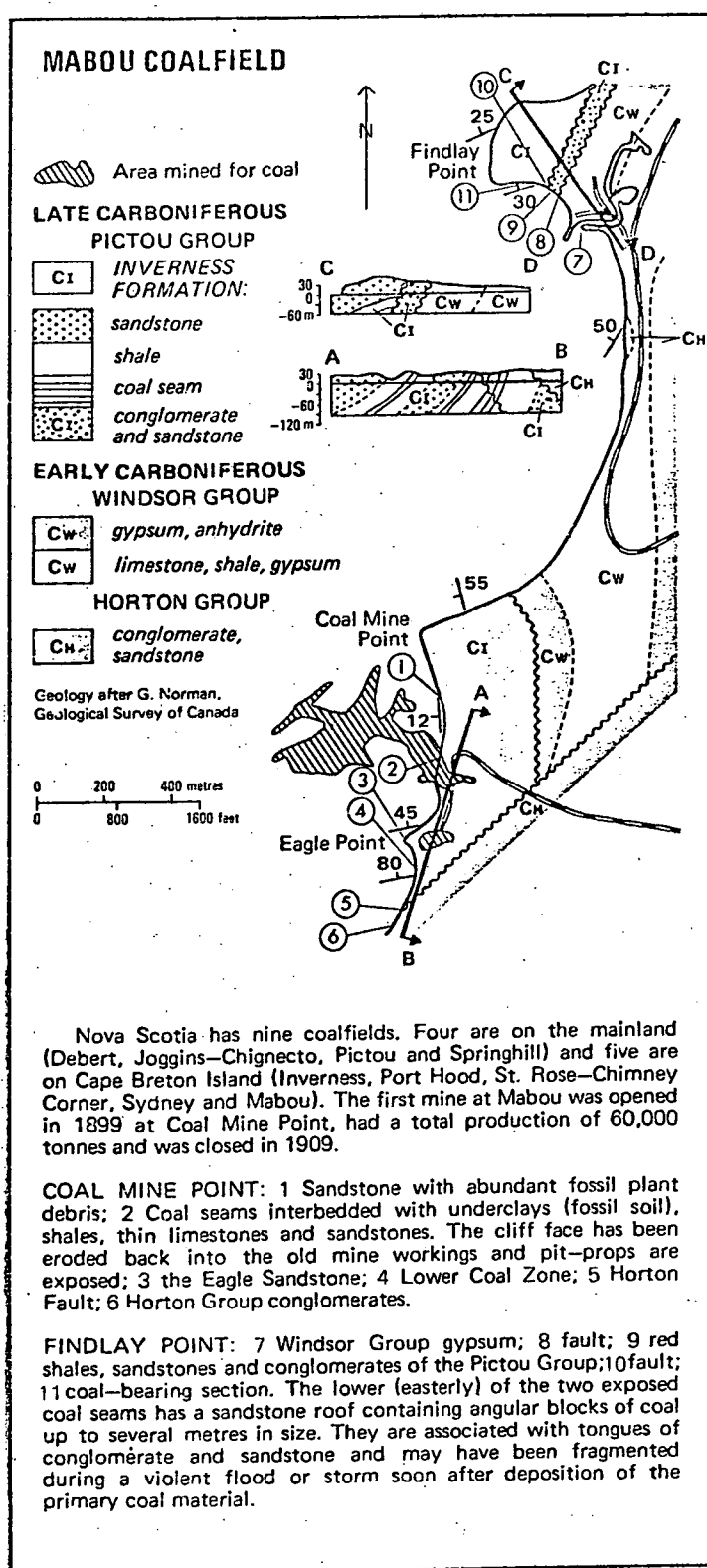


Fig. 7.2

On-shore geology of the Mabou Coalfield. (Bujak and Donohoe, 1980).

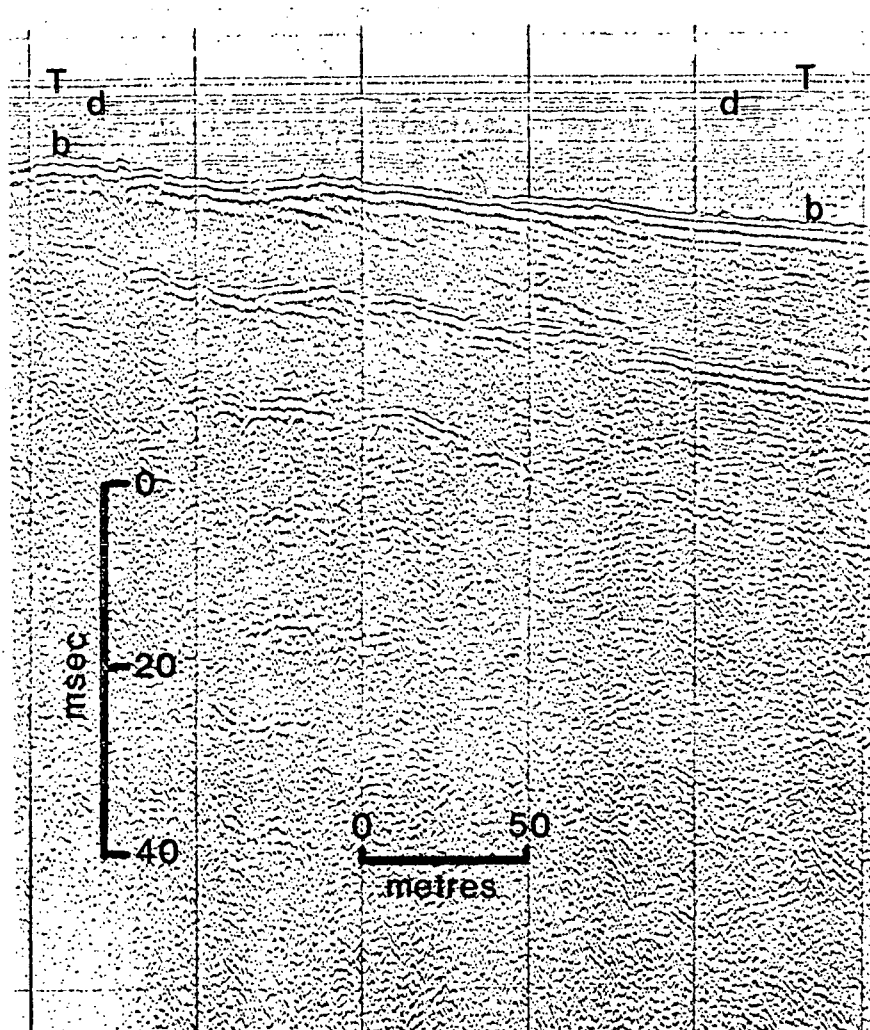


Fig. 7.3

Near-shore section made with multi-tip sparker and single-channel hydrophone.

- T - trigger event
- d - direct wave arrival
- b - bottom reflection

most of the coalfield lies under the sea. A part of it had been mined during the late 19th century, but the workings were abandoned in 1909 when they flooded. A reconnaissance survey was done with a multitip sparker and single channel receiver to try to define the offshore rock structure by looking at the dip of subbottom reflectors such as that which can be seen clearly in fig. 7.3, (Mckay, 1977). A suitable grid of lines allows true strike and dip to be calculated at line intersections (fig. 7.4) provided the lines do not intersect at too acute an angle and provided a value for the velocity of sound in the rock is known.

Maries and Beckmann (1961) described a method of gathering data suitable for interpretation as an unreversed refraction line by allowing a hydrophone on a long wire to be payed out astern of a boat carrying a pulsed sparker source. Howells and McKay (1977) appear to be the first to have pointed out that a simple logistical modification of this technique can lead to the speedy gathering of data suitable for interpretation as a reversed refraction line. This modification merely requires that as soon as the trailing hydrophone is near to reaching the extent of its travel, the boat is stopped by a quick burst of full-astern. The hydrophone cable is then hauled in smoothly at a rate as close to that at which it ran out, as nearly as can be conveniently judged.

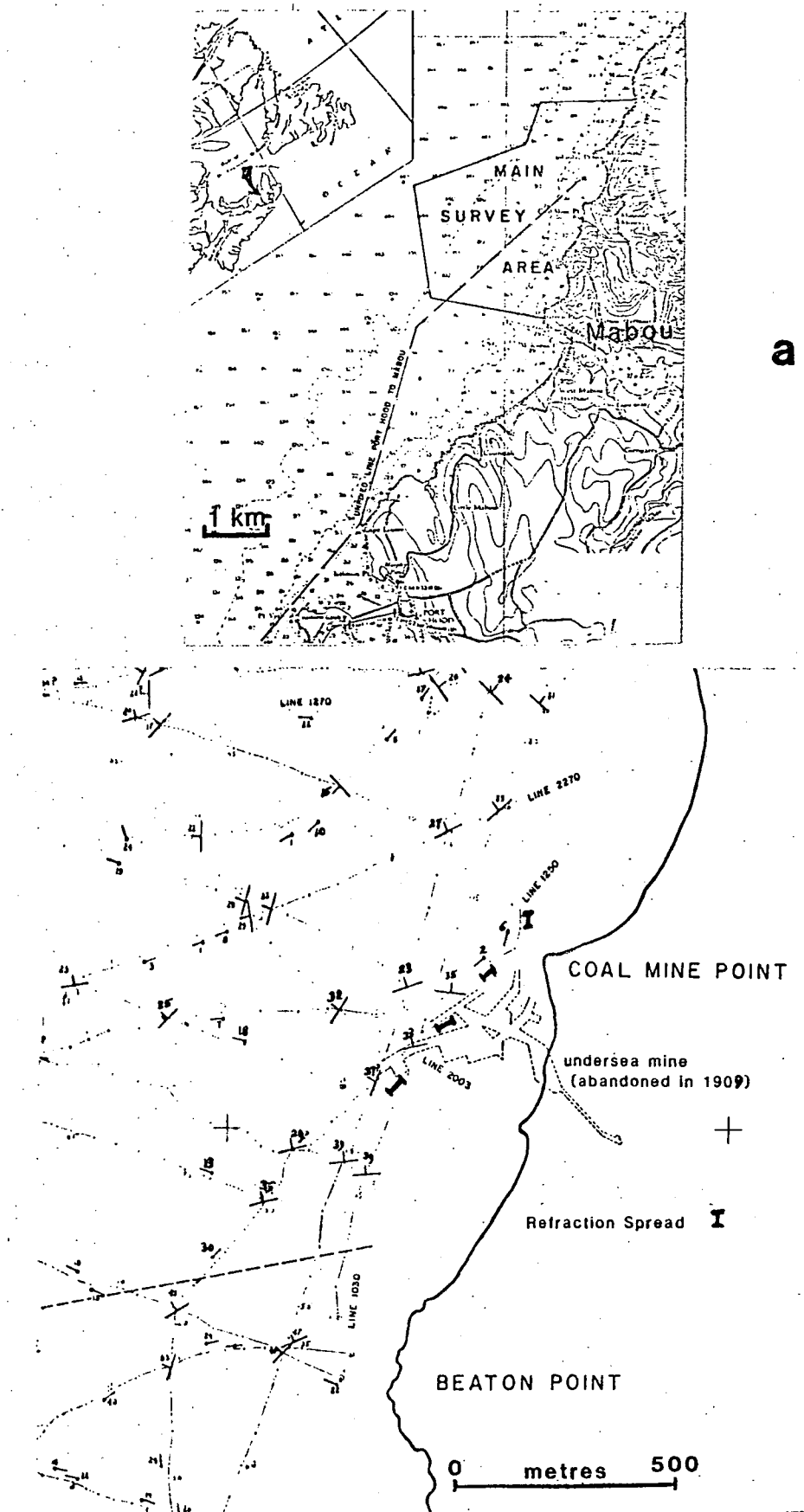


Fig. 7.4

a. Location map of survey area at Mabou.

b. Part of track plot of survey showing apparent dips along survey lines and true strike and dip at line intersections, as obtained from the subbottom profiles.

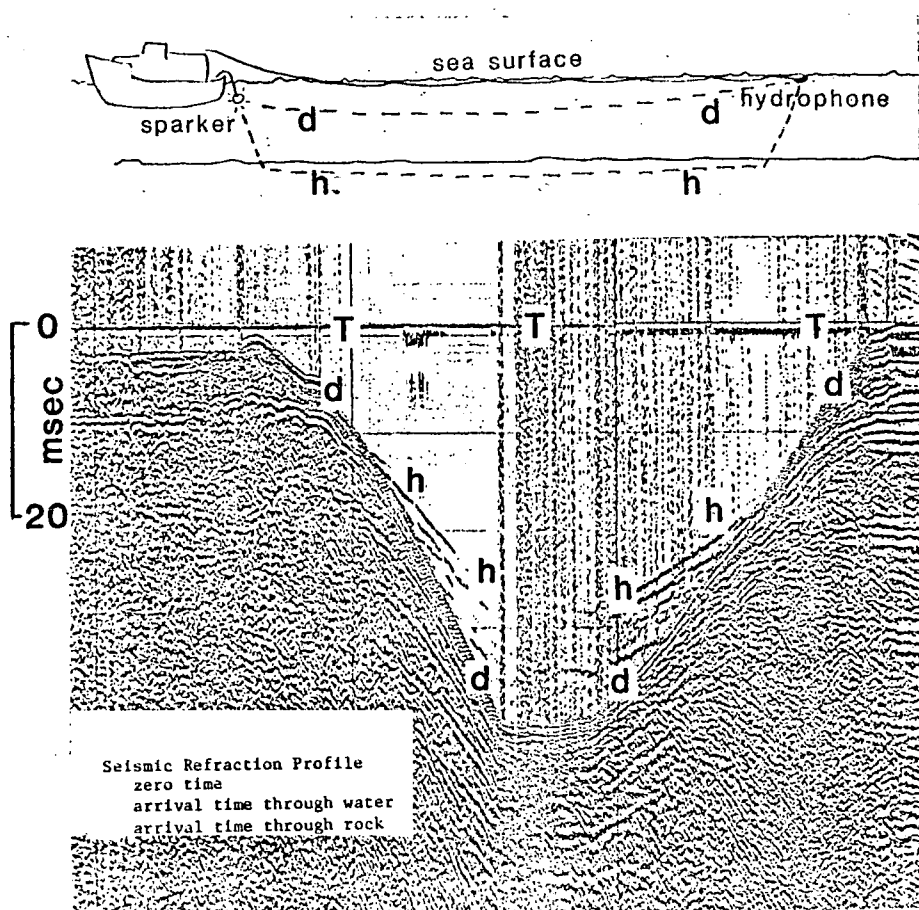


Fig. 7.5

Refraction profile.

T - trigger event
 d - direct arrival
 h - headwave

One example of such a reversed profile is given in fig.8 of Howells and McKay (1977), although some difficulty was experienced with the boat's drifting during the recovery of the hydrophone while such data was gathered as part of the investigations in the Miramichi Estuary. An example of a reversed profile made in favourably calm sea conditions (sea state ca. 1-2) is shown in fig. 7.5. This data was gathered in water about 7 metres deep in the vicinity of the site of the abandoned subsea coalmine (fig. 7.4). Travel times are plotted in figure 7.6.

A conventional graphical analysis of this reversed profile yields gradients of the slope of headwave arrival times as:

$$\text{down-dip slope, } m_d = 0.565 \pm 0.017$$

$$\text{up-dip slope, } m_u = 0.465 \pm 0.015$$

This gives the critical angle of refraction as:

$$\begin{aligned} i_c &= (\sin^{-1} m_d + \sin^{-1} m_u) / 2 \\ &= 31.05^\circ \pm 1.08^\circ \end{aligned}$$

With the seawater velocity found to be 1.460 ± 0.003 km/sec. by measurement with a calibration frame used in conjunction with an Atlas Deso-10 echo sounder transducer, this leads to the acoustic velocity in the lower refracting medium as:

$$v = 2.83 \pm 0.085 \text{ km/sec.}$$

The slope of the interface is given by

$$\alpha = (\sin^{-1} m_d - \sin^{-1} m_u)$$

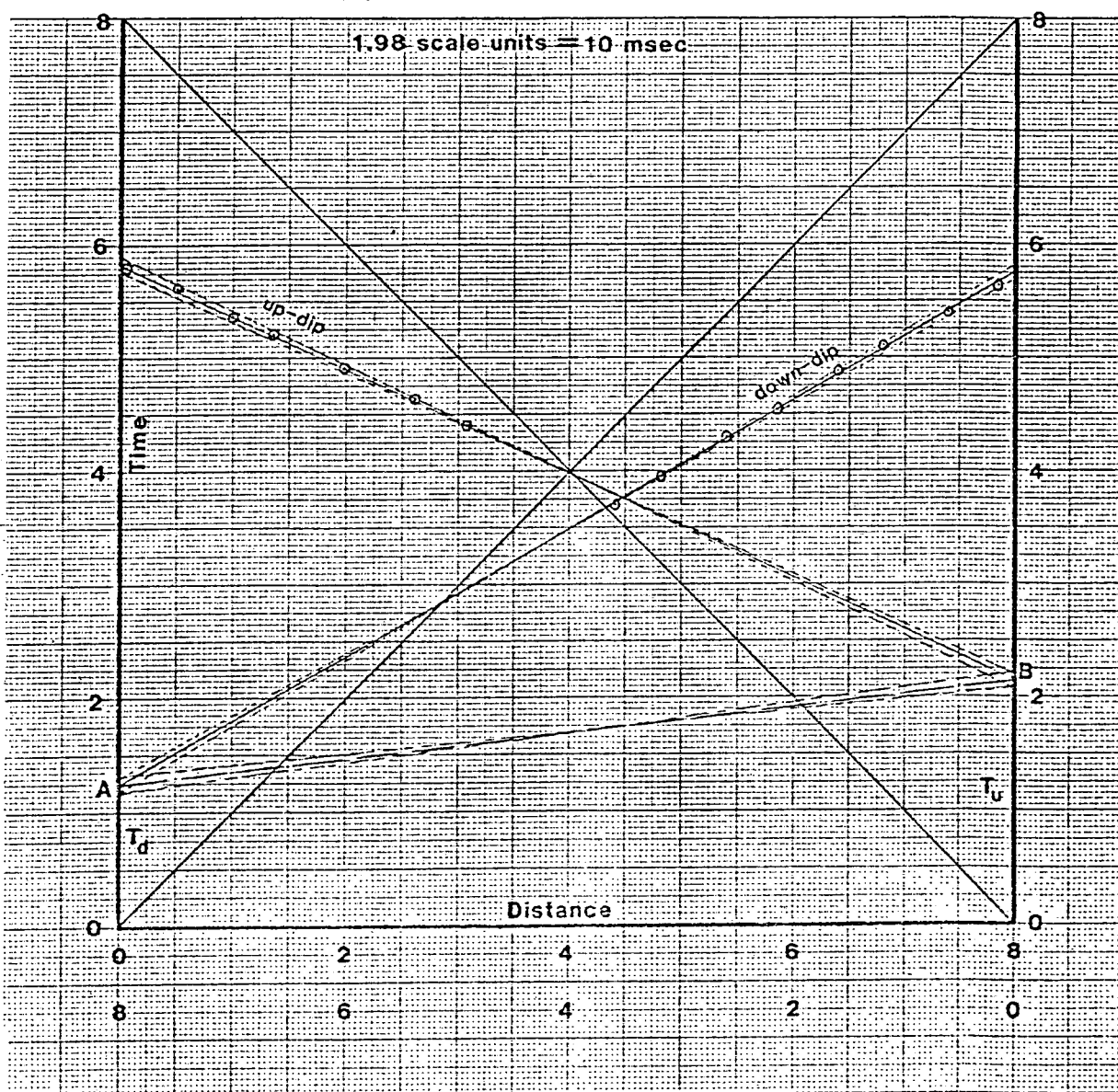


Fig. 7.6

A plot of headwave arrival times for both paying out and hauling in the hydrophone cable.

$$\alpha = 3.34^\circ \pm 2.17^\circ$$

which corresponds to a slope of the refracting interface of 0.059 ± 0.037 .

As a verification that the up-dip and down-dip sections of the profile do refer to the same part of the seabed, it is useful to compare the gradient of the refracting interface which is derived from the difference in the intercept times T_u and T_d . It is found that the difference in depth of that interface is:

$$V_{\text{water}} \cdot (T_u - T_d) / 2 \cos i_c = 3.99 \pm 0.64 \text{ metres}$$

By considering the maximum travel time of the direct water wave, the separation between the points where T_u and T_d are determined is found to be $58.9 \pm 0.3 \text{ m}$. This leads to a slope of the refracting interface of 0.068 ± 0.12 , which is in agreement with what was found (above) from the up-dip and down-dip slopes, within the limits of the experimental error.

Some confirmation that this value for the velocity is valid comes from plans of the old mine workings. These have been examined by P. Hacquebard (Geological Survey of Canada) who is of the opinion (personal communication) that the results of the seismic study at Mabou are in agreement with the old mine plans, when dips are calculated on the basis of the velocity value given above. The extent to which this observation is of worth, however,

is limited by the relatively small number of dip angles determined over the old mine area and also by the reliability of the old plans, given that the mine eventually closed in curious circumstances by supposed accidental flooding, but with no loss of life because, unusually, no-one was underground at the time.

Although the technique used at Mabou and at Miramichi was essentially the same, a comparison of fig. 7.5 with fig. 8 of Howells and McKay (1977) will show that the frequency content of the acoustic source is different, being about 500Hz at Miramichi and 1.3KHz at Mabou, as observed from the appearance of seabed reflexions. This difference comes not from a change in the voltage or energy storage of the driving high voltage source which were the same in each case (8000 volts, 400 Joules), but from the design of the sparker transducer. The lower frequency was output by a single-electrode sparker whereas the higher frequency came from a 48-electrode sparker which was 3.5m long and was deployed with its long axis athwart the stern of the boat. This technique is useful for raising the frequency of the acoustic source and providing some degree of directionality to its output to offset in part the higher attenuation experienced at higher frequency. It is naturally important that such a source should remain at right angles to the line between it and the long-tow hydrophone to avoid having different

path lengths between the hydrophone and different parts of the acoustic source. With a long source rigged athwart the stern, there is some risk of this being turned to different angles while the boat is stopped and the hydrophone is being pulled in. A careful comparison of the left and right parts of fig. 7.5 indicates a small difference in source signature which is probably attributable to this effect. For this reason, it may often be logistically simpler to use a point source for work which requires that the towing boat be stopped. For techniques where a steady uninterrupted tow speed is used (to be discussed later in sections 7.3 to 7.6) and stable tow conditions can be realised, an athwart-towed line source of the type described above may be preferable.

7.2 SYDNEY COALFIELD - NEAR-SEABED TECHNIQUE

(DISCONTINUOUS).

In principle, all that is required to extend the technique of section 7.1 to deeper water is to make provision for greater separation between the source and receiver, and to use a lower acoustic frequency so that attenuation processes do not reduce the transmitted signal to unobservably low levels before reaching the offset distances required for the appearance of head waves as first arrivals. There are two reasons why it is improper

to pursue this approach. The first concerns the desirability of siting any experiment aimed at elucidating a subbottom velocity profile over an area where structural interfaces are simple - preferably plane horizontal surfaces. The larger the area which is required to satisfy this condition, the less chance there is of finding one which is suitable. The second reason is one of logistics. The conventional equipment used for investigating the geological features of the near-subbottom comprises pingers, boomers and low-energy sparkers. It is convenient to be able to use these same acoustic sources for velocity determination. The need to mobilise lower frequency sources such as airguns or high energy sparkers for the purpose of velocity determination by headwave observations would be costly, and at times when small boats are in use as survey vessels, impossible. The receiving and recording electronic equipment would not necessarily be appropriate for lower frequency work either. The alternative is to keep the scale of the spacing between the source and receiver similar to that used for the shallow-water (ca. 10 metres deep) work described in section 7.1, by arranging for source and receiver to be deployed deep in the water column. This approach was first suggested (McKay, 1975) and implemented (McKay, 1978) for obtaining data suitable for reduction by the wide-angle reflexion travel-time techniques (section 7.5). In this section, the extension of the refraction

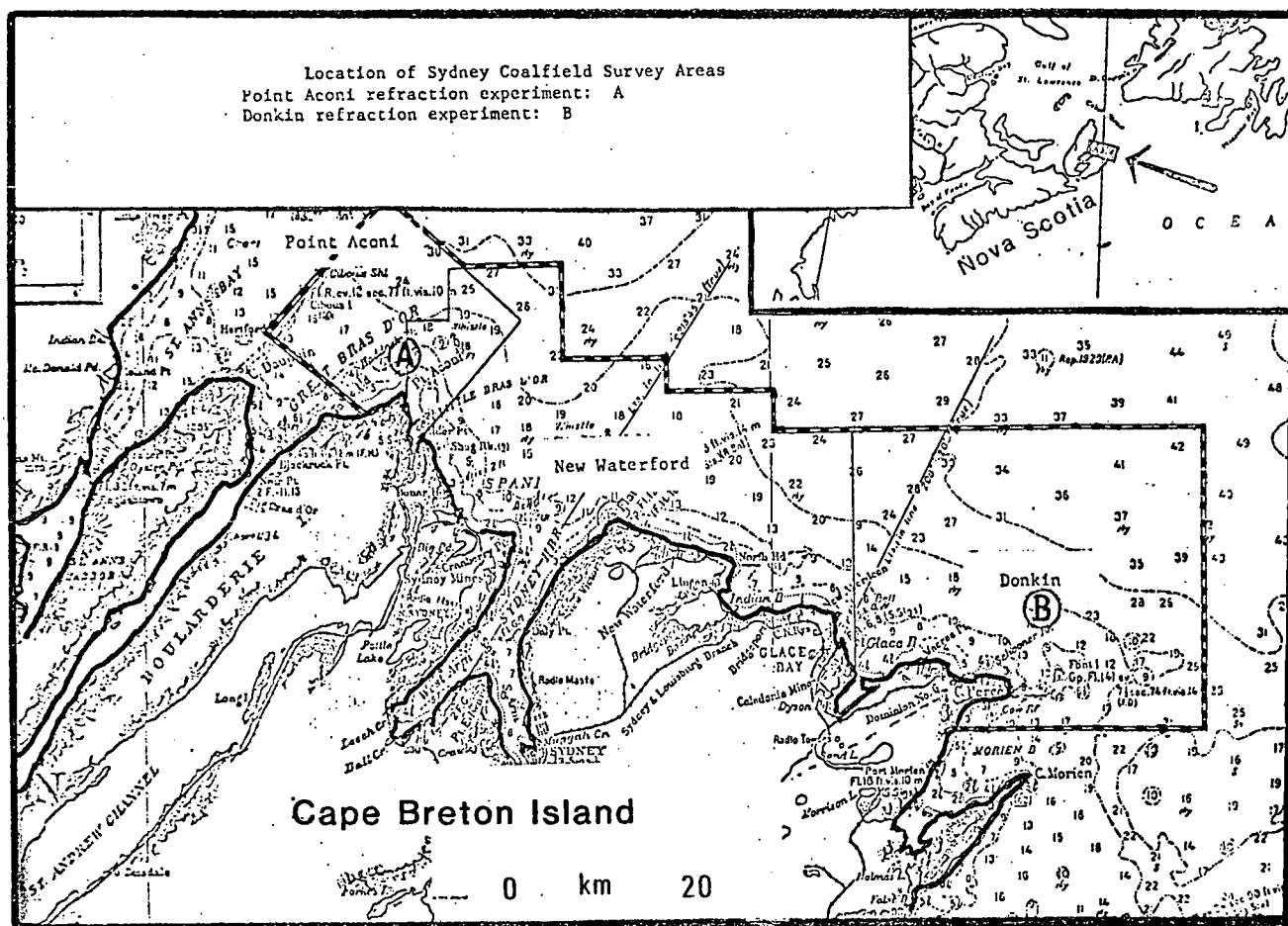


Fig. 7.7

Location map of offshore survey of Sydney coalfield. Positions of refraction experiments are at "A" and "B".

technique of 7.1 to deeper water is considered in relation to investigations carried out over another coalfield off the shore of Cape Breton Island, that of the Sydney Basin (figs. 7.1, 7.1b).

The area investigated in the Sydney Basin is shown in fig 7.7. The basic survey equipment was similar to that used at Mabou, and comprised a Teledyne high voltage source fired 2 or 4 times per second at 200Joules into a Teledyne multi-tip spark array trailed line-astern as the acoustic transmitter, while the receiver was a 7 metre long single channel hydrophone array. Variable density display of data on an E.P.C. 4600 recorder resulted in profiles of the type shown in fig. 7.8. Over 1000km of lines were run with a 40 ft. survey boat on a square grid of 1.5km spacing, navigated by means of a Del-Norte Instruments "Trisponder" radar-frequency navigation system with ancillary equipment giving the facility to plot the boat's position every 10 seconds in real time.

The survey results generally confirmed notions about the structure of the basin based on on-shore geology and information from currently operating sub-sea mines, namely that of a sequence of regularly dipping beds with an E-W strike and dip angle varying from about 10° in the east of the area to 2° or 3° in the west. The cusped syncline and anticline structure of fig. 7.8b a few km. off Donkin

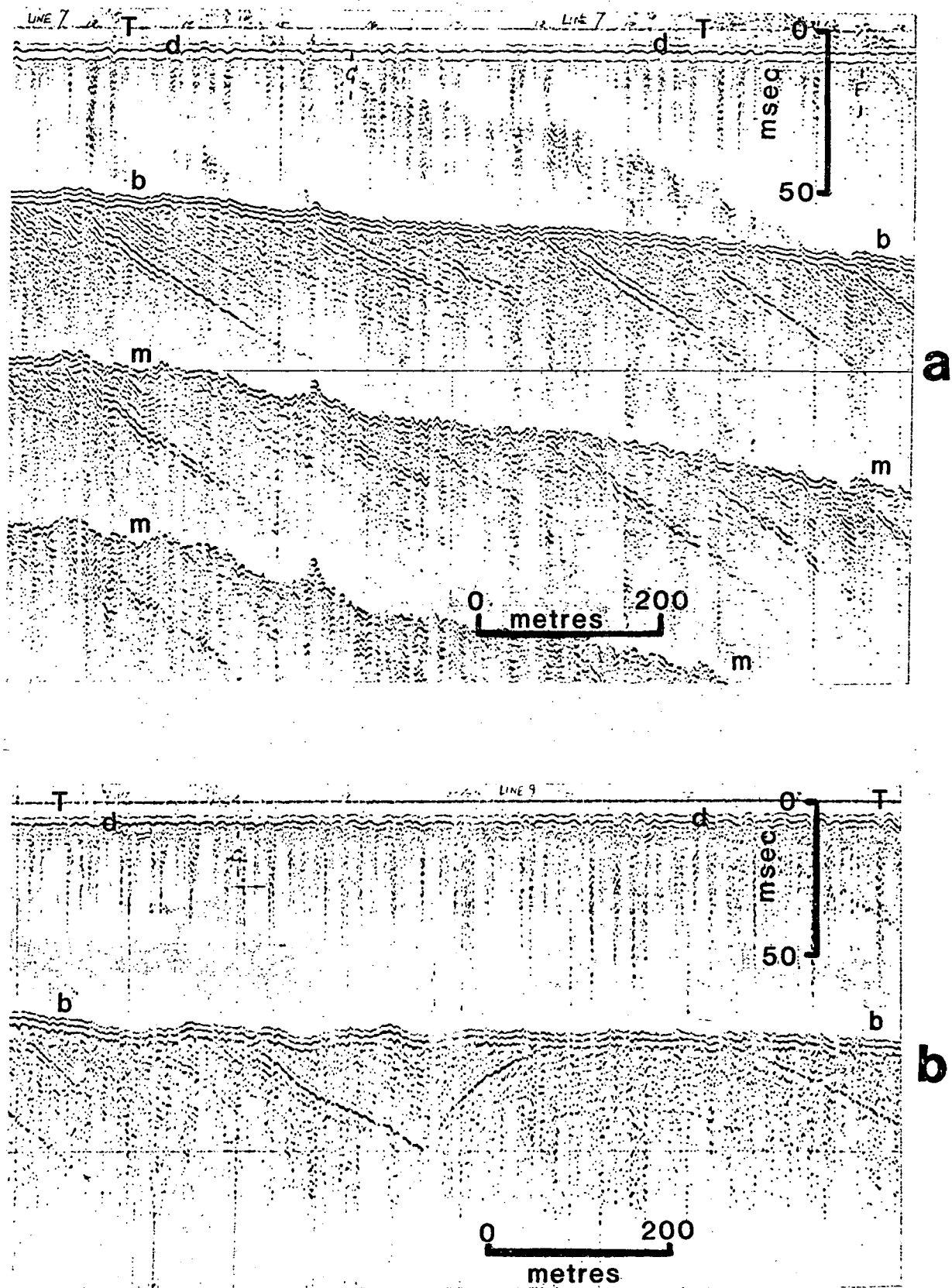
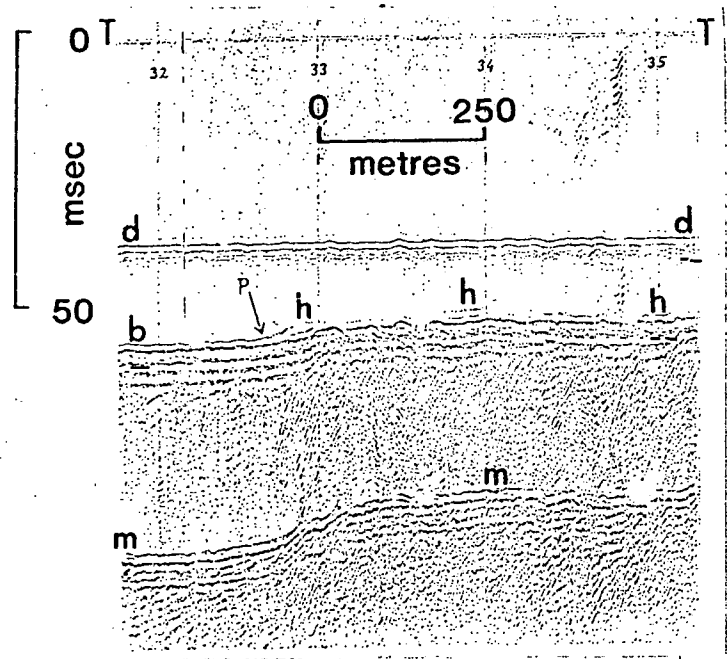


Fig. 7.8

Representative subbottom profiles over the eastern (Donkin) area of the Sydney coalfield.

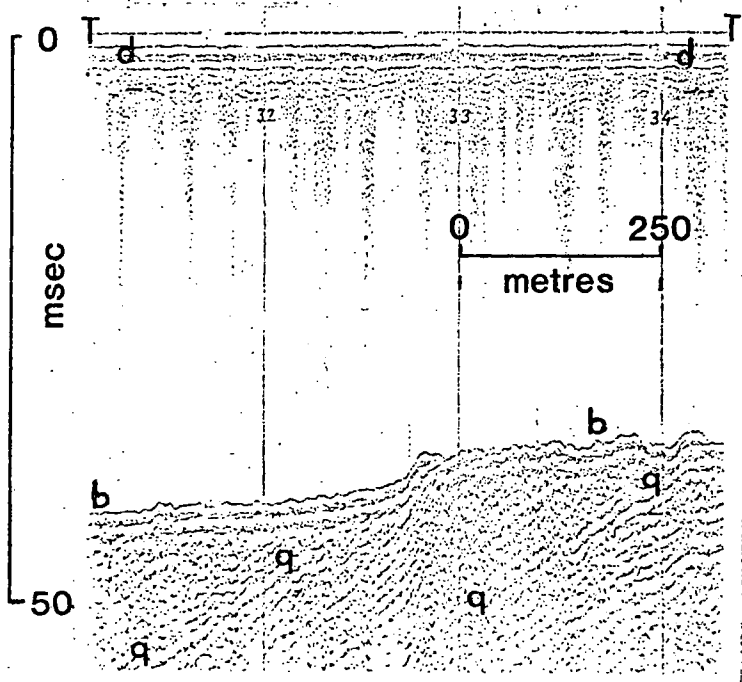
a. Evenly dipping beds (angle ca. 8°)

b. Cuspate syncline and anticline structures, which were important in planning a new mine in this area.



Refracted events observed on the signal from a short array trailed 100 m astern. Refraction and bottom reflection times coincide at point P.

T Trigger event	d Direct arrival
h Refraction	b Bottom reflection
m First multiple of bottom reflection	



Vertical incidence record corresponding

T Trigger event	d Direct arrival
b Seabed reflection	q Subbottom rock structure

Fig. 7.9

Recognition of a headwave coincident in time with the seabed reflecti

was not foreseen from the land geology and shed light on the results of boreholes sunk in that area (neighbourhood of 'B', fig.7.7). These boreholes had been sited without seismic control on the assumption of the continuity of on-shore trends.

Water depths over the survey area were mostly between 20 and 50 metres. Although water 20 metres deep still offers some possibility of using the technique employed at Mabou, greater depths make the method impracticable for the reasons given above. A related surface method was successful in a water depth of ca. 40 metres, namely that of identifying the coincidence in travel time between a headwave and the seabed reflexion where both of these occur sometime after the direct water wave arrival. This is illustrated in fig. 7.9. The profile of fig. 7.9a results from the trailing of a second 7 metre long hydrophone array 55 metres astern of the position of the acoustic source and the vertical-incidence hydrophone array which gave the profile of fig. 7.9b. A measurement at the point P (where bottom reflexion b and headwave h coincide) of the direct water wave and bottom reflected travel times defines the critical angle, from which the acoustic velocity in the subbottom material can be derived. It is found that

$$\theta_c = \sin^{-1} (0.687 \pm 0.062)$$

which gives

$$v_p = 2.17 \pm 0.20 \text{ km/sec.}$$

The main source of error is in deciding whether the refracting interface really lies at the seabed or some little way beneath it. Fig. 7.9b shows a seabed which changes in nature from right to left, the change occurring relatively abruptly at the centre. The section to the right has the appearance of rock outcropping at the seabed, whereas the part to the left seems to be smoother, indicating a veneer of sediment lying over the rock. The observed headwave (fig. 7.9a) arrival parallels the rock surface and there is no reason to suppose that it is not associated with the rock. However, at the point where seabed reflexion and headwave are coincident in travel time, the rock interface appears to be below the seabed. The length of the signature of the seabed reflexion as received by the 7 metre array gives rise to uncertainty of 4 to 5 msec. as to how far below the seabed the interface lies. An improvement on this technique would be to lay out a trailing hydrophone or short array so that the relative angular relationships were about the same as those obtained with the deployment which gave rise to fig. 7.9 and then slowly move the position of this trailing hydrophone so that many more transition points such as that at P could be observed. If these occurred at points where, on the vertical incidence profile, the refracting interface could be identified more precisely than in the example presented here, greater precision

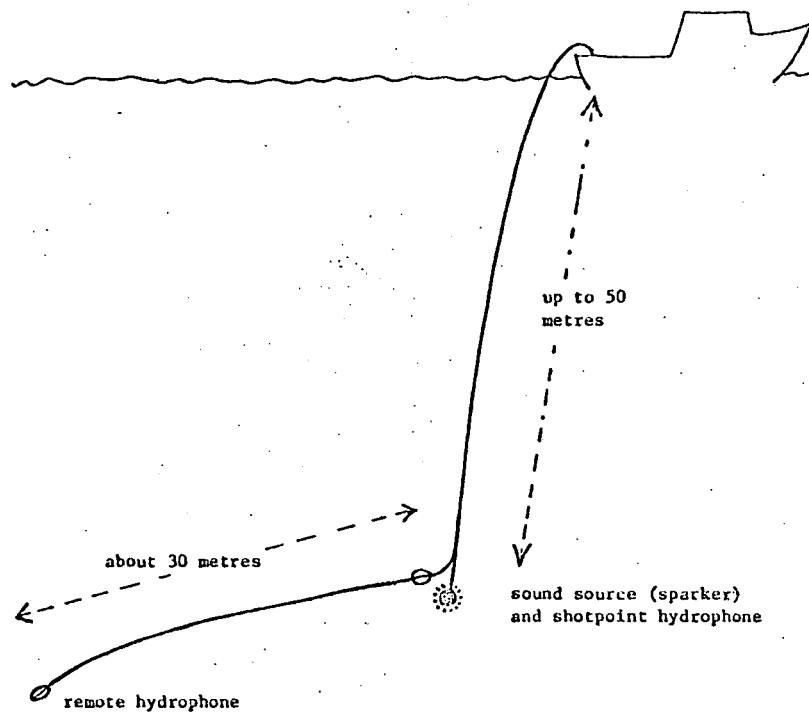


Diagram of the equipment layout used for the making of refraction profiles.

Fig. 7.10

Equipment deployment for refraction experiments.

would be possible in giving a value for the acoustic velocity in the seabed material. As yet, the opportunity has not arisen for putting such a method to the test.

The observation of headwaves as first arrivals during the Sydney coalfield survey required the devising of a new equipment deployment which would enable a sparker and hydrophone to be placed near the seabed. A modification of equipment of the type described in section 2.3 and used for headwave observations at Mabou was appropriate, the main alteration being the addition of weights to take it deep in the water column. This deployment is shown in fig. 7.10, and its use is discussed in McKay and McKay (1982, p.875). In its first configuration it used a single-electrode sparker with standard N.S. Research Foundation hydrophones, one immediately adjacent to the sparker and one trailed 46 metres astern of it. A 15kg weight was attached at the sparker and the whole assembly was lowered in the water column by first deploying it astern and then slowing the boat to near a standstill so that source and receivers fell in the water column until they were near the seabed. The importance of the hydrophone adjacent to the acoustic source lies in that it is the only way of determining where in the water column the acoustic source lies. By observing reflexions from either the seabed or the sea surface, the position of the sparker can be accurately located. The providing for a

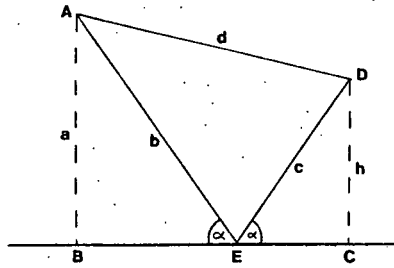


Diagram to show the geometrical relationships from which Eq.7.1) is derived. Upper-case letters denote points, lower case denote the length of straight lines. Physically, all lines shown are in the water column, with the exception of the horizontal lines through E which can represent either the sea surface or the seabed.

The acoustic source is at A and the receiving hydrophone is at D. We know the distance a from the vertical incidence reflection time, the separation d of the source and hydrophone, and the total distance $(b+c)$ along the reflected wave path. We require the distance h in order to locate the receiving hydrophone. From $\triangle DAE$

$$d^2 = b^2 + c^2 - 2bc \cos 2\alpha,$$

$$d^2 = (b+c)^2 - 2bc(1 - \cos 2\alpha).$$

But

$$1 - \cos 2\alpha = 2 \sin^2 \alpha,$$

$$d^2 = (b+c)^2 - 4bc \sin^2 \alpha,$$

$$\sin \alpha = a/b,$$

$$d^2 = (b+c)^2 - 4a^2 c/b,$$

since $\triangle ABE$ is similar to $\triangle ECD$,

$$c/b = h/a,$$

$$d^2 = (b+c)^2 - 4ah,$$

hence

$$h = [(b+c)^2 - d^2]/4a.$$

Fig. 7.11

Derivation of triangulation method for locating a trailing hydrophone (McKay and McKay, 1982).

means of locating the position of the acoustic source at the instant of firing is a fundamental pre-requisite for sub-surface acoustic work and a shotpoint hydrophone is the simplest method of doing this. When the position of the source is known, the position of other receivers can be determined by the modified triangulation procedure derived by P.M.McKay (personal communication, 1975), quoted by McKay (1975) and explained in McKay and McKay (1982). This derivation is reproduced in fig. 7.11. The relationship which allows easy calculation of what are essentially 'static corrections' in the language of land seismics is:

$$h = \{(b + c)^2 - d^2\}/4a \quad \dots\dots\dots (7.1)$$

where the parameters are defined in fig. 7.11.

The importance of this expression lies in its simplicity. Bennell (personal communication, 1979) has noted other ways of triangulating the position of an acoustic receiver which involve much more cumbersome expressions (which have the particular drawback of seriously complicating the derivation of an error analysis).

The various progress reports on the making of a multichannel deep-tow seismic system by the U.S. Navy indicate a slow realisation that a shotpoint hydrophone is important and that static corrections can often be made in the simple manner noted above. In reporting on this development programme, aimed at putting a 24 channel

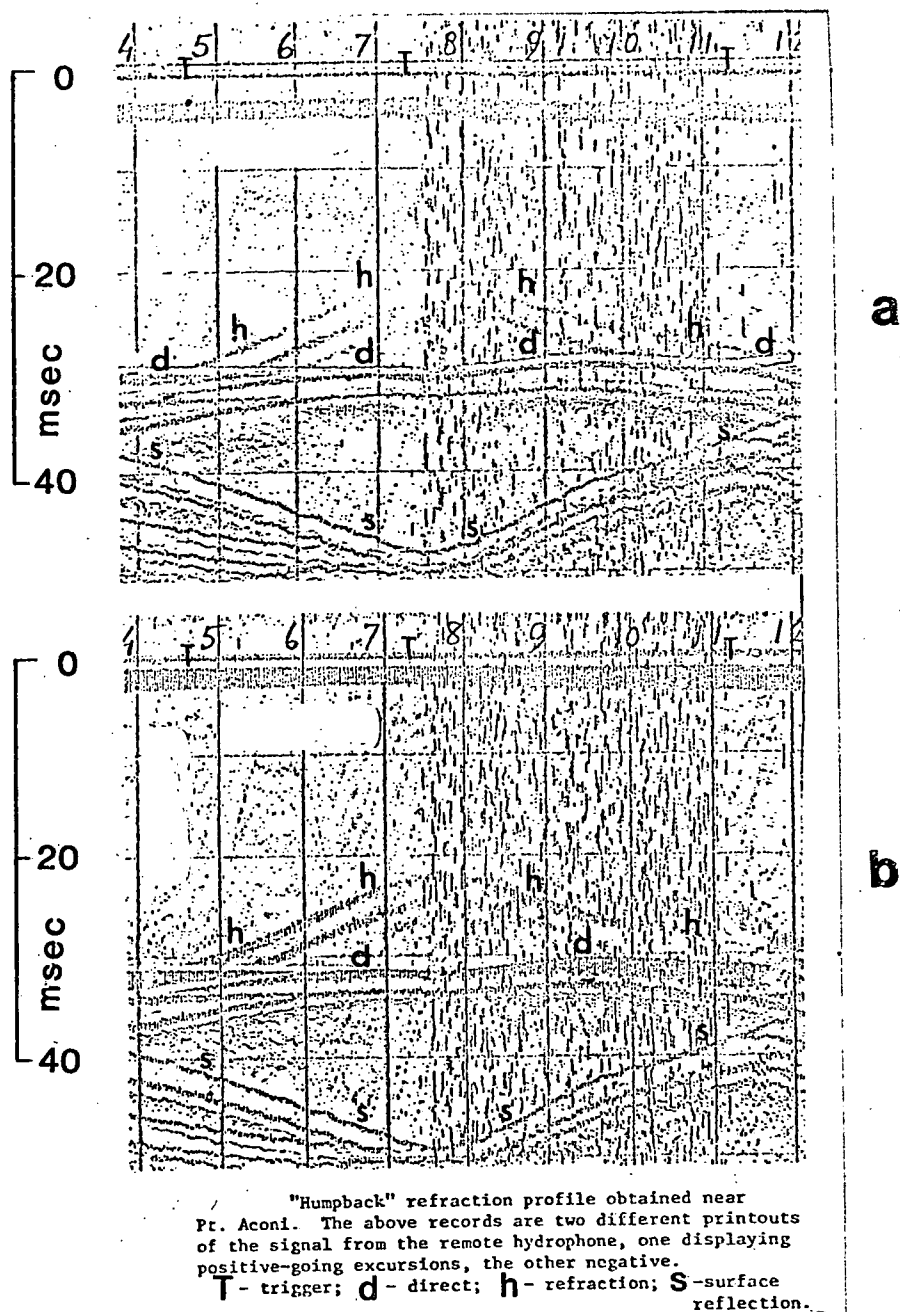


Fig. 7.12

Headwave observations from the location "A".

a. Print of positive-going signal.

b. Print of negative-going signal.

receiving array together with an acoustic source of about 600Hz to tow about 100m off bottom in depths to 3000 metres, Fagot and Eckstein (1979) omit all reference to a shotpoint hydrophone, which it must be presumed was not then part of their original design, and include an analysis of errors which will arise in measured sub-seabed velocities if the 24-channel array should happen to tilt up or down from the horizontal, from which it must be concluded that they had no realisation of the ease with which such a tilt could be measured and allowed for in the data analysis. Subsequent papers about the same project by Winokur and Gonzalez (1982), by Fagot et al. (1982), and by Fagot (1983) have done little to indicate that the points mentioned above have been picked up, but Fagot (personal communication, 1983) has indicated that a shotpoint hydrophone is now to be incorporated in the design of the system, which is currently scheduled to produce field results by mid 1984.

The first headwave observations with the near-seabed equipment were made off Point Aconi at location 'A' (fig. 7.7) with the results shown in fig. 7.12. The method of reducing such data is given in McKay and McKay (1982). It is normally convenient to have available two variable density displays of the rectified signal, such as are shown in fig. 7.12a and b, the one displaying positive-going events, the other negative-going. The

reason for this is that it is easier to pick a travel-time to the leading edge of a black line on such a display, and arrivals such as those reflected at the sea-surface which are phase shifted by 180° are more easily picked on the inverse of the display on which arrivals such as the direct water wave and the bottom reflexion are measured. The Point Aconi data (fig. 7.12) suffered from the failure of the shotpoint hydrophone on deployment, when an underwater connector pulled loose. From what has been written above, it would be presumed that any resulting data would, in consequence of this missing hydrophone, be useless. This would be so but for one feature of the data shown in fig. 7.12: from about the point on the record where the headwave arrival reaches its minimum travel time, the noise increases, and remains at a higher level until just before the end of the record displayed in the figure. This noise is taken as an indication that the trailing hydrophone (fig. 7.10) was being dragged along the seabed, and that consequently its position is known. It is thus possible in this instance to use the equation (7.1) to locate the acoustic source and so proceed to a derivation of the seabed acoustic velocity as outlined in McKay and McKay (1982,p875). In this instance the method is akin to the method of Bryan (1980) who used source and receiver on the seabed.

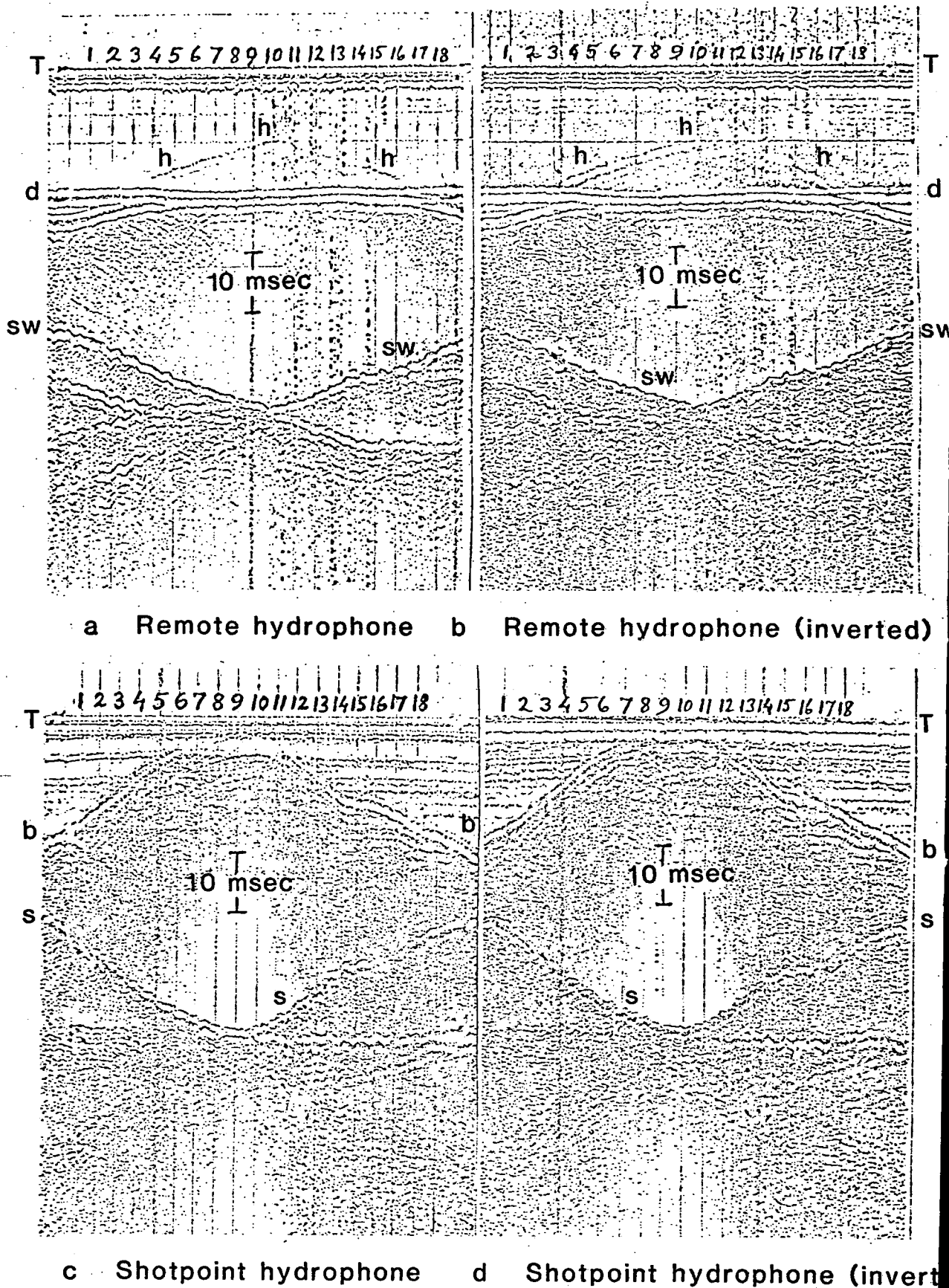


Fig. 7.13

Records from the remote and shotpoint hydrophones for a successful refraction profile. (Area 'B')

T	Trigger	d	Direct arrival
h	Refracted wave	sw	Surface reflection to remote
s	Surface reflection to shotpoint	R	Reflection from bottom and surface
		b	bottom reflection (vertical incidence)

Data obtained from the Donkin area (location 'B' of fig. 7.7) is shown in figs. 7.13 and 7.14. One difference from that of fig. 7. 12 is that the shotpoint hydrophone output was obtained here, after the subsurface assembly had been completely re-rigged to allow greater flexibility of deployment by providing more cable length on the section between the boat and the depressing weight. Another difference lies in the slightly increased acoustic frequency of the acoustic source from ca. 0.9KHz to ca. 1.3KHz. The difference is due in part to the replacement of the single-tip sparker with one having 6 electrodes (in a ring of 25cm. diameter to preserve a point-like character to the source) and in part to the pressure effects noted in section 2.1 (fig.2.2), given that the water depth in the vicinity of location 'B' is ca. 12 metres deeper than at location 'A'. The various arrival times necessary for the velocity computation can be obtained from profiles of the type shown in fig. 7.13. In addition, the flat horizontal reflector apparent in figs. 7.13c and d tangential with the reflector S at its turning point is formed by acoustic arrivals reflected at both sea surface and seabed and indicates a horizontal seabed. Results are shown in table 7.1. The experimental errors are derived by making the travel time measurements at five points along the profile and working out a velocity for each point.

Results of refraction experiments.

Number	Locality		Rock velocity (km/s)	Water depth (m)	Method of derivation
	Lat. N.	Long. W			
1	46°23'	60°17'	2.20 ± 0.05	29	Travel time
1	46°23'	60°17'	2.17 ± 0.12	29	Gradient of travel time
2	46°14'	59°49'	2.00 ± 0.15	43	Travel time
3	46°14'	59°48'	2.13 ± 0.07	41	Travel time
4	46°14'	59°48'	3.00 ± 0.10	42	Travel time
5	46°14'	59°48'	2.86 ± 0.15	41	Travel time
5	46°14'	59°48'	2.60 ± 0.21	41	Gradient of travel time

Table 7.1

Results of the near-seabed refraction experiments carried out in the Point Aconi and Donkin areas.

Identifying numbers in the above table refer to the field data shown in other figures as follows:

- 1 - fig. 7.12
- 3 - fig. 7.13
- 5 - fig. 7.14

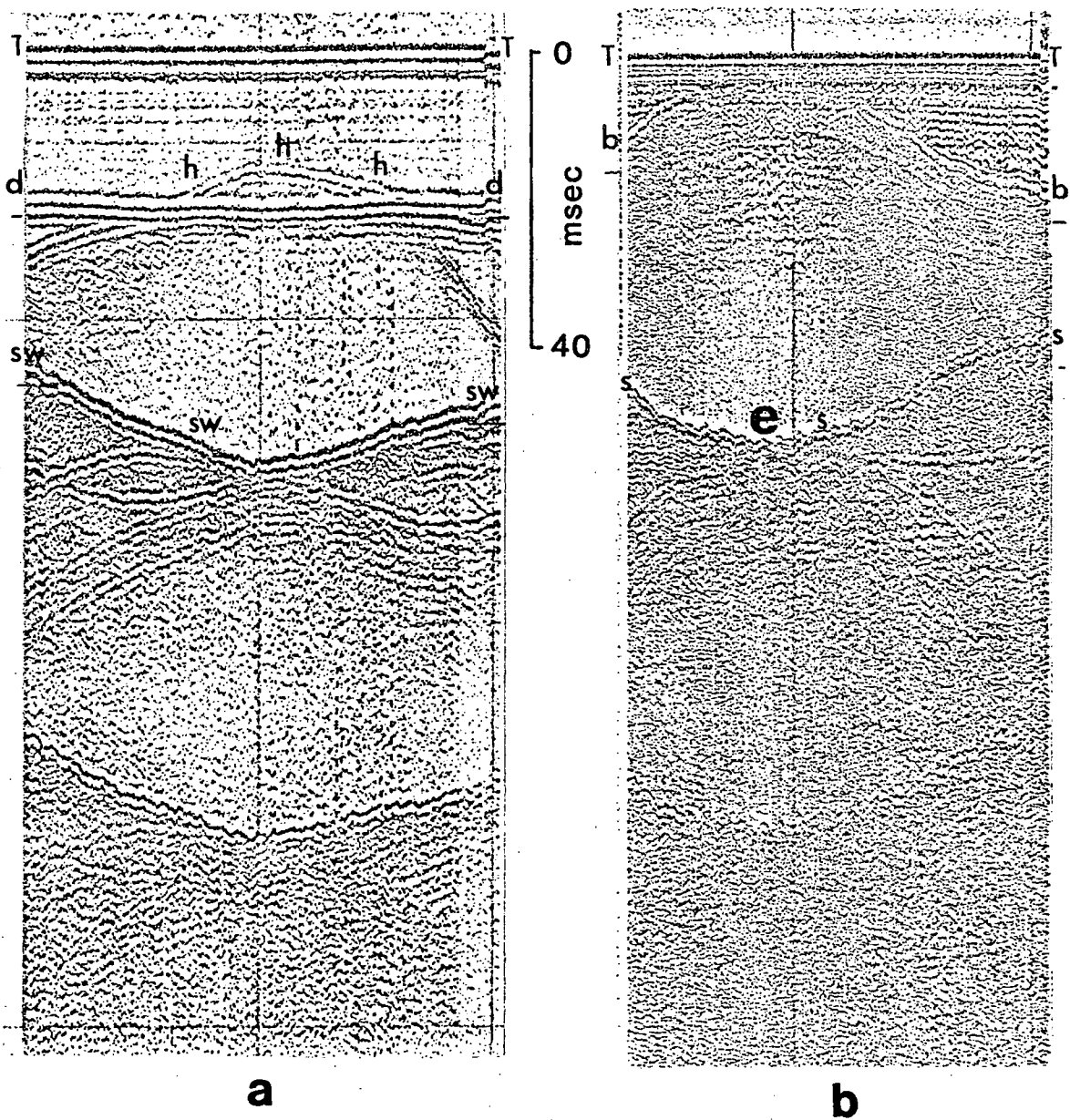


Fig. 7.14

Headwave observations from location "B" offering the possibility of a gradient method of interpretation.

- a. Signals from remote hydrophone.
- b. Signals from shotpoint hydrophone.

A possible objection to this technique lies in the possibility that the first break of the headwave has not been identified and that a later phase has been picked. This arises because the method relies on the measurement of the travel time of the headwave rather than on its gradient, which is the usual measurement to make in the conventional method exemplified in section 7.1. Consequently the picking of a second phase would lead to a systematic error which would not show up in the scatter analysis of velocities found from measurements at different points. One way around this is to take care in correlating the headwave signature with that of the direct wave to ensure that all phases are present and that the first break has been identified. Another way is to derive a differential method for use in the circumstances which permit it, namely when a turning point exists in the flying height of the acoustic source provided that this occurs while the gradient of the arrival time of the headwave is non-zero. The appropriate expression is given in McKay and McKay (1982), and fig. 7.14 is an example of data suitable for treatment by this method. In fig. 7.14b, the point 'e' marks the turning point of the reflector s, the vertical incidence surface reflexion which indicates the position of the acoustic source. This point occurs before the time mark shown on both figs. 7.14a and b and it can be seen that in fig. 7.14a at the point corresponding to 'e', the headwave h has not reached

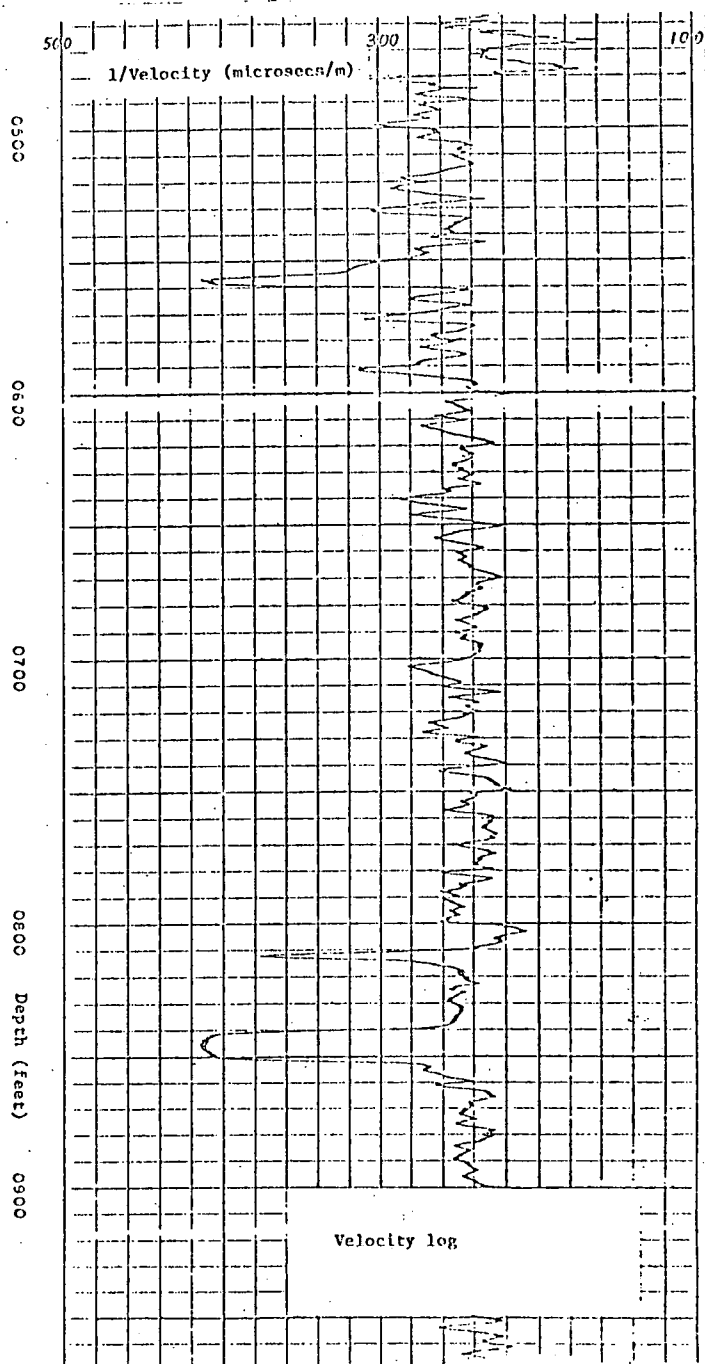


Fig. 7.15

Velocity log from offshore drillhole, near location "B".

its turning point so that the gradient of its arrival time is non-zero. Results from this procedure are given in table 7.1 (no. 5). The method was also applied to the profile of fig. 7.12, since a hydrophone trailing on the seabed is also clearly at a stationary point in the water column.

Other methods of estimating the rock velocities were available for the Sydney coalfield. A number of holes had been drilled in the Donkin area and off Point Aconi, and lithological and acoustic logs were available from these. An acoustic log from the Donkin area is shown in fig. 7.15. This indicates a velocity of $4.2 \pm 0.4 \text{ km/sec}$ in the region between 200m and 300m subbottom. Another estimate of velocity was obtained by correlating the difference in depth of a lithological horizon (typically a coal seam) which could be identified in adjacent drill holes with the difference in travel time to some prominent reflector on such seismic records as passed over or very close to these drill hole locations. The limited subbottom penetration shown in fig. 7.8 meant that in most cases, this correlation could be made only by extrapolating downwards the near-surface information as in fig. 7.16. The results of this work, carried out by Prentiss et al. (1979) are shown in Table 7.2.

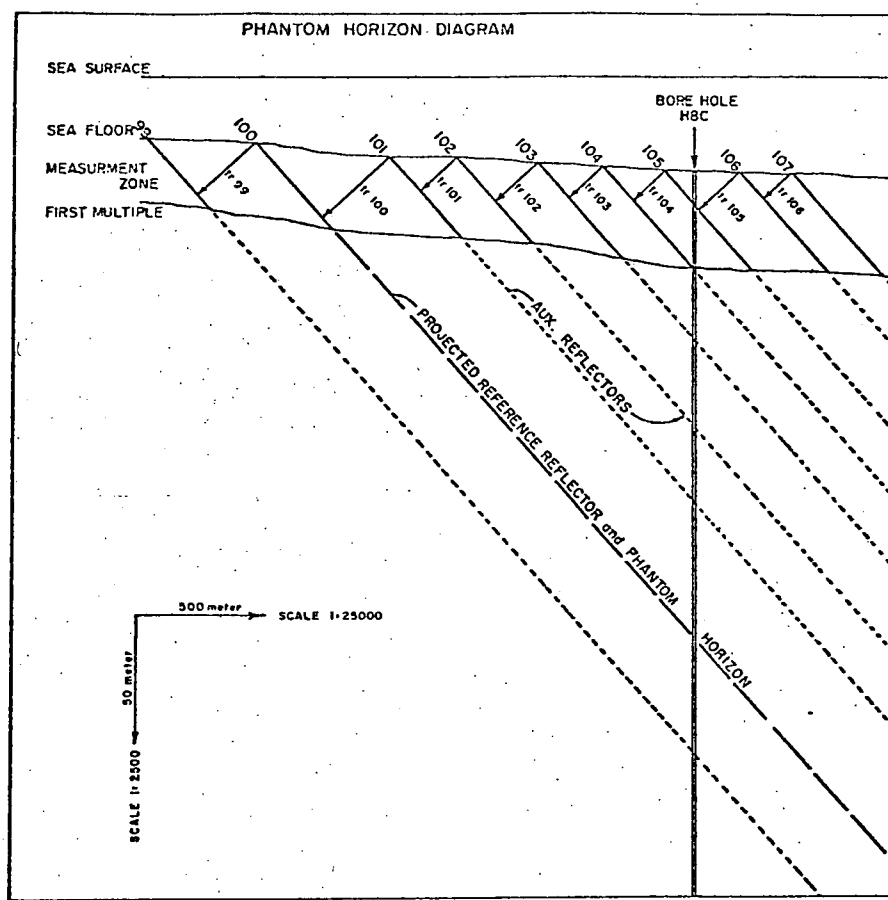


Fig. 7.16

Method of extrapolating observed dip of the coal measures for calculation of seismic velocity by correlation between strata identified in adjacent boreholes.

Velocities measured by comparing seismic phantom and borehole depths to various coal seams.

<u>Area</u>	<u>Seam</u>	<u>Velocity(km/sec)</u>
Donkin	Phalen	3.26
	Harbour	3.29
	Hub	3.29
	Lloyd Cove	3.28
	Point Aconi	3.42
	bottom Lloyd Cove	3.31
	bottom Cannel	3.30
Point Aconi	Hub	3.64
	Lloyd Cove	3.77
	bottom Lloyd Cove	3.74
	bottom Cannel	3.61
Average for Donkin area		3.306 ± 0.050
Average for Point Aconi area		3.689 ± 0.075
	<u>Borehole</u>	
New Waterford	H-1A	3.52 ± 0.071
	H-2	3.65 ± 0.088
Average for New Waterford		3.585 ± 0.080

Table 7.2

Results of measurements of velocity by the method of correlating strata in boreholes.

A comparison of the near seabed refraction results (table 7.1), the horizon correlation results (table 7.2) and the acoustic log (above) suggest increasing velocity with depth, with some of the near seabed values appearing very low. There may be reasons why low values could occur, the main possibilities being either thin deposits of glacial till overlying the coal measures at the seabed or, more likely, a significantly weathered zone resulting from subaerial exposure and freeze-thaw action prior to submergence within the last 10,000 years. Whereas at Mabou, the seabed investigated was clearly a recently eroded wave-cut platform at the base of a cliff, the manner of submergence of the areas off Sydney is less clear, and may have occurred in places without such severe coastal erosion.

7.3 HEADWAVE OBSERVATIONS IN THE NERLERK AREA OF THE BEAUFORT SEA.

Although for geological reasons it appeared to have been a failure for the accurate prediction of acoustic velocities for the determination of dip-angles of the coal measures in the Sydney Basin (section 7.2) the near-seabed method had been demonstrated as a viable way of observing headwaves as first arrivals by the use of frequencies of ca. 1KHz and source-receiver separations of ca. 40 metres,

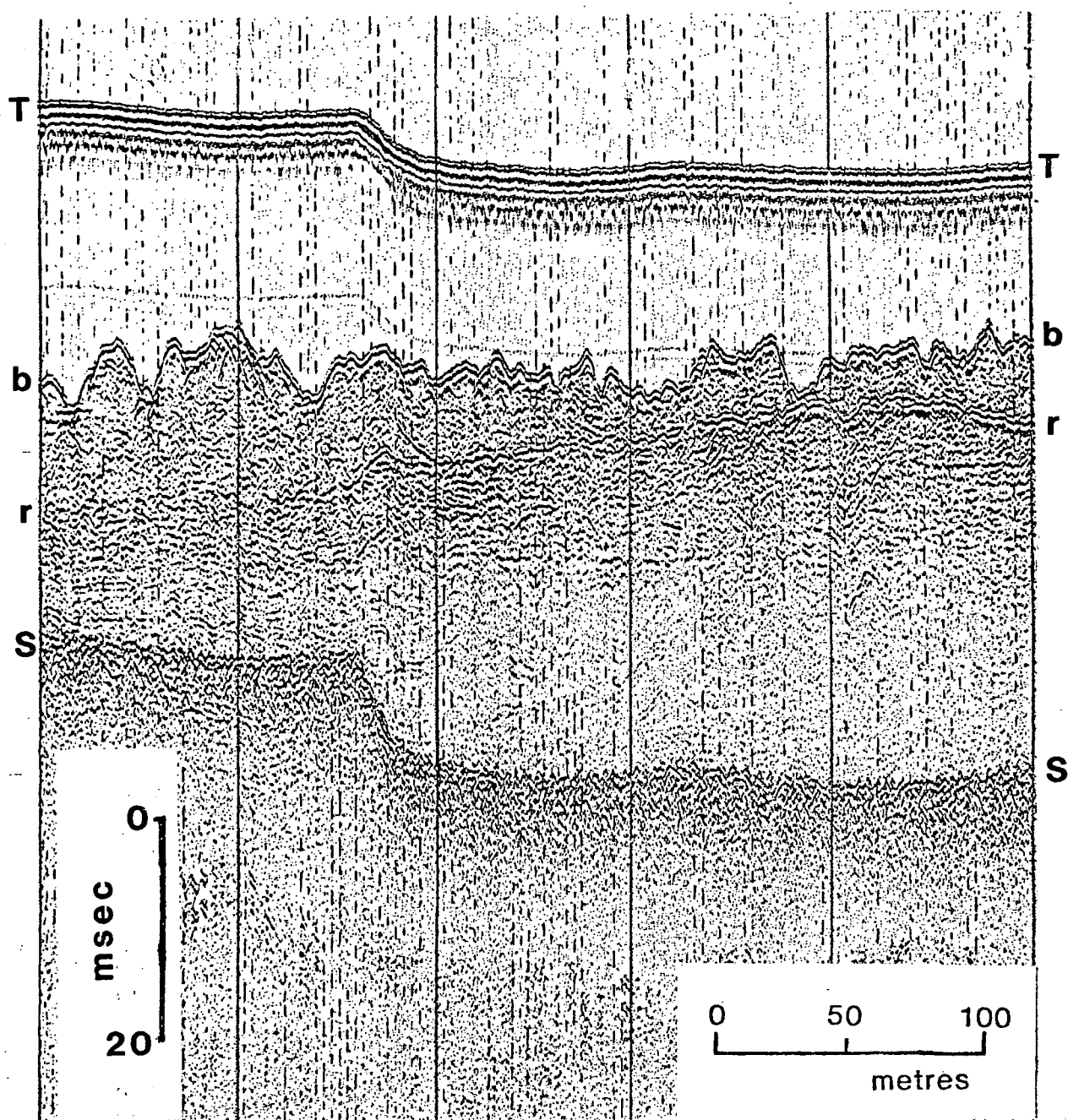


Fig. 7.16b

Illustration of how the masking effect which the surface reflexion S can have on subbottom information depends on the height at which the fish is towed above the seabed.

an appropriate scale on which to make measurements over small areas of the seabed.

The requirement by a subbottom profiling contractor to tow equipment of the Nova Scotia Research Foundation close to the seabed of the Beaufort Sea in order to avoid the over-writing by sea-surface reflexions of subbottom information (fig. 7.16b) led to a situation during fieldwork where it became possible to gather near-bottom data suitable for the recognition of headwaves as first arrivals on a continuous basis while the conventional vertical incidence reflexion data was being obtained, by the expedient of making a minor field adaptation, namely the addition of a hydrophone trailed ca. 40 metres behind the towfish of the basic deep-tow system described in section 2.1. The good remote-control facilities for winching the main tow cable in and out allowed the towfish to be flown within a few metres of the seabed, sufficiently close that headwaves as first arrivals could be detected frequently as low amplitude events occurring just before the arrival of the direct water wave (fig. 7.17; fig. 6 of McKay, 1983).

The calculation of velocities from this data is based on the same travel-time expressions used for that obtained over the Sydney coalfield (section 7.2), as given in McKay and McKay (1982) and discussed in McKay (1983). These

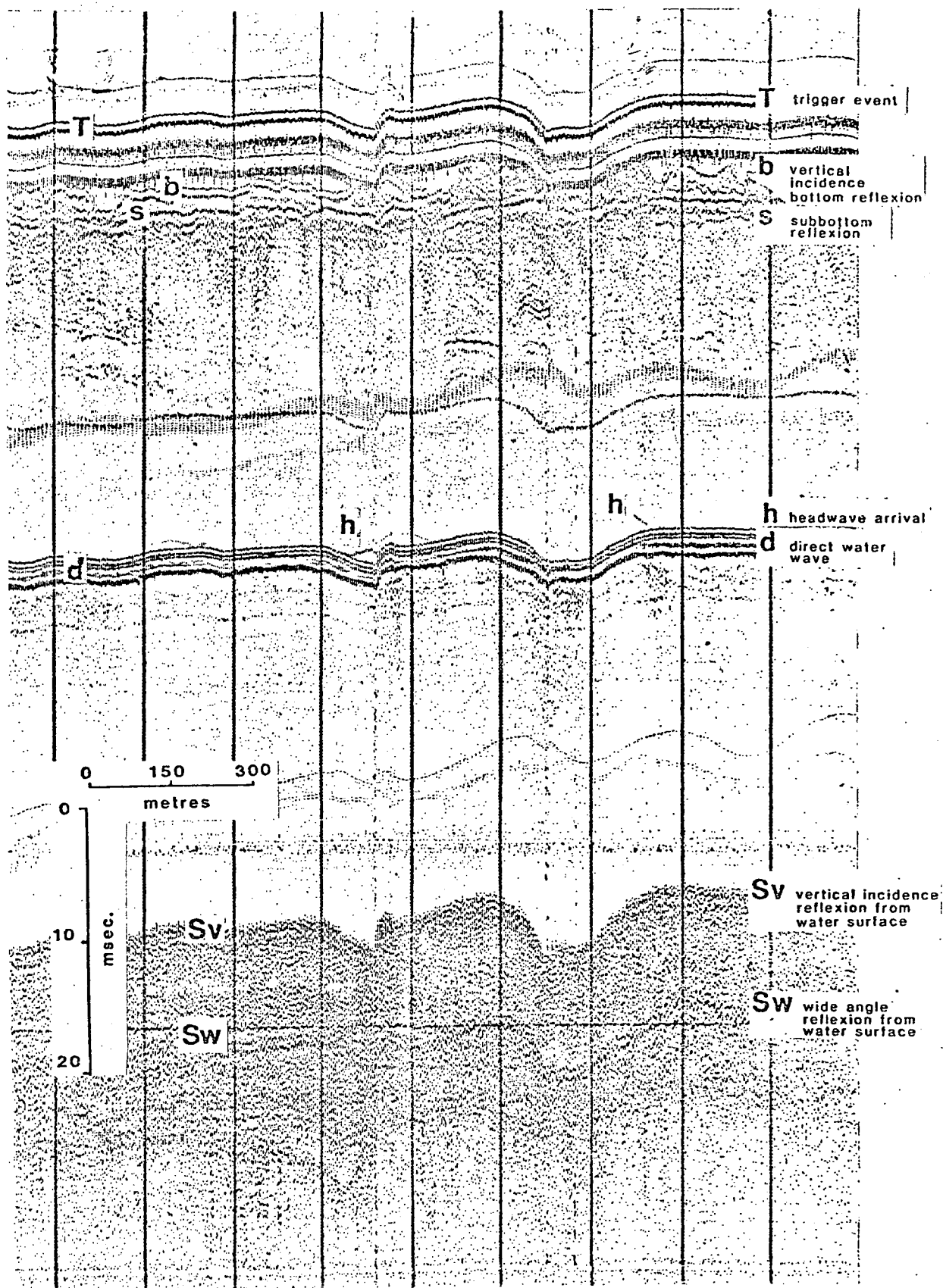


Fig. 7.17

Headwave observations from a shallow subbottom interface over the Nerlerk drill-site in the Beaufort Sea.

calculations can proceed only on the assumption that some specific interface seen on the vertical incidence reflexion record is associated with the appearance of the headwave. In the case of the Sydney data, this interface was identified as the seabed, but in the Nerlerk area, it is taken to be the strong horizontal reflector 'S' (fig. 7.17) occurring about 2 metres below the seabed. There are various reasons for making this identification in preference to relating the headwave to the seabed interface:

(i) Other acoustic profiling systems were in use during the Nerlerk survey, specifically a 33KHz Atlas Deso echo sounder, mounted to the ship's hull, and a Klein 3.5KHz pinger, mounted on a fish together with side-scan sonar transducers. Both sounder and pinger always defined the base of the thin uppermost sediment layer, but the sounder gave no deeper information and the pinger seldom gave deeper information. The experience of King (1967), who, with an echo sounder operating in a similar frequency band, correlated the appearance of sonogrammes with bottom sediment type, leads to the identification of this uppermost layer as being predominantly of clay-sized material.

(ii) Pelletier (1975, 1979) has identified the seabed material as being a clayey mud in this region, on the basis of bottom grab samples.

(iii) Wide angle reflexion calculations on data obtained in

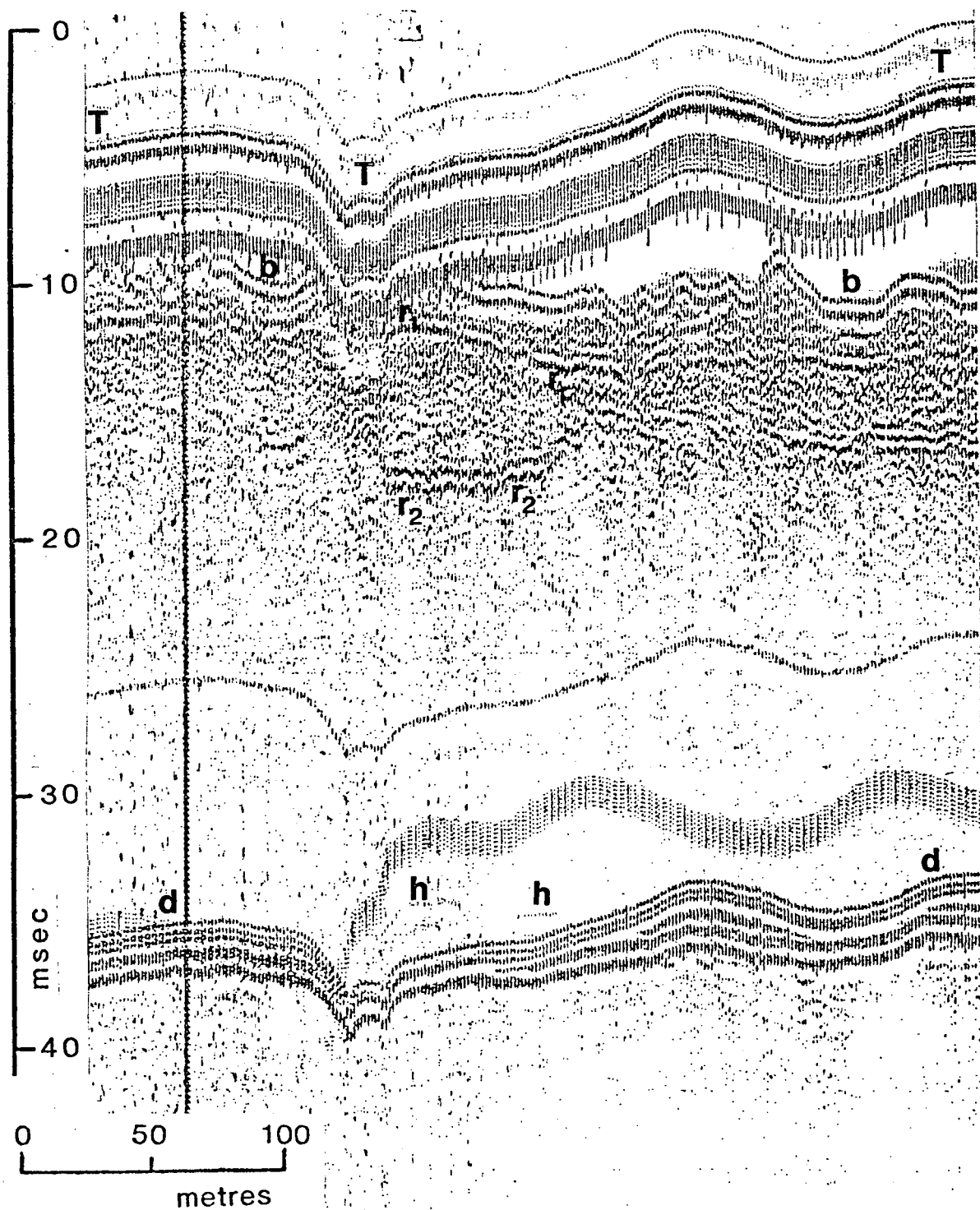


Fig. 7.18

Headwave observations with the difference that an ambiguity arises from the presence of two subbottom reflectors r_1 and r_2 , either of which may be associated with the headwave h .

the nearby region of Aiverk where the layer is ca. 10 metres thick suggest that the acoustic velocity in this layer is the same as that of the overlying seawater, to within an error of about 4%.

(iv) There is no correlation in the behaviour of the arrival time of the headwave with the severe undulations which occur in the seabed as a result of ice scouring.

Of these reasons, the last is perhaps the most satisfactory in that it arises from the same data as is used in the headwave analysis to yield velocities. None of the reasons given above rules out the possibility that the headwave is associated with an interface deeper than that at the base of the surficial layer of mud. In most instances, there is no other interface apparent on the vertical incidence profile with which it would be possible to associate the headwave, but in a few cases, there does exist a deeper reflector. When this occurs (see fig. 7.18) an ambiguity arises, and two velocities have been worked out, one on the basis of each reflector. All the results are shown in table 7.3. In cases where a smaller ambiguity arises as to the exact location of the subbottom reflector 's', two results have been calculated, and the suffix 'A' is appended to the first number shown in table 7.3 to denote the second result. The velocities found are reasonably consistent, and seem to indicate that the material of the lower layer has a value of about 1.13

Compressional-wave velocities from headwave arrivals

Line No.	Fix No.	Velocity	Standard Error	Line No.	Fix No.	Velocity	Standard Error
B1116	3975.2	1.710	.035	B2102	5850.5	1.631	.049
	3975.8	1.647	.029		5850.25	1.650	.033
	4100	1.598	.024		5850	1.659	.052
	4100.5	1.631	.044		5825.8	1.670	.051
	4125.2	1.681	.065		5750	1.619	.050
	4125.8	1.601	.038		5725.7	1.604	.041
	4175.5	1.688	.063		5725	1.637	.068
	4175.5A	1.628	.056		5725A	1.656	.071
	4175.8	1.627	.050		5700.5	1.607	.063
	4250.5	1.683	.046		5700.5A	1.628	.066
B1112	4500	1.617	.048		5700	1.616	.047
	3999	1.666	.037		5700A	1.635	.050
1104	3550	1.682	.045		5675	1.651	.063
	2000	1.611	.047		5675A	1.661	.065
	1950	1.645	.031		5650	1.641	.066
B2108	*7080*	1.619	.039		5625	1.608	.052
	7025.5	1.630	.036		5575	1.658	.055
	7025	1.583	.045		5550.5	1.660	.031
	6875.2	1.624	.034		5550	1.603	.044
	6875	1.645	.040		5475	1.655	.052
	6850.7	1.680	.042		5425	1.613	.038
	6725	1.660	.060		*5400*	1.667	.050
	6700.7	1.688	.057		*5300.3*	1.645	.041
	6700.2	1.669	.059		*5300*	1.556	.049
	6700	1.664	.056		*5200.5*	1.637	.071
B2112	6400.4	1.600	.059	The following four measurements were made on the assumption that a deeper interface is associated with the observed headwave			
	6400.2	1.668	.059				
	6225.3	1.614	.049				
	6225.1	1.657	.054				
	6225	1.653	.053				
	6200.5	1.605	.048				
	6200.2	1.590	.043				
	6175.9	1.621	.053				
	6175.5	1.621	.041				
B2102	*6000.2*	1.589	.072				
	6000	1.591	.060	B1112	*3999*A	2.029	.065
	*6000*A	1.645	.070		3550A	1.836	.058
	6000.1	1.566	.056	B2102	*5400*A	1.916	.074
	*6000.1*A	1.631	.069	B2102	*5200.5*A	1.915	.101
	5990.9	1.635	.044				
	*5990.9*A	1.681	.049				
	5990.7	1.600	.054				
	6000.8	1.574	.043				
	*6000.8*A	1.642	.052				
	5925	1.648	.053				

Table 7.3

Compressional wave velocities of the seabed material beneath the surficial layer of holocene mud. Area of the Nerlerk drilling site.

times that of the seawater with an experimental error of ca. 3%. This value correlates with what is known about the nature of the deeper layer from drilling, namely that it is a sandy silt (I. Watson, Dome Petroleum Geotechnical Division, personal communication, 1983) but does not indicate that there is anything unusual about its properties. Since there is no particularly good reason for associating the observed headwaves with the occasional deeper reflector (fig. 7.18), it is reasonable to conclude that ice-bonding is absent at this depth in the Nerlerk area. Doubts which may remain about the instances where a deeper reflector is present could be resolved with equipment having the ability to record more than one offset channel. Details of the error analysis and the computer programme used to derive the results of table 7.3 are given in Appendix 2.

Although the approach to headwave observation described above does lead to a more consistent gathering of useful data than the method described in section 7.2, observations are very dependent on the flying height of the acoustic source and receiver, and the consistent development of a headwave as a first arrival is difficult to achieve. As noted in McKay (1983), the procedure would not be possible in appreciably deeper water than that of the Nerlerk area (45m) without the implementation of automated bottom-referencing tow bodies. A considerably

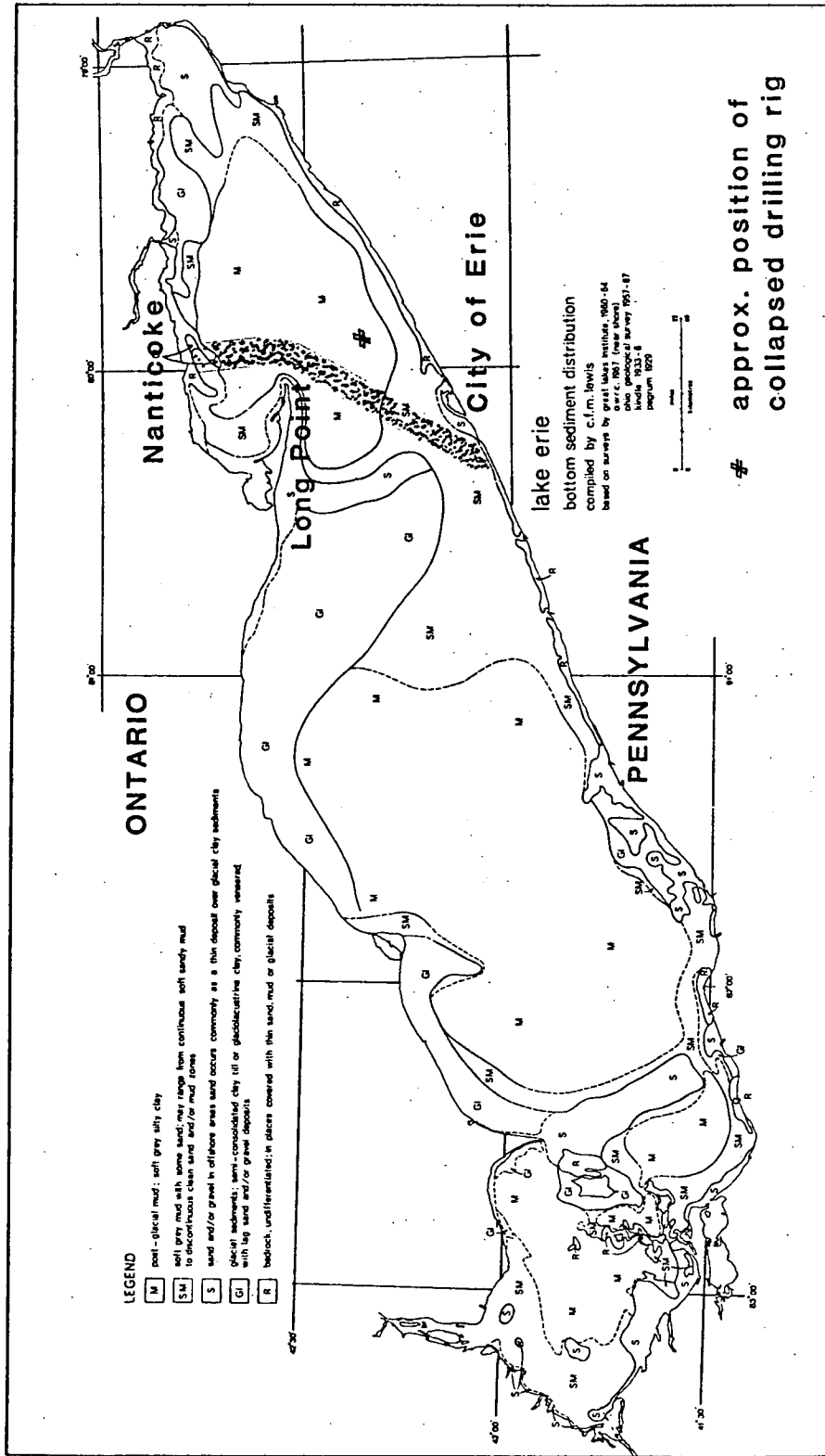
greater quantity of data of this type was gathered from other parts of the Beaufort Sea, but its analysis has not yet been permitted by Gulf Oil.

7.4 WIDE ANGLE REFLEXION METHOD: SURFACE - LAKE ERIE.

The relative proximity of the Nanticoke coal-fired electrical generating station whose output was not required in Ontario, to the state of Pennsylvania which had recently been deprived of generating capacity as a result of the Three Mile Island nuclear reactor incident led to the making of a plan to lay cables across Lake Erie, not so much because this was technically the easiest route, but because the alternative way over-land would involve much legal entanglement with numerous local authorities and the possibility of taxation by the State of New York on the electricity transmitted across its territory. Such a project required knowledge about the lakebed conditions in detail greater than was known from earlier work. Investigations were carried out with sidescan sonar, low-frequency echo sounder and boomer for vertical incidence reflexion profiling. This profiling equipment offered the possibility of investigating the velocity structure in the lake-bed sediments by the use of a technique almost identical with that described in section 7.1 where equipment handling is concerned, but

aimed at making observations of acoustic arrivals to be dealt with in quite a different way, namely the method of wide angle reflexion (section 6.1). A similar buoyed hydrophone on a long (ca. 100m.) cable was deployed and streamed aft at points along survey lines where it seemed that the sediment layering was likely to remain simple during the streaming process. Unlike the procedure of section 7.1, no interruption was made to the progress of the boat, which continued all the time along its survey lines gathering the continuous vertical incidence reflexion and side-scan profiles. In this sense, the method is identical to that described by Le Pichon et al. (1968) apart from a difference of about two orders of magnitude in the scale on which it is done and the way in which acoustic signals are brought to the recording vessel by wire rather than by radio link from a sonobuoy. Bennell (1978) adopted a similar method for gathering information suitable for interpretation by the wide angle reflexion method in Tremadog Bay, with the difference that he recorded the data while the trailing hydrophone was hauled in towards the recording vessel. His approach has the following two disadvantages over that described above:

- (i) The hydrophone is moving through the water while the recording is being made, leading to greater noise levels than can be obtained when the hydrophone is nearly stationary.
- (ii) It is a method which is less appropriate for use



Distribution of surficial sediments in Lake Erie.

Fig. 7.19 Surficial sediment distribution in Lake Erie (Sly and Lewis, 1972). the location of the survey for investigating the proposed cable crossing is shown as the stippled band towards the eastern end of the lake, between Nanticoke and the city of Erie.

during the sailing of survey lines, because of the much greater length of line which would be covered during the retrieval of the trailing hydrophone, with an associated increase in the difficulty of having the gathering of the wide-angle data coincide with the passage of the survey boat over an area of simple sediment layering.

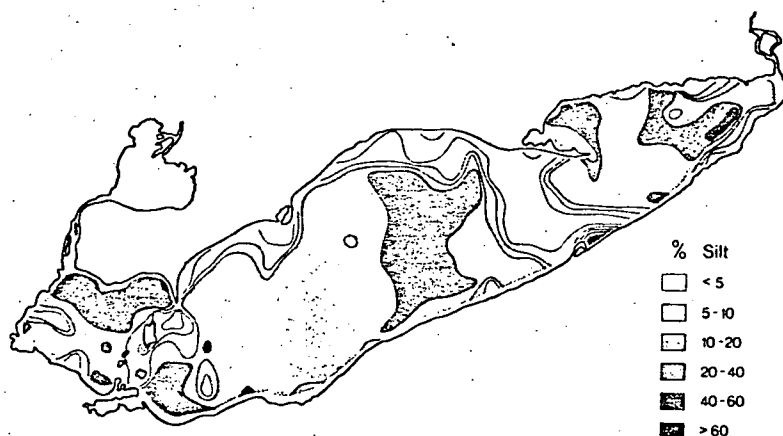
The method of recording while cable is being payed out does suffer from the difficulty that the cable must be handled skillfully to maintain its free running. This problem can be overcome with practice.

The location of the Lake Erie survey area is shown in fig. 7.19 as the band between Nanticoke on the northern (Ontario) shore and the city of Erie on the southern (Pennsylvania) shore, taking a dog-leg to avoid Long Point. The distribution of surficial sediments in Lake Erie is shown in fig. 7.19 as given by Sly and Lewis (1972) and results of grain-size analyses on bottom samples taken on a 10km. grid by Thomas et al. (1976) are shown in figs. 7.20a,b and c. The names used for areas of the lake are shown in fig. 7.21a and the thickness of recent sediment is given in fig. 7.21b. Thomas et al. recognised three major sediment units on the basis of echo sounding and grab sampling: glacial till, glaciolacustrine clay and postglacial muds, representing the late glacial and post-glacial stages of the evolution of the lake basin. Residual glacial moraines divide the basin into



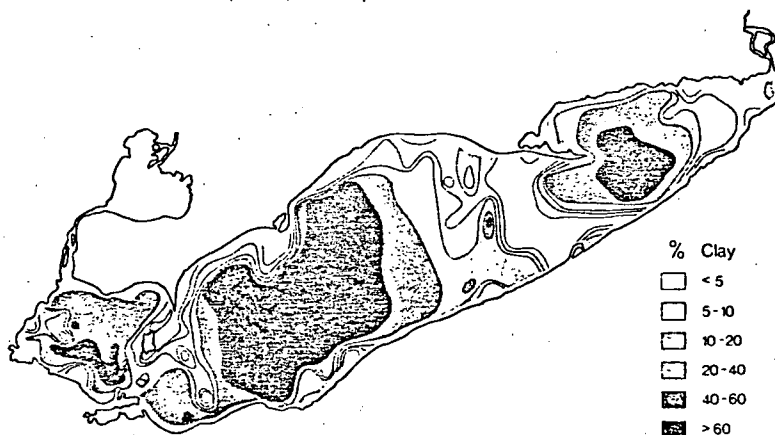
a

Distribution of percent sand-size fraction in the surficial sediment of Lake Erie.



b

Distribution of percent silt-size fraction in the surficial sediments of Lake Erie.

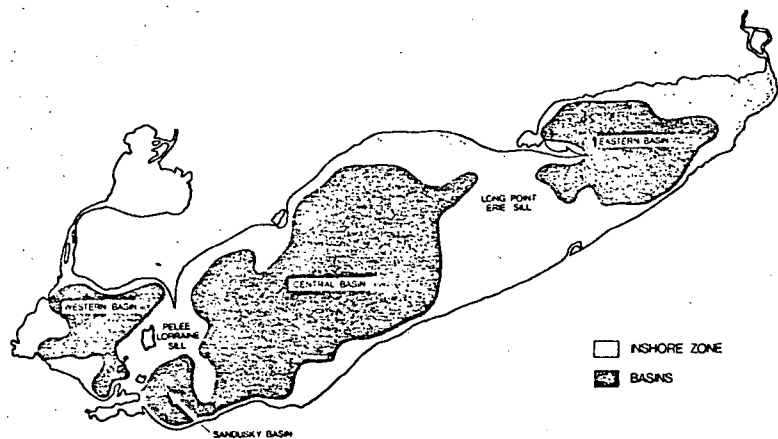


c

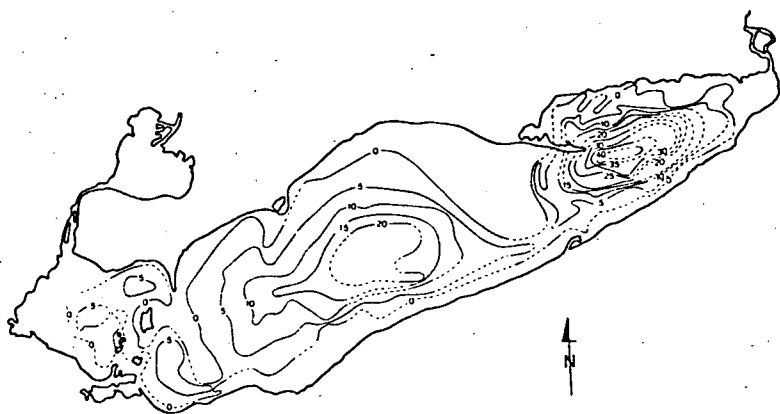
Distribution of percent clay-size fraction in the surficial sediments of Lake Erie.

Fig. 7.20

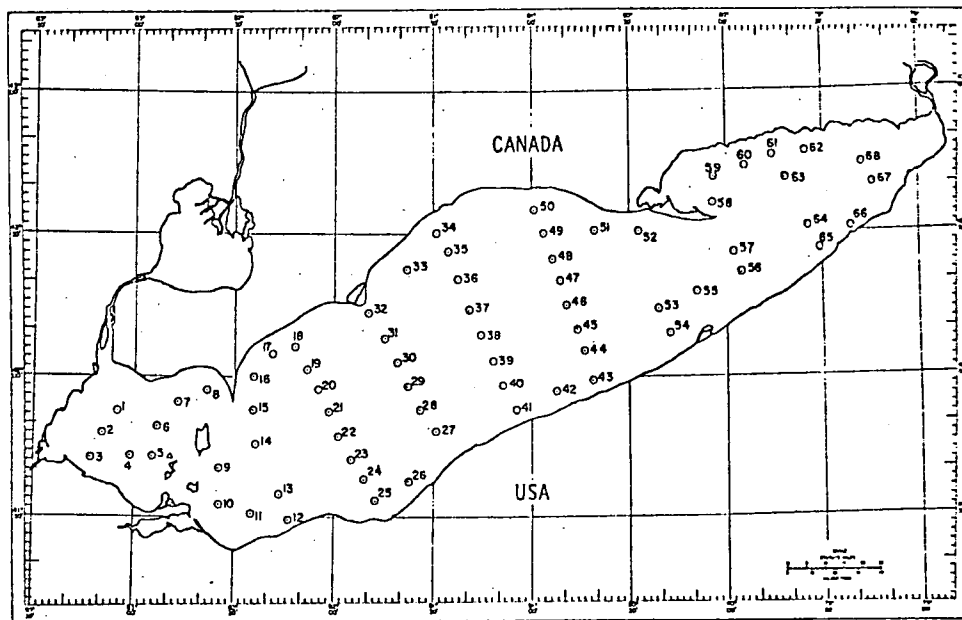
Distribution of sand, silt and clay fractions in the surficial sediments of Lake Erie (Thomas et al., 1976).



Basin distribution and nomenclature including cross-lake sills and inshore zone.



Isopachyte map of recent sediment in Lake Erie (in meters).



Lake Erie map showing locations of cores.

Fig. 7.21

- a. Names of different basins of Lake Erie (Thomas et al., 1976).
- b. Thickness of recent sediments. (Thomas et al., 1976).
- c. Location of cores taken by Morgan (1969).



Fig. 7.22

Wide-angle reflexion data from the approaches to the American shore. Time scale is omitted to emphasise the fact that measurements are made relative to the water velocity and are independent of the running speed of the graphic recorder.

the four sub-basins named in fig. 7.21a. The cross-lake cable route survey traverses the Eastern Basin which with soundings to over 70 metres, is the deepest part of the lake. Thomas et al. note that the sedimentation in this basin is modified by the influx of "substantial" quantities of silt-size material derived from shoreline erosion along the north shore of the Central Basin. This material is transported eastwards by water currents which deposit the coarser fractions on the Long Point sand spit and carry the finer particles into the Eastern Basin to give a net rate of sediment accumulation there of about 0.3cm/yr (Sly and Lewis, 1972). About 40 wide-angle reflexion profiles were obtained during the course of the survey. Some were obtained with a hydrophone on a cable which was insufficiently long to give good moveout and many others were located over insufficiently simple geology for ease of interpretation. Those which were found to occur over simple horizontal layering were interpreted graphically in the field by the simple $T^2 - X^2$ method of Green (1938). Measurements, in arbitrary scale units were made of the direct water wave and reflected arrival times to obtain a velocity which is the average value for the layers above the reflector identified. The profile shown in fig. 7.22 from the approaches to the American shore is dealt with in this way as shown in fig. 7.23. Measurement of the gradient of the line fitted to the data gives, for the average velocity in sediment and

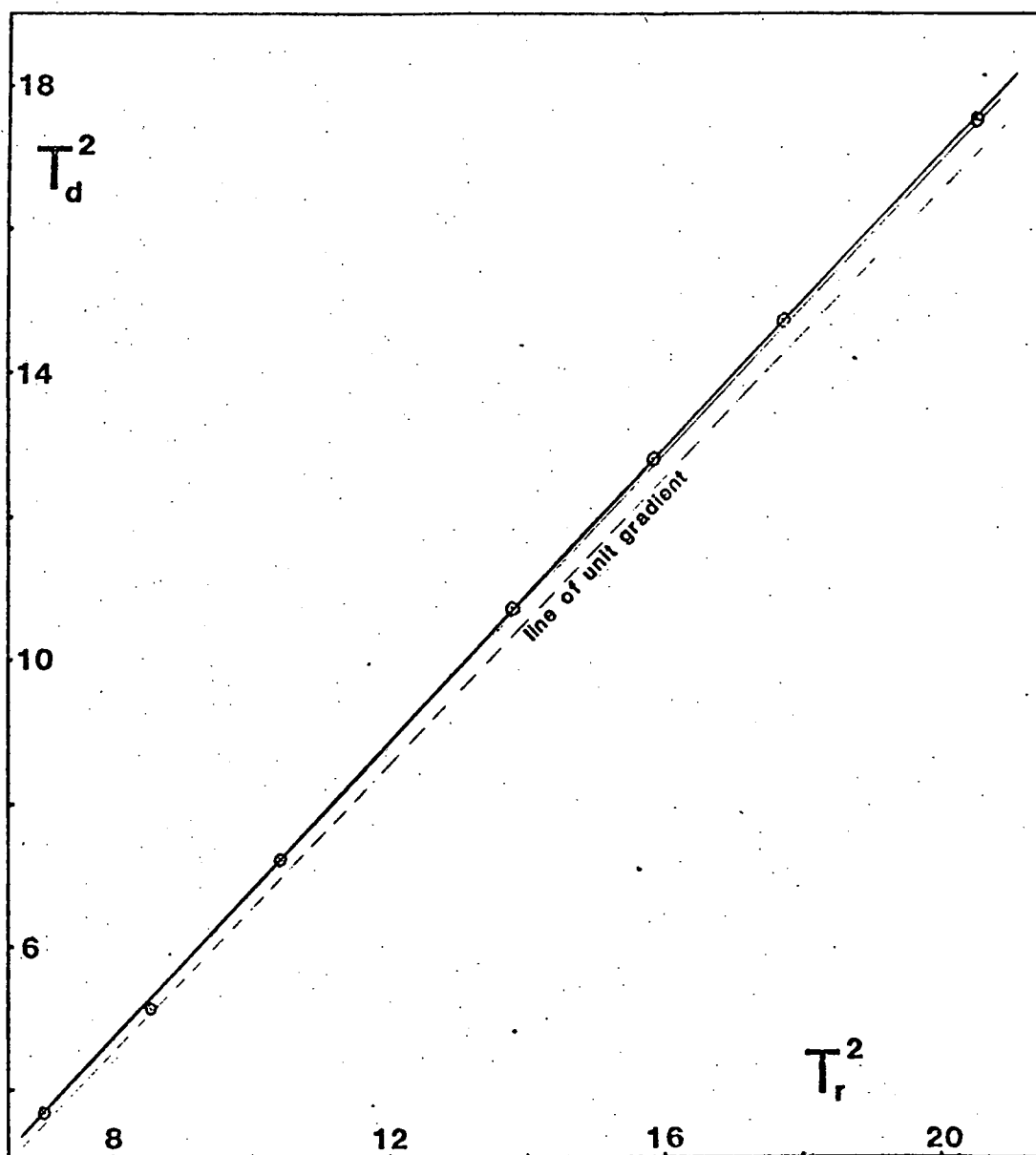


Fig. 7.23

Conventional travel-time method applied to the profile from the American Approaches.

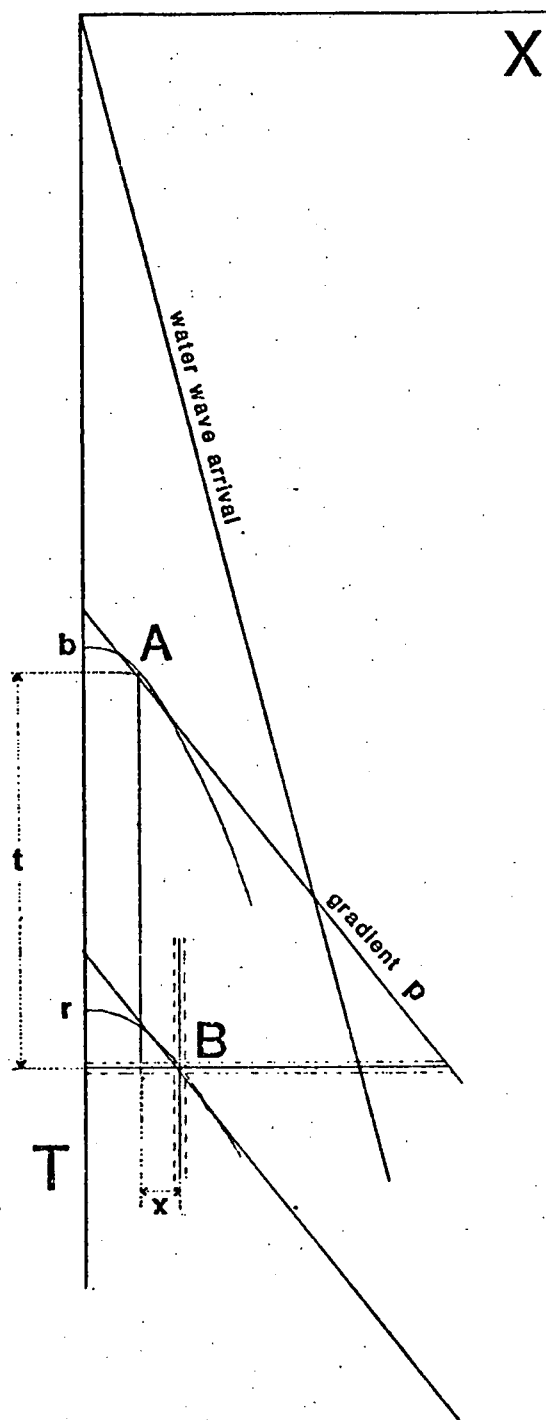


Fig. 7.24

The ray-parameter travel-time method applied to the profile from the American Approaches.

water layer, the value:

$$v_a = (1.024 \pm 0.003) v_w$$

where v_w is the acoustic velocity in the water.

This leads to an interval velocity in the subbottom layer as:

$$v_{sed} = (T_r v_a - T_b v_w) / (T_r - T_b)$$

where

T_r is the reflexion time to the subbottom reflector and

T_b is the reflexion time to the lakebed.

which can be evaluated as:

$$\begin{aligned} v_{sed} &= (1.075 \pm 0.009) v_w \\ &= 1560 \pm 13 \text{ m/sec for a value for } v_w \text{ of } 1460 \text{ m/sec.} \end{aligned}$$

The alternative "ray Parameter" method (Bryan, 1974) may also be applied to derive a lakebed sediment velocity. This is illustrated in fig. 7.24, and relies essentially on the identification of points of equal slope on the reflections b (bottom) and r (subbottom reflector). The error arising in this method is dominated by the uncertainty ' Δx ' in finding the value of the horizontal displacement ' x ' between the point of equal gradient A and B. In arbitrary scale units, it is found that:

$$dT/dx = 7.73/6.25$$

$$x = 0.69 \pm 0.10$$

$$t = 6.75 \pm 0.10$$

both dx and x require converting from the dimension of time to distance by the factor $17.89/4.72$ which is the

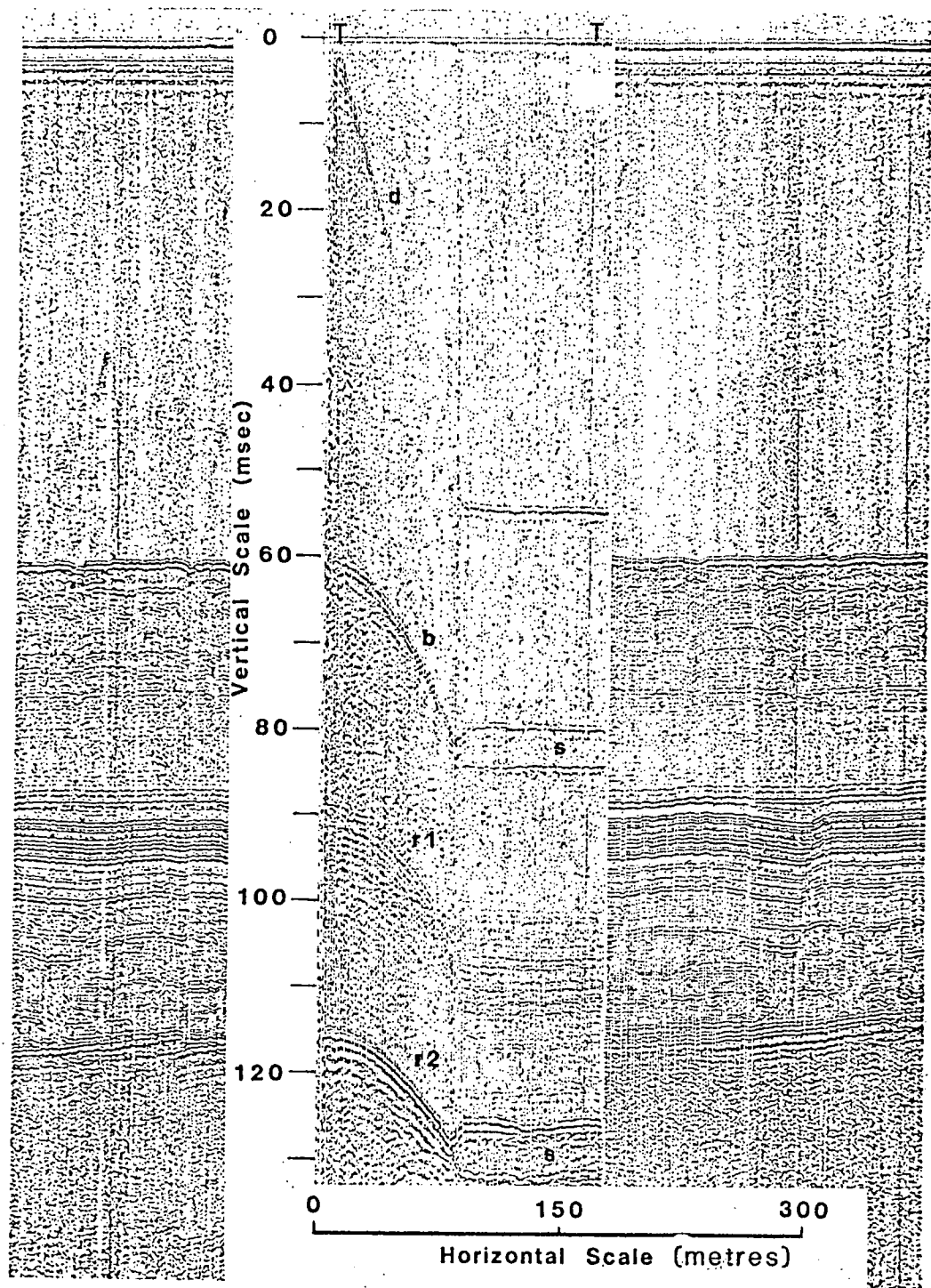


Fig. 7.25

Wide angle reflection profile, Lake Erie.

- T - trigger event
- d - direct arrival
- b - bottom reflection
- r1 - first subbottom reflector
- r2 - second subbottom reflector

gradient of the direct water wave arrival. This leads to a value:

$$\begin{aligned} v_{\text{sed}} &= (1.19 \pm 0.17) v_w \\ &= 1737 \pm 248 \text{ m/sec.} \end{aligned}$$

for a value for v of 1460 m/sec.

The ray parameter method thus yields less precise results. The main reason for this lies in the fact that the subbottom velocity is not appreciably different from that of the water. This means that the two hyperbolic curves of reflexion arrivals from the top and bottom of the sediment layer are more nearly alike than if the velocity were significantly greater as would be the case for consolidated sediments. In consequence the points of equal gradient, A and B occur with small offset 'x'. The main power of the ray parameter method lies in its freedom from dependence on a well defined time and distance origin, because all the measurements required are differential. The Lake Erie data does have a well defined origin and it appears that the simple T^2-X^2 method gives the more precise result.

Data reduction by the T^2-X^2 method was carried out on several profiles in the northern part of the Eastern Basin (fig. 7.21a). One of these, shown in fig. 7.25, was made over an area of the lake bed shown in vertical-incidence section in fig. 7.26 as a line tracing (original vertical incidence profiles and detailed navigation covering the

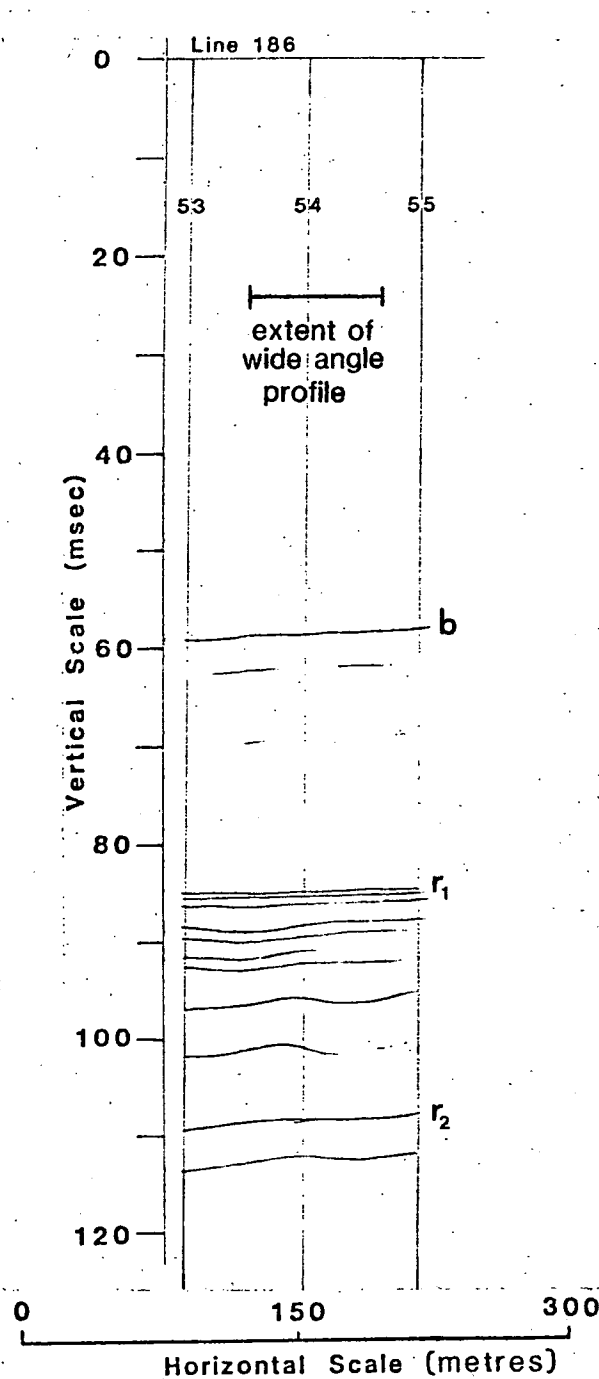


Fig. 7.26

Tracing of reflecting horizons from vertical incidence reflection profile made simultaneously with the wide-angle profile.

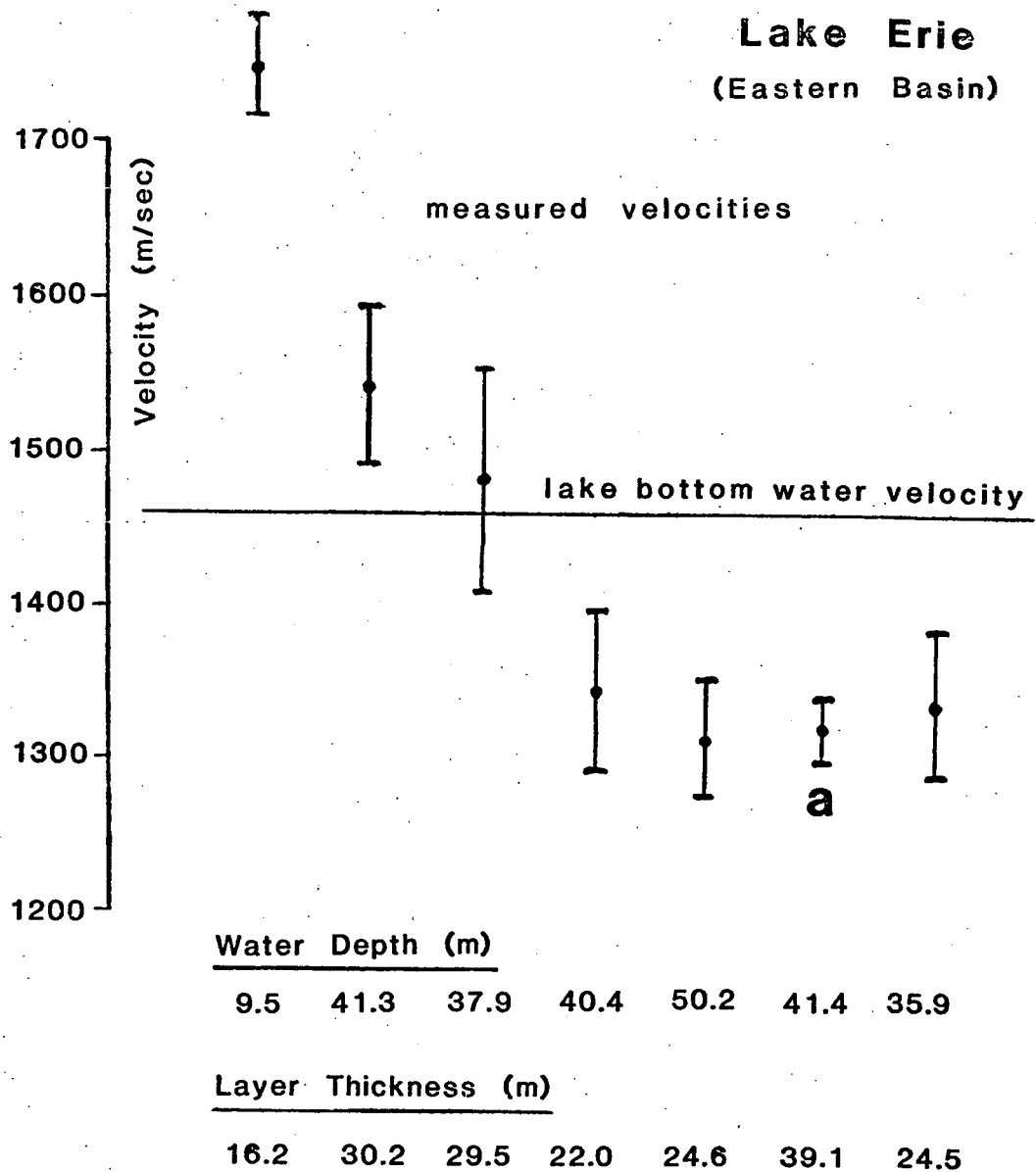


Fig. 7.27

Sediment velocity results from the Eastern Basin of Lake Erie. The thickness of the layer is shown in which the measurement was made. The result 'a' is from the profile of fig. 7.25. The results for 'a' can be refined by deriving interval velocities for an upper (1.326+20m/sec.) and lower layer (1.270+48m/sec.) each ca. 19m. thick, indicating a possible decrease in velocity with depth.

areas of the wide angle profiles have not been made available by Ontario Hydro). The results from this and from nearby profiles are shown in fig. 7.27. The higher velocities occur closer to the Nanticoke shore; the lower velocities are from farther out in the basin, but exact navigation is not available. The results from this and from most nearby profiles indicate a low velocity bottom material (fig. 7.27). These results may be compared with those of Morgan (1969) who studied in the laboratory core samples taken from within 0.5m of the lakebed at the locations shown in fig. 7.21c. Morgan discussed the values which he determined only in terms of the depression below the velocity value for water which can occur in a very porous sediment according to Hamilton's (1956) application of Wood's equation (equation 3.1, section 3.4), and does not consider the possible influence of gas on the acoustic velocities, although its presence in the cores taken at many of the sites (fig. 7.21c) was known (C.F.M. Lewis, Geological Survey of Canada, personal communication, 1983). This is odd, when it is remembered that Wood's equation was originally used to describe the behaviour of emulsions. A section of vertical incidence profile which is available from the northern part of the Eastern Basin (fig. 7.28) shows features in the water column which are considered (K. MacMillan, Canadian Centre for Inland Waters, personal communication, 1980) to be gas seeps. Layering in the lower of the two major

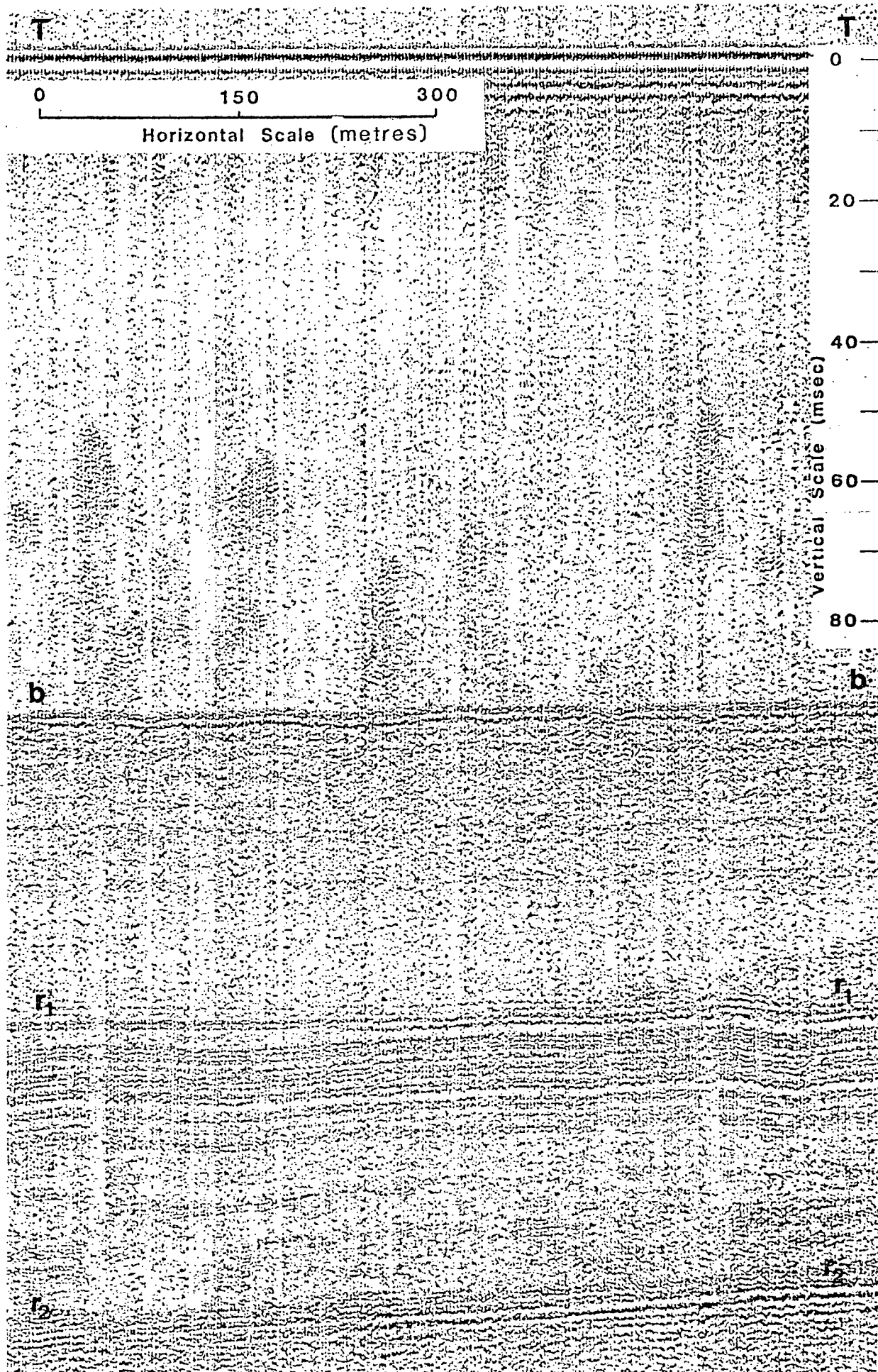


Fig. 7.28

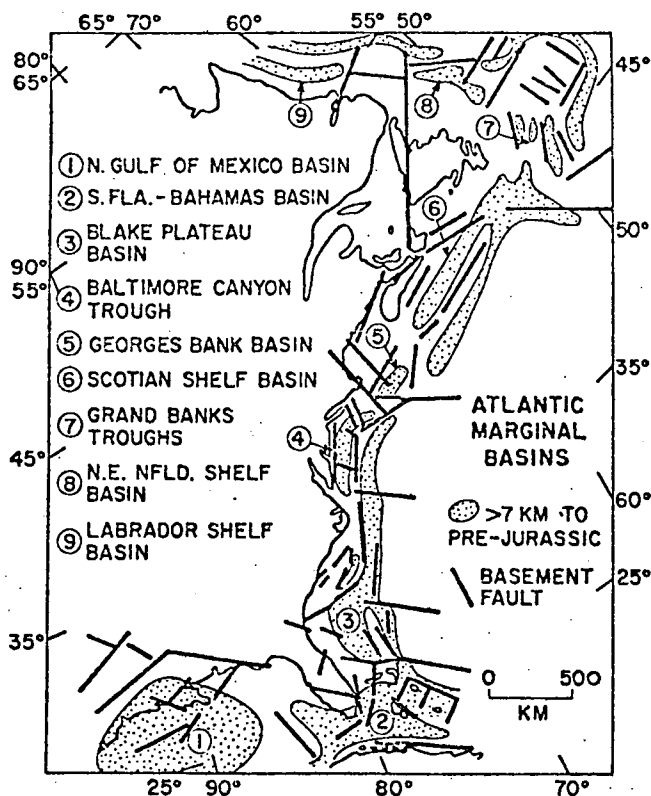
Vertical incidence section from the northern part of the Eastern Basin to show possible aeration in the water column, small pockmark-like indentations in the bottom and patchiness in the strata in the region between reflectors r_1 and r_2 .

sediment units appears patchy and small depressions (also evident on simultaneous sidescan sonar records) can be seen in the lakebed. With the same reservations expressed in section 3.8 (fig. 3.27) about observations of McQuillan and Fannin (1979), these features of fig. 7.28 may be taken to indicate possible sediment gasification. Other velocity values obtained in parts of the Eastern Basin by the graphical T^2-X^2 method (fig. 7.23) ranged from about the same value as the lake water to about 0.9 of that value, when the two layers seen in figs. 7.25 and 7.28 were taken as one unit.

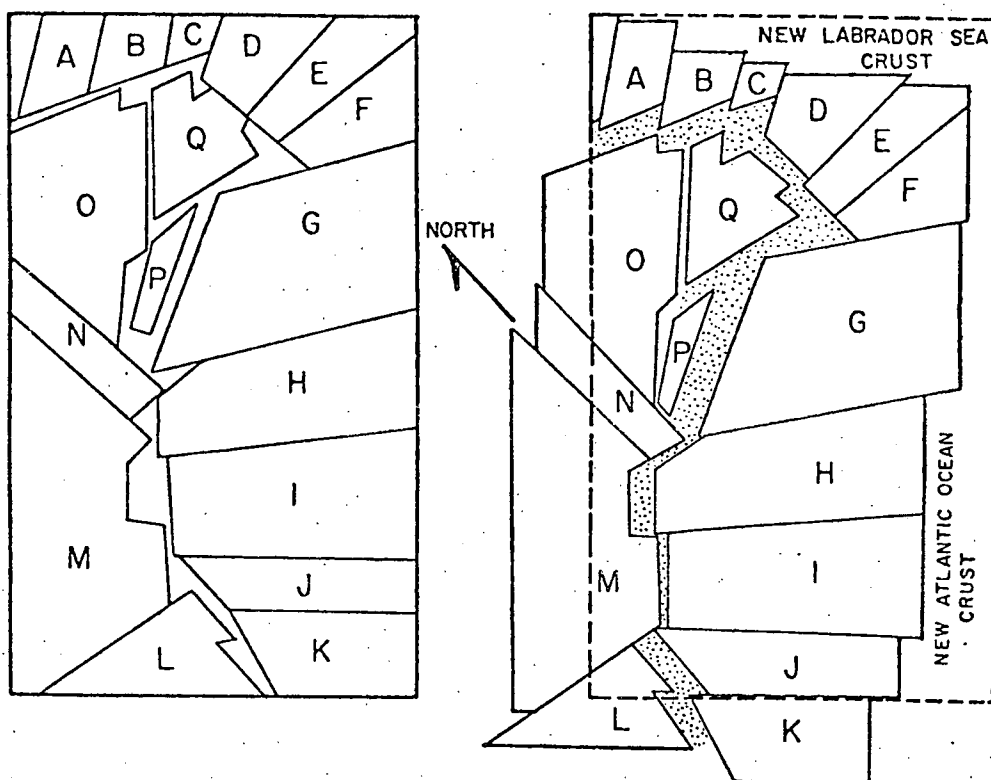
The low velocity results merit some discussion in terms of what little is known about the engineering properties of the sediment. Penetrometer readings were taken by C.F.M. Lewis (personal communication, 1983) during coring at a site in 32m of water in the northern part of the Eastern Basin at $42^{\circ}39.5'N$, $80^{\circ}03.0'W$. He found no measurable resistance to penetration through the top 7m of recent muds but measured a resistance of 0.15kg/cm^2 while penetrating a 15cm layer of "firm brown clay" underlying the muds. Further penetration was made into the underlying glaciolacustrine clay again without measurable resistance. The glaciolacustrine clays represent the earliest sediments of the Lake Erie basin and were deposited in a proto-lake ponded by the ice mass to the north (Thomas et al., 1976). Crustal tilting (Andrews and

Barnett, 1972) and the erosion of drainage sills during de-glaciation resulted in the subaerial exposure of these early deposits at around 12,000 years ago (Bryson et al., 1969; Sly, 1976). Lewis (cited above) attributes this drying out to the formation of the indurated zone which he encountered at the base of the recent muds. The possible negative velocity gradient suggested by the results (see caption to fig. 7.27) from the profile of fig. 7.25 may be interpreted in the light of the above findings as arising from the partial entrapment of gas migrating upwards from deeper known petrogenic strata (from which natural gas is tapped commercially) by a thin indurated layer separating the two sediment types. A possible difficulty with this hypothesis is that the interface between the sediment layers in fig. 7.25 is almost 20m deeper than the level which is suggested for the low-water stand by Sly and Lewis (1972).

Low velocities, indicative of sediment gasification and the possibility of negative velocity gradients suggest a lakebed giving poor foundation characteristics. This was confirmed by a small jack-up drilling platform which, during the summer of 1980, while working in the southern part of the Eastern Basin suffered severe subsidence which caused damage sufficiently serious to require the mounting of a salvage operation.



Major structural basins of the North American Atlantic continental margin geosyncline, showing interpreted basement faults (after Sheridan, 1974a).



Reconstruction of hypothetical unit structural block of Atlantic continental margin geosyncline and clockwise plate rotation which produces the interpreted basement fault motions (Sheridan, 1974b). The speckled pattern represents the depositional graben of the geosyncline.

Fig. 7.29

Tectonics of basin formation on the eastern margin of North America (Sheridan, 1976).

7.5 NEAR BOTTOM WIDE-ANGLE REFLEXION ON THE SCOTIAN SHELF.

The derivation of the acoustic velocity in a sediment layer from the average velocity in water and sediment column together was carried out in section 7.4 in discussing the data of fig. 7.23. It will be apparent from this derivation that the greater the ratio of the thickness of the water column to that of the sediment layer of interest, the greater will become the error in the sediment velocity obtained. In order to maintain a favourable ratio, it is necessary in water deeper than that investigated in Lake Erie to find ways of deploying the acoustic source and receiver close to the seabed. Some details of how this may be done are given by McKay and McKay (1982). The equipment adaptations necessary are described in section 2.4 and Appendix 1. Some field results with this equipment were obtained in parts of the Scotian Shelf, a part of the eastern continental margin of N. America whose general tectonic setting is shown in fig. 7.29, a setting which gives rise to the clastic wedge of fig. 7.30 along the outer margin (Jansa and Wade, 1975) and to intra-shelf basins such as the Emerald and LaHave Basins (fig. 7.31) where Pleistocene sediments overlie Cretaceous strata which has subsided in response to the movements associated with continental break-up.

Diagrammatic Cross Section of
Scotian Shelf Stratigraphy

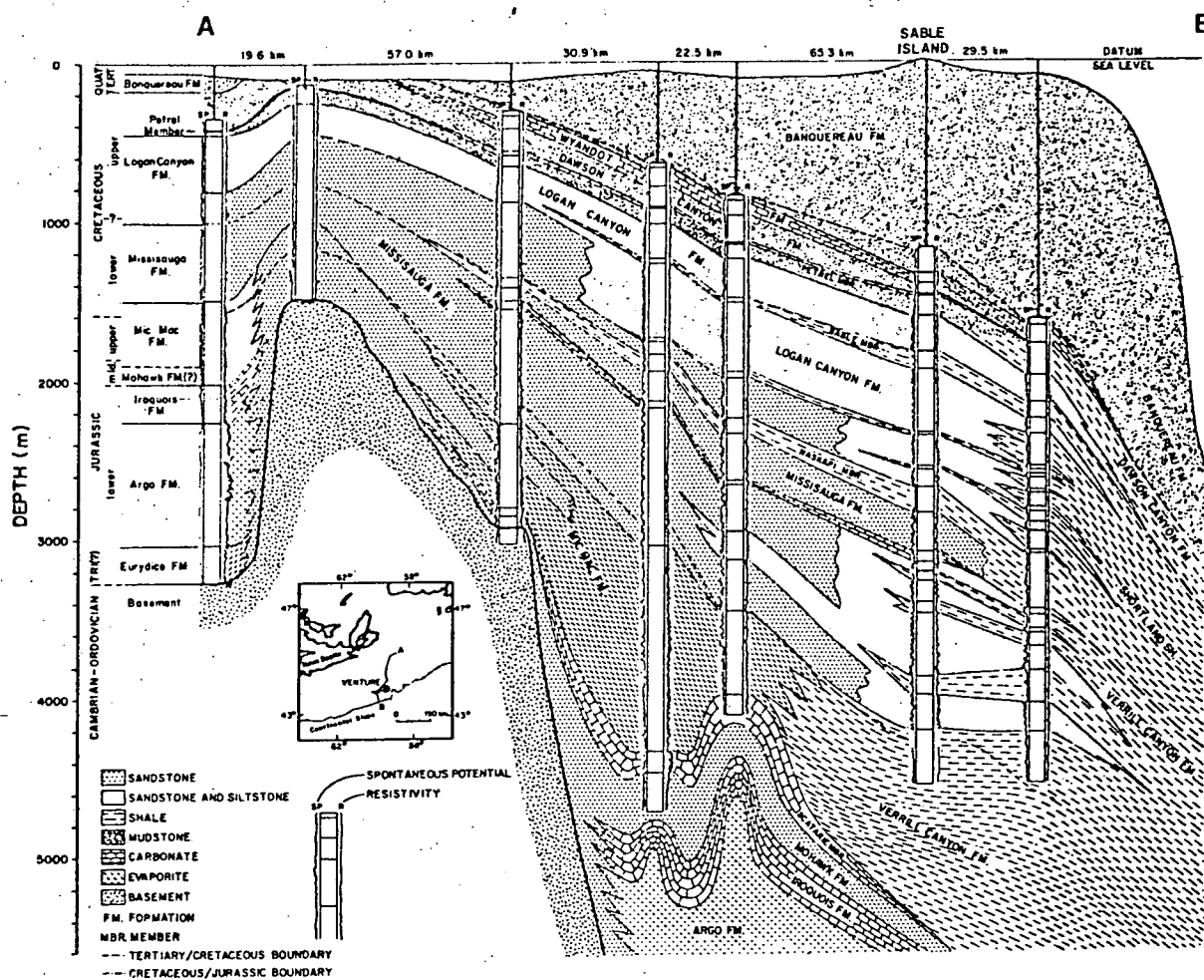


Fig. 7.30

General section of the clastic wedge on the ocean margin of the Scotian Shelf (Anon., 1983c).

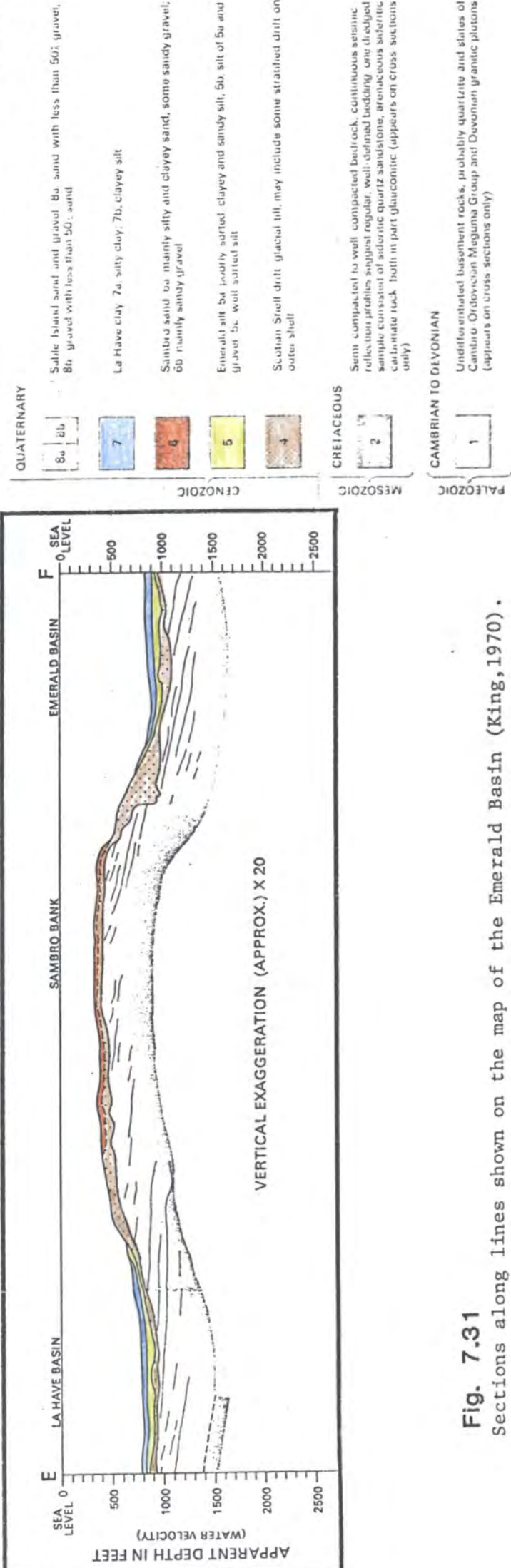
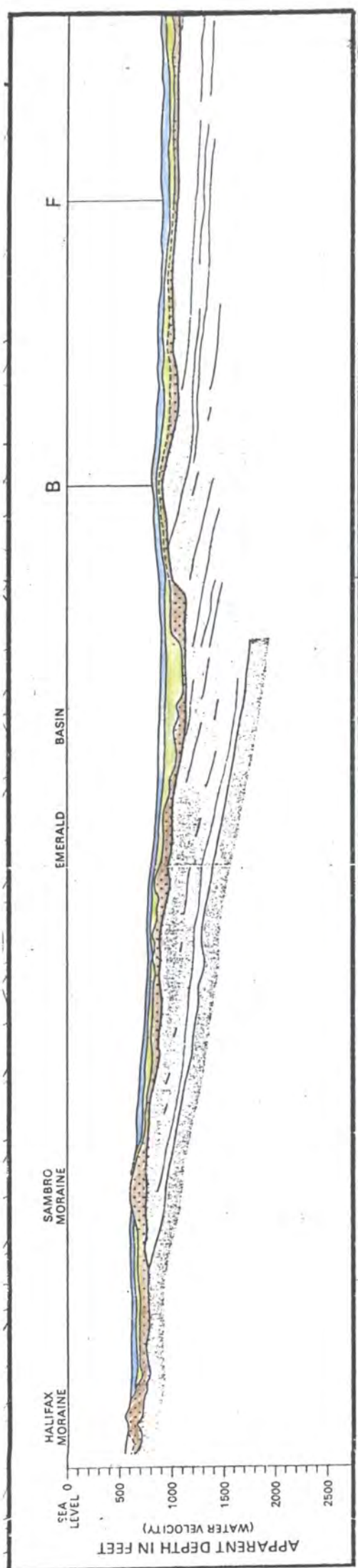


Fig. 7.31
Sections along lines shown on the map of the Emerald Basin (King, 1970).

The deep-tow wide angle reflexion seismic equipment was deployed over small sections of the Emerald Basin (fig. 7.32) and Sable Island Bank (fig. 7.33). Sections of the results obtained are shown in fig. 7.34 (vertical incidence) and 7.35 (switching wide angle). An examination of the vertical incidence profile (fig. 7.34) shows that irregularities in reflector interfaces occur through all the switching sequences, except in the case of no. 5, which exhibits good horizontal layering. This was chosen for interpretation, while the others were discarded. The interpretation method is basically the $T^2 - X^2$ method described in section 7.4 with the additional requirement that corrections are calculated for the height of receivers above the seabed (McKay and McKay, 1982). A programme designed to carry out this data reduction appears in Appendix 3, while a $T^2 - X^2$ plot for the interfaces 1 to 5 shown in fig. 7.34 is given in fig. 7.36 and the output of the data reduction programme appears in fig. 7.37. The programme includes a facility to reject successively the point which is the most deviant from the straight line fitted in $T^2 - X^2$ space, which is the reason why several lines of figures appear under the headings 'average velocity', 'interval velocity' and 'average (velocity) to this reflector'. Only the first line in each set of figures refers to a line fitted through all points picked, and only it should be regarded as the proper result. It is apparent that the interval

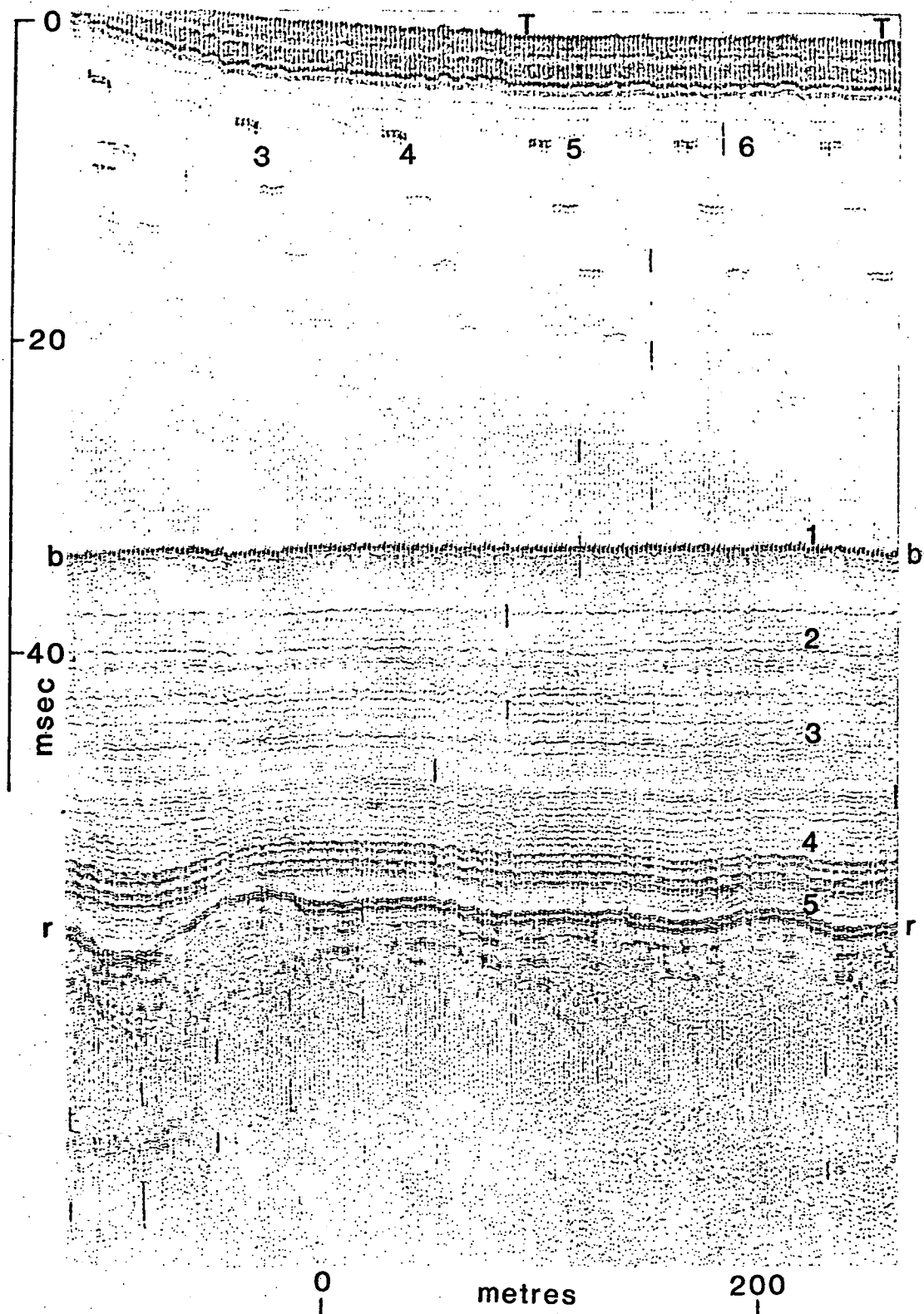


Fig. 7.34

Vertical incidence profile over area of deployment of the 5-channel switching hydrophone array in the Emerald Basin.

- T - trigger event
- b - bottom reflection
- 1, 2, 3, 4, 5 subbottom reflectors

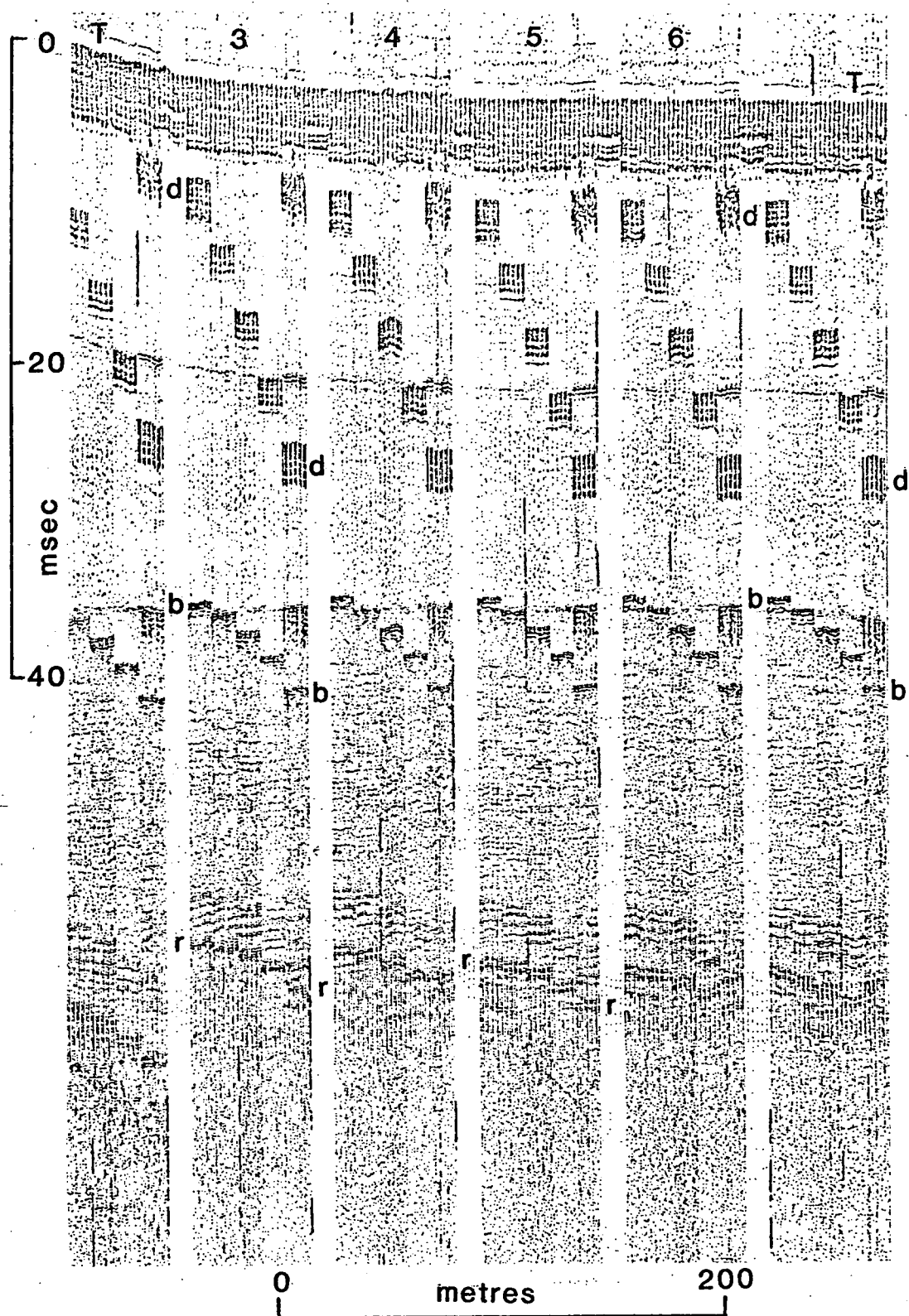


Fig. 7.35

Output of switching hydrophone array (Emerald Basin).

- T - trigger event
- b - bottom reflection
- r - subbottom reflection

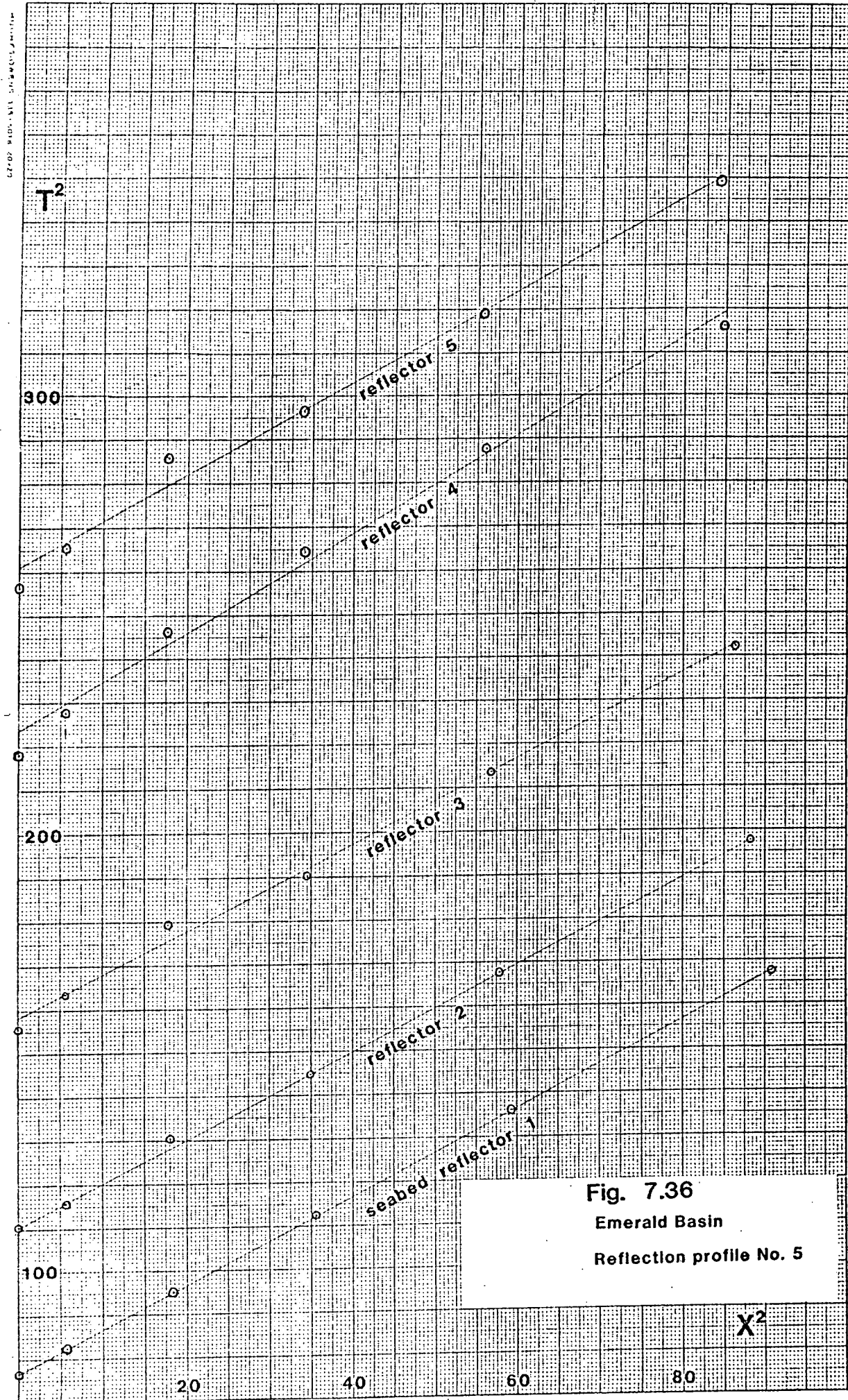


Fig. 7.36
Emerald Basin
Reflection profile No. 5

velocities determined in this way are subject to quite wide errors, and the true value of the results are perhaps better seen by comparing the 'average velocity' (in water and sediment) for reflector no.5 with the 'average (velocity) to this reflector' (in the sediment, with the effect of the water column removed). It can be seen that the error in the sediment velocity is only about twice that of the average velocity - because the sediment and water columns are about equal in length. Since the total water column at this location in the Emerald Basin is ca. 260m and the sediment layer ca. 15m, even if an average velocity (in water column and sediment) could be determined with the same precision of ca. $\pm 2.5\%$ by measurement from the sea surface, such a measurement would lead to no better accuracy than $\pm 43\%$ for the velocity in the layer between the seabed and reflector 5. It is thus easy to see the power of the deep tow approach to this problem.

Although the approach of travel time analysis is essentially kinematic, consideration of the physical model of the sediment used is of consequence in assessing the contribution to errors arising from phase shifts on reflexion at interfaces. This is discussed in McKay and McKay (1982) in terms of the Rayleigh-Brekhovskikh reflexion coefficient derived for a lossy elastic model, and does not appear to be a serious problem when the

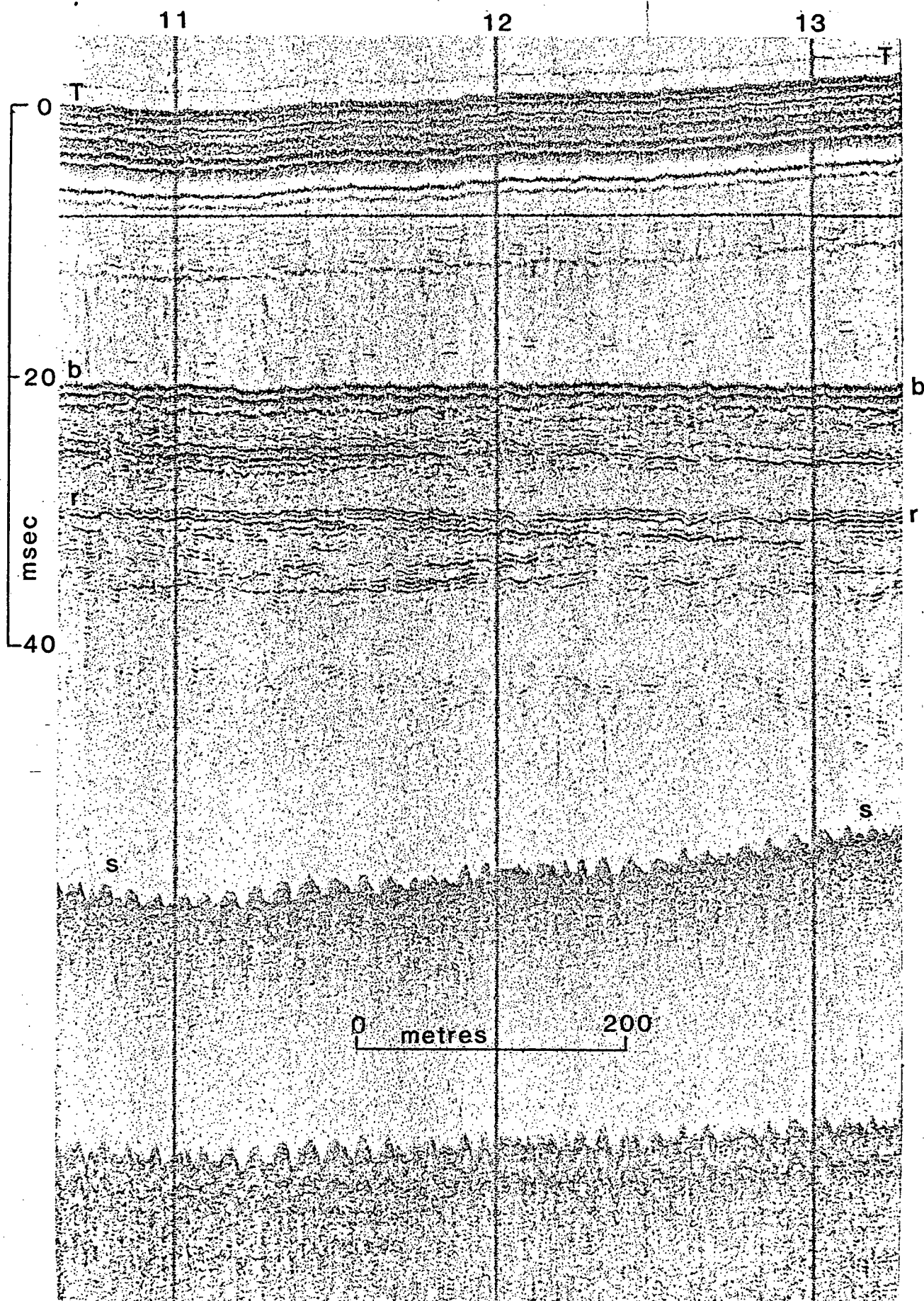


Fig. 7.38

Vertical incidence profile over area of deployment of the 5-channel switching hydrophone array on Sable Island Bank.

- T - trigger event
- b - bottom reflection
- r - subbottom reflection
- s - surface reflection

experimental precision of time measurement is no better than has yet been achieved with deep tow equipment. Even if very different phase shifts were predicted by the Biot-Stoll porous model, the natural limit of an error of 180° in phase does not lead to corrections which would greatly alter the results at the accuracies obtained. Bennell (1978) in working in shallow water obtained better time resolution and claimed accuracies about an order of magnitude better than those of fig. 7.37. At this level of precision, the phase shifts and physical model would become important, but Bennell appears to have disregarded them.

Most of the data obtained by the switching deep-tow equipment has been dealt with by treating as a single layer that part of the sediment column which is bounded by the deepest readily discernible horizontal reflector (in the case of fig. 7.34, reflector no.5). Data (figs. 7.38 and 7.39) obtained over Sable Island Bank (fig. 7.33) has been treated in a similar way. Results are shown in fig 7.40. An alternative related method for gathering wide angle information used three hydrophones whose output was recorded continuously instead of being switched on and off. This method has the advantage that common mid-points can be selected to avoid the rather stringent requirement of the very simple patterns of sediment layering needed by the switching method. Data collected in this way is shown

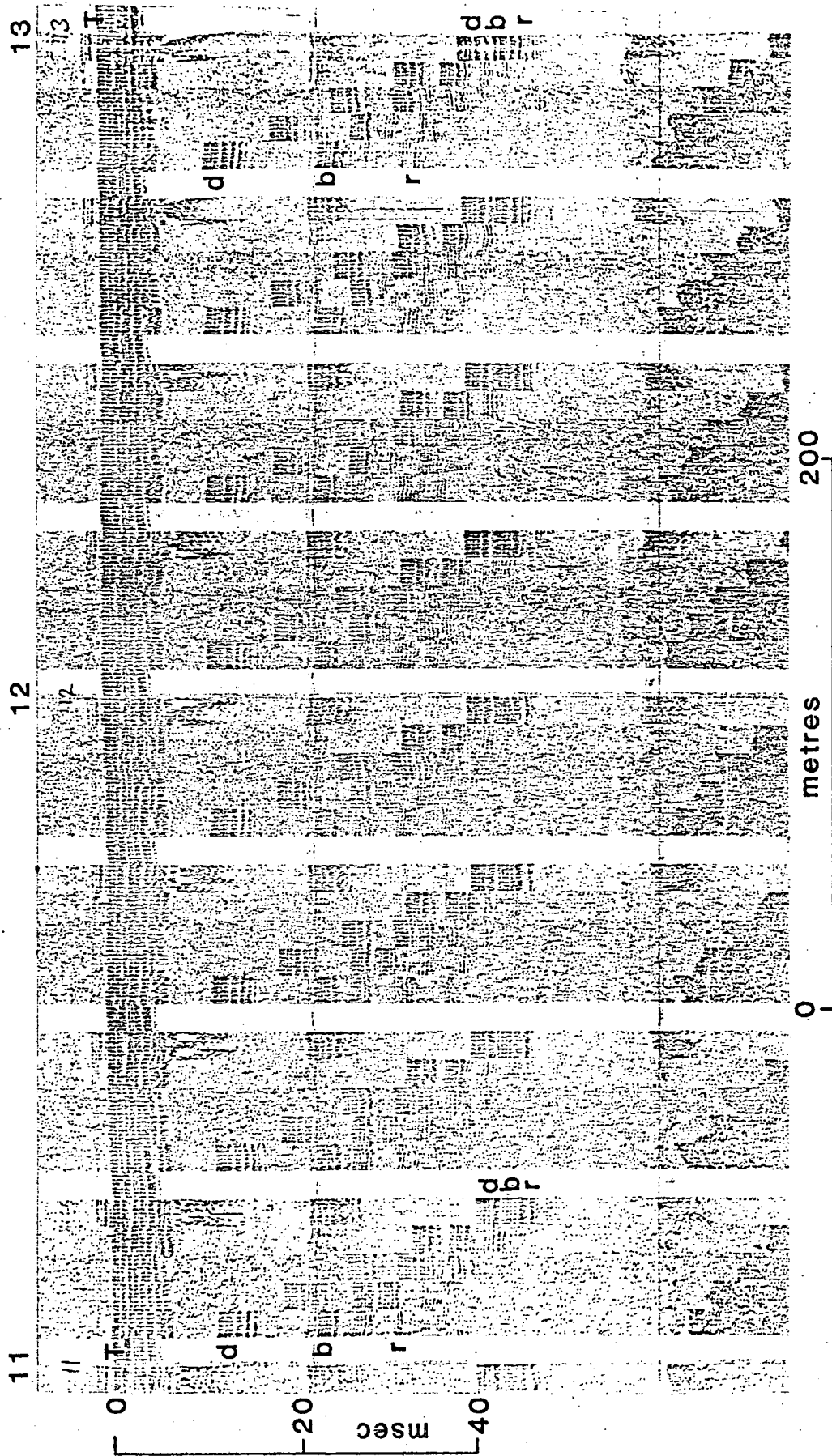


Fig. 7.39

Output of switching hydrophone array (Sable Island Bank).

- T - trigger event
- b - bottom reflection
- r - subbottom reflection

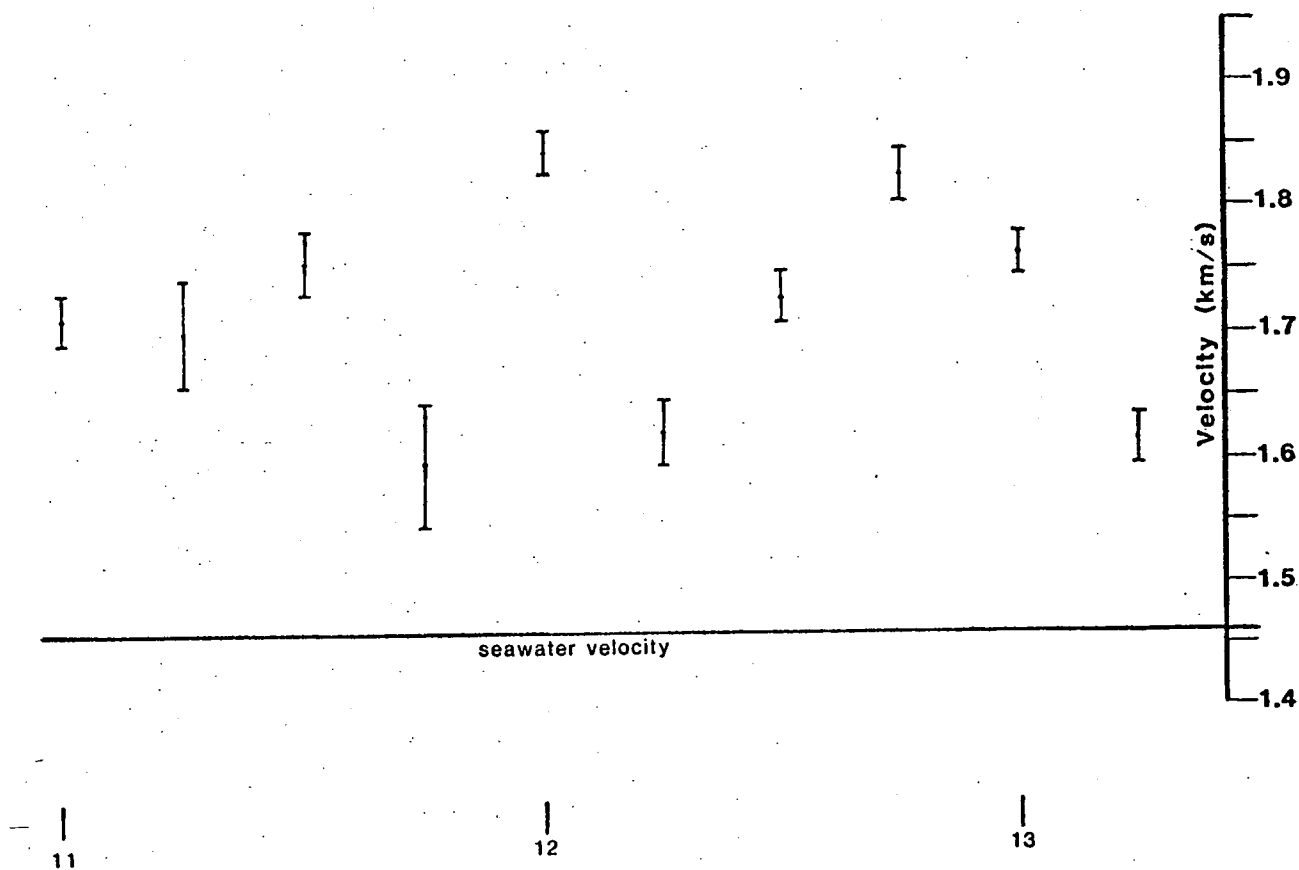


Fig. 7.40

Velocity values determined from deployment of switching array on Sable Island Bank.

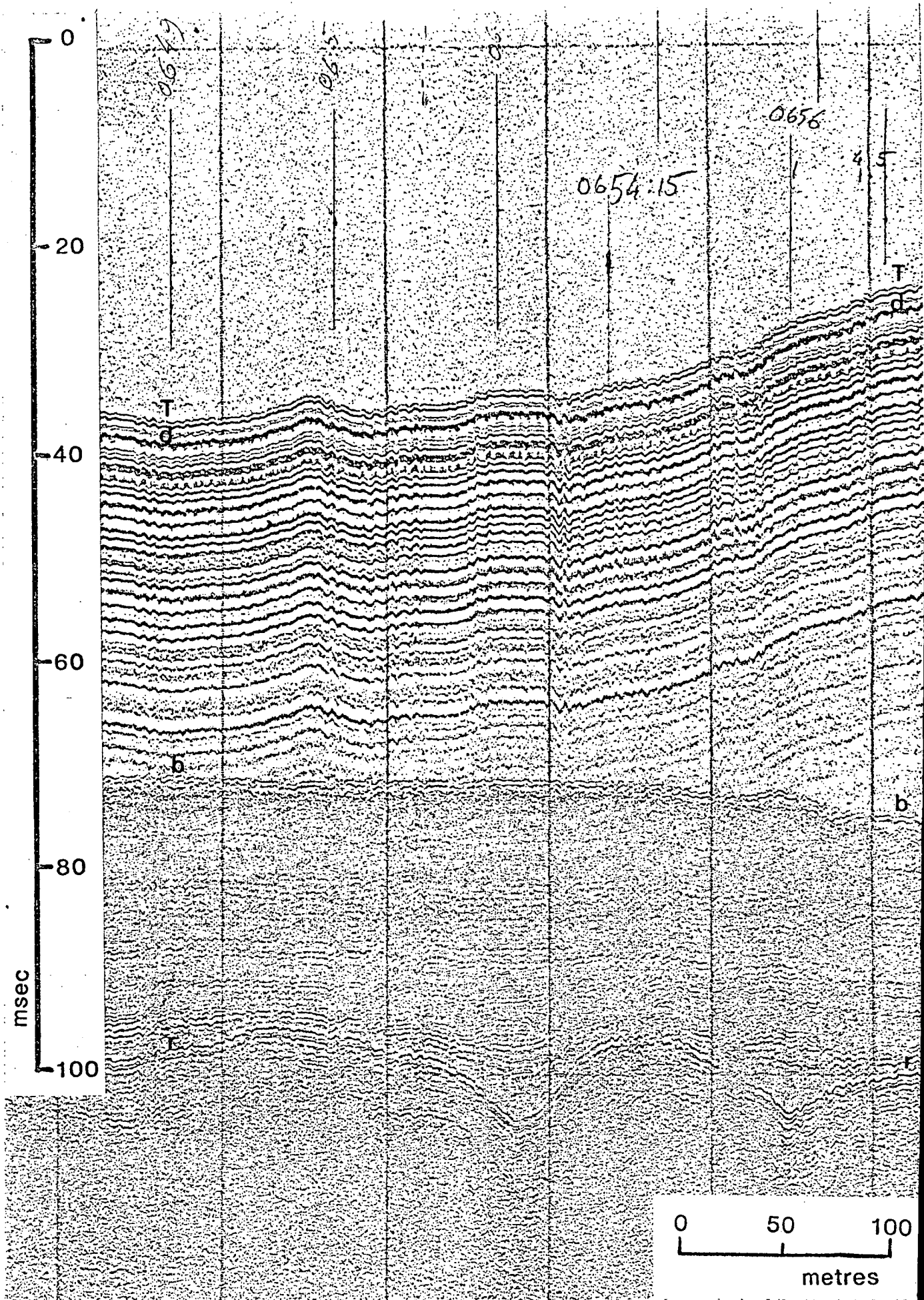


Fig. 7.41

Vertical incidence hydrophone of 3-hydrophone assembly. (Emerald Basin)

in figs. 7.41, 7.42 and 7.43, and results are displayed in fig. 7.44.

No other velocity results are known for the Emerald Basin sediments, but Hunter et al. (1980) has carried out seabed refraction studies ^{on Sable Island Bank} on an appropriate scale to define the velocities in surficial sediments. The locations at which he carried out these studies are shown in fig. 7.33, and the results he obtained appear in fig. 7.45. There is at least broad agreement between the two sets of results.

It is interesting to compare the results from the Emerald Basin with other information which is available about that location. Vilks and Rashid (1976) who took cores in the Emerald Basin to study the foraminifera and organic geochemistry found at a location in the N.E. of the basin, concentrations of methane of up to 1.4% by volume (at atmospheric pressure), with no guarantee that none escaped during the sampling procedure. Restored to a pressure of 25 atmospheres, which would correspond to the ambient pressure at the seabed, this fractional volume would be reduced to ca. 0.06%, a volume which would produce (section 3.4) a substantial decrease in the velocity of sound propagation to about 75% of the value in a water-saturated sediment. In the light of the velocity findings of fig. 7.37 which suggest a negative velocity

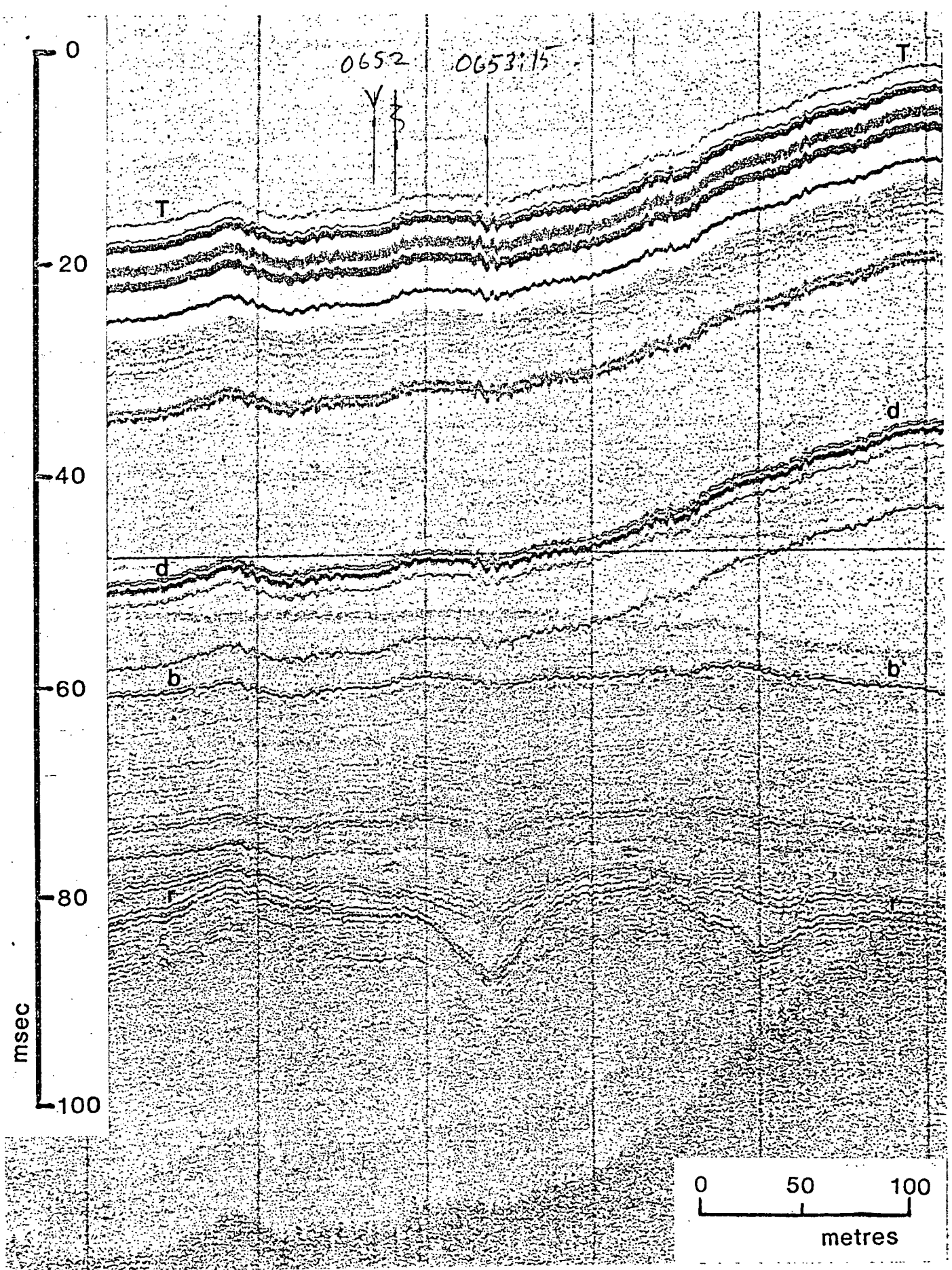


Fig. 7.42
Middle hydrophone of 3-hydrophone assembly. (Emerald Basin)

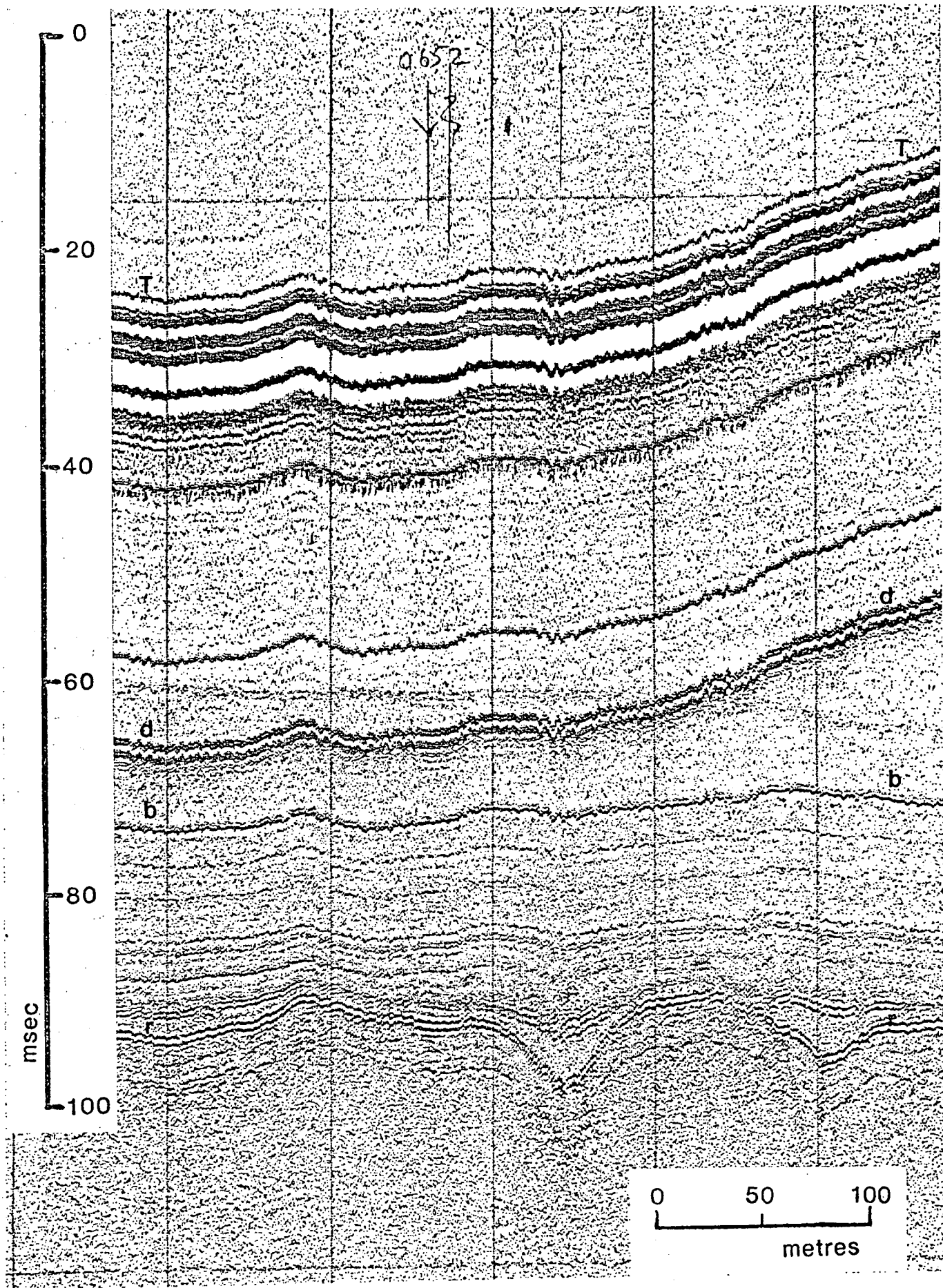


Fig. 7.43

Trailing hydrophone of 3-hydrophone assembly. (Emerald Basin)

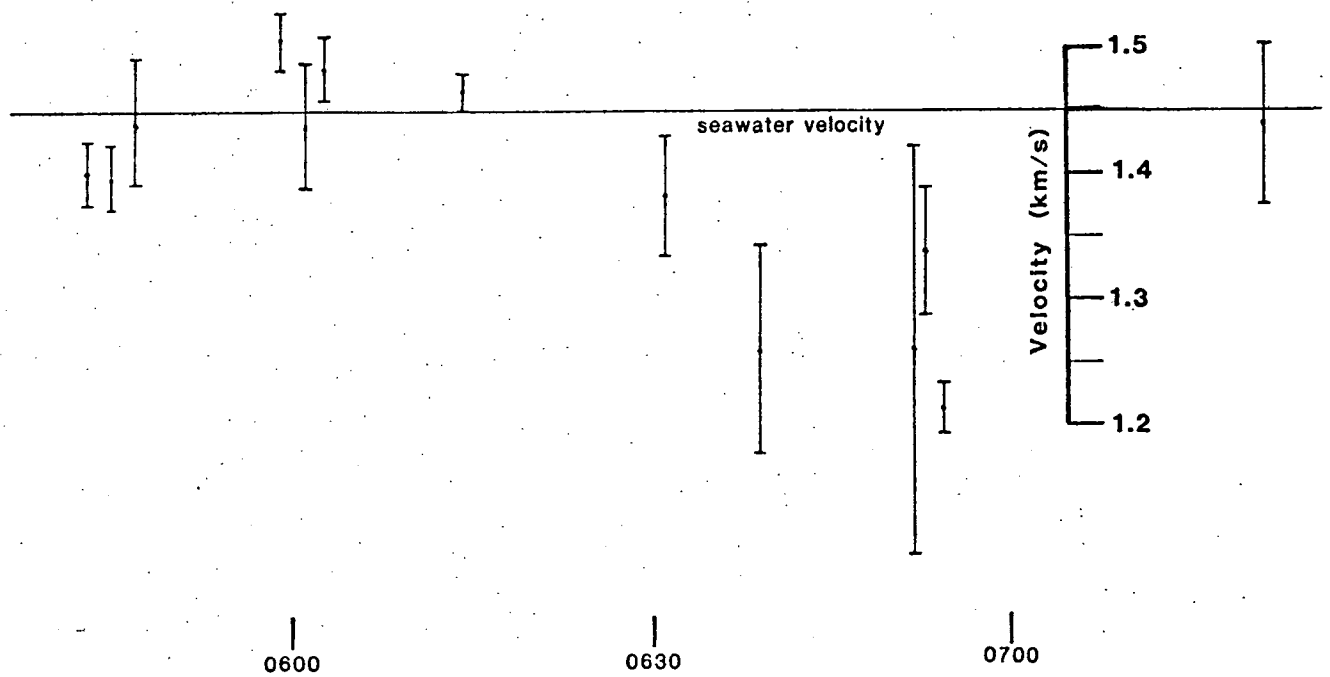


Fig. 7.44

Velocity values determined from deployment of the 3-hydrophone array in the Emerald Basin.

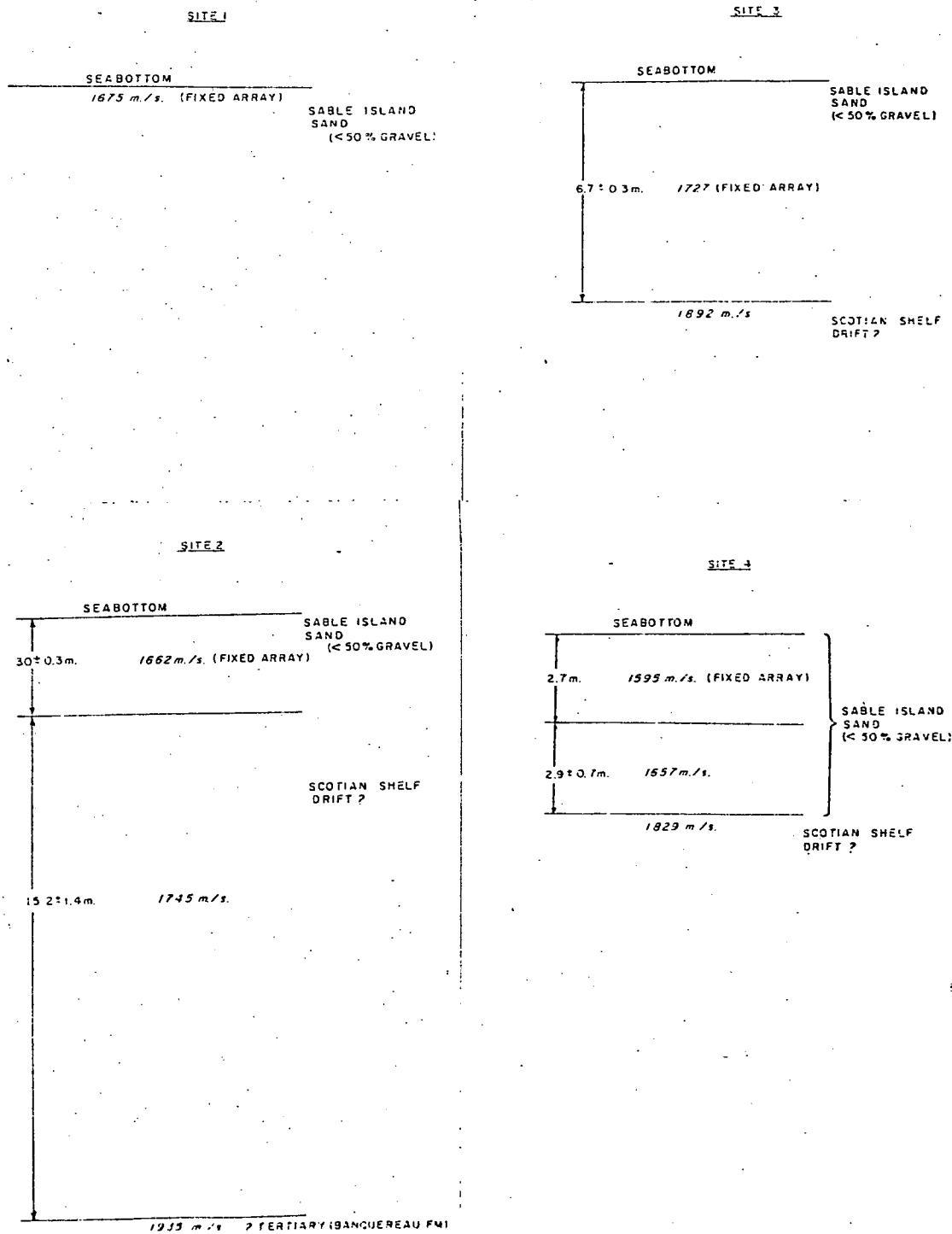


Fig. 7.45

Results of the refraction measurements made by Hunter et al. (1980) at the sites 1, 2, 3 and 4 on the Sable Island Bank.

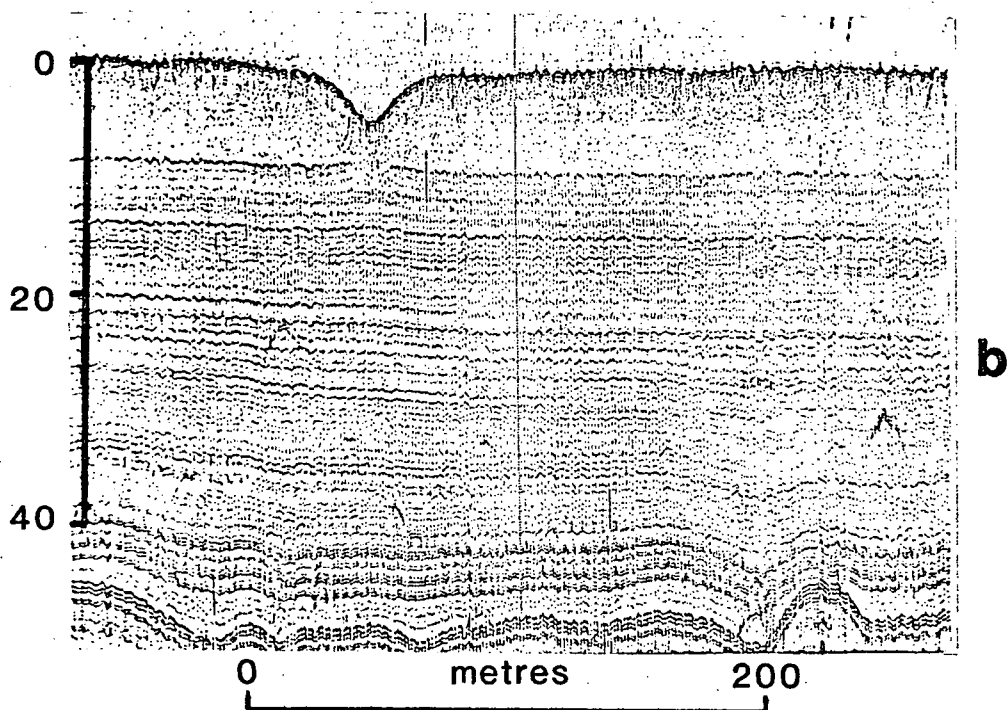
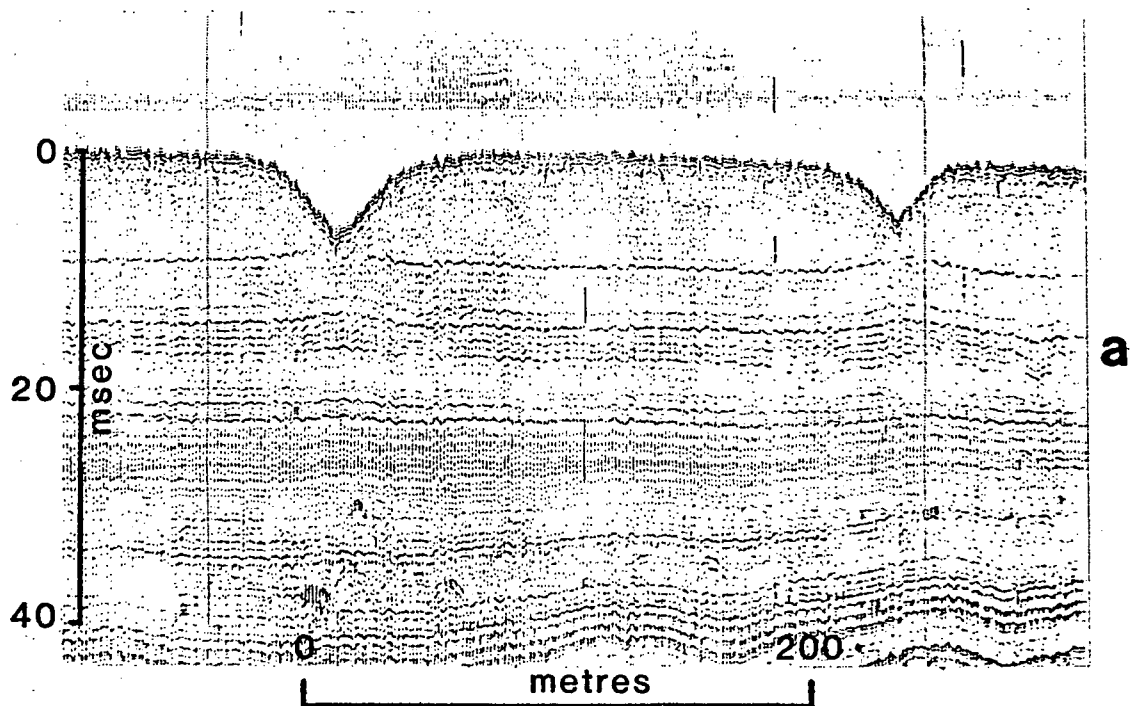
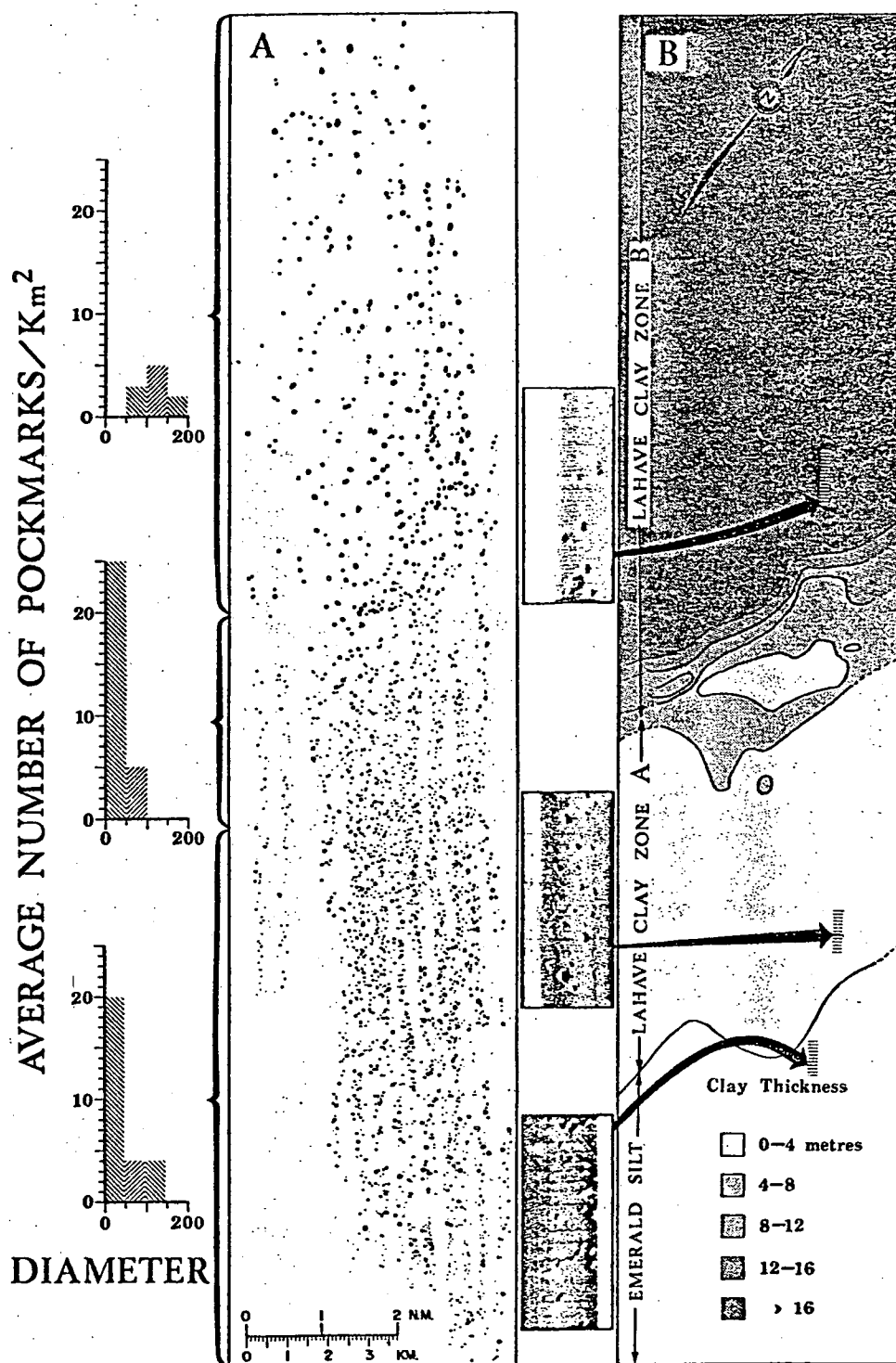


Fig. 7.46

Pockmarks in the Emerald Basin.

gradient, it is interesting to re-examine vertical incidence profiles such as those of 7.46 obtained within two kilometres of the location of fig. 7.34 where the velocity measurement was obtained. Fig. 7.46 shows well developed pock marks, such as have been described (see figs. 7.47 and 7.48) by Josehans et al. (1978) who appear to have assumed that the disturbances below pockmarks in the sediment layering seen on acoustic profiles was necessarily real. An alternative possibility is that the upward flexures in the first clear subbottom reflector (fig. 7.46) arise because of a low velocity in the sediment layer just beneath the seabed. This is not in accord with the higher velocity measured (fig. 7.37) in the immediate subbottom at the location of fig. 7.34, but is compatible with the low velocities observed in fig. 7.44. Although the measurements are insufficient to allow certainty, it may be proposed that pockmarks of the type shown in fig. 7.46 form not merely by the gradual steady leakage of gas from the seafloor as originally proposed by King and MacLean (1970) and accepted by Josenhans et al. (1978) but that they form in a more catastrophic manner by the buildup and sudden release of gas, somewhat in the manner of the formation of the beach pockmarks described in section 3.6. Such a buildup of gas requires the existence of a less permeable layer overlying a more permeable layer so that gas can accumulate in the lower permeable layer. The velocity results of fig. 7.37

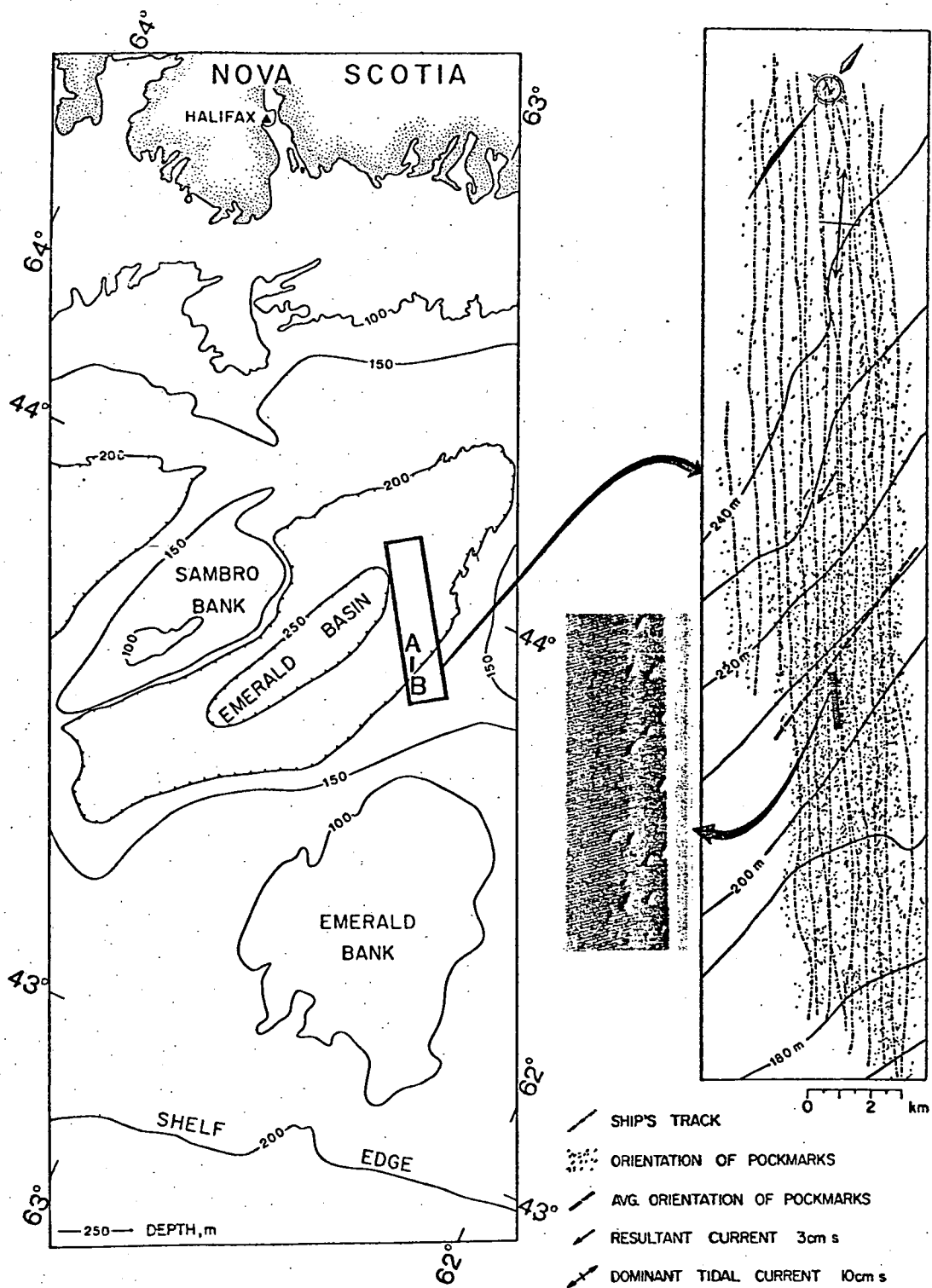
POCKMARK MOSAIC CLAY THICKNESS



Section A (see Fig. 7.48) shows a mosaic representing approximately 80% of the sea floor traced from scale-corrected side-scan sonar records. The linear gaps on the mosaic indicate areas of no coverage that resulted from difficulty with navigation and with guiding the ship. Note the three zones of different pockmark size and density delineated by the brackets and their correlation with clay thickness. The side-scan sonar records illustrate the typical pockmark size and distribution found within each of the three zones.

Fig. 7.47

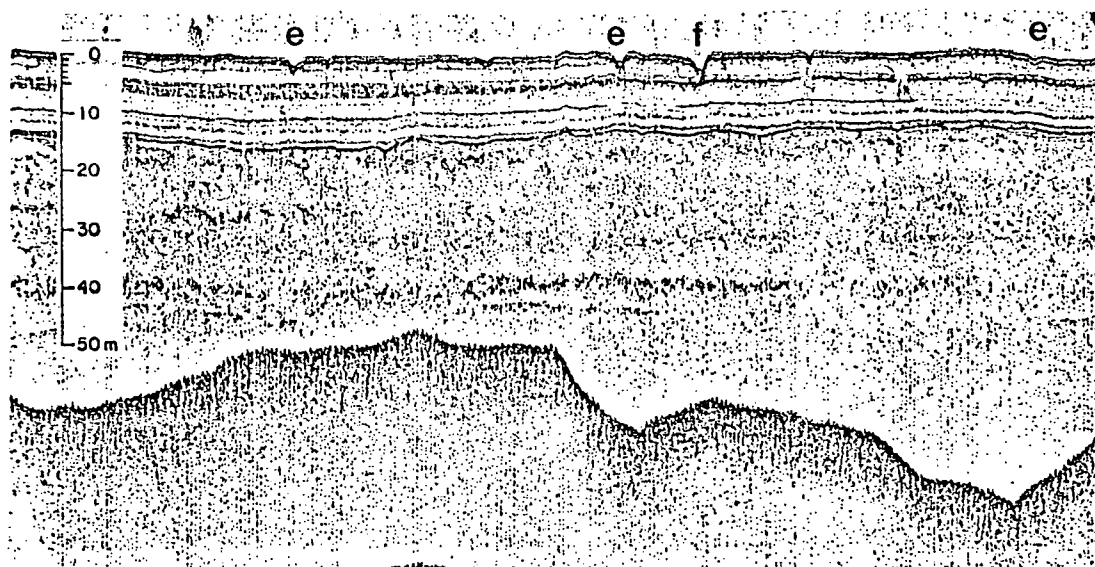
Pockmarks and clay thickness in the Emerald Basin
(Josenhans et al., 1978).



Index map. Enlarged view of study area showing bathymetry, ship's track, local bottom currents, orientation of pockmarks and a typical side-scan sonar record illustrating the orientation of pockmarks.

Fig. 7.48

Location of the work of Josenhans et al. (1978).



NOVA SCOTIA RESEARCH FOUNDATION CORPORATION

DEEP TOW PROFILING SYSTEM

Water Depth 130 meters.

Soft clay silt with modern and relict pock marks covers a sequence of glacial clays.

Location: Forties Area, North Sea

Tow Depth: 30-40 meters

Acoustic Source: Single electrode
sparker

Tape replay record copyright Institute of Geological Sciences.

Fig. 7.49

Pockmarks in the Forties area of the North Sea.

suggest that this may be occurring. Further evidence in support of this idea is given by Josenhans et al. (1978) who have observed that the size of pockmark which predominates in any one area depends on the thickness of clay cover at the seabed, although they offered no explanation for this variation. In terms of the gas-buildup idea, a partially permeable upper layer would be more effective at gas containment the thicker it became, so that pockmark size would be expected to follow such a pattern. This notion would also allow the possibility that the capping clay layer would change in its velocity characteristics depending on the gas pressure gradient across it, so that the highest velocity found in the immediate subbottom at the location of fig. 7.34 is not incompatible with a lower velocity in the same layer as suggested by the above interpretation of interface flexures beneath the pockmarks shown in fig 7.46.

Although no velocity information is available from the North Sea pockmark fields, some indication of flexures similar to those of fig. 7.46 can be seen in a figure of McQuillan and Fannin (1979) reproduced in fig. 7.49. Upward flexures are suggested below the pockmarks labelled 'e'. The largest pockmark in this profile, at 'f', does not, however, seem to exhibit this phenomenon.

7.6 SUMMARY

The travel-time methods described in this chapter allow conventional investigations of this type to be carried into deeper water by the application of the deep-tow principle, and do not require the acceptance of a particular mechanical model for the seabed material because the approach is purely kinematic.

Possible mechanical models are discussed in the next chapter with the aim of identifying one which would be suitable for a more detailed treatment of acoustic data obtained in surficial sediment studies.

8. SEDIMENT MODELS AND SEISMOGRAMME CONSTRUCTION.

8.0 INTRODUCTION.

A sediment is a complex substance comprising several fluid and solid phases which may be unstable and may transform from one to another by such processes as were described in chapters 3 and 4. While the mechanical behaviour of a mineral grain may be described elegantly in terms of its crystalline structure and the compressive response of a gaseous pore fluid may fit well with the kinetic theory, an amorphous assemblage of such substances together with water (which, in the liquid state can prove difficult to treat nicely even in its uncontaminated form) poses problems for the theorist. There is no universally accepted mechanical model for a sediment. Two alternative general approaches to the problem seem to be available: to assign elastic and loss parameters to the material on the basis of its bulk response without regard for the details of the processes occurring within its structure or else to try to model such processes and accept that empirical constants must be added to measurable parameters to complete the model. Neither approach is elegant, and the choice is best made on the basis of immediate utility.

8.1 MECHANICAL MODELS FOR UNCONSOLIDATED SEDIMENTS.

The extension of the classical model of an elastic continuum wherein energy dissipation is included by the mathematical device of allowing complex values for elastic moduli and wave numbers (e.g. Brekhovskikh, 1960), implying attenuation varying linearly with frequency (i.e. a 'quality' factor 'Q' independent of frequency) is the model often used in analysing the results of ultrasonic material testing (e.g. Folds and Loggins, 1977; Barnard et al., 1975) and normally accepted as valid for most solid-earth materials, e.g. Bullen (1965), Officer (1974), Ben-Menahem and Singh (1981), and is applied to unconsolidated sediments, e.g. Chapman and Ellis (1980), Hamilton (1971, 1976a and b, 1980), Hamilton and Bachman (1982), Fryer (1981) and Vidmar (1980a,b; 1983), although one recent author in dealing with partially consolidated sediments has proposed a simple pure elastic model with no account of losses (Walker, 1978).

Within the context of this model, the manner of treating boundaries between adjacent physically different parts of the propagating medium depends on the scale of the boundary. In the case of small scale boundaries such as those occurring between sand grains and interstitial fluid in a sediment or rock, the model ignores the local deformations occurring at these interfaces, allowing their

effects to be represented by the averaged elastic and loss parameters assigned to the medium as a whole. An example of such an averaging process is given by Wood's (1931) treatment of emulsions (see section 3.4). In the case of boundaries on a much larger scale such as separate stratified sedimentary units from one another, a mathematical treatment of reflexion and transmission becomes possible in the manner described by Ben-Menahem and Singh (1981, p.125-150) for plane waves or by Brekhovskikh (1980, p.286-298) and Ivanov (1975) for spherical waves, on the basis of earlier work by such as Knott (1899), Muscat and Meres (1940), Merkulova (1970), Thompson (1950) and Koefoed (1962) and the extensive work of Rayleigh (1955).

One of the difficulties with this model as applied to unconsolidated sediments has been pointed out by McCann (C., 1966), who noted that values of attenuation found by measurements made at different frequencies on laboratory samples of sediments having a mean grain diameter greater than 0.007cm. did not agree with the notion of a 'Q' which was independent of frequency. He attributed the observed dissipation to 'solid friction' or inter-granular scraping between the solid particles.

A further problem arises from the observation by Paterson (1956) and Plona (1980) of a second compressional

wave which propagates in porous materials like seabed sediments with a velocity about an order of magnitude lower than that of the ordinarily observed compressional wave. The model cannot accommodate this observation. The problem of the interaction of two continua with one another has been investigated by Truesdell and Toupin (1960) and elaborated by Green and Naghdi (1965) and Green and Steel (1966). This work relates to diffusion processes and includes thermodynamic effects and non-linearity in mechanical properties. In application to sediments or other porous media, it suffers from conceptual difficulties in the initial postulates that each point within the mixture of continua can be occupied simultaneously by each continuum. This leaves undefined the parameters of porosity and permeability, as normally used in characterising a sediment, a point which, coupled with the lack of any treatment by these authors of wave motion in the continua, renders this approach somewhat less than immediately applicable to the investigation of seabed sediment properties.

Preliminary attempts by Rivlin and Ericksen (1955) to re-formulate the general relations between stress and deformation were later developed by Bernstein (1960) and Noll (1958, 1972, 1973) who carried the formulation to a further degree of generality, incorporating his own (Noll, 1955) earlier work on the continuity of the solid and

fluid states. One of the principal differences in outlook from classical theory is in Noll's use of the principle of determinism which, in this context implies that the present stress is determined by the entire past history of deformation. This differs from the treatment of relaxation processes in visco-elastic materials. 'History' implies a knowledge of all deformations extending infinitely far into the past, rather than merely the relaxation time required for a transient strain to reach an equilibrium value, (Leitman and Mizel, 1970). Since the mathematical concept of an infinite past is itself unacceptable for real materials, modifications have been proposed by Noll (1973) to allow for finite past histories, in part by the device of introducing the idea of materials with fading memory. This approach does appear to have led to useful predictive models of polymers, commonly referred to in the literature as "rubber-like solids" or "rubber-like fluids" (e.g. Lodge, 1964). This field has proved to be amenable to the development of a theory (Treloar, 1958; Parrot, 1969) which is able to relate the bulk mechanical properties of a polymer to its molecular structure, on the basis of applying modern mathematical thinking on continua to the chemical model of such a substance, which is considered to be made up of long chains of monomers which interact very little with one another except at points of cross-linkage where they are strongly bonded either chemically, or by

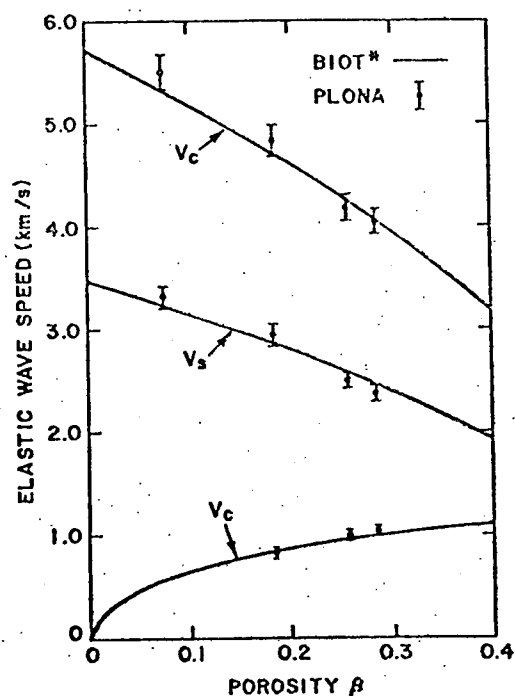
physical entanglement. Such materials are presently the most complex for which the useful prediction of bulk from microscopic properties can be made.

Noll's concept of 'fading memory' in a material is one which is well attuned to the general thinking on the consolidation of sediments and to the oedimeter testing of soils to establish such conventional engineering properties as over-consolidation (e.g. Terzaghi, 1944; Taylor, 1948) wherein the soil can be said to have remembered being loaded to a greater extent than at present. The memory is a fading one because all cycles of such possible loading (as in glaciated lands) and accompanying diagenetic and lithification processes (which may lead eventually to the transformation of the soil into a metamorphic rock) although contributing to the present state of the material cannot be deduced from the testing of its mechanical properties.

The extreme of seeking a mechanical model which would carry through the description of a continuum all the way from a suspended sediment to a phyllite carrying with it the necessary thermodynamic (including chemical) processes may seem fanciful, but Truesdell (1980) has pointed out that some rheologists would not think such an aim to be unattainable. These considerations serve mainly to show that the criteria for choosing a mechanical model to use

for a sediment are essentially subjective, and are to be governed by the usefulness of the model in bringing together such parameters as the modeller wishes to deal with, over time scales which are respectively appropriate for the continued existence of the sediment and the carrying out of the modelling process.

Although superficially analagous in moulding properties to some plastics, marine and lacustrine clays are quite different from these polymers in microscopic properties. It seems unlikely that the models of Treloar and of Parrot for rubber-like substances could be of use, but it may be that some relationships between a gross property like the plasticity index of soil mechanics and the electrical binding forces between clay particles could be found to be sufficiently analagous with the structure of a cross-linked polymer to make the chemists' models attractive for the treatment of cohesive sediments. Such an analogy seems quite impossible for coarser sediments and any detailed treatment of sands in particular, calls for a model which recognises the intergranular flow of pore water even on the time scales of the period of quite high frequency seismic signals. Such an approach was suggested by Biot (1956a,b) on the basis of earlier work on porous sound absorbing materials by Beranek (1947) and Morse (1952) and later further generalised by him (Biot, 1962a,b). The application to sound absorbent materials



Comparison of theoretical predictions and experimental results for speeds of the fast compressional wave, the shear wave, and the slow compressional wave. Input parameters to the model are $K_s=0.407$ Mb, $\mu_s=0.297$ Mb, $\rho_s=2.48$ g/cc, $K_f=0.022$ Mb, $\rho_f=1.00$ g/cc, $\eta=1.00$ cs, $f=500$ kHz, $\kappa_0=2 \mu\text{m}^2$, $a_0=0.02$ mm, and $\beta_0=0.20$.

Fig. 8.1

Comparisons between Biot (1956a,b) theory and the experimental laboratory data of Plona (1980) for a synthetic sediment as compiled by Berryman (1981). The parameters mentioned above are defined in table 8.1.

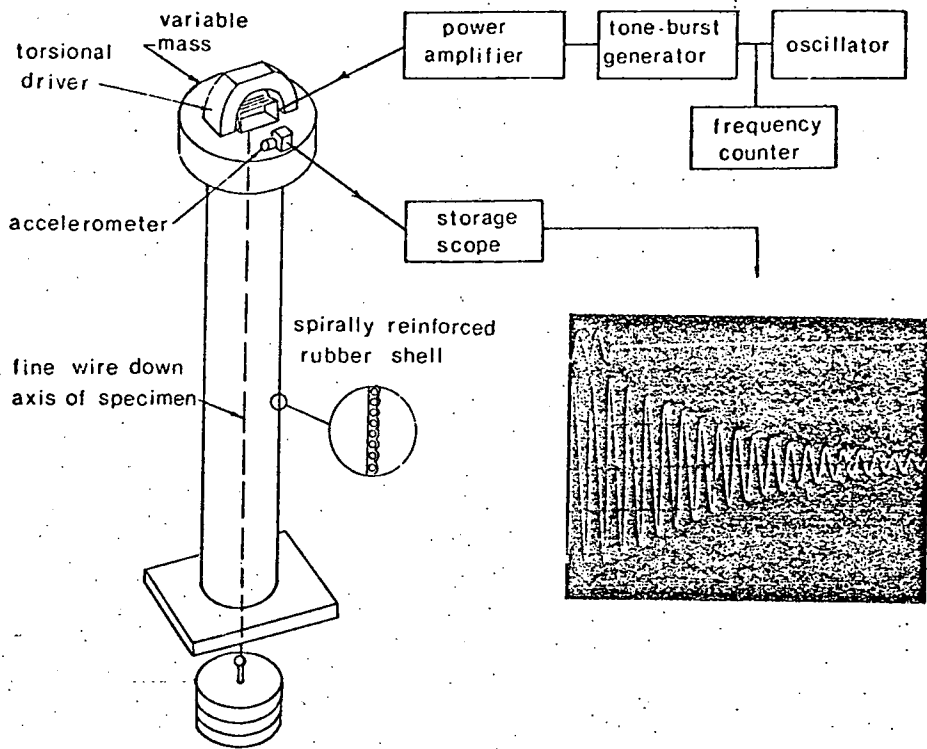
has been discussed recently by Bliss (1982) and by Attenborough (1983) while the application to sediments has been developed by Stoll (1974, 1977, 1980) and reviewed by Brunson and Molinelli (1982). The mathematical formulation of the theory is given in the Appendix to Stoll's paper (1980). The model predicts the existence of a second compressional wave (fig. 8.1) as has been observed in laboratory studies by Paterson (1956) and by Plona (1980). Stoll (1977, 1980) notes that this model allows mechanisms for energy dissipation in a sediment accounting for frictional losses at points of grain-to-grain contact (McCann, C., 1966) for viscous losses arising either from flow between the fluid field and skeletal frame as a whole or from local fluid motion caused by relative approach between adjacent particles (see also Palmer and Traviolia, 1980) and for inelasticity in bonds between particles, if any exists. Stoll (1979) has investigated attenuation experimentally in sediments by means of a resonance column method (fig. 8.2). This apparatus was designed to allow specimens to remain fully saturated and to be loaded axially in a stable manner. He found (fig. 8.2b) that the logarithmic decrement of column oscillation was amplitude-dependent at high strain, but was almost independent of strain at lower strain levels. This is significant in that it seems to indicate that amplitude-dependent losses (which are probably associated with inter-particle velocities) may not be significant at

BIOT/STOLL PHYSICAL PARAMETERS

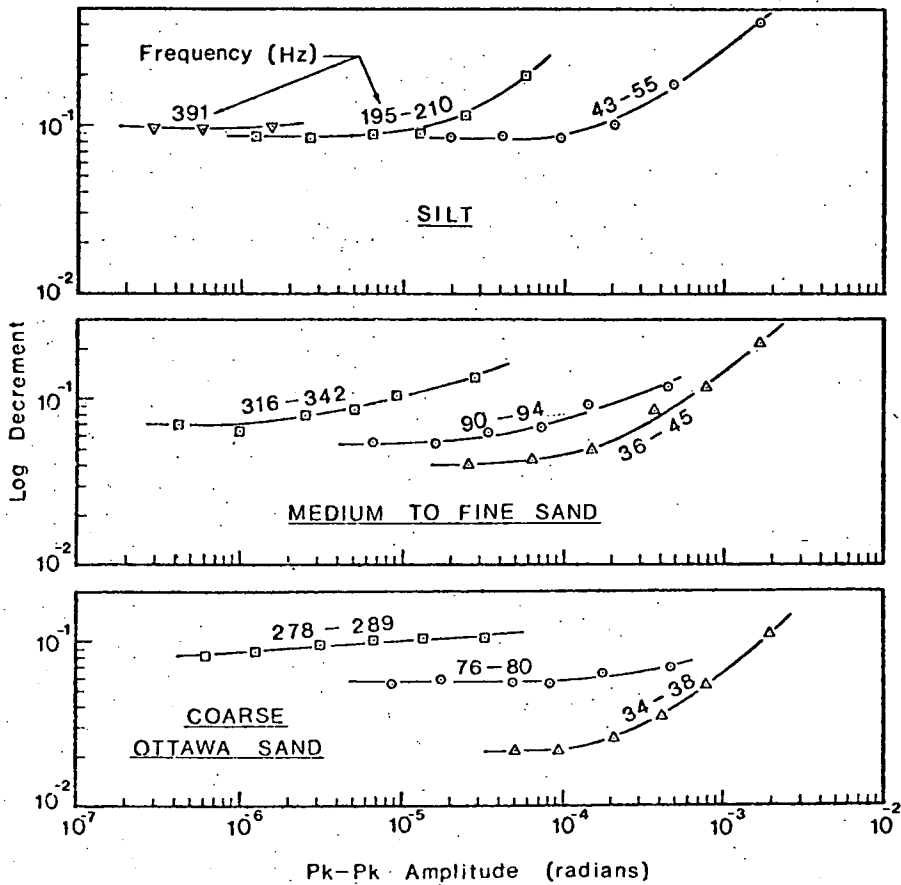
Symbol	Physical Property	Units
ρ_r	Density of sediment grains	$\text{kg}\cdot\text{m}^{-3}$
K_r	Bulk modulus of sediment grains	$\text{N}\cdot\text{m}^{-2}$
ρ_f	Density of pore fluid	$\text{kg}\cdot\text{m}^{-3}$
K_f	Bulk modulus of pore fluid	$\text{N}\cdot\text{m}^{-2}$
η	Viscosity of pore fluid	$\text{kg}\cdot\text{m}^{-1}\cdot\text{s}^{-1}$
β	Porosity	--
K	Permeability	m^2
a	Pore size parameter	m
α	Structure factor	--
μ_b	Shear modulus of frame (real part)	$\text{N}\cdot\text{m}^{-2}$
μ_b^*	Shear modulus of frame (imaginary part)	$\text{N}\cdot\text{m}^{-2}$
K_b	Bulk modulus of frame (real part)	$\text{N}\cdot\text{m}^{-2}$
K_b^*	Bulk modulus of frame (imaginary part)	$\text{N}\cdot\text{m}^{-2}$

Table 8.1

Parameters used in the Biot-Stoll sediment model. The 'pore-size parameter' and 'structure factor' are to some extent arbitrary and indicate the semi-empirical nature of this model.



a



b

Fig. 8.2

(a) Resonant column apparatus used by Stoll (1979) to investigate attenuation in sediments.

(b) Results of Stoll (1979) show that logarithmic decrement for the column oscillation depends on amplitude at higher torsional strain, but is almost independent of it at low strain. The specimens were water-saturated and the axial stress was 30kPa.

the strain levels which are encountered in seismic investigation (ca. 10^{-9} at frequencies of ca. 250Hz; McDonal et al. 1958; Arnold, 1977). Winkler and Nur (1982) have used resonating bars of sandstone in a similar way to study attenuation at frequencies between 500 and 9000Hz and find amplitude-dependent effects only at strain levels above 10^{-6} .

Hovem (1980) has compared an older sediment model of Urick (1948), which treats a sediment as a suspension of high concentration, with the Biot-Stoll model as applied to fine marine sediments of very high porosity and notes that both models predict that losses arising from viscous flow of the pore fluid should be proportional to the square of frequency at low frequency and to the square root at high frequency. Hovem suggests that such a variation is not in accord with observations (Hamilton, 1972) which tend to follow more closely a linear dependence on frequency and that this may indicate that loss mechanisms other than viscous attenuation are rather more important. Some experimental values suggest a variation exponent greater than 1.0 in the transitional to high frequency range (Buchan et al., 1966; McKay, 1967) so that there does seem to be a discrepancy with the Biot-Stoll theory which Hovem suggests is due to its ability to accommodate only one pore size parameter, which is normally adjusted to the mean grain size and could not

be expected to represent accurately an ill-sorted sediment.

From the considerations discussed above, it would seem that possible approaches to the modelling of the rheological characteristics of a sediment are:

- (i) Use a simple elastic model, with loss.
- (ii) Adapt the model evolved for complex polymers.
- (iii) Use the Biot-Stoll model.
- (iv) Develop modern mathematical thinking on continua in applying it to the specific case of a porous sediment.

Each of these offers advantages and suffers from difficulties. Computational facility is offered by (i) but frequency dependent loss is not considered. Loss arising from viscous flow is treated by (ii), but the analogy between a polymer and a sediment is not a close one and the treatment of acoustic wave motion by the use of this model has not been highly developed. The model (iii) has been evolved specifically for sediments but is semi-empirical in nature, is mathematically complex and may become more complex when the effect of grain size distribution is incorporated. A new approach (iv) might be rewarding, but current mathematical ideas on rheology do not appear to be immediately adaptable to the case of sediments.

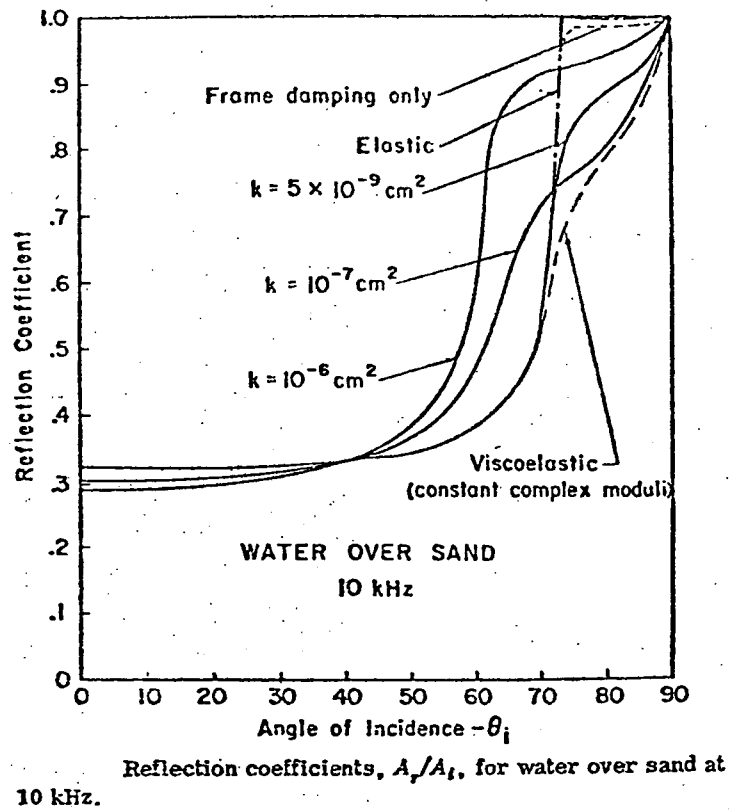


Fig. 8.3

A comparison between reflexion coefficients derived by Stoll and Kan (1981) on the basis of different sediment models. They compare a purely elastic model and a lossy elastic model (labelled "visco-elastic" on the diagram) with the porous model for various values of permeability (and for the limit of zero permeability where damping results only from losses in the skeletal frame of the sediment).

The simple model (i) in offering ease of calculation may be a good choice for modelling narrow-frequency band processes in short timescales where de-watering and consolidation do not occur.

8.2 SEISMOGRAMME SYNTHESIS.

Stoll and Kan (1981) and Hajra and Mukhopadhyay (1982) have investigated the behaviour of a plane wave incident at the boundary of a porous medium when it is modelled according to the Biot theory, thus opening the way for the construction of synthetic seismogrammes on the same theoretical basis. Although some initial work on seismogramme construction has been undertaken by Bennell (1983, personal communication) following the work by Hamdi and Taylor Smith (1982) on the porous model, experience of seismogramme synthesis in this way appears limited at present, and has probably been confined to modelling for vertical incidence. The mechanical model of assuming simple elasticity with loss as outlined in the last section has been used extensively in seismogramme modelling, as reviewed by Cervený (1983), in the fields of lithospheric and whole-earth seismology. A comparison of the reflexion coefficient when modelled in those different ways is shown in fig. 8.3 (Stoll and Kan, 1981). It can be seen that differences between the predictions of the

porous and the lossy elastic models decreases with the permeability. Since the field data to be modelled in the next chapter is from an area where ice-bonding (chapter 4) was anticipated, with consequent low permeability (Antsyferov et al., 1964), it seemed appropriate to take advantage of the available methods of seismogramme synthesis which have been used widely in other applications.

The concept of ray tracing has been discussed in some detail in chapter 5 in connexion with sound propagation in the water column and the conditions under which the method is valid have been noted (specifically in the 'geometrical optics' approximation, equation 5.22). Cerveny (1983) points out the limitations of this method, of which he regards the following to be of major importance:

- (i) The method is of limited accuracy in such singular regions of the reflected field as caustics (where many rays come together as can be seen in parts of fig. 5.10a) or the critical region for headwave development.
- (ii) The method is sensitive to approximations which may be introduced in specifying the layering or velocity structure to be modelled.

The principal alternative method to the ray method for modelling seismogrammes has been that termed the reflectivity method by Fuchs and Mueller (1971) wherein the plane wave reflexion coefficient is first evaluated

for a grid of incidence angle and frequency points and then integrated for each receiver position to yield a seismogramme by multiplication with the source spectrum and inverse Fourier transformation. Although not subject to the criticisms made about the ray method above, there are restrictions on the validity of the reflectivity method, apart from the limitation imposed by the practical requirement of defining the angle of incidence-frequency grid to be of sufficiently wide mesh to allow computation to proceed in a reasonable time. Fuchs and Mueller give these limitations as:

- (i) the restriction to real angles of incidence
- (ii) the use of the asymptotic form of Bessel functions.

Of these, the second is a computational device and need be of no greater concern than is the definition of the grid for integration already mentioned. The first may be of more consequence, and is a matter which has been discussed at length by Frisk (1979) who has shown that inhomogeneous waves can make a significant contribution to the reflected field. He considers this effect important when the combined height of source and receiver is less than one quarter wavelength from the reflecting boundary. It would seem that this effect, with the associated possibility of a reflexion coefficient which exceeds unity (Atalar, 1979, 1981) may be of importance in modelling data obtained with equipment on the seabed (e.g. Bryan, 1980) but should not pose problems when source and receiver are higher in the

water column. Some advantages of putting equipment close to, but above the bottom have been pointed out by McKay and McKay (1982). Here is another advantage. One further problem may arise with the Fuchs Mueller approach. This is the possibility of coupling between compressional and shear waves which may be produced by strong velocity gradients (Fryer, 1981). The coupling is strongest for the strongest gradients. Fryer has considered gradients up to 2/sec which is as high as any in the literature for normal sediments. At these gradients, coupling is small for frequencies above 1Hz, so that the problem is likely to be a small one.

With the above minor reservations, the reflectivity method of Fuchs and Mueller as implemented by Kennett (1974) has been applied to modelling seismogrammes obtained in the Beaufort Sea, as described in the following chapter.

9. MODELLING OF 12-CHANNEL DATA FROM THE BEAUFORT SEA.

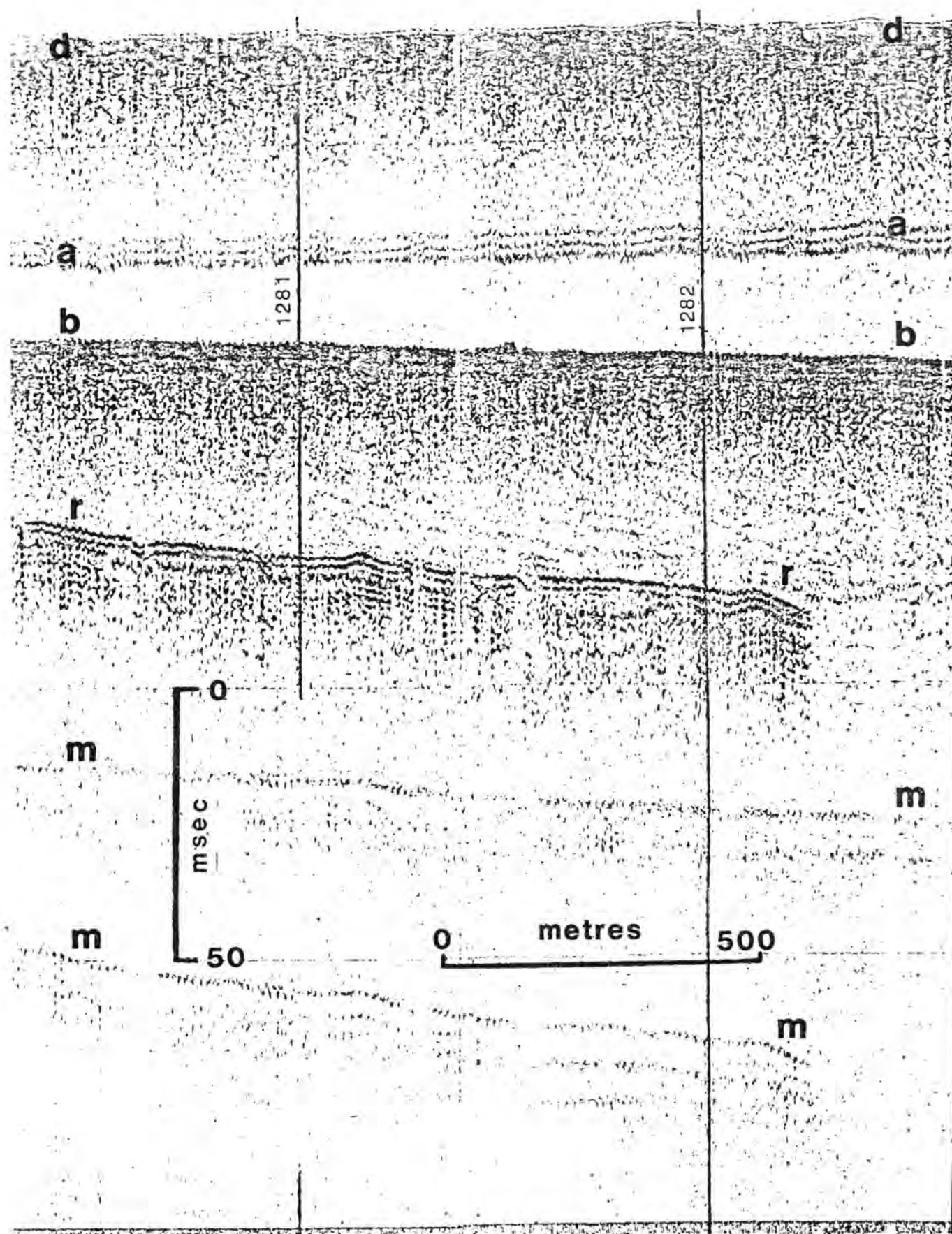


Fig. 9.1

Near-vertical incidence section in the neighbourhood of fixes 1281 and 1282 of line 8 of the Beaufort Sea work (see fig. 9.2).

d - direct arrival	r - subbottom reflexion
b - bottom reflexion	m - multiple reflexion
a - late phase of direct arrival	

9.0 INTRODUCTION.

In collaboration with J. Hunter of the Geological Survey of Canada, the 12-channel hydrophone array described in section 2.6 was deployed from H.M.C.C.G.S. 'Nahidik' along with a small Bolt airgun as acoustic source. The airgun was towed 2 metres below the surface and the nose of the array was from wire angles estimated to be 2.5m subsurface. The deployment was thus essentially a surface one. The airgun firing rate was one per two seconds at a pressure of ca. 1500psi. Channel number 1 of the array was offset 36.5 metres from the airgun and its output was displayed continuously in variable intensity format (fig. 9.1) to give a near-vertical incidence subbottom profile. Every two minutes (or in places every 5 minutes) the output of the entire 12-channel array was recorded on a Nimbus model ES-1210F stacking seismograph operating on a 200msec. sample length, so that the digitising interval for the set of 1024 numbers (each of 10-bit resolution) representing the output of one channel was 0.2msec. Generally a stack of either 2 or 3 airgun shots was recorded on a Nimbus cassette tape recorder and printed out on the Nimbus printer as a proof copy. The area where records were obtained is shown in fig. 9.2. These records were later moved to 5" floppy discs by personnel of the terrain geophysics group of the Geological Survey and stored in

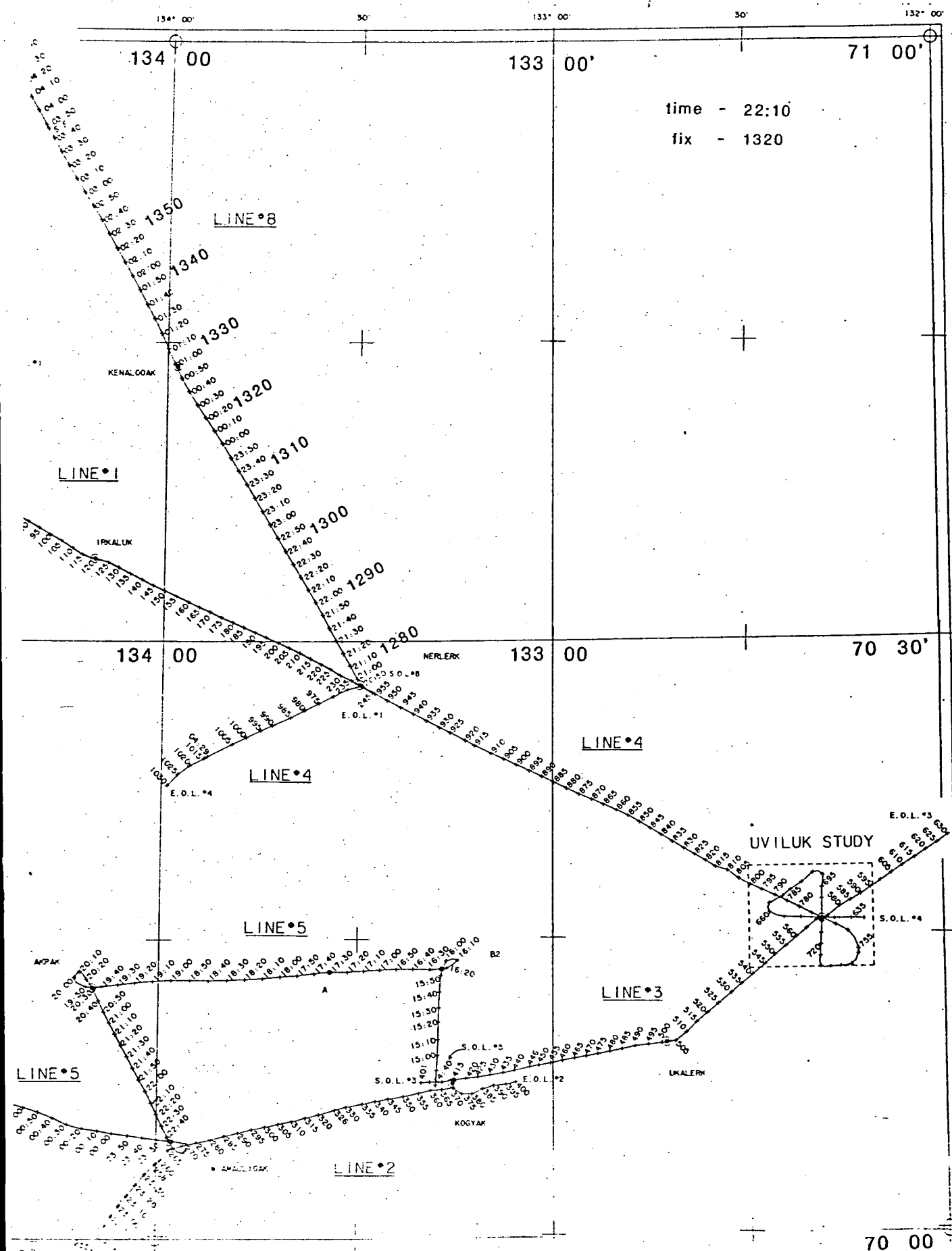


Fig. 9.2

Track plot. Beaufort Sea lines.

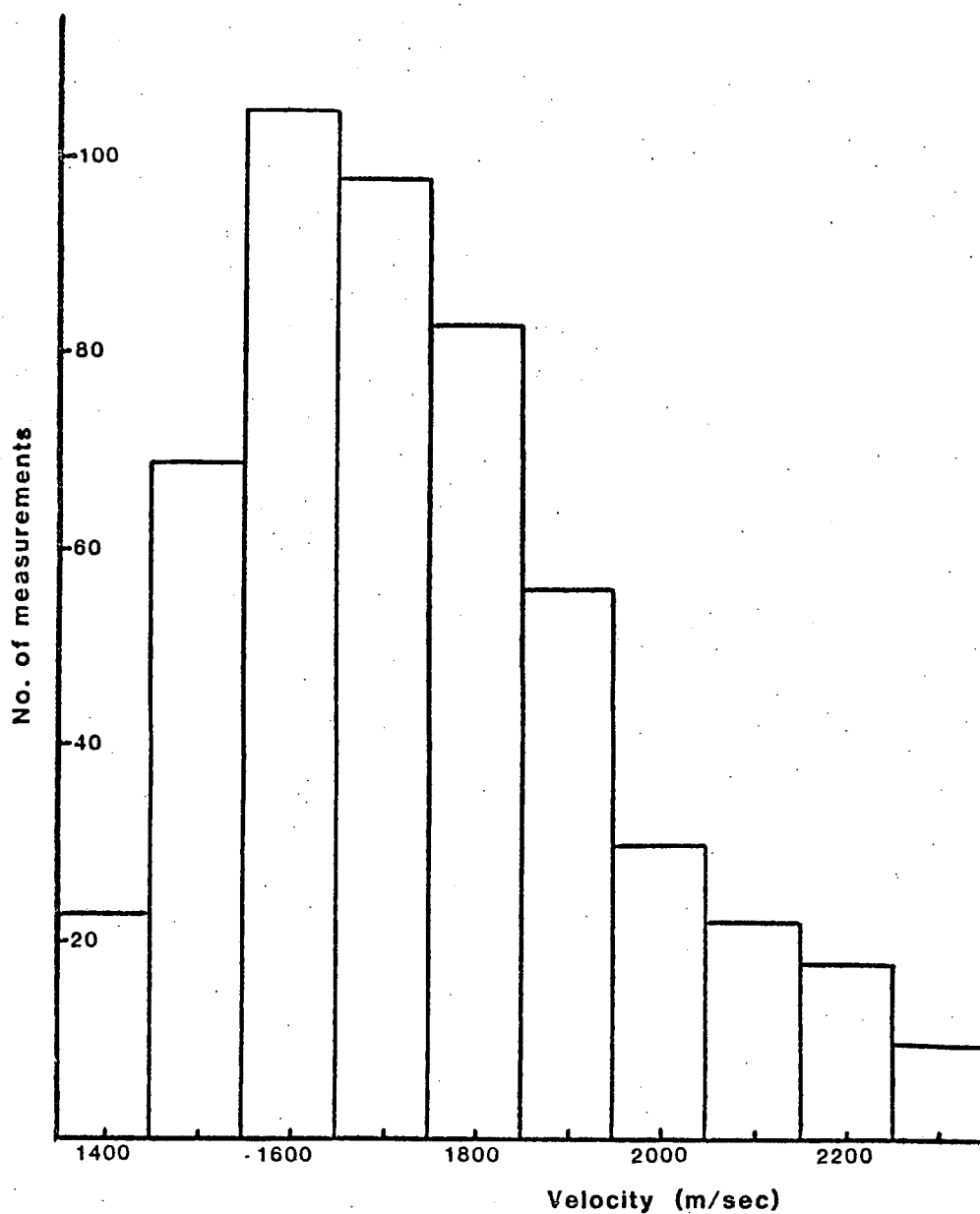


Fig. 9.3

Velocity histogramme for Beaufort Sea surficial sediments based on interval velocities observed from reflectors between 0 and 55m below the seabed, (after Hunter, 1983)

one of two different formats. The first of these formats (GSC) preserved only one in every four of the original data points and reduced the number precision from 10 to 8 bits. The second (Normanton) format preserved all of the original information. The reason for the rejection of much of the data in the G.S.C. disc format was that a system of archiving and picking travel times on hammer seismic profiles obtained on land had been set up by Hunter (1980) to use an Apple II micro-computer, and a density of about 250 points per trace was all that could conveniently be accommodated on the monitor display of this computer.

9.1 DISCUSSION OF TRAVEL TIME REDUCTION METHODS.

The G.S.C. format discs were used by H. MacAulay to derive interval velocities from the Beaufort Sea data and his results are summarized in the histogrammes (figs 9.3 to 9.6) and the generalised velocity-depth profile of fig. 9.7 (J. Hunter, personal communication, 1983). There are some shortcomings in this approach to dealing with the data:

(i) The 0.2msec time resolution of the original Nimbus records is an interval which is already a coarse one when compared with the criterion of Bennell (1978) for the time resolution required for interval velocity measurement (to

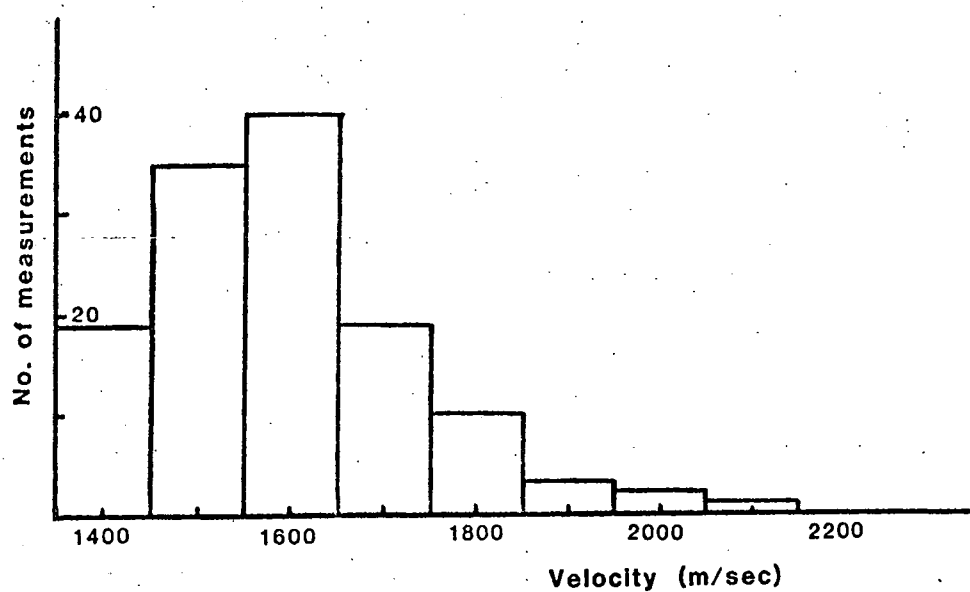


Fig. 9.4

Velocity histogramme for Beaufort Sea surficial sediments based on interval velocities observed from reflectors between 0 and 25m below the seabed, (after Hunter, 1983).

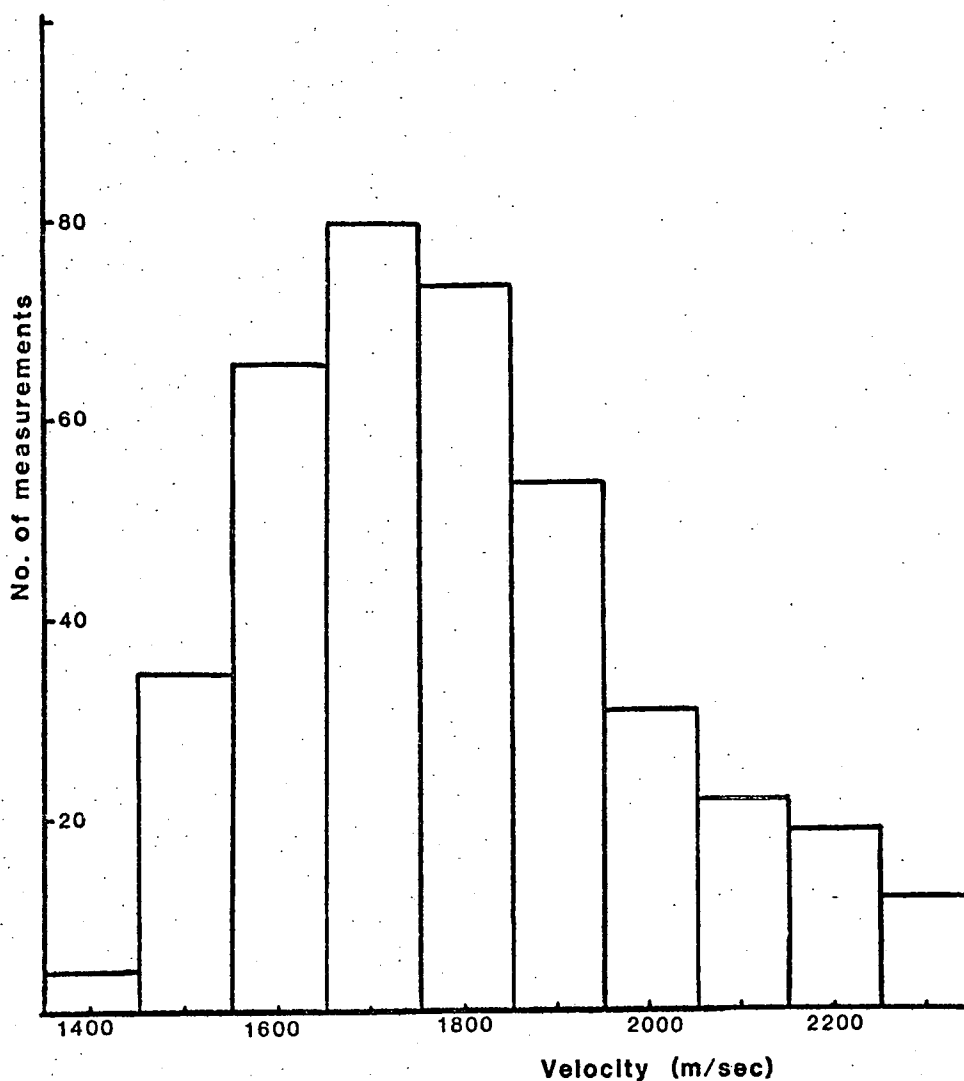


Fig. 9.5

Velocity histogramme for Beaufort Sea surficial sediments based on interval velocities obtained from reflectors between 25 and 55m below the seabed, (after Hunter, 1983).

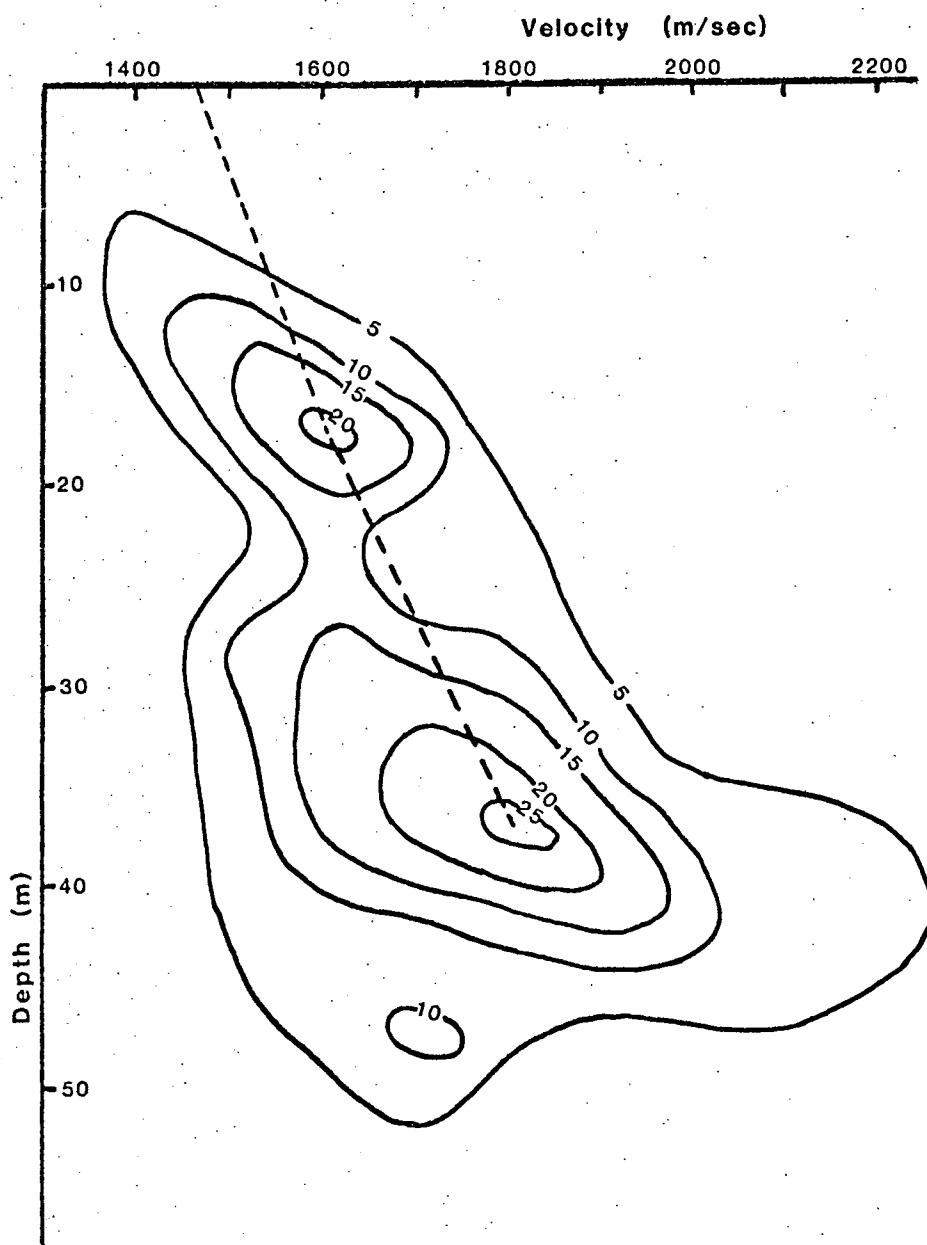


Fig. 9.6

Velocity-depth histogramme for 514 observations of surficial sediment acoustic velocity. (Hunter, 1983).

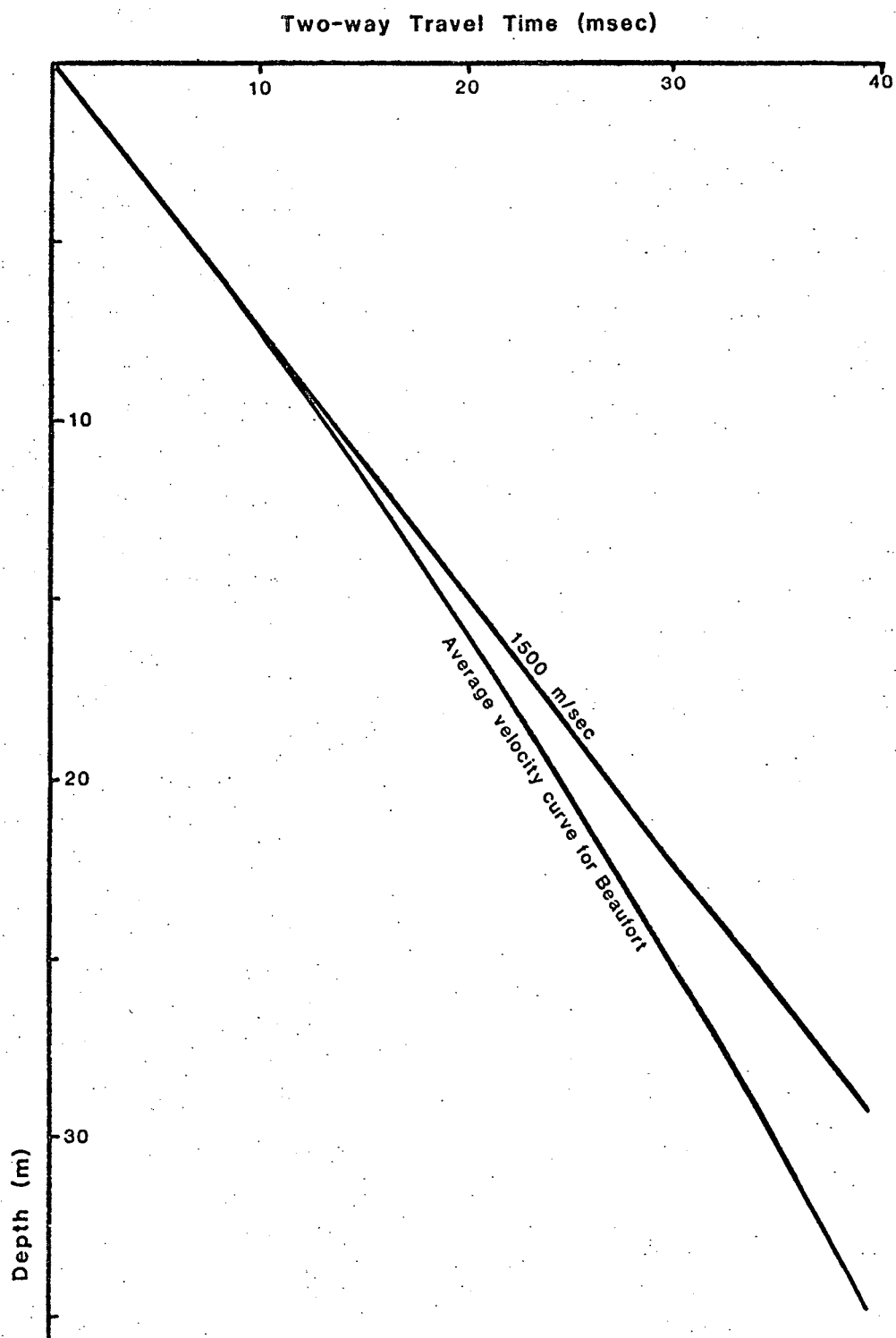


Fig. 9.7

Average velocity curve for surficial sediments of the Beaufort Sea compared with an arbitrary value of 1500m/sec which is often assumed in interpreting seismic sections.

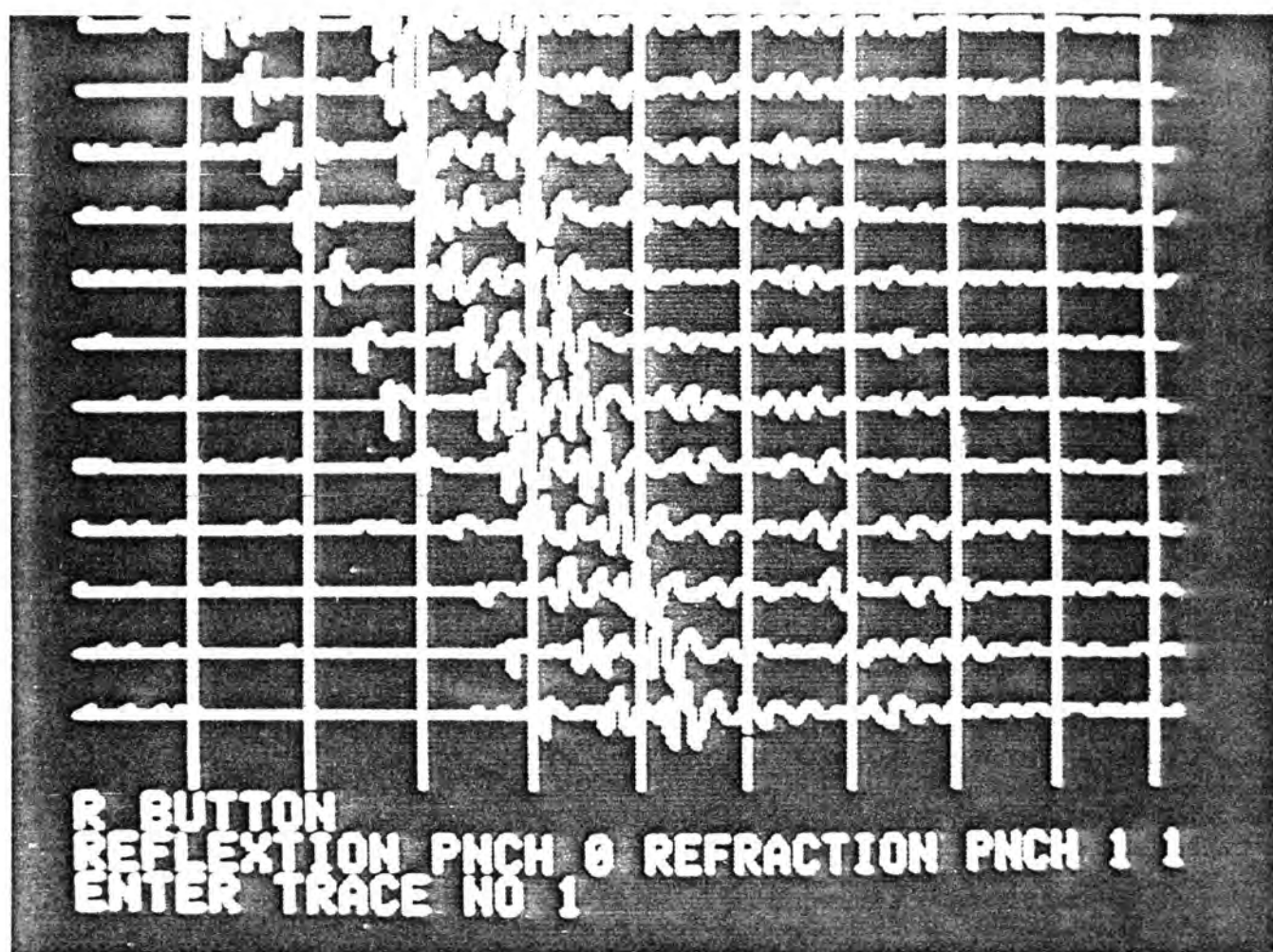


Fig. 9.8

Monitor screen display during running of the programme "P-PICKER."
(Before picking reflectors.)

better than 1%) in a situation of similar travel times - viz. ca. 0.02msec. Its further degradation to 0.8msec. is undesirable and leads to increased error.

(ii) The G.S.C. travel-time picking and velocity calculation programme "P-Picker" (Hunter, 1980) does not calculate the error on the derived T^2-X^2 value.

(iii) The delay time between the electrical triggering of the airgun (and digital recorder) and the moment of firing of the airgun has not been taken into account in MacAulay's processing.

The display of a seismogramme in G.S.C. format on the Apple II monitor screen is shown in fig. 9.8. The curves picked by use of the cursor and "games paddles" are shown in fig. 9.9. These figures may be compared with fig. 9.10 which shows the quality of record which can be obtained from recordings in the Normanton format, and with fig. 9.10a which shows a display of G.S.C. format data plotted for the same seismogramme. The display of fig. 9.10 was obtained by procedures outlined in Appendix 4.

9.2 SYNTHETIC SEISMOGRAMME MODELLING APPROACH.

The modelling programme used was one for constructing seismogrammes on the basis of the reflectivity method of Fuchs and Mueller (1971) as adapted by Kennett (1974) for

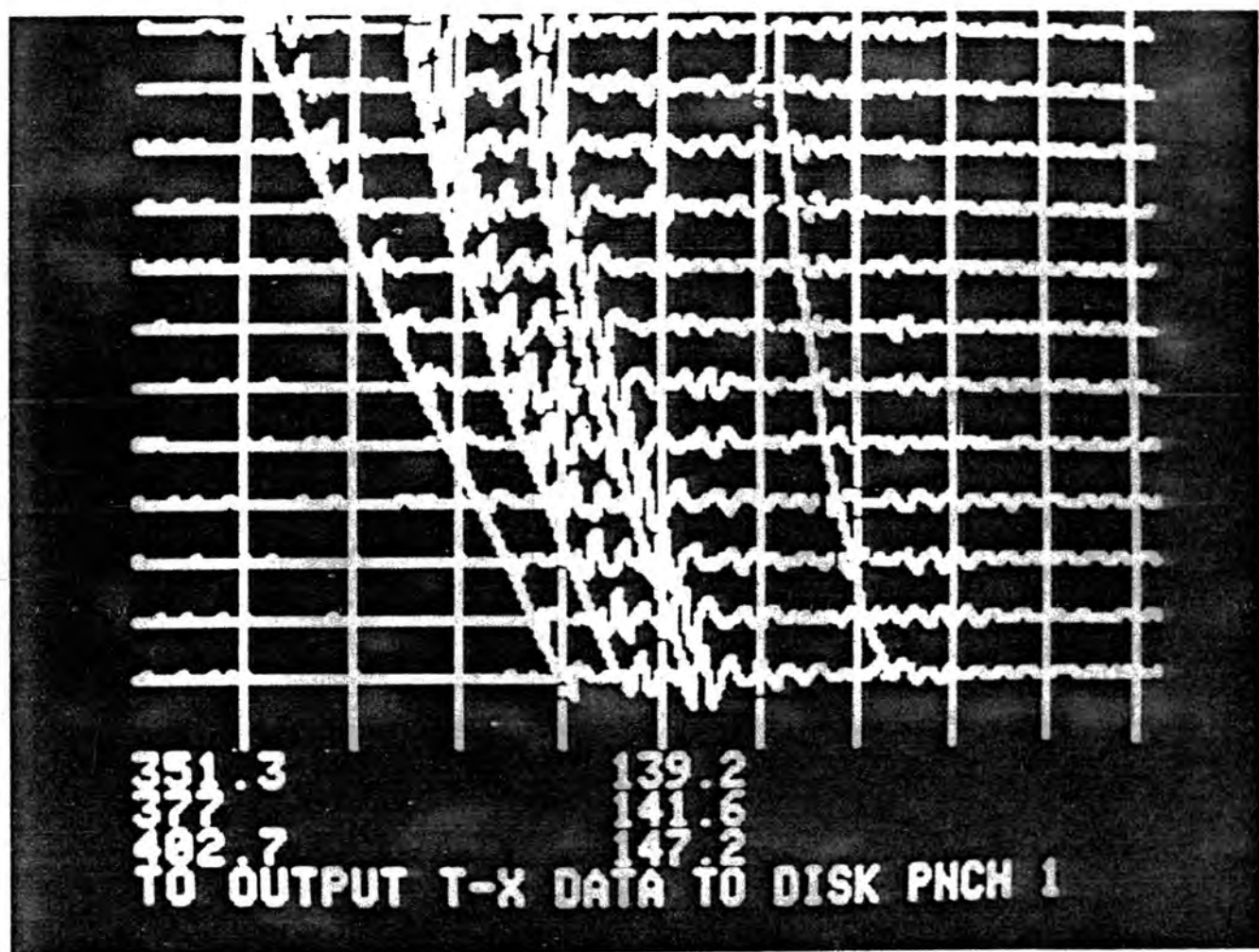


Fig. 9.9

Monitor screen display during running of the programme "P-PICKER."
(After picking reflectors.)

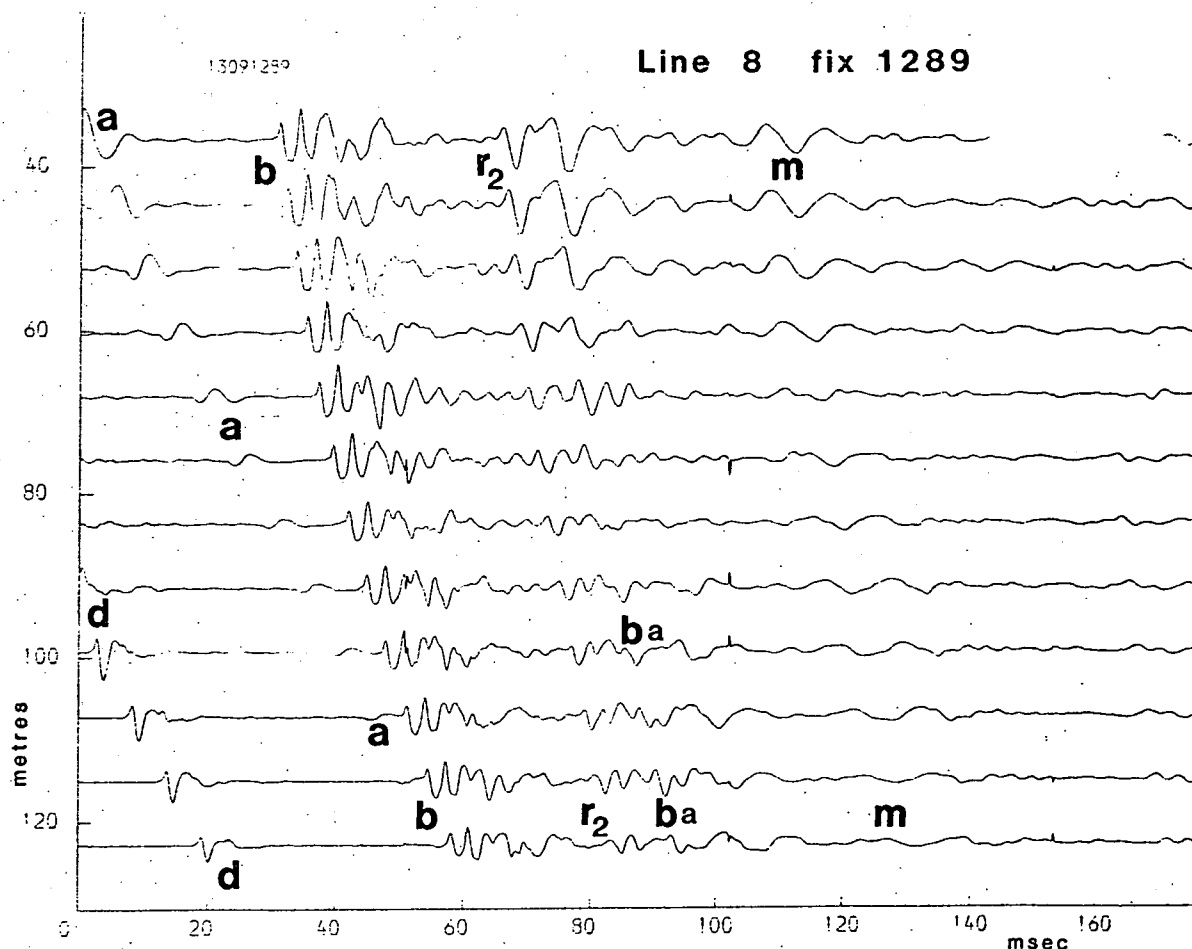


Fig. 9.10

A 12-channel seismogramme from an area to the N.W. of Nerlerk (see section 7.3 and fig. 9.2). The waveforms were stored as 10-bit fixed point numbers with a sampling interval of 0.2msec. Arrivals are identified as:

d - direct wave	r_2 - subbottom reflexion
a - late phase of direct wave (see text)	b_a - bottom reflexion associated with phase "a"
b - bottom reflexion	m - subbottom reflexion associated with phase "a"

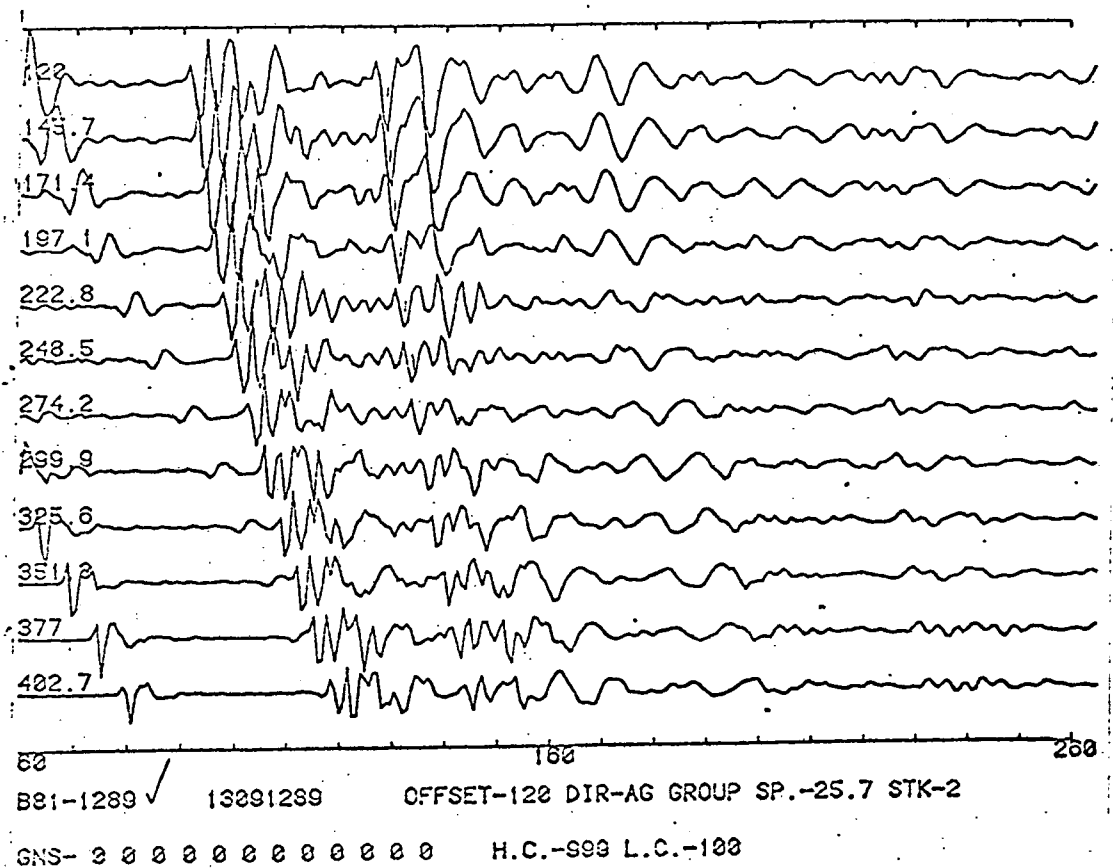


Fig. 9.10a

The same seismogramme as shown in fig. 9.10, but with the waveforms held as 8-bit numbers and the digitising interval 0.8msec.

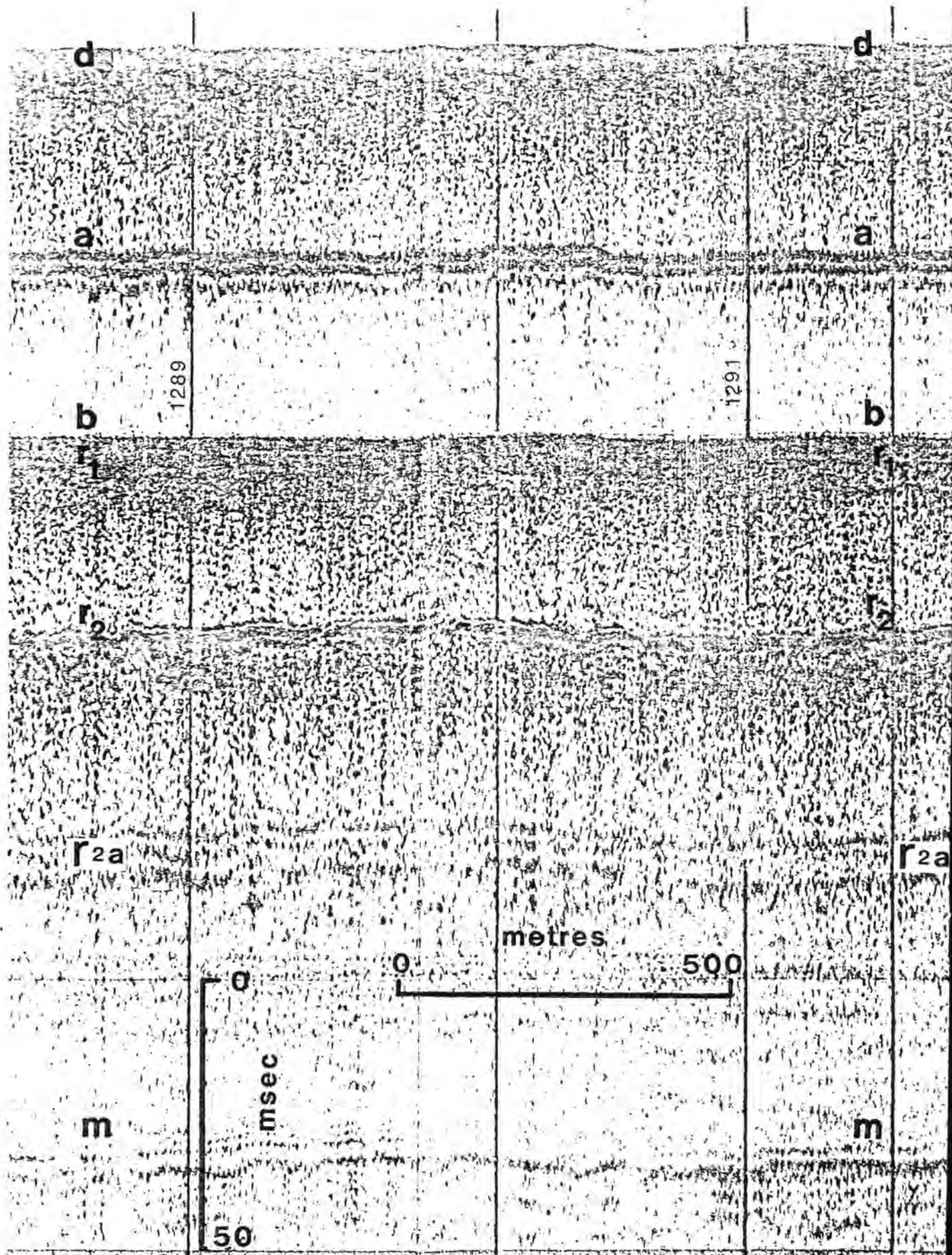


Fig. 9.11

Near-vertical incidence section in the neighbourhood of fixes 1289 to 1291 of line 8 (see fig. 9.2). The arrivals are identified as shown in fig. 9.10. The arrival 'ba' is overwritten by r2, but can be seen as a separate event in fig. 4.34, which (at a reduced scale) reproduces this section of profile together with a leftward extension of it.

use in multilayered media. The adaptation of this programme to the modelling of seismogrammes of the type described in section 9.1 entailed a change in scale of about two orders of magnitude in distances and acoustic frequencies from those encountered in the class of problem which it had originally been designed to treat, viz. lithospheric refraction lines. The main difficulty appeared to lie only in the format of the output, a problem overcome by writing a new output programme (Appendix 4). A further difference between the general experimental methods of crustal seismology and the acoustic investigation of surficial sediments led to the programme's calculating the pressure response at the seabed (the normal location for ocean-bottom seismometers) instead of higher in the water column (the normal location of a towed hydrophone array). This difference was easily overcome by specifying as the topmost layer in the layered structure of earth materials to be modelled, a layer of water corresponding in thickness to the depth of water between the towed acoustic transmitter and receiver, and the seabed.

The first of the field seismogrammes chosen to model in this way was that shown in fig. 9.10. A variable density raster display of the output of the first channel of the 12-channel receiver is shown in fig. 9.11 and of the twelfth channel in fig. 9.12. The principal

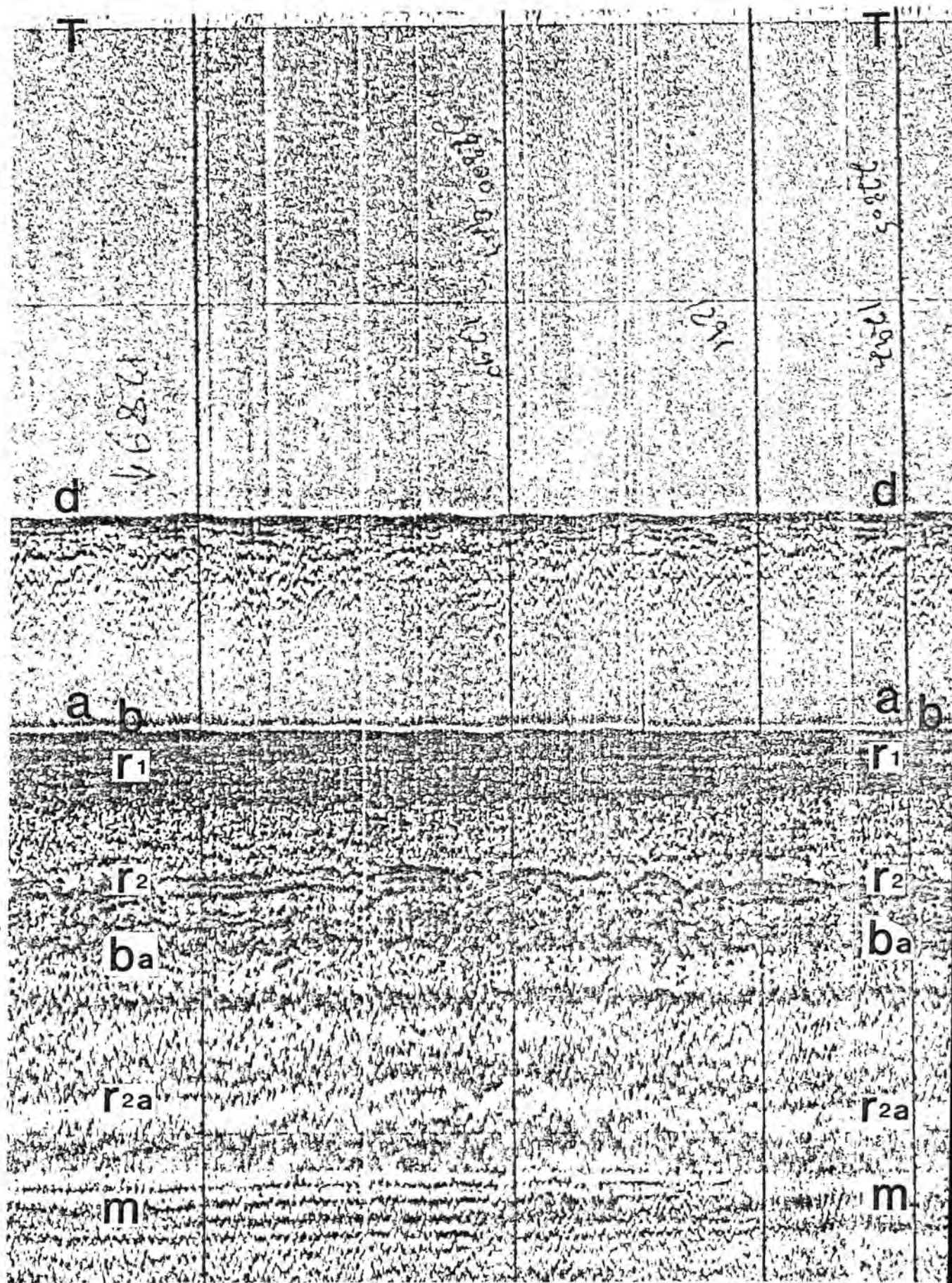


Fig. 9.12

A display of channel 12 of the array in the same format as shown in fig. 9.11. See figs. 9.10 and 9.11 for scales and labelling.

characteristics of these records are as follows:

- (i) Event "d" - the direct water wave arrival.
- (ii) Event "a" - originally thought to be associated with propagation of sound in the fabric of the seismic array, it has a moveout velocity identical to that of the direct water wave and thus seems to represent a secondary pulse from the sound source, possibly arising from the re-closing of the airgun.
- (iii) Event "b" - the seabed reflexion.
- (iv) Event "r1"- reflexion from a regular interface just below the seabed, just resolvable from the seabed reflexion.
- (v) Event "r2"- A strong, deeper subsurface reflexion identifying an interface which is much more irregular than that associated with "r1".
- (vi) Event "ba"- A repeat of the bottom reflexion resulting from event "a".
- (vii) Event "r2a"-A repeat of the event "r2" resulting from the event "a".
- (viii) Event "m" - Subbottom reflexion associated with phase "a".

The characteristics of the seismogramme of fig. 9.10 which it is reasonable to expect to fit with a synthetic

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	I/O P	I/O S	REV				
74.3	1.460	0.900	1.93	0.00910	0.00010	3				
29.5	1.950	0.350	1.90	0.00100	0.00200	3				
	4.500	2.500	1.98	0.00020	0.00040	3				
0.50	3.30	2.0	200.	250.	1100.	1150.	0.1E+09	0.960	0.0004	512

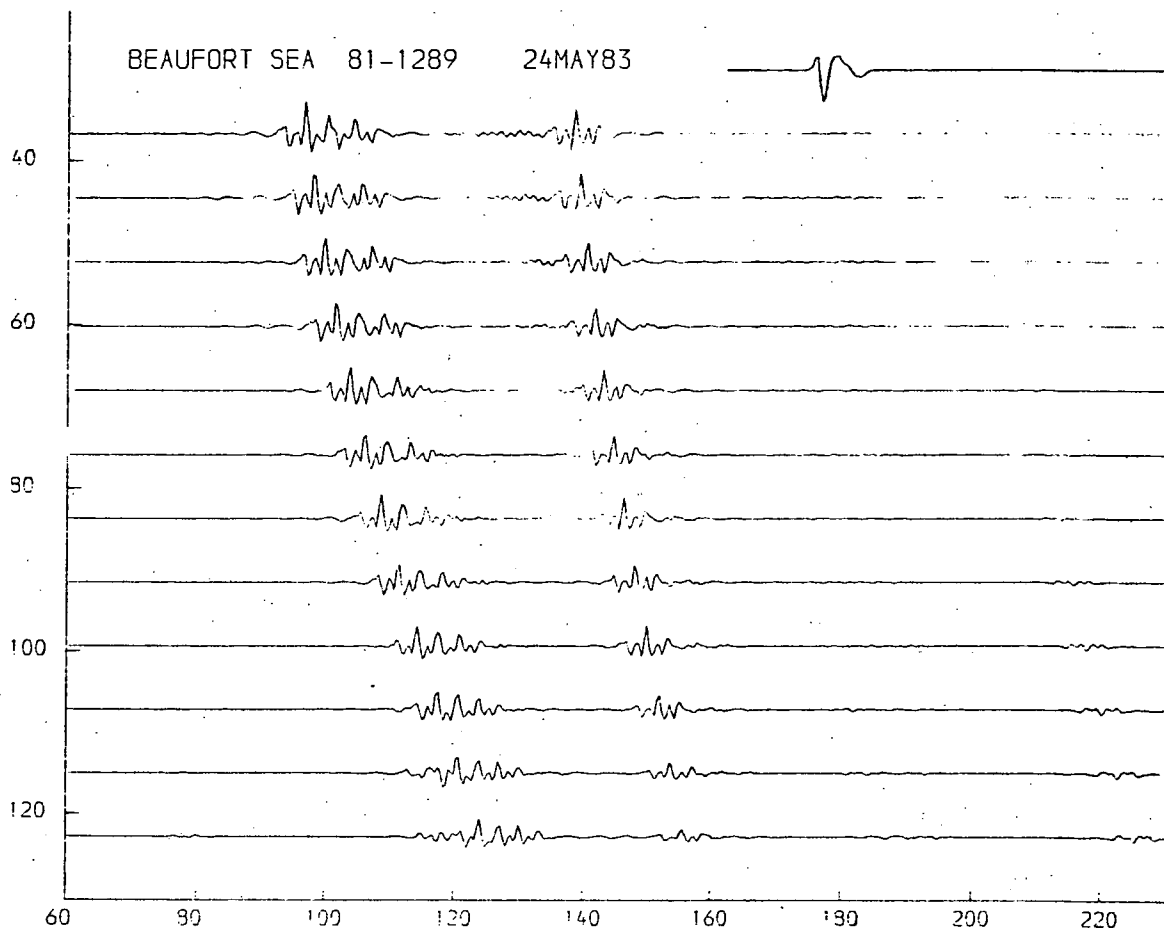


Fig. 9.13

Composite seismogramme which models surface reflexions by taking virtual source and receivers at points above the water surface. Source at 2.0m. depth; receivers at 2.5m.

seismogramme are limited by the following considerations:

(i) The towing of acoustic source and receiver beneath the water surface leads to surface ghosting effects which complicate the character of the bottom and subbottom reflexions.

(ii) Although the secondary acoustic event "a" appears as a direct arrival separated in time from the bottom reflexions of the primary event, the bottom-reflected arrival of this secondary event appears just after the "r2" reflexion of the primary event.

(iii) The manner of dealing with the water column as described above does not allow the proper modelling of the later multiple reflexions.

The consequence of (i) and (ii) above is that the features which it seemed most important to match up between the field and synthetic seismogrammes were the initial parts of the arrivals "b" and "r2" and the general amplitude of "r1", whose exact shape was clearly affected by a combination of the later parts of the waveform associated with reflexion "b" and by the surface reflexions at source and receiver. Some effort was applied to modelling the surface ghosting effects by means of adding together synthetic seismogrammes generated on the assumption that source and receiver were located at their reflected image positions above the water surface. The results of this procedure are shown in figs. 9.13 and 9.14 which are composite seismogrammes for slightly different hydrophone

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S	NREV
74.3	1.460	0.900	1.03	0.00010	0.00010	3
29.5	1.950	0.350	1.90	0.00100	0.00200	3
	4.500	2.500	1.98	0.00020	0.00040	3
0.50 3 30.0	2.0 200.	250. 1100. 1150.	0.1E+09	0.060	0.00040	512

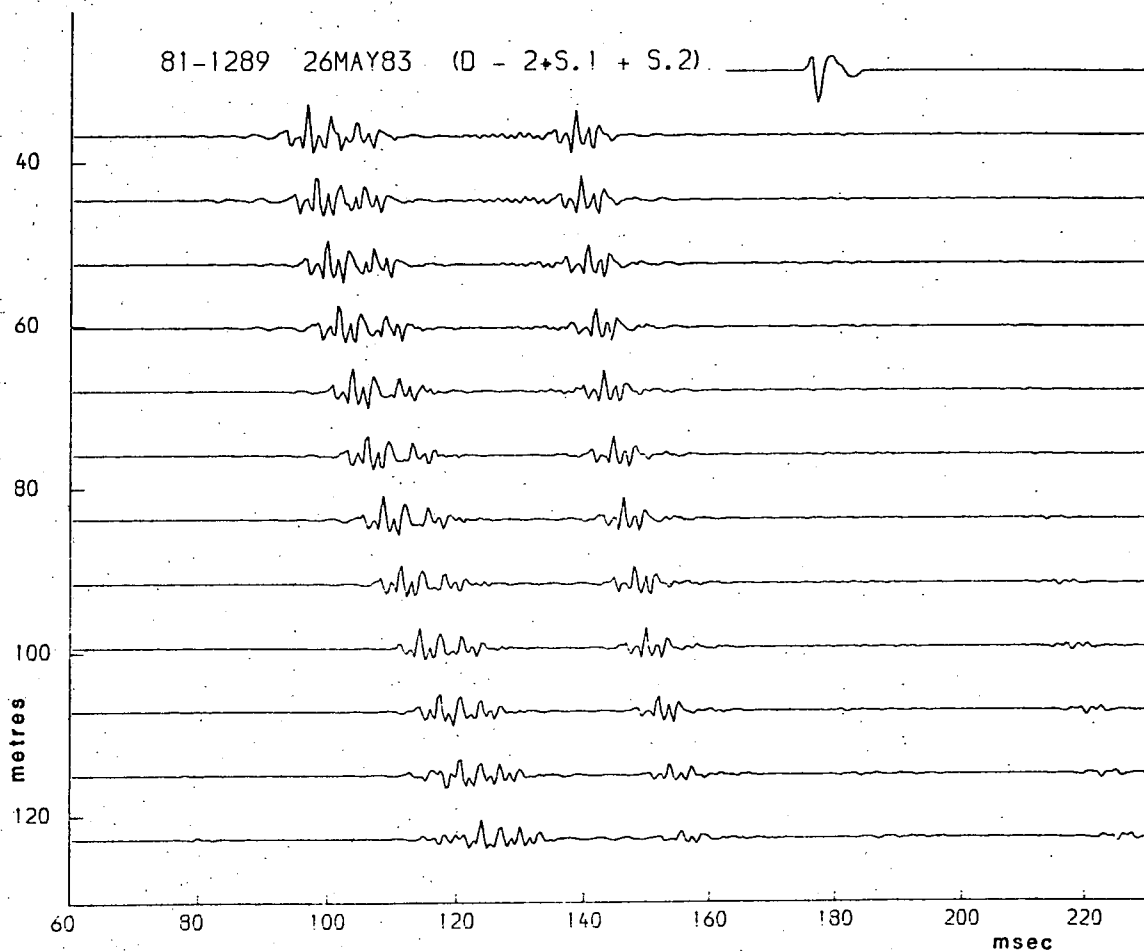


Fig. 9.14

9.14 Similar composite to fig. 9.13, but with both source and receiver at 2.0m.

THICKNESS (M.)	P. VELOCITY (KM/S.)	S. VELOCITY (KM/S.)	DENSITY (G/ML.)	1/Q P	1/Q S	NREV
70.3	1.530	0.000	1.03	0.00010	0.00010	3
6.2	1.552	0.210	1.50	0.00100	0.00200	3
32.6	1.900	0.400	1.90	0.00100	0.00200	3
500.0	4.500	2.500	1.99	0.00020	0.00040	3
0.50	3.000	2.000	250.0000	0.00000	0.00000	512

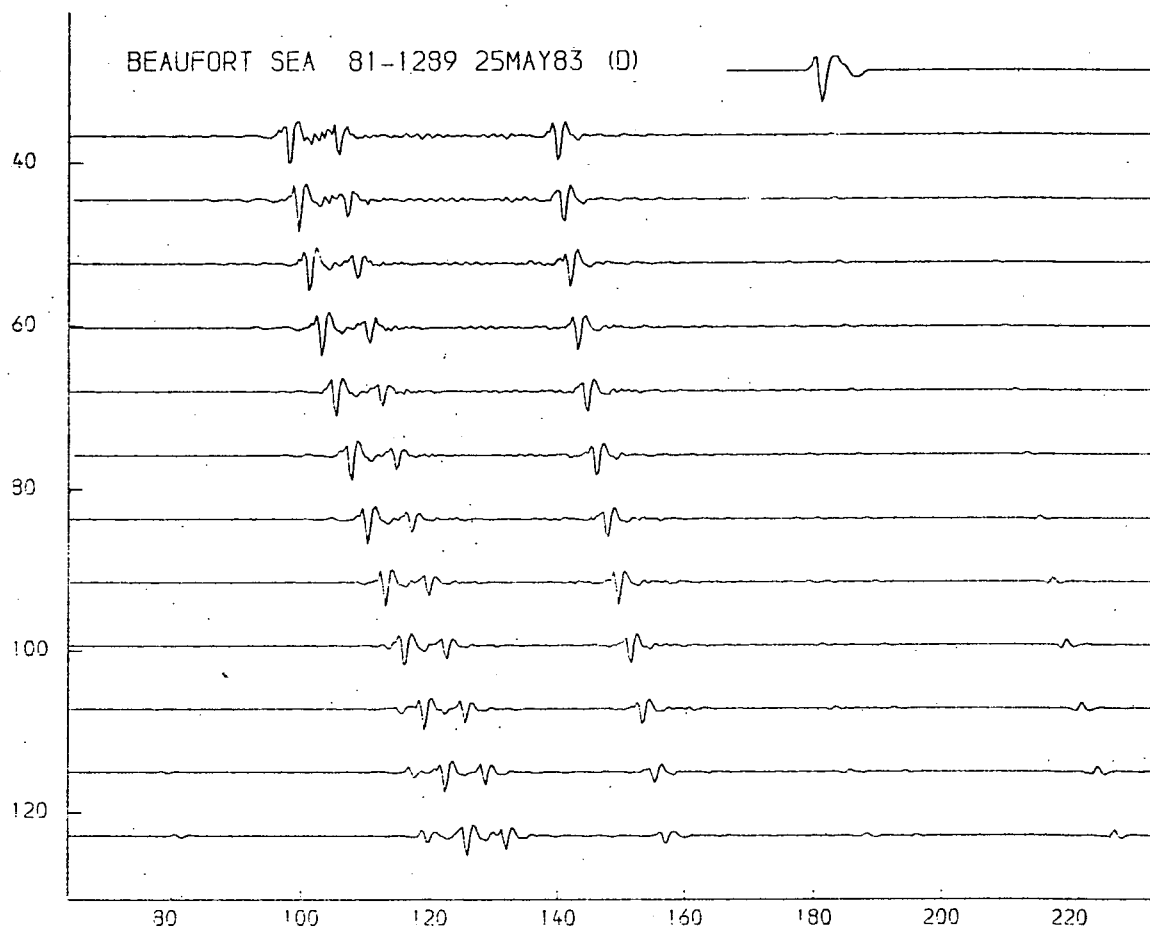


Fig. 9.15

Simple seismogramme which is constructed on the basis of the wavelet shown, in the frequency band 200 to 1150Hz.

towing depths and these may be compared with the simple synthetic seismogramme of fig. 9.15 which is of the type from which the others were compiled. The requirement to generate three separate seismogrammes and then add the results (with the phase change appropriate for a pressure release surface where appropriate) entailed more than tripling the time required in the computation of the seismogramme. Since this time requirement was already of concern in generating a simple seismogramme of the type shown in fig. 9.15, it was decided to abandon this more complex approach and concentrate on fitting only what were judged to be the principal characteristics mentioned above.

9.3 PARAMETER VALUES SUGGESTED IN THE LITERATURE.

A few general comments can be made about the anticipated properties of the seafloor materials on the basis of previously published information about sediment properties, and of what is known about the part of the Beaufort Sea investigated. The work of Hunter and others on the re-interpretation of oil industry seismic records to identify strongly ice-bonded sediment was summarised in section 4.4. This suggests the presence of a subbottom material having a compressional velocity of 3.0km/sec. or higher, but a density more like that typical of the sand

that it is, viz. ca. 1.85 to 2.00gm/ml. (Hamilton, 1978, 1980). If the reflector "r2" of fig. 9.11 represents the transition to a strongly ice-bonded material beneath it, then a small decrease in density would be expected. The thin layer of sediment between "b" and "r1" is known to be a recent mud (Pelletier, 1975, 1979, and section 7.3) and may be expected to have a density of ca. 1.6 to 1.7gm/ml. and a velocity close to that of the bottom water (Hamilton, 1974, 1980). The layer beneath "r1" is generally believed to be a fine sand (J.A. Hunter, S. Blasco (Geological Survey) and I. Watson (Dome Petroleum), personal communications) and appears to be the same stratum as was identified at the Nerlerk drill site (section 7.3) to have an acoustic velocity typical of such a sand.

Velocity gradients have been measured in situ by Tucholke and Shirley (1979) in pelagic sediments and turbidites in the North Atlantic by means of attaching a velocimeter to a piston corer. Their measurements, and others (mainly extrapolated from deeper data) reviewed by Hamilton (1976a, 1979a, 1980) suggest that gradients to be anticipated in a surficial material are of the order of 1m/sec/m. Attenuation values for near-seabed sands and muds have been compiled by Hamilton (1976b,c) who notes values in dB/m. at various frequencies which transform to $1/Q$ values as: 0.011 to 0.038 (150Hz) for a sandy clay in

Cardigan Bay (Li and Taylor Smith, 1969); 0.018 (30KHz) for fine sand off Corpus Christi (Muir and Adair, 1972), and 0.012 (15KHz) for silty sand off Santa Barbara (Igarashi, 1973). As noted in section 4.2, Antsyferov et al. (1964) measured in a laboratory sample of frozen sand, a value for $1/Q$ of 0.014 (380KHz) in association with compressional and shear wave velocities of 4650 and 2760m/sec. respectively. One of the lowest published values for p-attenuation ($1/Q_p=0.0004$) was found from an in-situ measurement at a station in the Hatteras Abyssal Plain by Frisk et al. (1981).

9.4 SOURCE FUNCTION, INTEGRATION INCREMENT AND TOWING

GEOMETRY.

The choosing of an appropriate source wavelet to use for the construction of a synthetic seismogramme is a common problem. The characteristics of impulsive sources of various kinds have long been of interest especially in the case of airguns where attention has focussed on the methods of constructing source arrays (e.g. Giles and Johnston, 1973). The seismic method of using a hydrophone array trailing from a body towed deep in the water column (chapter 7) would be ideally suited to the continuous observation of source characteristics by the monitoring of arrivals in the far field, well separated in time from

such interfering effects as surface reflexions, thus avoiding complications which arise in extrapolating from near-field data (Ziolkowski et al., 1982). This has not yet been accomplished because of the primitive recording methods available for the gathering of data treated in chapter 7. The 12-channel array used in the Beaufort Sea, although linked to an appropriate recorder, was not well adapted for deep-tow deployment. With the near-surface deployment used, the direct arrival through the water is essentially coincident in time with the surface reflexion (time difference ca. 0.1msec. at channel 1 and ca. 0.03msec. at channel 12), so that the arrival "d" of fig. 9.10 is really the sum of these two modes of arrival. At the near-grazing angles encountered with the surface deployment, it is not strictly accurate to think of surface interactions as reflexions. The generally irregular nature of the sea surface leads to scattering processes which are essentially diffraction effects. A surface having a statistically random deviation from a horizontal plane should produce scattering independent of frequency for incidence angles near grazing (Chapman, 1983). If the surface is "slightly rough" (Tolstoy, 1983), i.e. having some regular undulation with a superimposed random variation, the scattering process remains insensitive to frequency provided the velocity contrast across the boundary is large, as is the case for the air-water boundary. Consequently, no significant

distortion of the direct arrival is to be expected as a result of scattering processes at the surface, and it seems appropriate to use this arrival as the source wavelet for synthetic seismogramme construction. The late phase of this wavelet ("a" of figs. 9.1, 9.10, 9.11) can be seen on channel 8 of the seismogramme together with the initial wavelet. This is one of the very few places in the entire available data set where both "d" and "a" appear on the same trace, separated from other arrivals. Accordingly the "d" and "a" signal from this trace has been extracted and used as the source wavelet for most of the modelling. The amplitude of the late phase "a" has been arbitrarily increased by a factor of 1.5 as a compromise between an exact copy of the direct arrival on channel 8 and a better fit to the later parts of the seismogrammes. Such a wavelet was used in the making of models for the seismogrammes nos. 1281 and 1289 (below). In the part of the data set from which the field record no. 496 is drawn, the event "a" was not seen, and models for that seismogramme use the simple "d" signal.

The 12-channel array was carefully balanced for neutral buoyancy (see fig. 2.19) and is seen (by observation from a high point on the towing vessel) when towed in calm water just below the surface, to trail very near the horizontal throughout its length. The low drag characteristics of the array (Table 2.1) confirm a stable

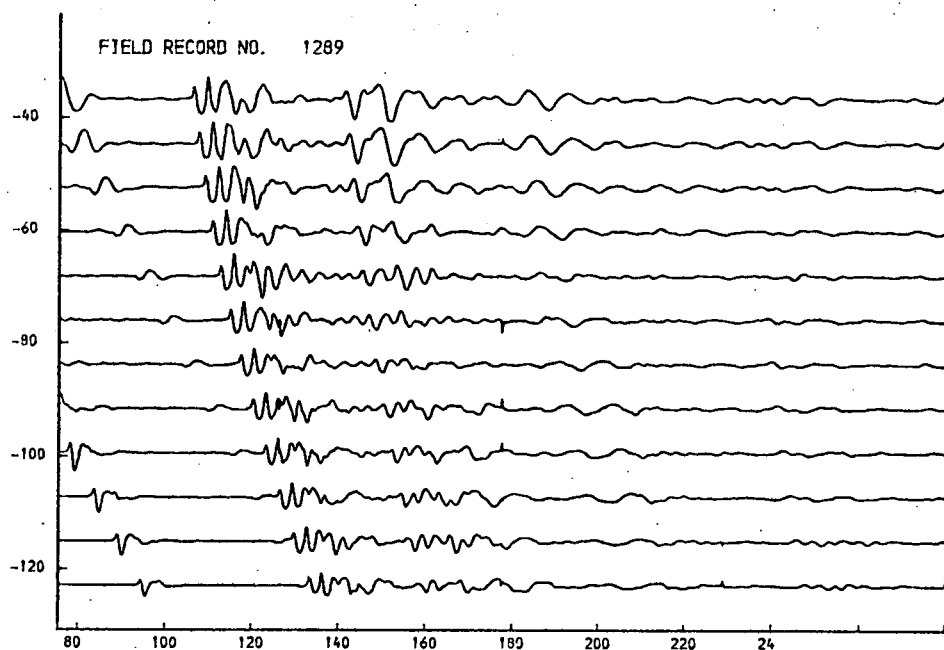


Fig. 9.16

This is the same seismogramme as appears in fig. 9.10. Here it is shown at the same scale as is used in the subsequent models.

horizontal towing attitude. As an indication of the deviations from the ideal of exactly equal towing depths for all the hydrophone groups, synthetic seismogrammes were constructed on the basis of deviations in towing depth of 0.5 to 2.0m. (figs. 9.17a and b). A comparison with the field record (fig. 9.16) shows that the array droops by ca. 0.5m. in the vicinity of the central hydrophone groups with respect to the nose and the tail, which are at the same level.

A check was carried out on the appropriate grid to specify for the construction of the array of values of the reflexion coefficient in the angle of incidence domain. The results are shown in figs. 9.18a and b. It can be seen that very little difference results from specifying an angle increment of 1° , 0.5° or 0.25° , but that at 2° , an instability appears in the model. Most of the modelling has been done on the basis of an increment of 0.5° or 0.6° . A similar check was carried out on the frequency bandwidth to be used (figs. 9.19a and b). It can be seen that the relative amplitudes in the various parts of the constructed seismogramme show very little variation with frequency. Most of the seismogramme synthesis has been carried out in a band 200-550Hz, as a compromise between modelling the full spectrum recorded in the field data which would require a bandwidth of ca. 50-1000Hz. and restricting the time required for

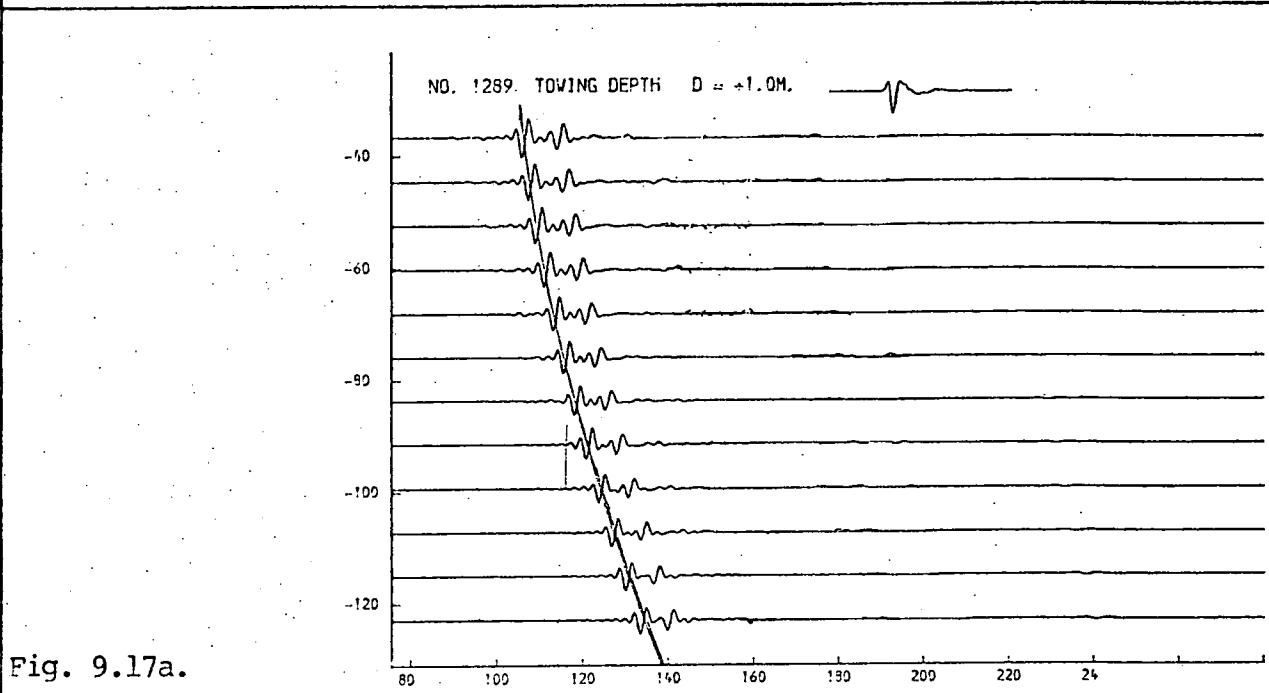
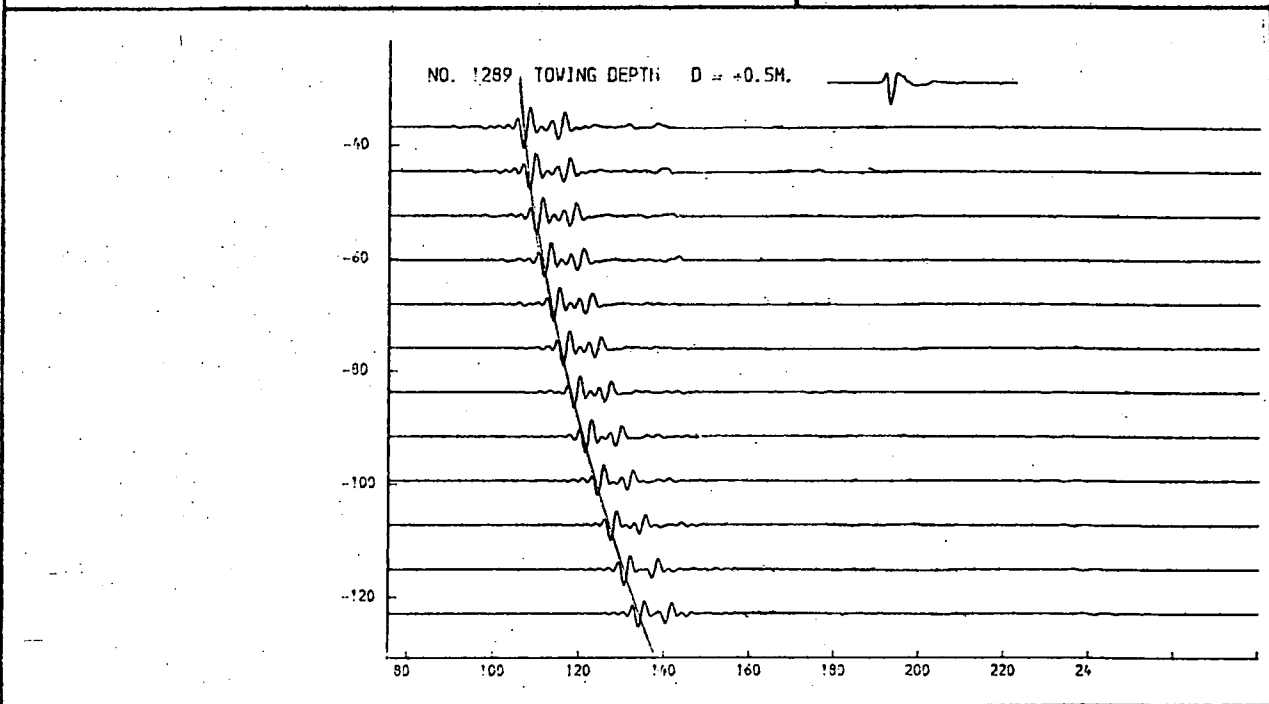
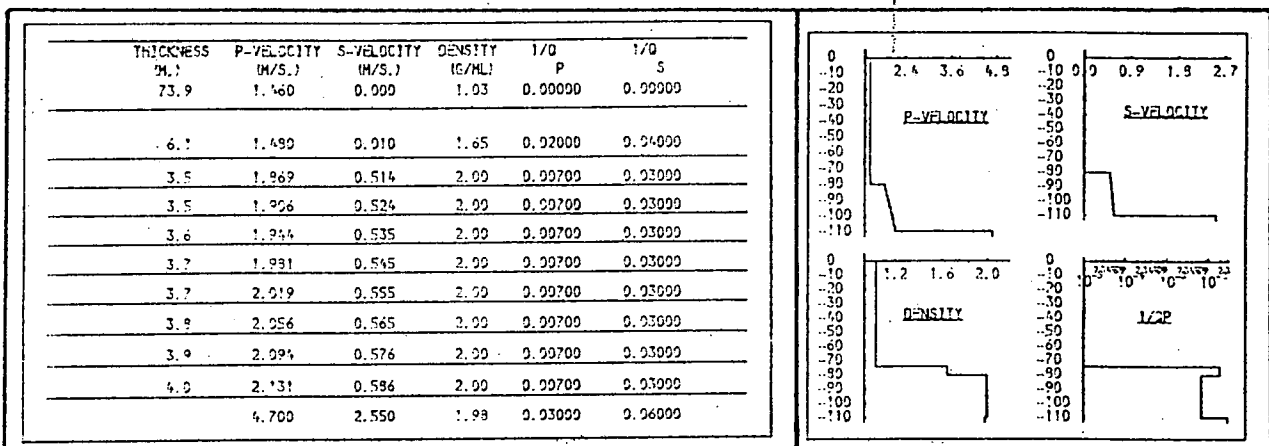


Fig. 9.17a.

The construction of a synthetic seismogram for different water depths simulates the expected position of the seabed reflexion for an array towing at different levels below the water surface. The curve superimposed on the seismogram is the envelope of the initial peaks of the field record. The coincidence of this field curve with the initial peaks of the central traces of a seismogram constructed for a towing depth 0.5m. greater than normal (upper figure) suggests a droop of ca. 0.5m. in the central part of the array.

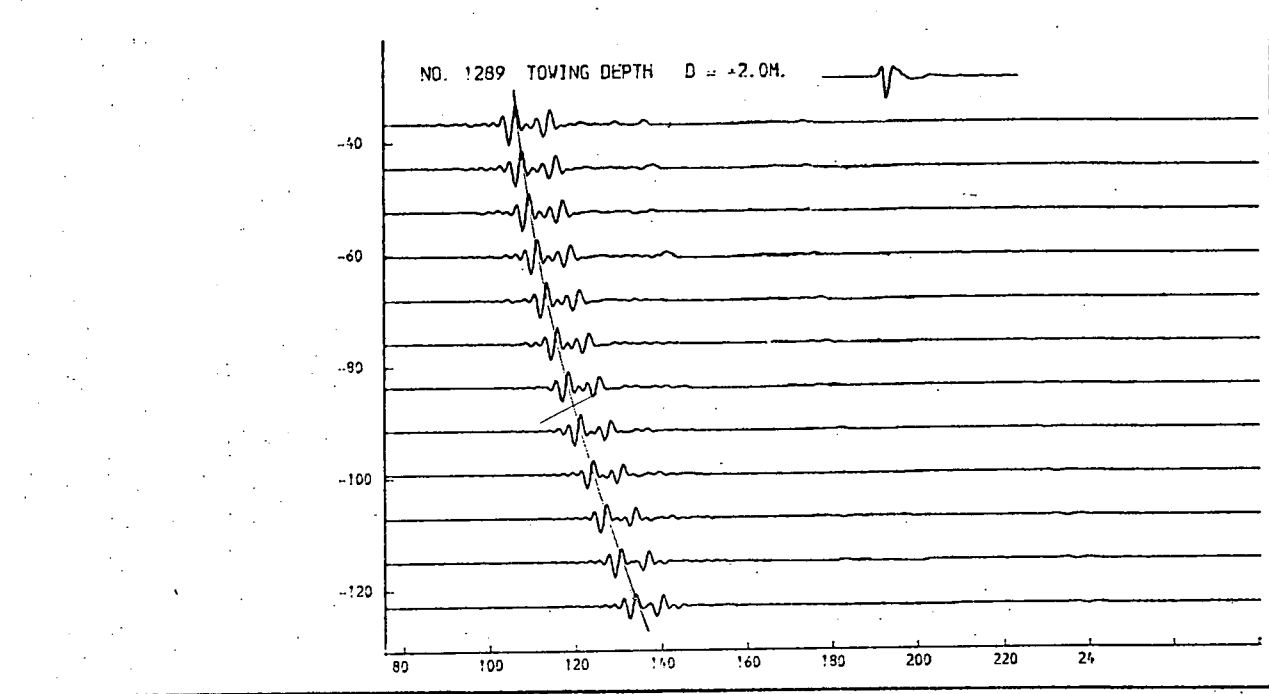
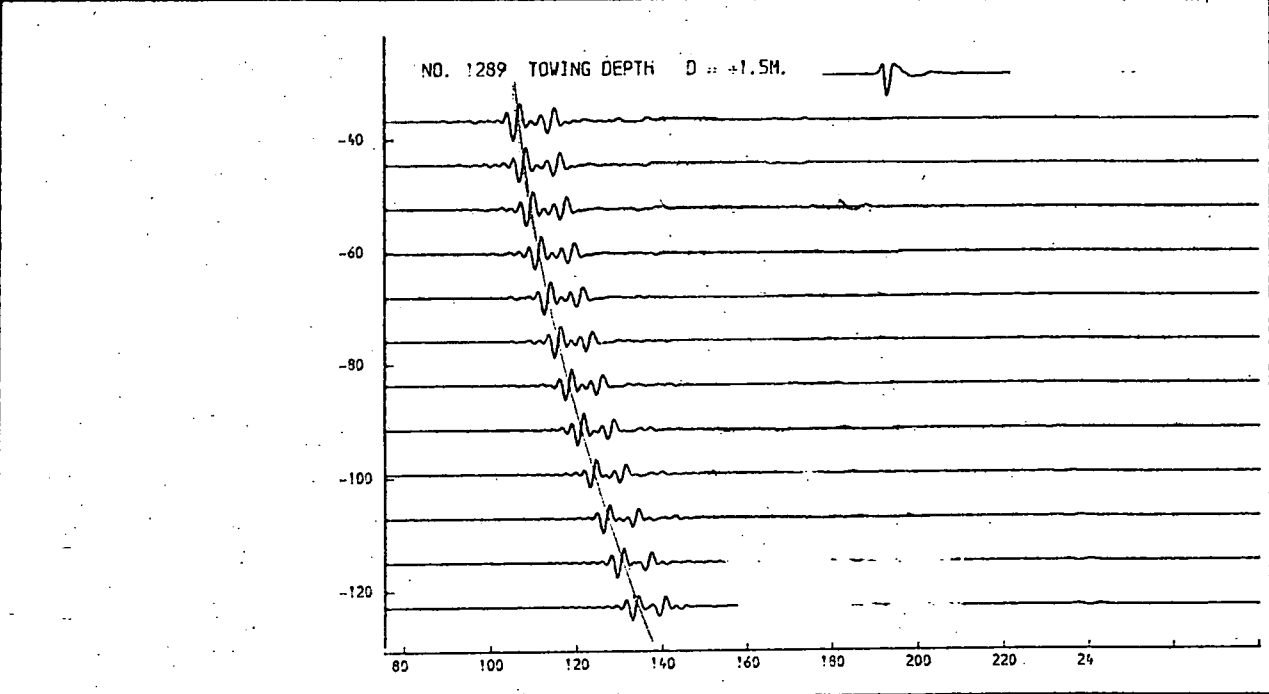
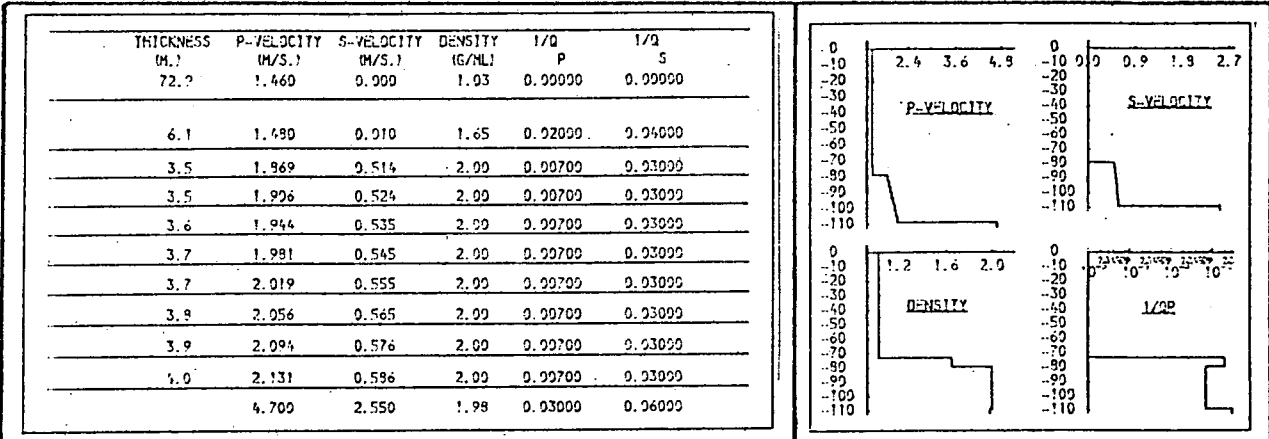


Fig. 9.17b
See 9.17a.

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S
74.4	1.460	0.300	1.03	0.00000	0.00000
6.1	1.480	0.010	1.65	0.02000	0.04000
29.7	2.000	0.550	2.00	0.00300	0.03000
4.700	2.550	1.98	0.03000	0.06000	

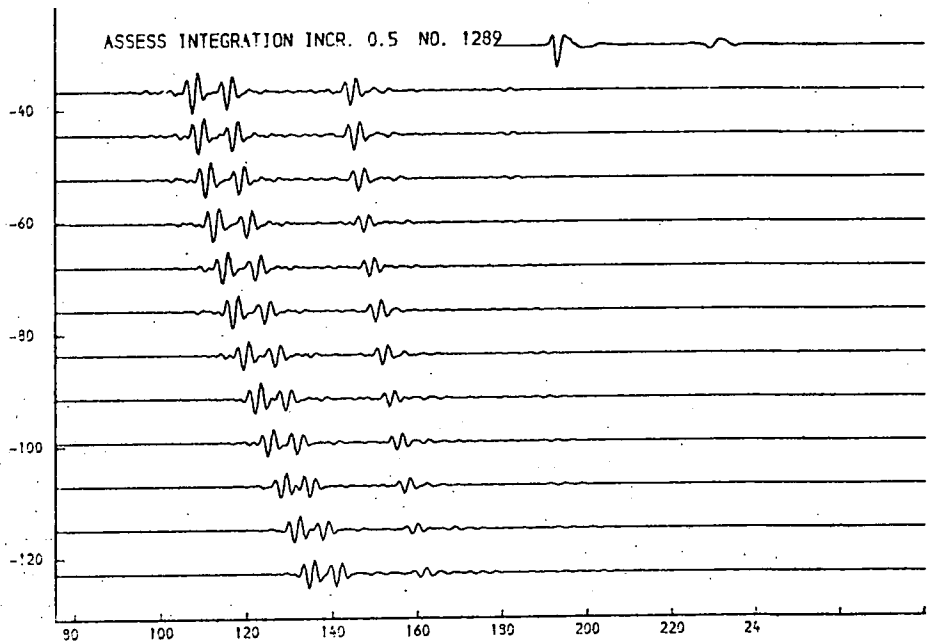
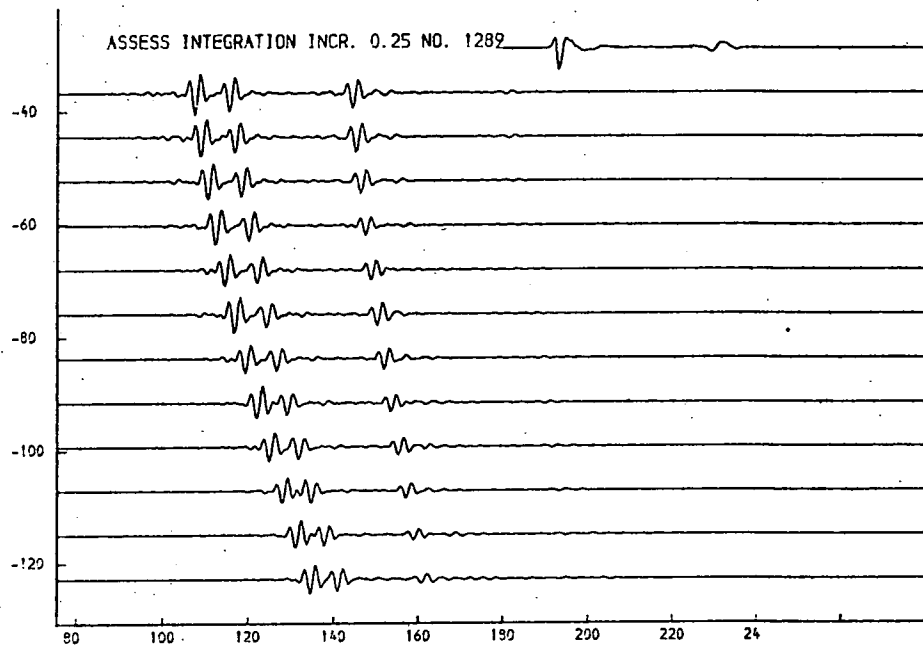
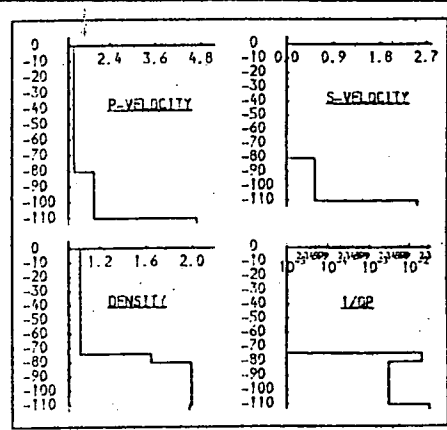


Fig. 9.18a

Variations are shown in the increment in angle of incidence used in generating the grid of values of the reflexion coefficient.

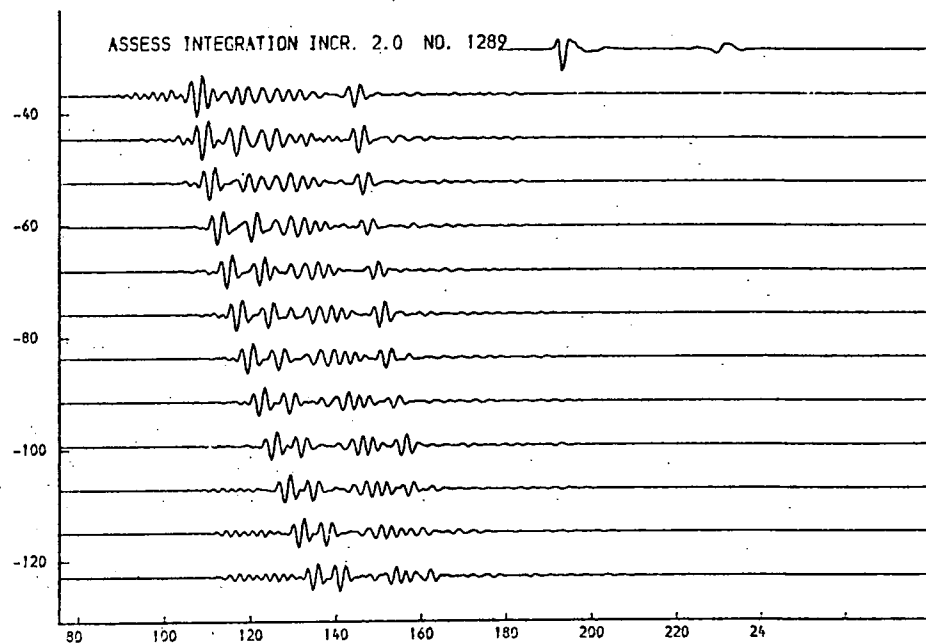
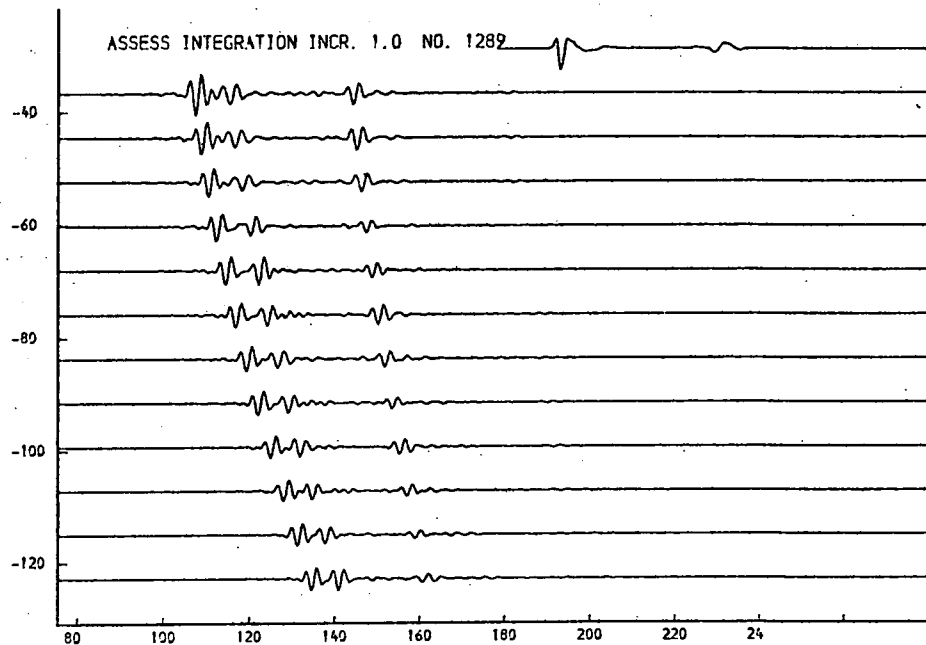
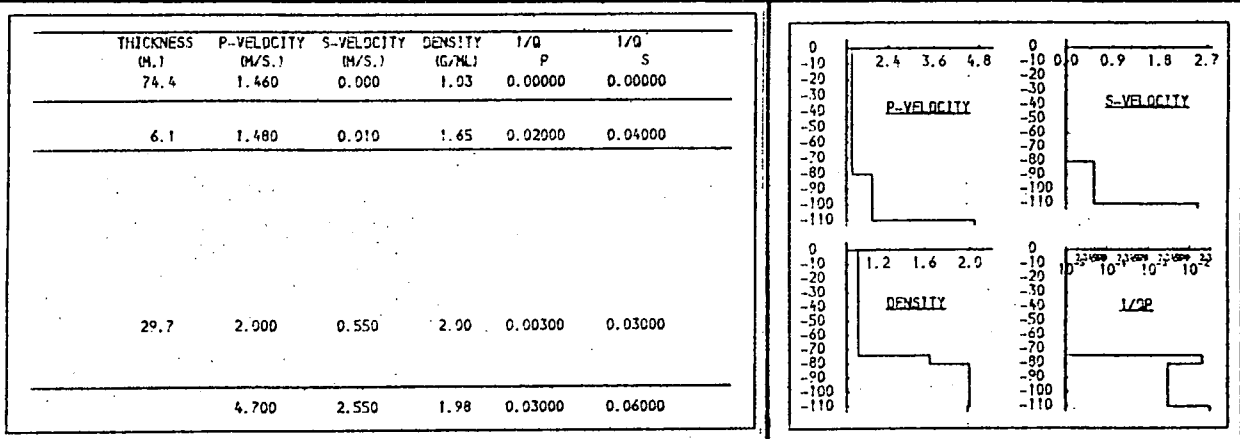


Fig. 9.18b

See 9.18a. Instability becomes obvious at an angle increment of 2 degrees.

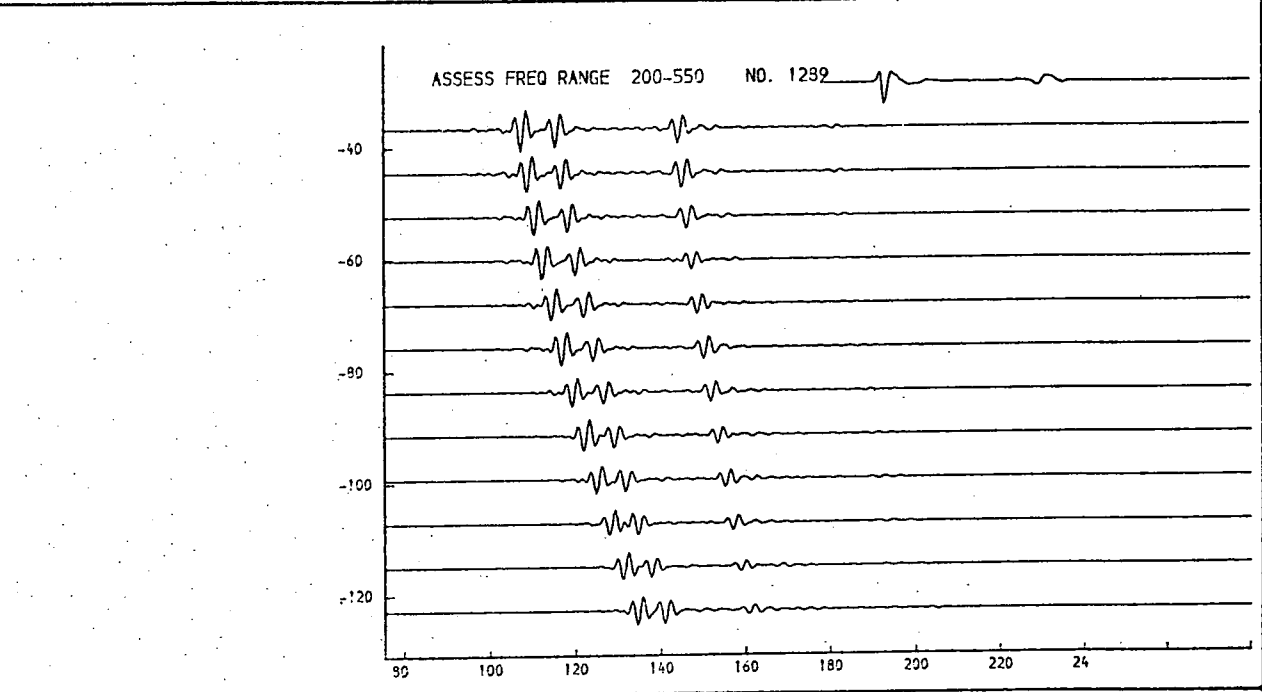
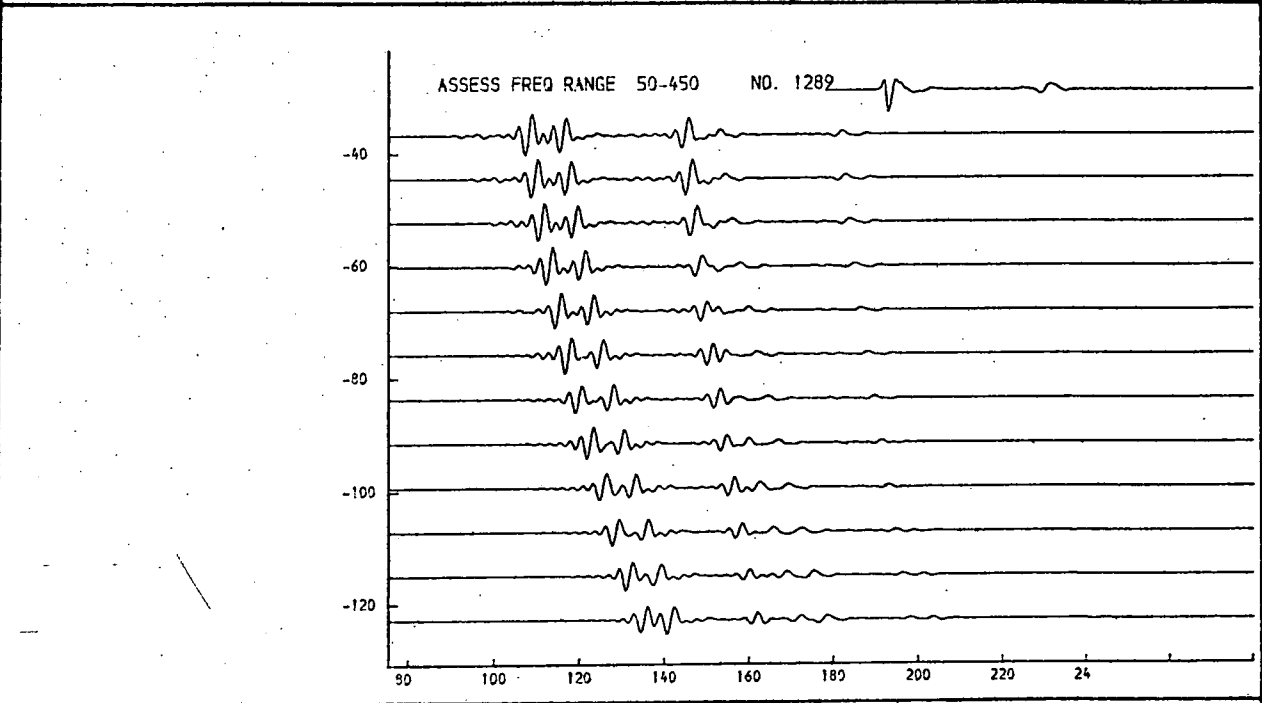
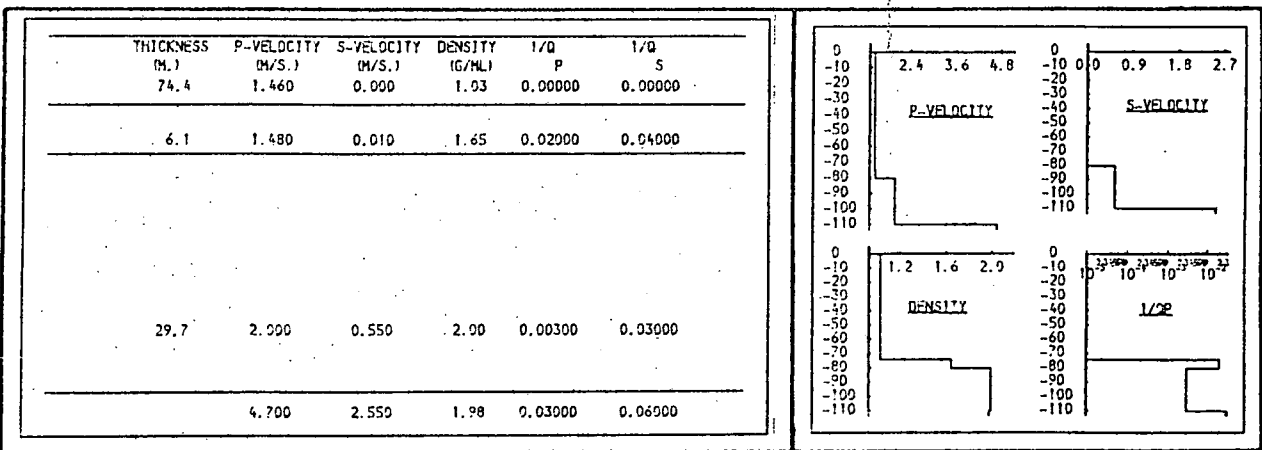


Fig. 9.19a

Although it is desirable to use a very wide frequency band for seismogramme construction (as is done in the model presented in fig. 9.24), the time required for computation can be decreased by restricting the bandwidth. The same model is constructed here for the different frequency bands shown.

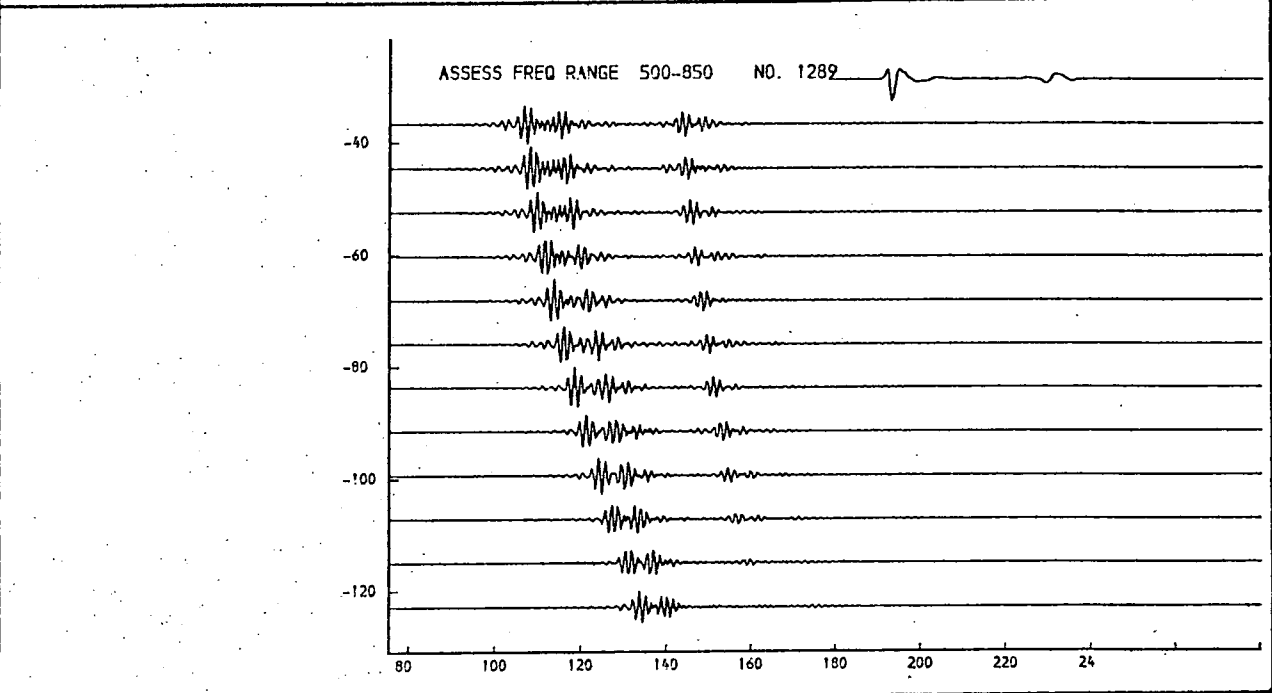
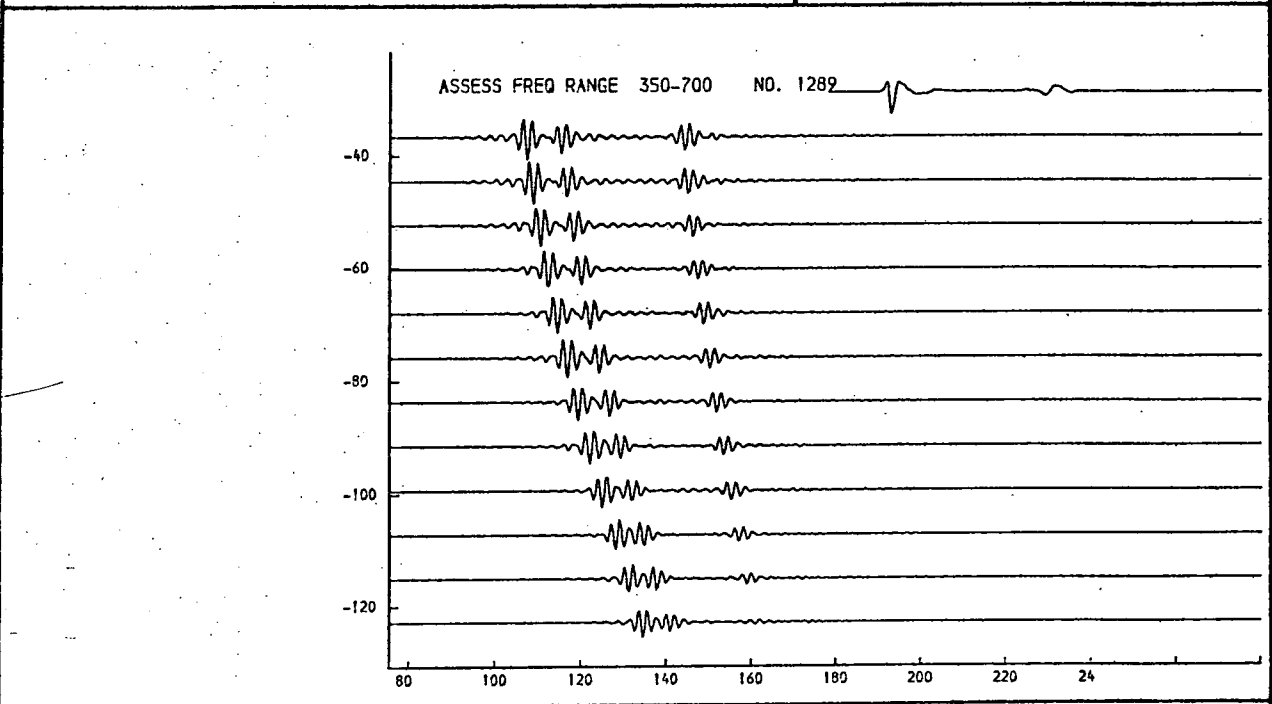
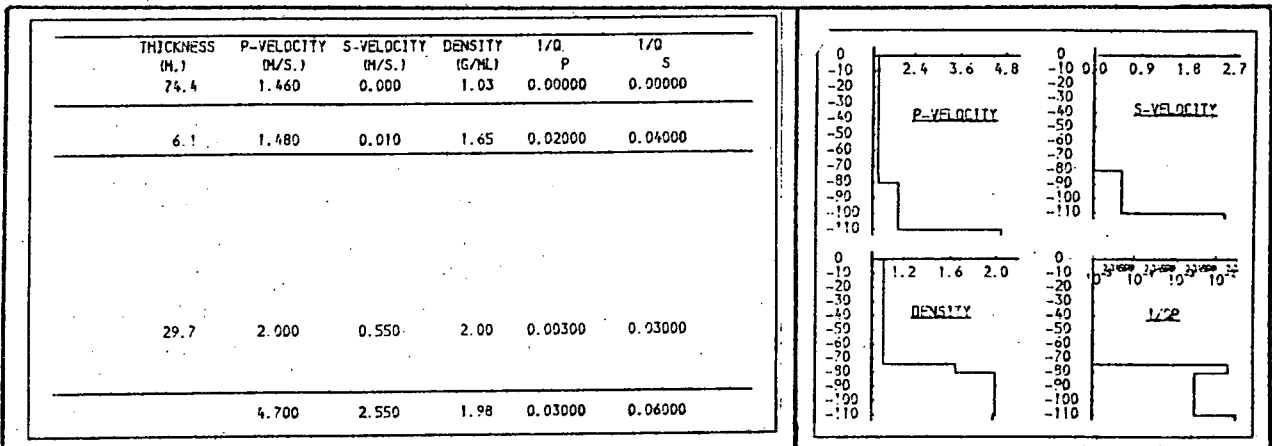


Fig. 9.19b

In this, and in the last figure, the characteristics of the model are seen not to be strongly dependent on frequency.

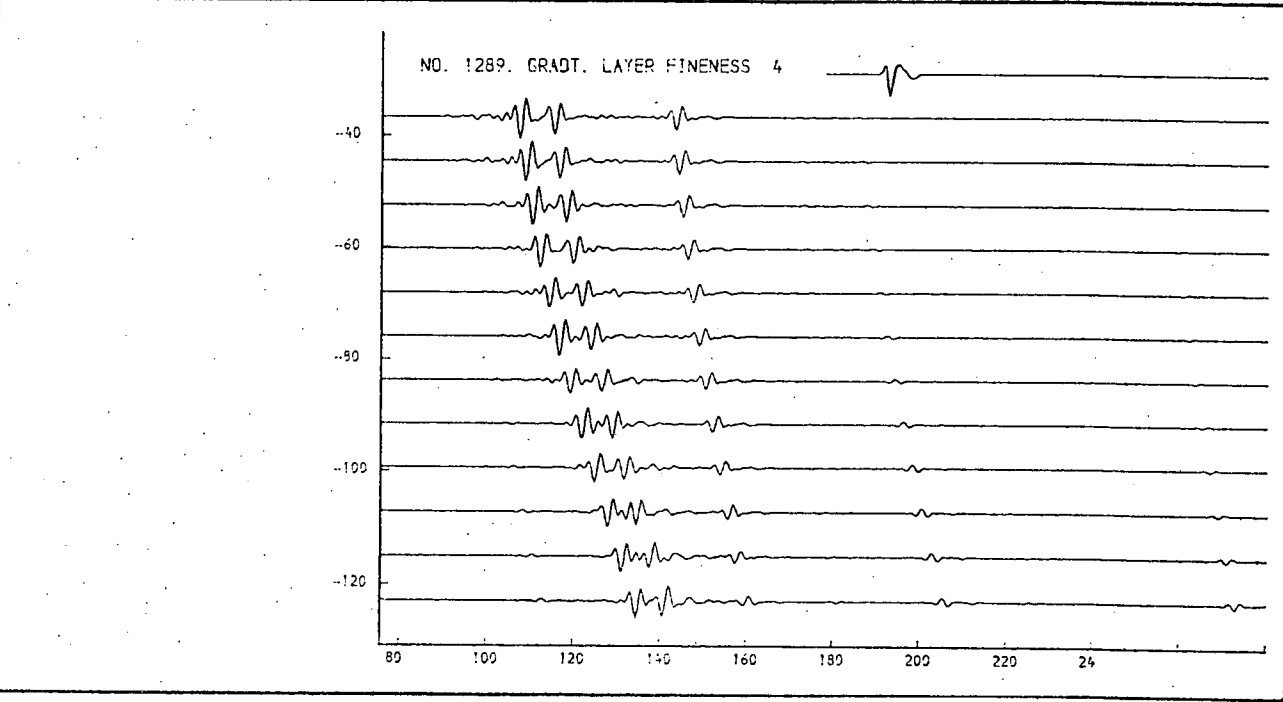
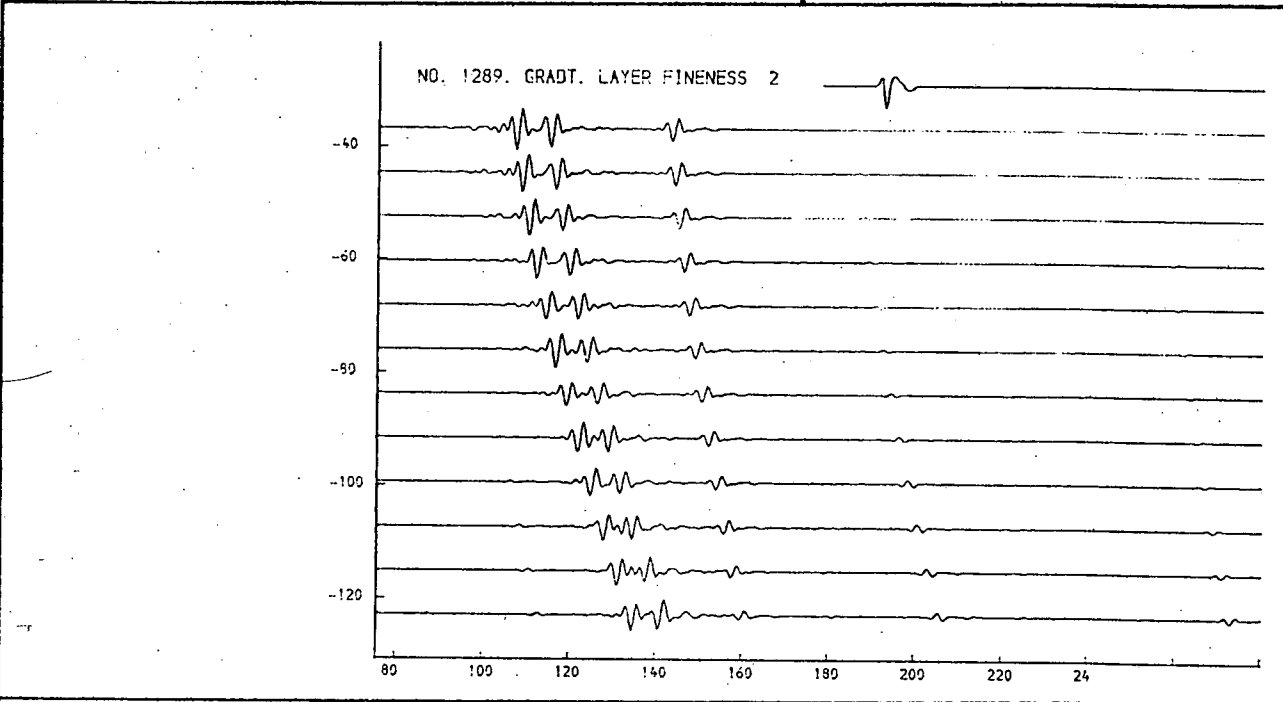
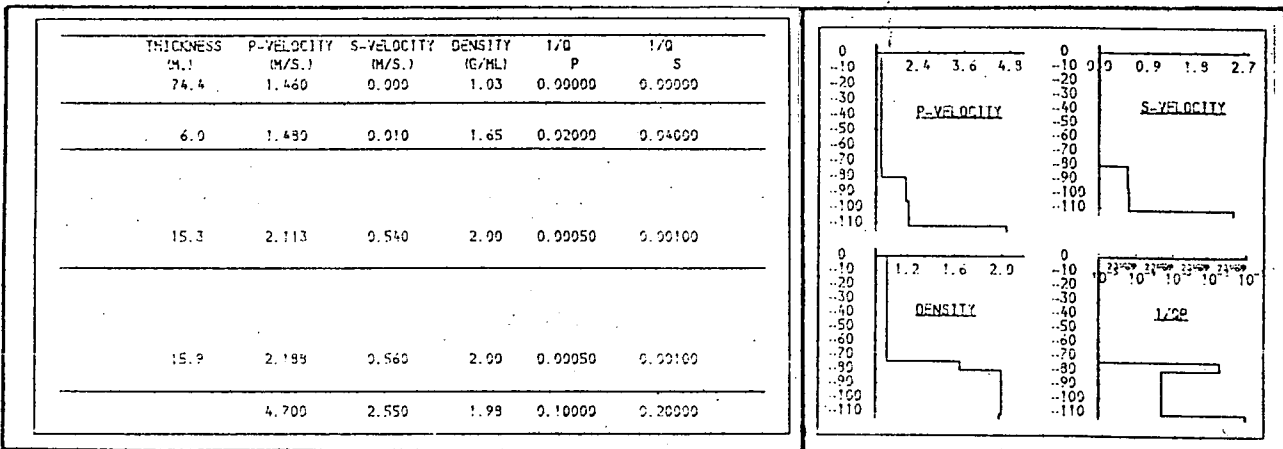


Fig. 9.21a

The Fuchs-Mueller method simulates gradients by dividing a layer into sub-layers. It is important to decide how fine a division is necessary. Division of layer 2 into 2 and 4 subdivisions is shown.

computation, which varies approximately as the square of the frequency bandwidth, to a reasonable value such as would allow many seismogrammes to be constructed. When a seemingly good model was found in this way, its construction was repeated with the full frequency bandwidth of 50-1000Hz.

A check was carried out on the method of modelling gradients used by the Fuchs-Mueller programme. Gradients are dealt with by dividing the layer in which it is desired to establish a velocity gradient into a number of layers in each of which the velocity is constant. It is of importance to discover if the procedure is sensitive to the fineness of layer subdivision. The results are shown in figs. 9.21a and b for a small gradient and in figs. 9.22a and b for a larger one. It appears that this modelling procedure is insensitive to the number of subdivisions when the increment in p-velocity between sub-layers is kept less than ca. 5% and in s-velocity to less than 20% of the average value in the whole layer. In the case of the seismogrammes modelled, this means that a subdivision into 8 layers is usually sufficient. As with increasing frequency bandwidth, increasing the structural complexity of a model by inserting more layers increases computation time. A more economical way of dealing with complex models is sought in the development of the "Gaussian beam" method of constructing seismogrammes

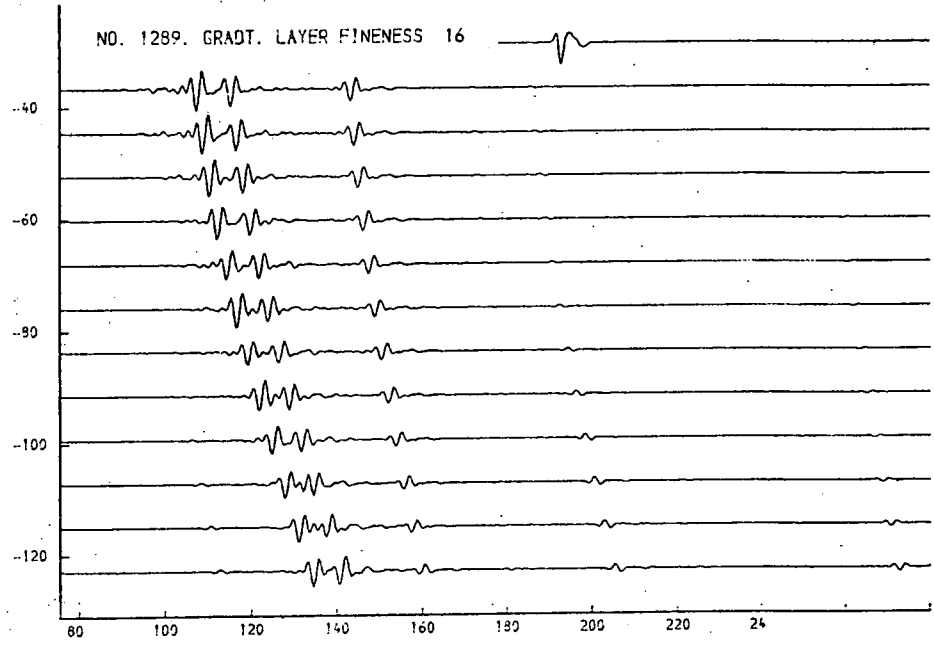
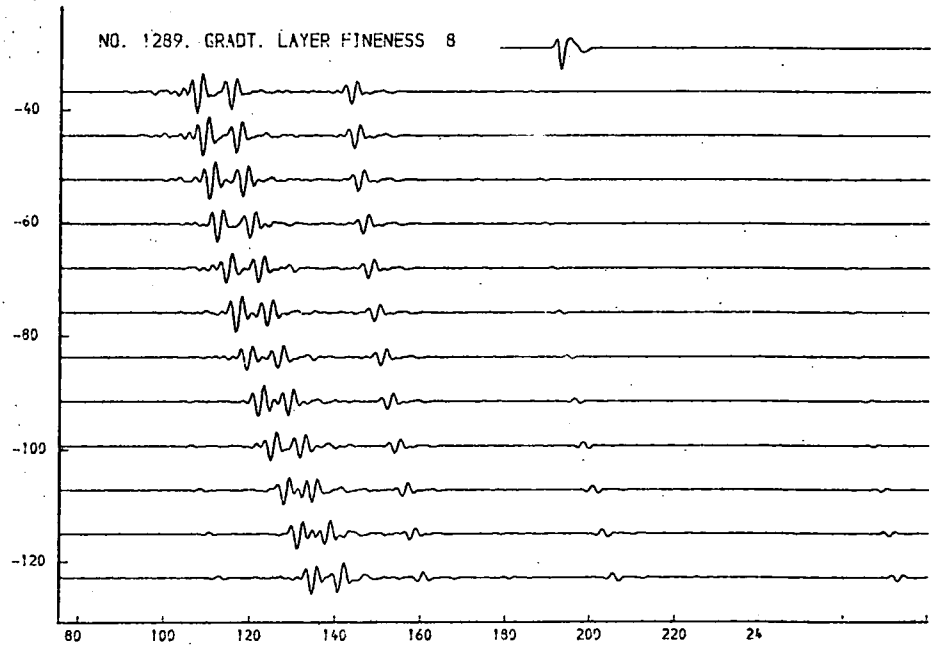
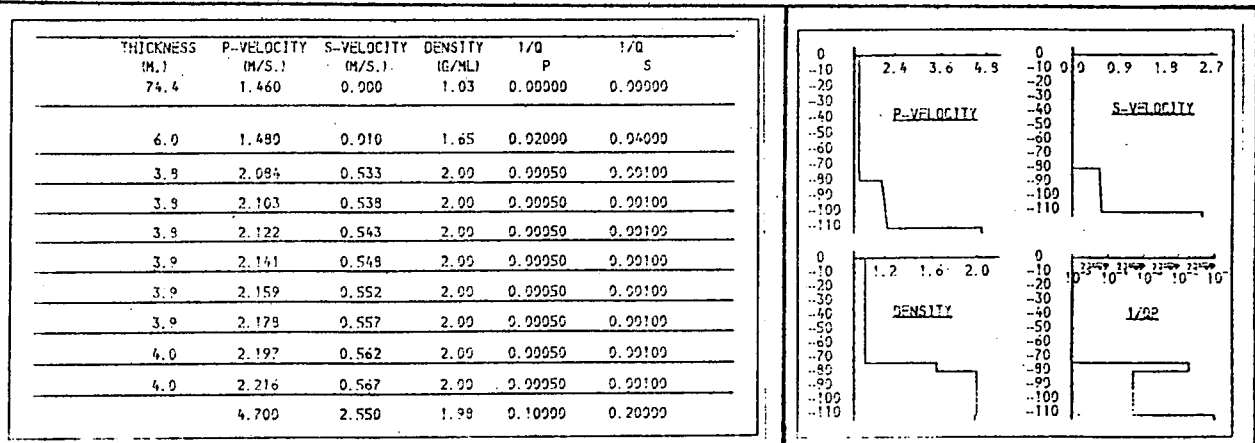


Fig. 9.21b

The Fuchs-Mueller method simulates gradients by dividing a layer into sub-layers. It is important to decide how fine a division is necessary. Division of layer 2 into 8 and 16 subdivisions is shown.

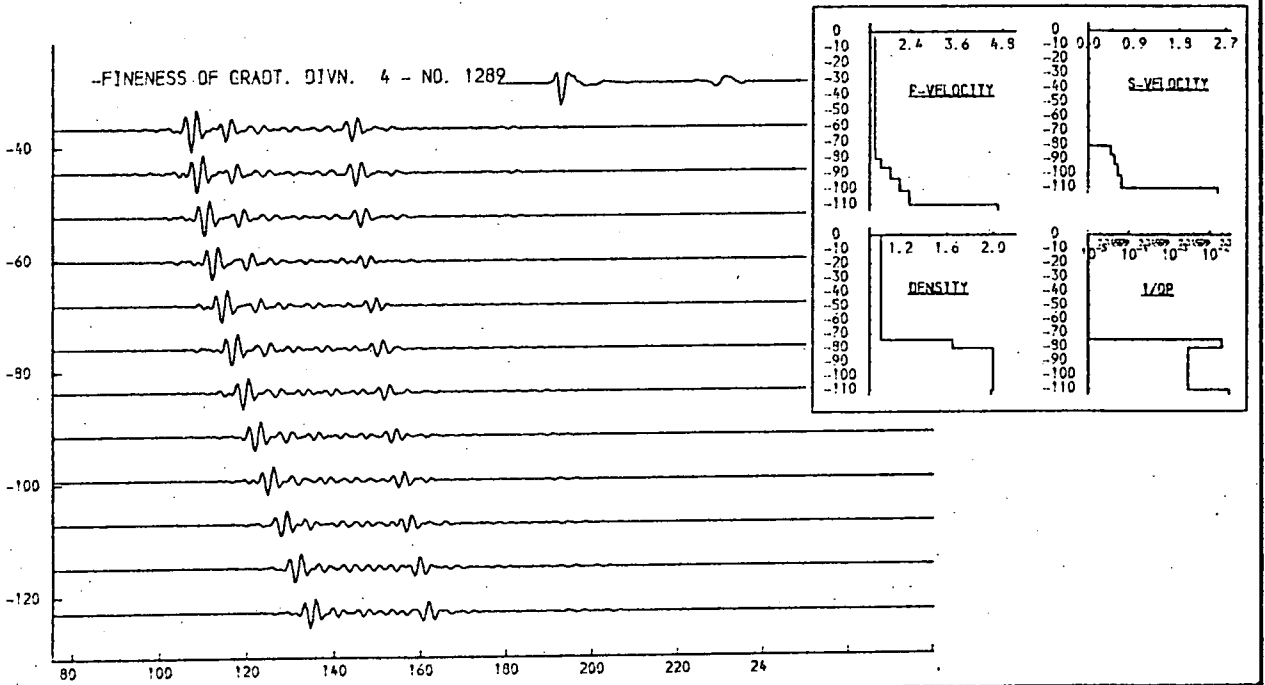
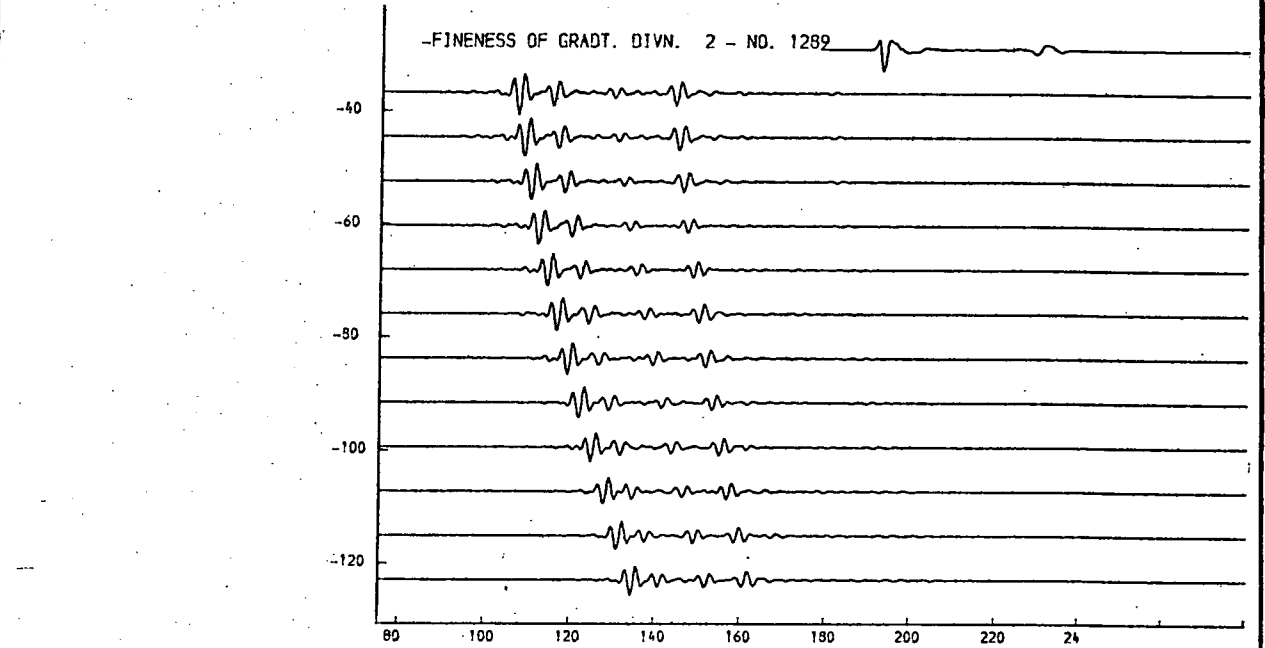
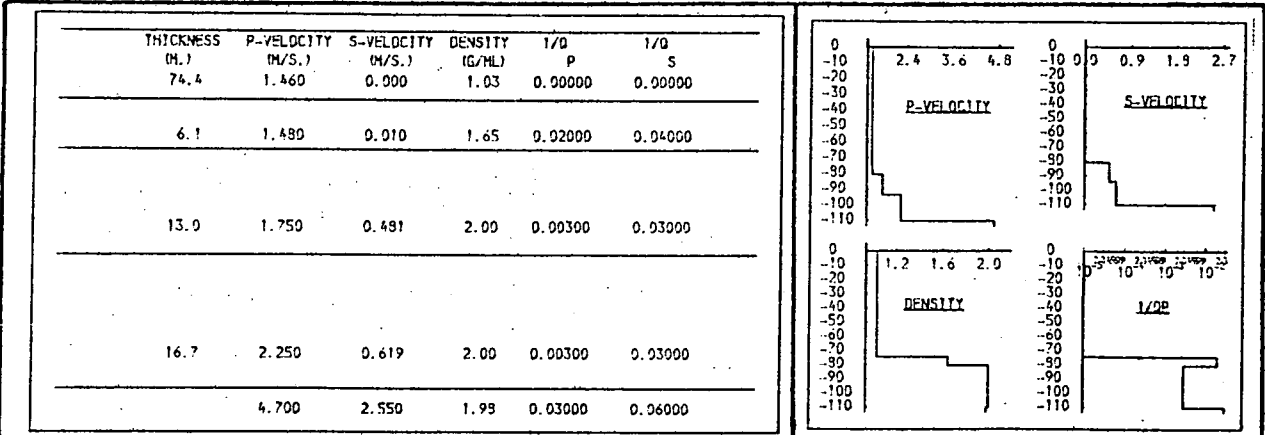


Fig. 9.22a

The effect of dividing into sub-layers to simulate a gradient is shown here and in the next figure for a larger velocity range than was depicted in figs. 9.21a and b.

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S
74.4	1.460	0.000	1.03	0.00000	0.00000
6.1	1.490	0.010	1.65	0.02000	0.04000
2.9	1.563	0.430	2.00	0.00300	0.03000
3.7	1.698	0.464	2.00	0.00300	0.03000
3.4	1.913	0.498	2.00	0.00300	0.03000
3.6	1.939	0.533	2.00	0.00300	0.03000
3.9	2.063	0.567	2.00	0.00300	0.03000
4.1	2.189	0.602	2.00	0.00300	0.03000
4.3	2.313	0.636	2.00	0.00300	0.03000
4.5	2.439	0.670	2.00	0.00300	0.03000
	4.700	2.550	1.98	0.03000	0.06000

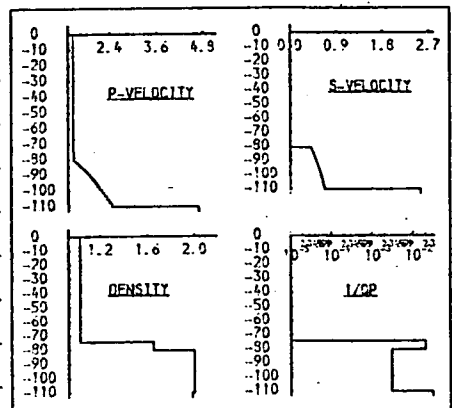
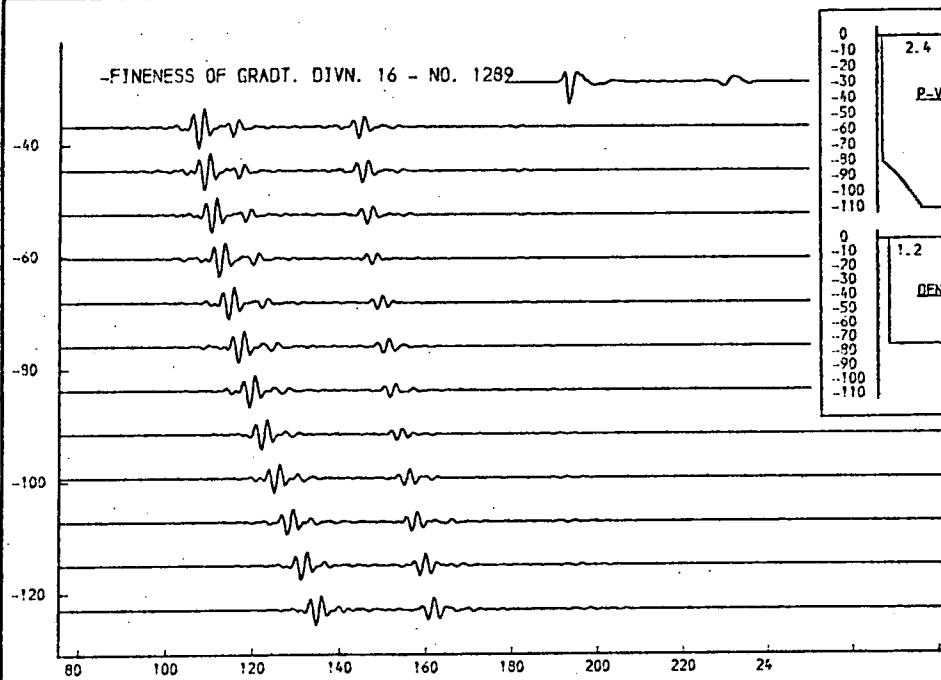
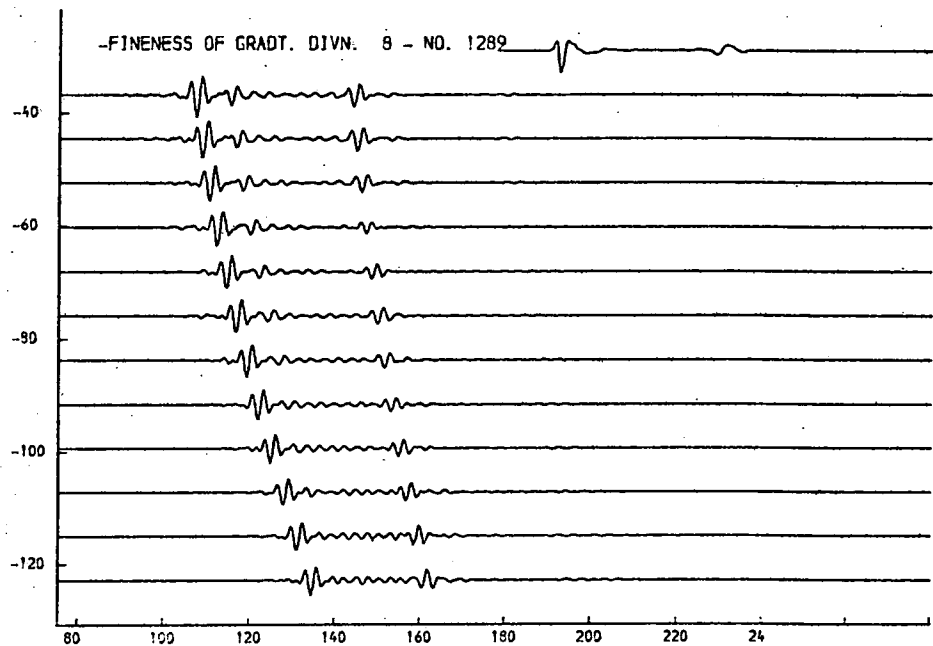
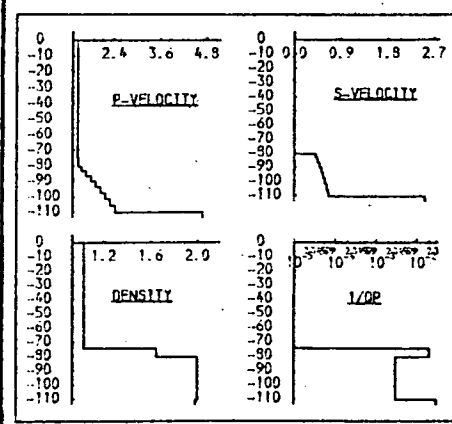


Fig. 9.22b

The effect of dividing into sub-layers to simulate a gradient is shown here and in the last figure for a larger velocity range than was depicted in figs. 9.21a and b.

(Cerveny et al., 1982), which aims to combine aspects of the ray-tracing (section 5.2) and the reflectivity (section 8.2) methods. The further development of such a technique and the continuing advances in computer hardware can be expected in the near future to ease computational constraints.

With the above considerations in mind, a number of model seismogrammes were constructed for selected field records from the Beaufort Sea.

9.5 DISCUSSION OF MODELLING SEISMOGRAMME NO. 1289.

Initial T^2-X^2 analysis gave an approximate value of 2.1km/sec for the compressional wave velocity in the layer between "r1" and "r2". This is in itself a surprising result and indicates that there is something rather unusual about the material in this layer, because the geological setting does not allow the plausible supposition of the occurrence of materials such as well-consolidated glacial tills which might normally be associated with such a velocity. The top sediment layer is too thin to be interpreted sensibly by the T^2-X^2 method, and has been assigned a velocity in the range appropriate for a water-saturated clayey mud. To eliminate the possibility that the reflector "r2" was

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S
74.4	1.460	0.900	1.03	0.00000	0.00000
6.1	1.490	0.910	1.65	0.02000	0.04000
3.7	1.934	0.532	2.00	0.00100	0.03000
3.9	1.953	0.537	2.00	0.00100	0.03000
3.9	1.972	0.542	2.00	0.00100	0.03000
3.9	1.991	0.547	2.00	0.00100	0.03000
3.9	2.009	0.553	2.00	0.00100	0.03000
3.9	2.029	0.559	2.00	0.00100	0.03000
3.9	2.047	0.563	2.00	0.00100	0.03000
4.0	2.066	0.569	2.00	0.00100	0.03000
	1.400	0.910	1.98	0.03000	0.06000

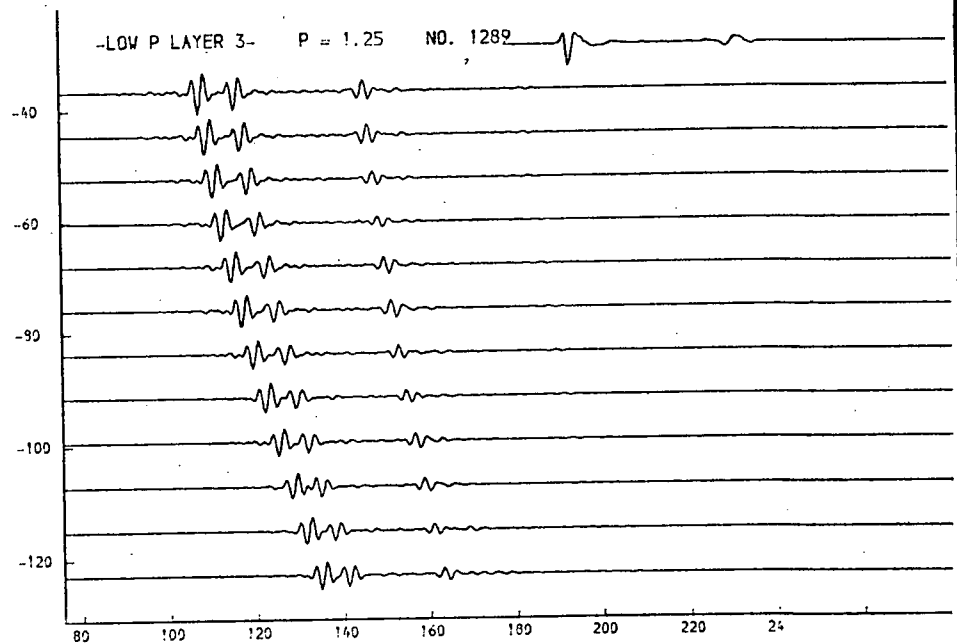
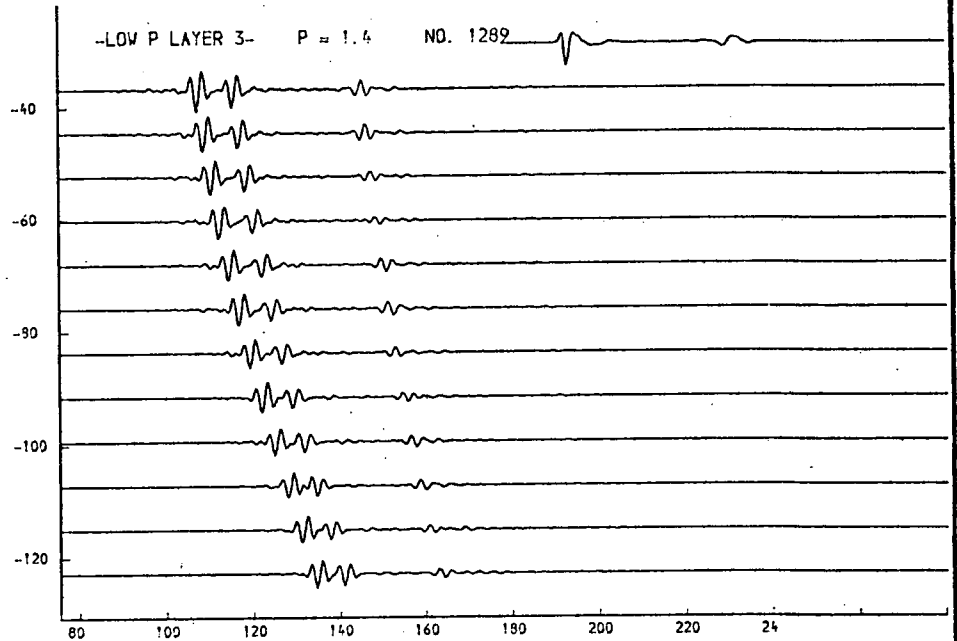
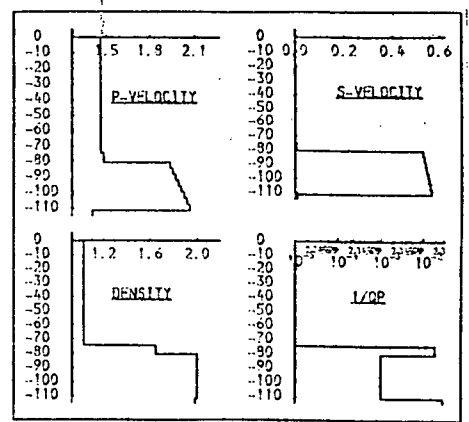


Fig. 9.23a

A low p-velocity in the material below reflector r2 does not produce a good model.

associated with a layer having a very low velocity due to gasification of the sediment, the models of fig. 9.23a and b were constructed with a low-velocity substrate. The character of the reflector "r2" in the models is clearly quite different from that of the field seismogramme (fig. 9.16) and has, in particular, its phase inverted. Such a model is consequently rejected.

One of the important contributions which the synthetic seismogramme method has made in other fields lies in its ability to handle velocity gradients. The region where such gradients become of the greatest importance is that in which source and receiver subtend a wide angle at the reflecting zone. In the particular field seismogramme treated here, incidence angles are confined to less than 45° so that velocity gradient effects are not expected to be large. In view of this and of the relatively low values which have been reported for gradients in sediments (see section 9.3), the modelling approach taken was to treat uniform layers in the manner outlined below. Possible anomalous velocity gradients were considered subsequently.

Eventually by trial and error the model shown in fig. 9.24 was constructed and is proposed as a reasonable fit to the field seismogramme of fig. 9.16. The parameters used in the construction of this model are

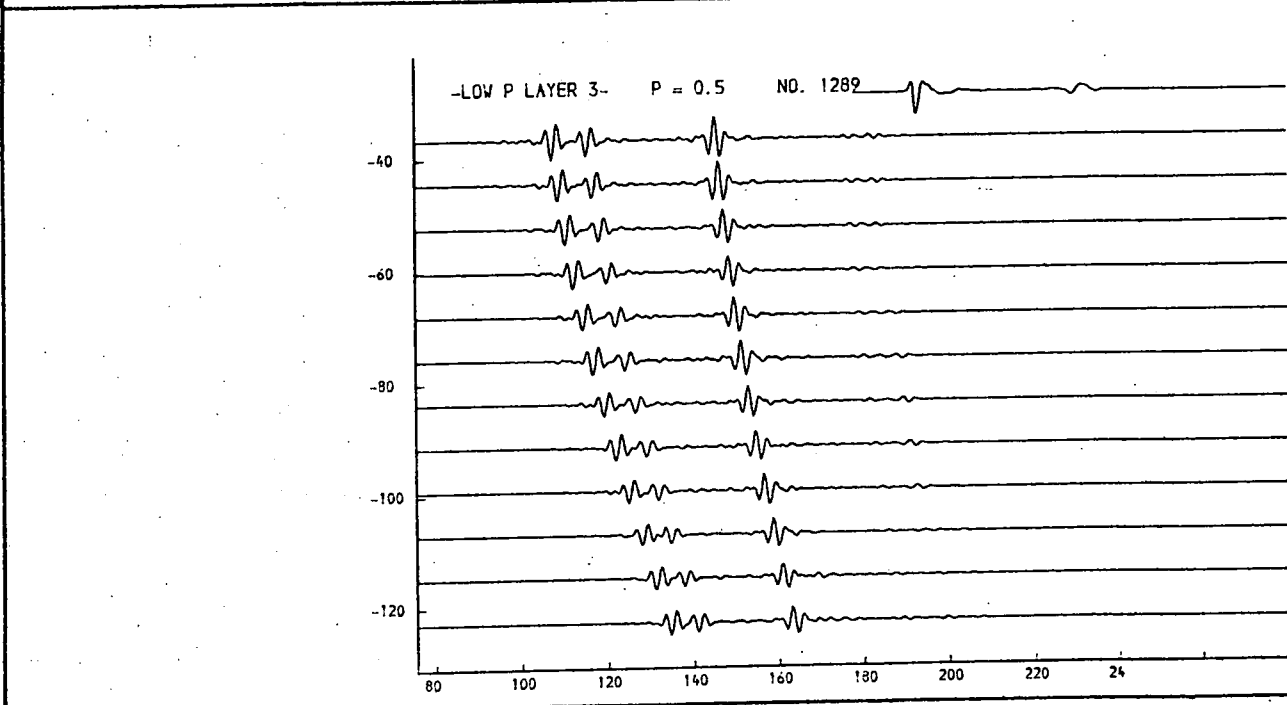
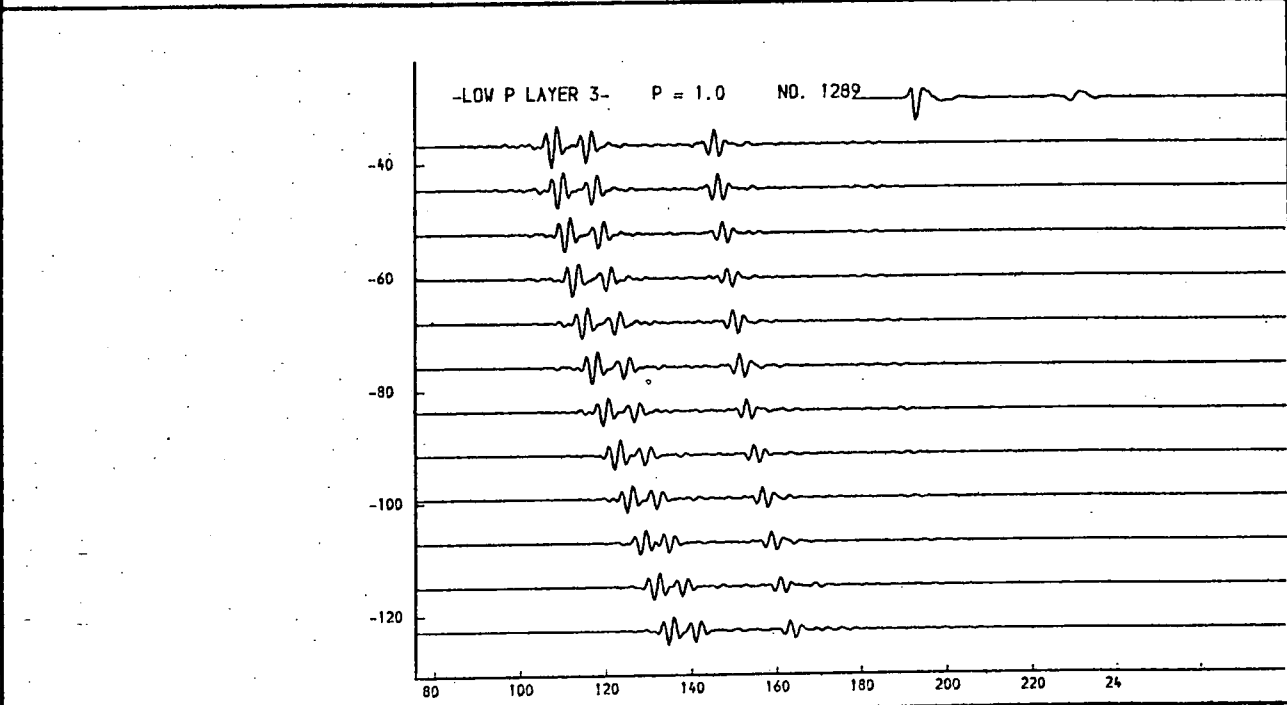
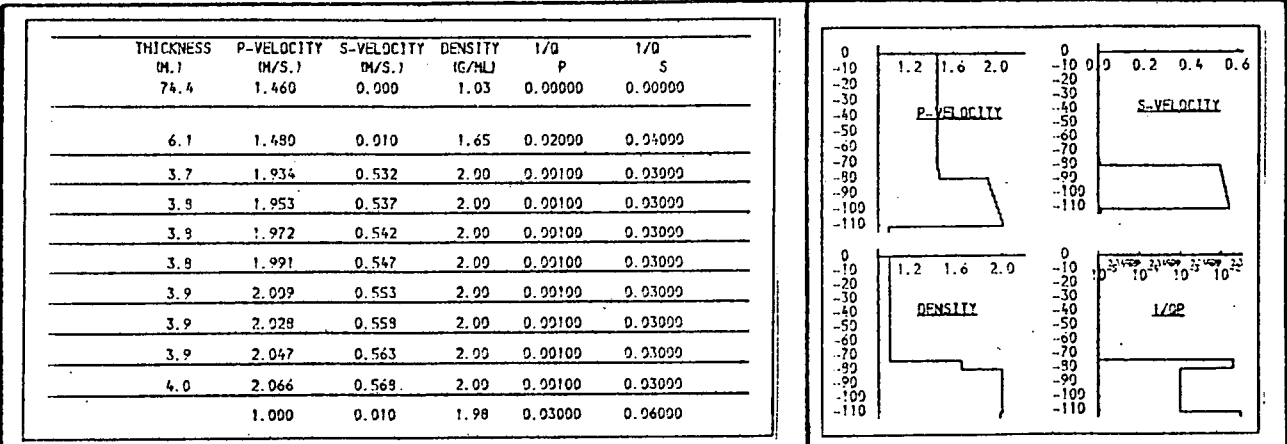


Fig. 9.23b

A low p-velocity in the material below reflector r2 does not produce a good model.

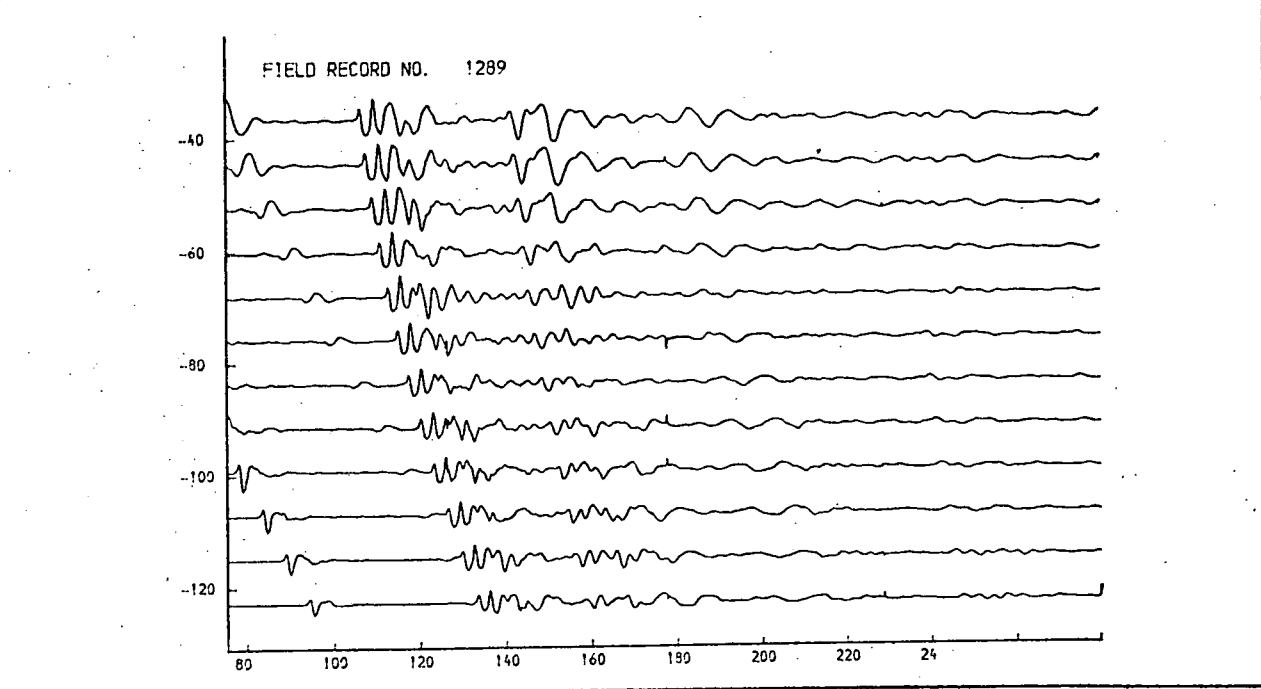
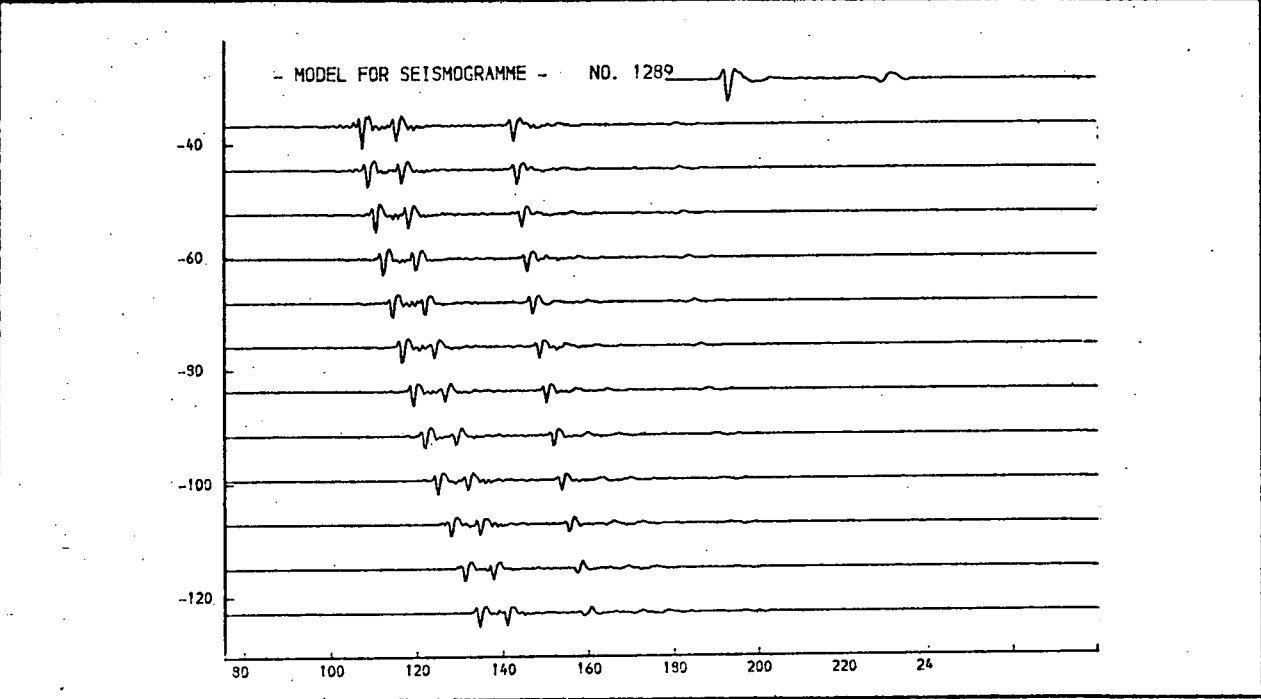
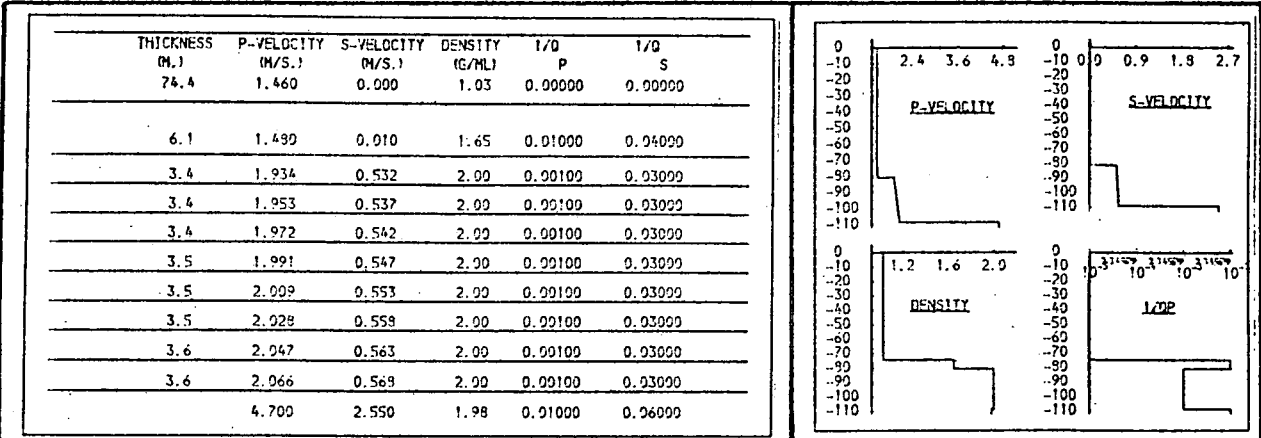


Fig. 9.24. Model for No. 1289.

Suggested model for seismogramme no. 1289, constructed in a wide frequency band (50 to 1000 Hz.). The possible value ranges for the parameters are discussed in the text and illustrated by the models shown in figs. 9.25 to 9.44 (see volume 2).

TABLE 9.1RESULTS OF MODELLING SEISMOGRAMME NO. 1289

	p-velo. (km/s.)	s-velo. (km/s.)	density (gm/ml)	1/Qp	1/Qs
Layer 1	1.46-1.55 (1.48)	0.0-0.2 (0.01)	1.60-1.70 (1.65)	<0.02 (0.02)	not found
Layer 2	1.95-2.05 (2.00)	0.5-0.9 (0.55)	1.95-2.05 (2.00)	<0.002 (0.001)	>0.02 (0.03)
Layer 3	4.2-4.9 (4.7)	2.5-2.8 (2.55)	1.95-2.05 (1.98)	>0.01 (0.01)	not found

Above are tabulated the parameter ranges which are possible in fitting the field record with a simple layered model. Suggested values about which variations have been considered are shown in brackets.

given in table 9.1. Figures referred to in the following detailed discussion of variations in the parameters have been placed in the accompanying volume for greater ease of reference between text and figures. Variations in the parameters influence the constructed seismogramme as follows:

(i) Layer 1, p-velocity (figs. 9.25a,b).

This uppermost layer is barely resolved by the field record because of surface ghosting effects, and attempts to match travel times are not feasible. A low velocity (fig. 9.25a, upper) does not seem to fit well. The amplitude of "r1" rises too much in channels 11 and 12. Velocities near that of seawater and above are possible, but the decrease in amplitude of "r1" for $p=1.55$ and 1.60 suggests that the value should not greatly exceed 1.50 .

(ii) Layer 1, s-velocity (figs. 9.26a,b).

The model is insensitive to changes in the s-velocity in layer 1 between 1 and 200 m/sec.

(iii) Layer 1, density (figs. 9.27a,b).

Density influences the amplitude of reflectors "r1" and "r2", but does not influence the variation of amplitude with

For figures 9.25a to 9.44d,

see second volume.

offset, the lower densities (1.55, 1.60) give too high an amplitude for "r1", while the higher (1.70, 1.75) give too low an amplitude for "r2".

(iv) Layer 1, attenuation - p (figs. 9.28a,b,c,d).

Very high attenuation in layer 1 suppresses both reflectors "r1" and "r2". A value of $1/Q$ below ca. 0.02 appears possible. This value cannot be significantly increased by considering gradients above "r2" (figs. 9.28e,f).

(v) Layer 1, attenuation - s (figs. 9.29a,b).

The model is insensitive to variation in this parameter.

(vi) Layer 2, p -velocity (figs. 9.30a,b).

By fitting arrivals on the basis of travel times, the velocity in this layer is found to be 2.00 ± 0.05 km/sec.

(vii) Layer 2, s -velocity (figs. 9.31a,b,c,d,e).

Low values of this parameter are associated with an amplitude for "r1" in channels 11 and 12 which is too large. High values give rise to a later arrival whose arrival

time is dependent on the s-velocity in this layer. This appears to be a p-s converted wave which does not appear on the field record. Its amplitude can be controlled by varying the s-attenuation in this layer. The higher values for the s-velocity in this layer are also associated with a rise in the amplitude of reflector "r2" towards greater offset distances. This is not what is observed, although it can be controlled by varying the values of Q_p and Q_s in layer 3 (q.v.). Such interactions make it difficult to determine the s-velocity precisely, but it seems to lie between ca. 0.5 and 0.9 km/sec.

(viii) Layer 2, density (figs. 9.32a,b).

This parameter has a small effect on the amplitude of both the reflexions "r1" and "r2". A range between ca. 1.95 and 2.05 appears possible.

(ix) Layer 2, attenuation-p (figs. 9.23a,b,c,d)

Higher values of attenuation in layer 2 reduce the amplitude of "r2" too much. Very low values are possible. It appears that $1/Q_p$ cannot be greater than ca. 0.002. The possible effect of a velocity gradient above the reflector "r2" was investigated. Such a phenomenon is known to increase the amplitude of a reflexion in the modelling of crustal seismogrammes (R. Long, personal

communication). This type of model is shown in figs. 9.34a and b. It does not allow significant increase in the attenuation in layer 2 without decrease in the amplitude of "r2". The gradient also reduces significantly the velocity contrast at the upper interface, bringing about too great a decrease in the amplitude of "r1". This can be avoided by using a non-linear gradient such as is shown in fig. 9.35a,b. This restores the reflector "r1", but "r2" remains too large towards channels 11 and 12 relative to the earlier channels (see also fig. 9.34a and b), even when a very high value for the attenuation in channel 3 (fig. 9.35a,b) is used to help control this rise.

(x) Layer 2, attenuation-s (figs. 9.36a,b; 9.37a,b)

As noted above in connexion with the discussion about the value for the s-velocity in layer 2, the value of $1/Q_s$ in this same layer is of importance in controlling the appearance of the late arrival, interpreted as a p-s conversion. Large changes in $1/Q_s$ (figs. 9.36a,b) show that a value of 0.04 or greater will suppress this event. (The rather odd effects which occur for $1/Q_s=4.0$ may be a result of using a value for attenuation which stretches the theoretical basis for the seismogramme construction beyond its limits. Barnard et al. (1975) have stated that the materials modelled must be "low loss", but Folds

and Loggins (1977) find that the lossy elastic treatment of Brekhovskikh (1960), which is the basis for the Fuchs-Mueller method of modelling, is exact even for very high attenuation, provided the signal used approximates the $\exp(i\omega t)$ time dependence. Although the airgun signature is clearly not too ill-behaved, it is necessary to exercise some circumspection in dealing with very high values of attenuation. Accordingly, the validity of the model for $1/Q_s=4.0$ (fig. 9.36b) is doubted.)

Finer increments in $1/Q_s$ (fig. 9.37a,b) show that its value must be no less than 0.02, if the late converted arrival is to be suppressed. (A stronger conversion such as would be associated with an s-velocity in the higher part of the range thought possible (see (vii) above) might require a slightly higher minimum value for $1/Q_s$ in layer 2, as shown in figs. 9.37c and d.

(xi) Layer 3, p-velocity (figs. 9.38a,b,c,d).

No deeper reflector than "r2" is identified. This makes it impossible to evaluate the p-velocity in this layer on the basis of travel time. Instead its influence on the amplitude of "r2" is investigated. This shows up in the overall amplitude of "r2", and a range from ca. 4.2 to 4.9 km/sec. appears possible.

(xii) Layer 3, s-velocity (figs. 9.39a,b,c,d).

The s-velocity in layer 3 influences the increasing offset of the reflector "r2". Values between ca. 2.50 and 2.80 km/sec. appear possible.

(xiii) Layer 3, density (figs. 9.40a,b).

As with the densities in the other layers, the effect on reflector amplitude varies imperceptibly with offset. A range of 1.95 to 2.05 appears possible.

(xiv) Layer 3, attenuation-p (figs. 9.41a,b,c,d).

This parameter has some influence on the reflector "r2", but the influence is not a strong one. If higher values are used from the range thought possible for the s-velocity in layer 2, then a relatively high value for $1/Q_p$, ca. 0.03 is required to control the rise of "r2" at high offset (figs. 9.42a,b,c,d). For lower values of s-velocity in layer 2, a value of 0.01 may be sufficient.

(xv) Layer 3, attenuation-s (figs. 9.43a,b,c,d)

This parameter does not appear to have a significant influence on the seismogramme. The odd appearance of the reflector "r2" for $1/Q_s=1.0$ is attributed to the doubtful

validity of the modelling method for very high attenuation.

(xvi) Velocity gradients (figs. 9.44a,b,c).

Gradients have already been considered in connexion with their possible influence on the limiting values for attenuation--(see (ix) above). A rapid change in velocity just above "r2" does not give a good model (see fig. 9.35a and the discussion on variation in $1/Q_p$ of layer 2), but a more limited change in velocity throughout layer 2 may bring about a small improvement (figs. 9.44a,b). A linear gradient corresponding to a total velocity range of up to ca. 150m/sec. is possible (i.e. a gradient of ca. 4m/sec/m.).

A possible gradient in s-velocity in layer 2 has been considered (figs. 9.44c and d). In these models, the value of $1/Q_s$ (layer 2) has been kept low ($1/Q_s=0.001$). Comparison with figs. 9.36a,b and 9.37a-d and the above discussion on $1/Q_s$ (layer 2) will show that such a gradient in the s-velocity can control the unwanted p-s conversion ("i" in figs. 9.31b-d), although for the highest gradients, the fit with the amplitude of "r2" in the farthest offset channels is not very good.

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S
63.0	1.460	0.990	1.03	0.00001	0.00001
4.9	1.934	0.532	2.00	0.00200	0.02000
4.9	1.953	0.537	2.00	0.00200	0.02000
4.9	1.972	0.542	2.00	0.00200	0.02000
5.0	1.991	0.547	2.00	0.00200	0.02000
5.0	2.009	0.553	2.00	0.00200	0.02000
5.1	2.028	0.558	2.00	0.00200	0.02000
5.1	2.047	0.563	2.00	0.00200	0.02000
5.2	2.066	0.568	2.00	0.00200	0.02000
	4.700	2.900	1.99	0.02000	0.04000

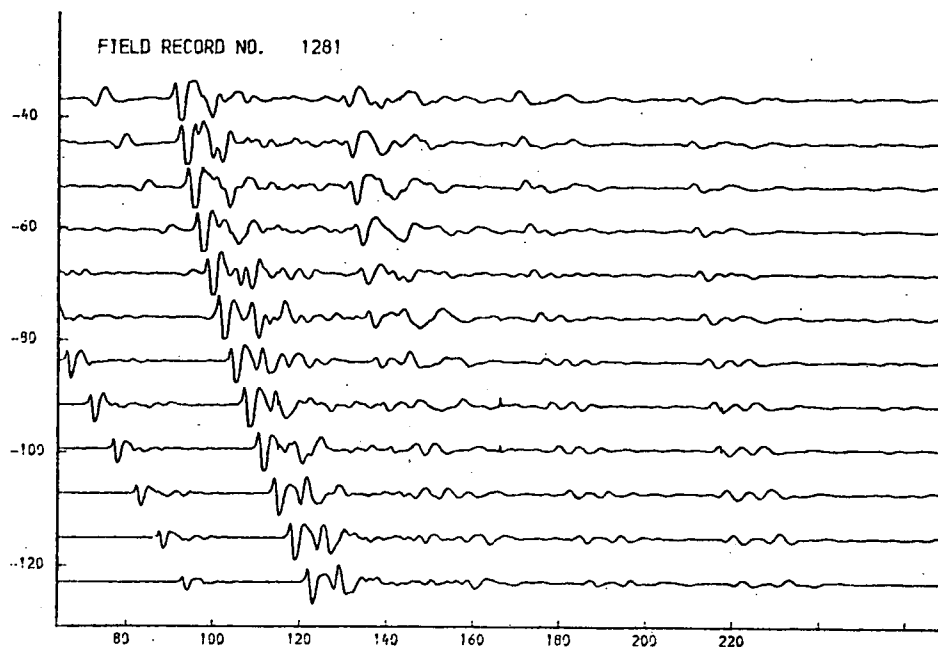
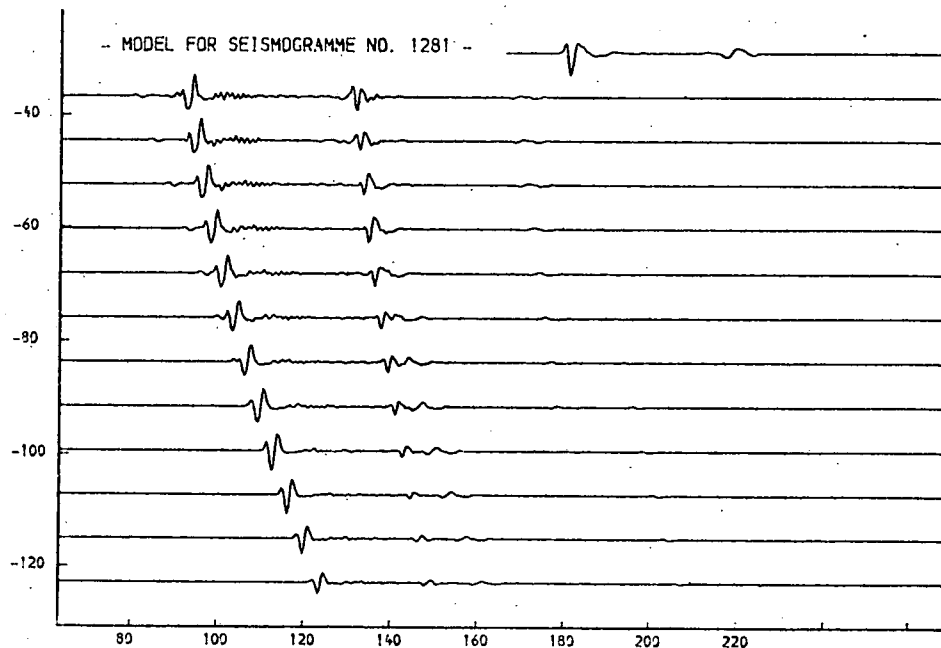
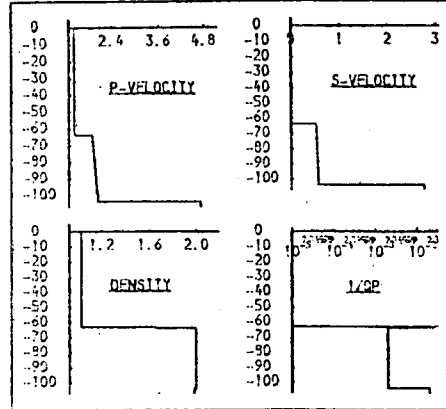


Fig. 9.45. Model for No. 1281.

Suggested model for seismogramme no. 1281, constructed in a wide frequency band (50 to 1000 Hz.). The possible value ranges for the parameters are discussed in the text and illustrated by the models shown in figs. 9.46 to 9.62 (see volume 2).

9.6 DISCUSSION OF MODELLING SEISMOGRAMME NO. 1281.

This field record (figs. 9.1 and 9.45) is from the same area as that discussed in section 9.5, and has some features in common with it. The deepest reflector seen ("r" of fig. 9.1) appears similar both on the subbottom profiles (c.f. figs. 9.1 and 9.11) and on the 12-channel seismogrammes (c.f. figs. 9.10, 9.16 and 9.45). This reflector is not continuous between the two records - an abrupt break in it can be seen clearly in fig. 9.1. The modelling process to be described will suggest that the reflector "r" of fig. 9.1 separates layers with similar properties to those separated by "r2" of the field record no. 1289. In anticipation of this, the reflector is referred to as "r2" in the following discussion. The reflector "r1", barely resolved in record no. 1289, is not resolved at all in record no. 1281, but is suggested by pinger and echo sounder records. The thickness of this layer is considered first in the modelling procedure. Figures referred to in the following discussion are found in the accompanying volume.

(i) Thickness of layer at seabed (figs. 9.46a,b,c,d,e).

A thickness of ca. 0.7m. appears to give the best relative amplitudes between the seabed reflexion and "r2". A thinner layer makes "r2" too small; a thicker layer

For figures 9.46a to 9.73d,

see second volume.

makes it too large.

(ii) Layer 1, p-velocity (figs. 9.47; 9.48a,b).

This parameter influences the amplitude of the reflector "r2". Low values, such as might be associated with a highly gasified sediment do not appear possible and above the seawater velocity the fit with the field record degrades by the reduction in amplitude of "r2" at small offset distance. A velocity close to that of seawater is suggested, but the range could extend from ca. 1.4 to 1.55km/sec.

(iii) Layer 1, s-velocity (figs. 9.49a,b).

Variation in this parameter, over quite a wide range, does not appear to influence the model.

(iv) Layer 1, density (figs. 9.50a,b).

The density of this layer seems to have very little influence on the model, and remains indeterminate.

(v) Layer 1, attenuation-p (figs. 9.51a,b).

Extremely high values for this parameter (fig. 9.51a) influence the model, but more reasonable values

(fig. 9.51b) seem to have an insignificant influence over a wide range.

(vi) Layer 1, attenuation-s (figs. 9.52a,b).

Over a very wide range of values, this parameter has no influence on the model.

(vii) Layer 2, p-velocity (figs. 9.53a,b)

The average p-velocity in this layer affects both the arrival time and the amplitude of "r2". A value of 2.0km/sec. appears optimal, with a spread of up to 50m/sec., although on the basis of amplitude, it may be that this error could be set at a slightly lower value.

(viii) Layer 2, s-velocity (figs. 9.54a,b,c,d).

In the discussion of seismogramme no. 1289, it was indicated that the effect of this parameter interacts particularly with s-attenuation in this layer in influencing the model. Similar arguments apply here, and it can be seen in the figure 9.54d that for very high s-velocity, the p-s converted arrival produced in the model breaks through even when $1/Q_s$ (layer 2) is held at the high value of 0.02. Quite a wide range of values for the s-velocity may be possible (ca. 0.3 to 0.8 km/sec.).

(ix) Layer 2, density (figs. 9.55a,b).

The influence of density is seen in the amplitude change in the channels at small offset. Values higher than 2.0 produce a fall-off in these events, but the lower values of 1.9 and 1.8 seem possible.

(x) Layer 2, attenuation-p (figs. 9.56a,b,c).

Values for $1/Q_p$ above ca. 0.002 do not appear possible.

(xi) Layer 2, attenuation-s (figs. 9.57a,b,c,d,e).

As with the seismogramme 1289, this parameter is used to control the p-s converted arrival. For an s-velocity in layer 2 averaging 0.55km/sec. (as illustrated), $1/Q_s$ must be greater than 0.01 to suppress the converted arrival.

(xii) Layer 3, p-velocity (figs. 9.58a,b,c,d).

As with the seismogramme no. 1289, no deeper reflector than "r2" can be distinguished so that it is impossible to evaluate this parameter on the basis of arrival times. It does influence the amplitude of "r2" and a range of values from ca. 4.1 to 5.0 km/sec. can be accommodated by the model.

(xiii) Layer 3, s-velocity (figs. 9.59a,b,c,d).

This parameter affects the rate of change in the amplitude of "r2" with increasing offset distance. Values from ca. 2.6 to 3.0 km/sec. seem to give the best fit.

(xiv) Layer 3, density (figs. 9.60a,b).

This density affects the amplitude of "r2", and a value between 1.90 and 2.10 seems to give the best fit.

(xv) Layer 3, p-attenuation (figs. 9.61a,b,c,d).

Very high values of $1/Q_p$ in this layer cause too rapid a fall-off in "r2" with increasing offset distance. Interaction with the effect of the s-velocity in this layer (see fig. 9.59) makes it difficult to assign an upper limit to $1/Q_p$, but a value as high as 0.1 does not seem impossible.

(xvi) Layer 3, s-attenuation (figs. 9.62a,b)

This parameter has no obvious influence on the model, except at very high values, where the validity of the modelling process is in doubt (see section 9.5(x)).

(xvii) Velocity gradients.

As with the modelling for seismogramme no. 1289, it can be seen (fig. 9.63a,b) that the effect of increasing the p-velocity gradient in layer 2 is to cause a rise in the amplitude of the reflector "r2" at high offset distance. This effect becomes apparent at a gradient of 7m/sec/m. If the value of the parameter $1/Q_p$ in layer 3 is set very high ($1/Q_p=0.1$), then this effect of an increasing gradient is considerably suppressed, becoming apparent only when the gradient rises to the very high value of 20m/sec/m. (see figs. 9.64a,b).

The possibility of suppressing the p-s converted arrival "i" (c.f. figs. 9.54c,d) by invoking an s-velocity gradient is illustrated in figs. 9.65a,b. This effect was shown for seismogramme no. 1289 in figs. 9.44c,d where the value of $1/Q_s$ was held low to clarify the effect. In figs. 9.65a,b, $1/Q_s$ remains at the relatively high value (0.02) and the varying s-velocity gradient is seen to suppress the residual converted arrival "i" when the size of the gradient approaches 20m/sec/m. Another influence of the high s-velocity gradient (fig. 9.65b) is to increase the amplitude of "r2" at high offset, an effect which is more pronounced when the attenuation in layer 3 is lower (figs. 9.66a,b).

VARIATIONS FOR HIGH GRADIENT AND HIGH ATTENUATION.

The variation of individual parameters (as in (viii) to (xiii) above) in the presence of a high p-velocity gradient in layer 2 (ca. 22m/sec/m.) and high attenuation in layer 3 ($1/Q_p=0.1$; $1/Q_s=0.2$) was considered and is discussed in the following text. The presence of a velocity gradient decreases the impedance contrast at the interface associated with the reflexion "r2" and in consequence is expected to decrease its amplitude. Effects other than this are given as follows:

(xviii) Layer 2, s-velocity (figs. 9.67a,b,c,d).

For low values of s-velocity, the presence of the high p-velocity gradient does not produce a large effect (compare figs. 9.67a and 9.54a), but the influence of the gradient increases with increasing s-velocity (compare figs. 9.67b,c with 9.54b,c), producing an increase in the amplitude of "r2" at high offset. At the highest values of s-velocity which were considered, viz. 1.0 and 1.2km/sec. (c.f. figs. 9.67d and 9.54d), the influence of the gradient decreases.

(xix) Layer 2, p-attenuation (figs. 9.68a,b).

The presence of a high p-velocity gradient does not greatly affect the influence of this parameter on the reflexion "r2" (c.f. figs. 9.56a,b,c).

TABLE 9.2

RESULTS OF MODELLING SEISMOGRAMME NO. 1281

	p-velo. (km/s.)	s-velo. (km/s.)	density (gm/ml)	1/Qp	1/Qs
Layer 1	not found	not found	not found	<0.3	not found
Layer 2	1.95-2.05 (2.00)	<0.7 (0.55)	1.9-2.1 (2.0)	<0.003 (0.002)	>0.02 (0.03)
Layer 3	4.3-4.9 (4.7)	2.7-3.1 (2.8)	1.9-2.1 (1.98)	>0.03 (0.03)	not found

Above are tabulated the parameter ranges which are possible in fitting the field record with a simple layered model. Suggested values about which variations have been considered are shown in brackets.

(xx) Layer 2, s-attenuation (figs. 9.69a,b,c,d).

The presence of the velocity gradient does not appear to influence the effects of this parameter (c.f. figs. 9.57a-e).

(xxi) Layer 3, p-velocity (figs. 9.70a,b,c,d).

The presence of the velocity gradient does not seem to have an influence on the effects of this parameter (c.f. figs. 9.58a,b,c,d).

(xxii) Layer 3, s-velocity (figs. 9.71a,b,c,d).

At the higher values which are illustrated for this parameter, the amplitude of "r2" at high offset is greater in the presence of the p-velocity gradient (c.f. figs. 9.59a,b,c,d).

(xxiii) Layer 3, p-attenuation (figs. 9.72a,b,c,d).

As this parameter is varied it can be seen that the relative amplitude of "r2" at high offset is always bigger in the presence of the p-velocity gradient (c.f. figs. 9.61a,b,c,d).

(xxiv) Layer 3, s-attenuation (figs. 9.73a,b,c,d).

The relative increase in the reflexion "r2" at high offset is again seen with this parameter (c.f. figs. 9.62a,b). Some of the very high values for $1/Q_s$

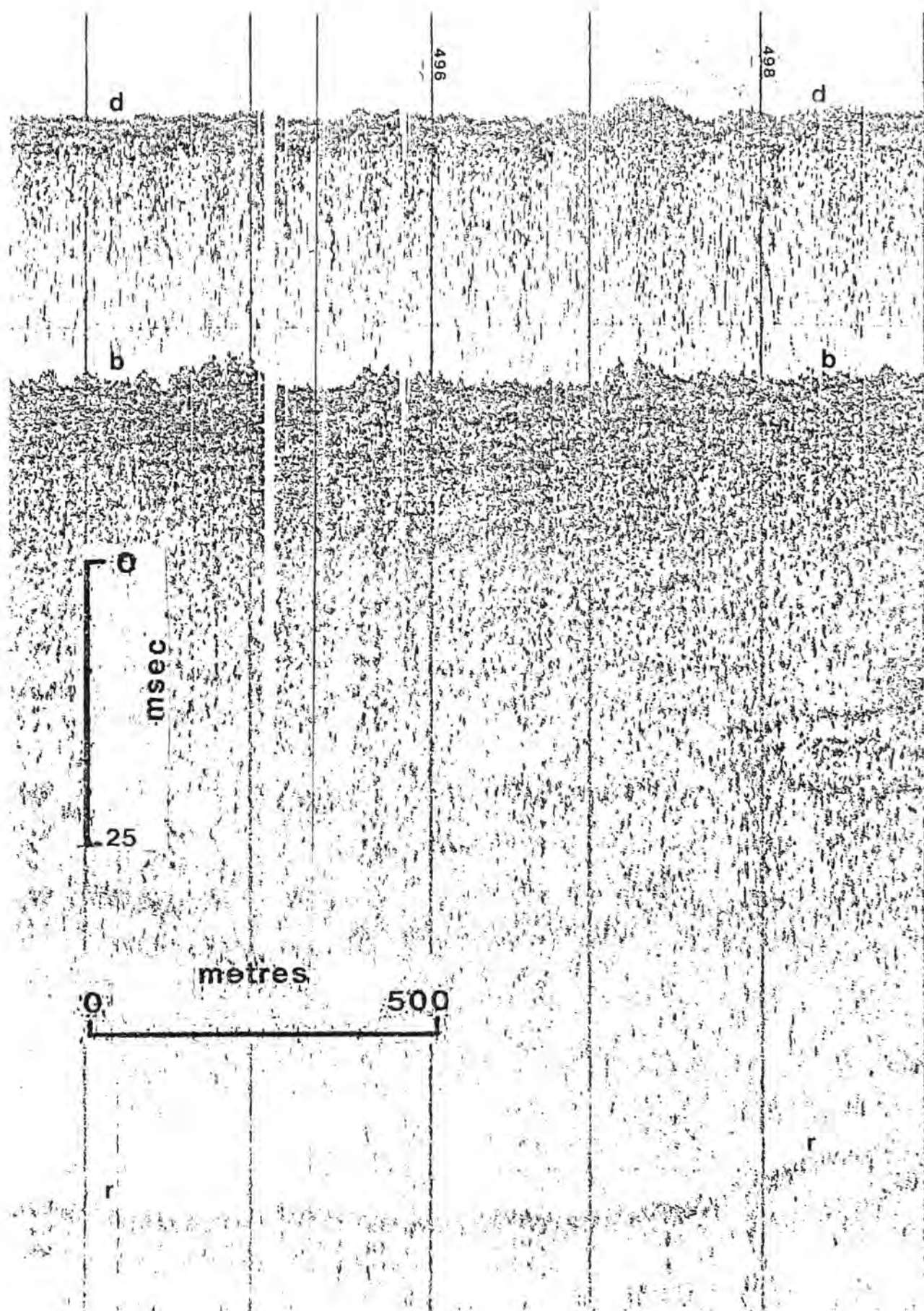


Fig. 9.75

Near-vertical incidence section in the neighbourhood of fix 496 of line 3 of the Beaufort Sea work (see fig. 9.2).

d - direct arrival r - subbottom reflexion
b - bottom reflexion

shown (figs. 9.73c,d) illustrate effects considered to be artifacts produced by using the modelling process beyond the range of its ability to give physically meaningful results.

9.7 DISCUSSION OF MODELLING SEISMOGRAMME NO. 496.

Whereas the field seismogrammes 1289 and 1281 discussed in sections 9.5 and 9.6 were recorded in a water depth of ca. 80 metres, to the north-west of the Nerlerk drill-site, seismogramme no. 496 is from a shallower water (ca. 26 metres) area in the neighbourhood of Ukalerk (see fig. 9.2). The vertical incidence profile in the area of record 496 can be seen in fig. 9.75. It shows very little. A reflector appears at ca. 70msec. below the seabed. This reflector "r" is seen clearly on the 12-channel seismogramme (fig. 9.76) and has been interpreted by the T^2-X^2 method to give a velocity of ca. 1.77km/sec. in the layer between the seabed and that reflector. This is the starting point for a trial model seismogramme. Figures from no. 9.77 onwards, which relate to the modelling process are to be found in the companion volume.

In addition to the reflector "r", the 12-channel record suggests a reflector "y", ca. 12msec. below the seabed, a

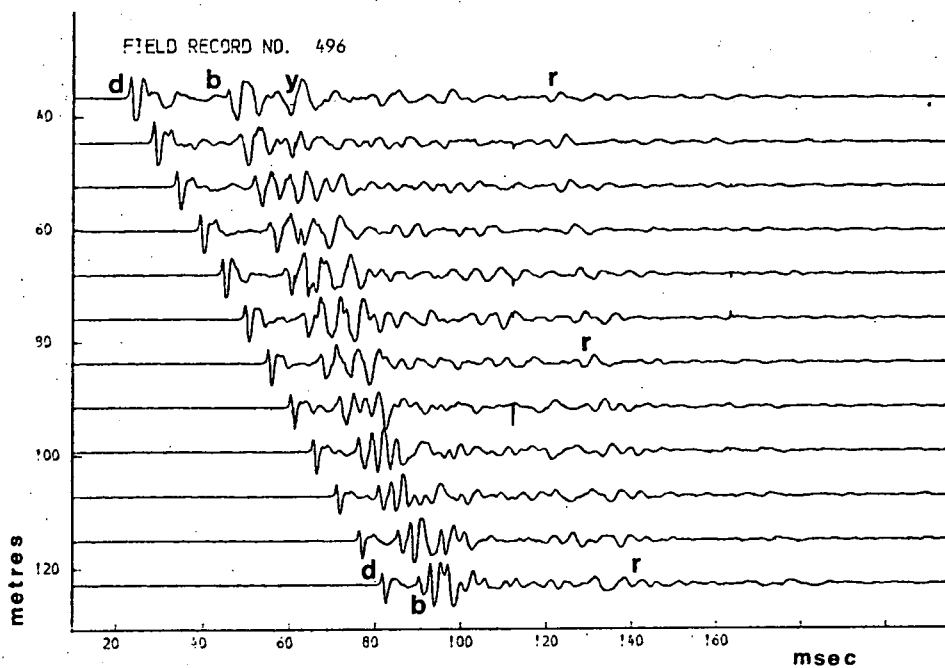


Fig. 9.76

The 12-channel seismogramme at the fix point no. 496 (see fig. 9.2).

d - direct arrival y - subbottom reflexion
b - bottom reflexion r - subbottom reflexion

For figures 9.77a to 9.123b,

see second volume.

reflector which is not well-resolved on the section shown in fig. 9.75. The possible presence of an interface at this level is borne in mind during the carrying out of the modelling process described below.

The first model constructed was based on a simple one-layer structure, viz. a thickness of ca. 70m. of material with a uniform velocity in the region of 1.77km/sec. These models (figs. 9.77a,b) suggest that the average velocity should be reduced to a value nearer 1.70km/sec., but, much more strikingly, give synthetic seismogrammes whose characteristics relating to those arrivals associated with seabed and near-seabed structure are quite unlike those of the field seismogramme (fig. 9.76). A modification of the proposed model was required, rather than mere adjustments of the parameters to optimise the position and amplitude of the reflector "r".

VARIATIONS IN SIMPLE LAYERED STRUCTURE.

Various parameter variations within the layout of a simple 2-layered structure were tried, and the results appear in figs. 9.78a,b, 9.79a,b, 9.80a,b, 9.81a,b and 9.82a,b. Little improvement in the model was found other than the suggestion that a higher p-velocity in layer 2

directed a little more acoustic energy towards the high-offset receivers. It appeared appropriate to investigate models having velocity gradients.

VARIATIONS IN MODELS WITH SIMPLE VELOCITY GRADIENTS.

The effects of many different gradients in p- and s-velocity and density, coupled with variations in individual parameters and in the thickness of an inferred surficial layer are seen in the range of figures from fig. 9.83a to fig. 9.95b. None of these models approaches the appearance of the field seismogramme (fig. 9.76). In constructing the set of models shown in this range of figures, no particular attention has been given to locating the reflector "r" correctly, because the main problem was to model those earlier acoustic arrivals of the field record related to bottom and near-bottom reflexion. It is noted, however, in figs. 9.92a,b, that when gradients in p-velocity are used to try to enhance the shallow events, a compensating decrease in velocity at greater depth may be necessary to preserve the required travel time characteristics of the deeper reflector "r". It will be seen that some emphasis has been given throughout this set of figures (9.83a to 9.95b) to the investigation of the effect of a thin surficial layer, such as was suggested in the interpretation of the

seismogramme no. 1281. This was done, because such a layer may affect the amplitude of deeper events and it was thought that the presence of a surficial layer with suitable properties might improve the model. This did not prove to be the case. As can be seen from figs. 9.84a-d; 9.85a-e; 9.86a-d; 9.87a,b; 9.88a,b; and 9.91a,b, variations in thickness and properties of the thin surficial layer can have rather more influence on the amplitude of the deeper reflector "r" than they do on the character of the earlier arrivals.

Models of the types illustrated in figs. 9.83a to 9.95b appear to offer little advantage over the simpler models (discussed earlier) in explaining the appearance of the field seismogramme. In view of the difficulty in modelling the earlier arrivals of the seismogramme, it seemed appropriate to try such an unlikely feature as a gradient in the bottom water. This is shown in fig. 9.96, and clearly does not help. A very strong gradient in the immediate subbottom (fig. 9.97a,b) produces more interesting results when the thickness of the high velocity zone produced by the introduction of the gradient exceeds 2m. (fig. 9.97b). Particularly in the lower model of fig. 9.97b, it can be seen that reasonable amplitudes are associated with the bottom and near-bottom reflexions at the higher offset distances. The high-velocity zones produced by the models of fig. 9.97a

appear to be too thin to produce this effect. In subsequent modelling, the idea of a relatively high p-velocity gradient in the immediate subbottom has been preserved.

VARIATIONS IN MODELS WITH MORE COMPLEX VELOCITY GRADIENTS.

The results from the introduction of the high p-velocity gradient in the near-seabed sediments suggests that velocity profiles of the general type shown in figs. 9.98a,b are required: the increase in velocity being required to model the characteristics of the near-seabed reflexions; the decrease in velocity at greater depth being required to preserve the arrival time of the reflector "r". The rate of velocity increase shown in figs. 9.98a and b does not appear to be great enough, and in figs. 9.99a-d, 9.100, 9.101a,b, 9.102a,b, 9.103a,b,c, and 9.104, various types of velocity profile with higher near-seabed gradients are shown. A problem with many of these models is that the continuity in the character of the first arrival of the group of near-seabed reflexions, although clear on the field record, is broken around channels 3,4 or 5 of the models. A more realistic representation of this feature is given by the models in fig. 9.99d, although these models do not represent the slightly later arrivals particularly well. An attempt was

made to model the arrival which appears ca. 12msec. behind the bottom reflexion on channel 1 of the field record by constructing models with a stepped velocity increase (figs. 9.105 to 9.108). The difficulty with this type of model is that when the relative amplitudes of the bottom and near-bottom arrivals approach that of the field record (e.g. figs. 9.106a, 9.107a,b), channels at higher offset exhibit unwanted amplitude peaks in the second arrival. Something of a compromise is shown in figs. 9.108a and b, where the presence of high velocity gradients above and below a near-seabed layer of velocity 2.2km/sec. produces a model with some of the characteristics of the field record, but the amplitude of the near bottom reflexion is much too low at small offset, and at high offset.

Another possibility is to introduce a low velocity layer as shown in figs. 9.109 to 9.111. Although this may produce results which are very different from the field record (e.g. fig. 9.109, 9.110b,c), a model like that of 9.110d appears to show more features characteristic of the field record, particularly in the region where bottom and subbottom reflexions begin to combine around channels 8 and 9. The change in thickness of a surficial layer (fig. 9.112) suggests that a thinner layer is better. Increase in gradient from a value of 400m/sec/m. to 800m/sec/m. does not produce any great difference in the

model (see fig. 9.113). Variation in thickness of the upper high-velocity layer (fig. 9.114a-d) suggests that the layer should be set at ca. 10-12m. thick. Reduction of the peak velocity in the lower high-velocity layer as shown in figs. 9.115a-c suggests that the velocity should remain at or above ca. 2.4km/sec. Variation in the thickness of the low-velocity layer (see fig. 9.116a,b) does not significantly affect the arrivals at high offset which remain too low in amplitude. If the upper velocity layer is to remain at ca. 12m. thick however, the underlying low-velocity layer must remain thin so that the arrival associated with the velocity gradient forming the upper boundary of the lower layer is not retarded unacceptably. These considerations suggest a convergence of the models which do and do not contain a low-velocity layer. The models of fig. 9.108a and b are similar to those of fig. 9.117a and b, although in the latter, the presence of the thin low-velocity layer helps to enhance the second arrivals at small offset distance, and a small reduction in the maximum p-velocity in the lower layer from 3.0 to 2.8 km/sec. results in more appropriate amplitudes at high offset distance. The possibility of velocities below that of the seawater in the low velocity layer is considered in fig. 9.117c. This has the effect of splitting the first subbottom reflector (ca. 20msec. behind the bottom reflector in fig. 9.117c) into two events ca. 4msec. apart, a characteristic not found in the

field record. Models which have a less abrupt transition to the minimum in the upper low-velocity layer result in near-bottom arrivals at high offset which are too late in time and too low in amplitude (figs. 9.117d and e).

If the low velocity layer is removed again (fig. 9.118a,b), some characteristics of the model remain, and appear unchanged by variation in s-velocity (fig. 9.119), but the amplitude of the second arrival (ca. 12msec. subbottom) is not represented as well in the absence of the low velocity layer. Subdivision of the stepped increase in velocity suppresses the second arrival altogether and advances too much the arrival times at high offset distance.

ADJUSTMENT OF VELOCITY IN DEEPER LAYERS.

In order to model correctly the arrival time of the reflector "r" (see fig. 9.76), it is necessary to have a low velocity zone immediately above that reflector. It is a matter of some interest to know just how low a velocity appears necessary for this. When the upper part of the sediment column is modelled as shown in figs. 9.121a and b, even as low a velocity as 0.9km/sec. in the deepest layer does not quite restore the travel-times for the reflector "r" to their observed values. An alternative to

suggesting such a low velocity is to make thinner the deeper of the two high-velocity layers. This procedure was adopted and the results are shown in figs. 9.122a and b and 9.123a and b. When the deeper high-velocity layer is reduced in thickness to that shown in figs. 9.123a and b, a p-velocity of 1.25km/sec. produces a good fit with the reflector "r". As can be seen from the upper model of fig. 9.123a, a p-velocity of 1.50km/sec. is too high, while the values suggested in fig. 9.123b are too low. The thinning of the deeper high-velocity layer also appears to have had the effect of reducing the amplitude of the 12msec. subbottom arrival at near-offset, so that there is some reason to retain a greater thickness for this layer and accept that a p-velocity significantly below that of the seawater is required in the zone just above the reflector "r".

9.8 REVIEW OF MODELS FOR SEISMOGRAMMES 1281 AND 1289.

In the light of what had been found by the modelling of seismogramme no. 496, it seemed appropriate to look again at the other seismogrammes which had been modelled to see if similar velocity structures could be accommodated by them. Figs. 9.124a and b show that such a model is possible, and that p-velocities in the material below layer 2 of this model are at least 1km/sec. lower than

For figures 9.124a to 9.131b,

see second volume.

those found by the treatment discussed in section 9.5. Variation of s-velocity (figs. 9.125a and b) suggests that this parameter takes correspondingly lower values, but that it remains at or above 2.0km/sec. Similar observations can be made about the same parameters for the modelling of no. 1281 when p-velocity is allowed to vary in layer 2, as shown in figs. 9.126a and b (where a p-velocity of ca. 3.0km/sec. in the underlying layer is seen to be appropriate), and in figs. 9.127a and b (from which s-velocity in the underlying layer is deduced to be ca. 2.0). Another possible way of producing a reasonable amplitude for the r2 reflector (see fig. 9.10) when the p-velocity in the underlying layer is lower is to insert a thin low-velocity layer just above this reflector (figs. 9.128a and b). The model so produced is less satisfactory than those of figs. 9.124 to 9.127 which incorporate smoother velocity gradients in that a spurious precursor to the reflector r2 is generated (fig. 9.128a and b).

Both of the seismogrammes 1281 and 1289 (particularly the latter) exhibit a characteristic in the reflector r2 which has not yet been considered. This is its "doublet" nature in having what appears to be another arrival following closely (ca. 0.8msec) behind it. Two possibilities for modelling this are shown in fig. 9.129. The one, in the light of what was seen in fig. 9.128, interposes a low velocity layer between layers 2 and 3;

the other uses a velocity step and gradient as illustrated in the lower model of fig. 9.129. Of these, the first gives an inappropriate inverted phase for the earlier part of r_2 and the wrong variation in amplitude with offset. The second model (fig. 9.129, lower) preserves the correct phase relationship between the two parts of the r_2 doublet and gives a reasonable variation in amplitude with offset.

It is necessary to ask whether it is possible to distinguish between the two rather different models which seem to be possible as interpretations of the seismogrammes nos. 1289 (figs. 9.24 and 9.124) and 1281 (figs. 9.45 and 9.126). The simple layered models discussed in sections 9.5 and 9.6 give as close a representation to the overall picture as do the models just discussed. Other information (see section 4.4) from the field area (Neave et al., 1978; and MacAulay and Hunter, 1982) about velocities to be expected in fully ice-bonded sediment layers suggests that a value for p -velocity of ca. 3.0km/sec. is more appropriate than one of 4.6km/sec. and that a shear wave velocity of ca. 2.0km/sec is to be preferred to one of ca. 2.7km/sec. This suggests that the model with a varying velocity structure within layer 2 is the more appropriate.

It would be useful if field seismogrammes could be generated which would distinguish between the possible

models without recourse to other information. A way of doing this is suggested in figs. 9.130a,b,c and d. Although the expected seismogrammes from the two different models are similar when the water layer has that thickness encountered in the field, a reduction in its thickness to 50m. and 25m. suggests that the models could quite readily be distinguished on the basis of data gathered at a towing height of ca. 25m. above the seabed. McKay, 1983 (see section 7.3) towed equipment much closer to the seabed than 25m., so that it should be relatively easy to arrange to deploy equipment at the height above the seabed suggested here.

The models given in figs. 9.131a and b were constructed in order to find out if the presence of velocity variations within layer 2 would affect significantly the value of p-attenuation which is possible in that layer. It appears that there is little difference: the amplitude of r_2 still begins to fall off when $1/Q_p$ exceeds 0.003 (c.f. table 9.2 and figs. 9.56a,b,c).

9.9 OTHER FIELD RECORDS.

Time constraints have not permitted modelling of more field seismogrammes, but a few of those gathered can be examined briefly in terms of what has been deduced from

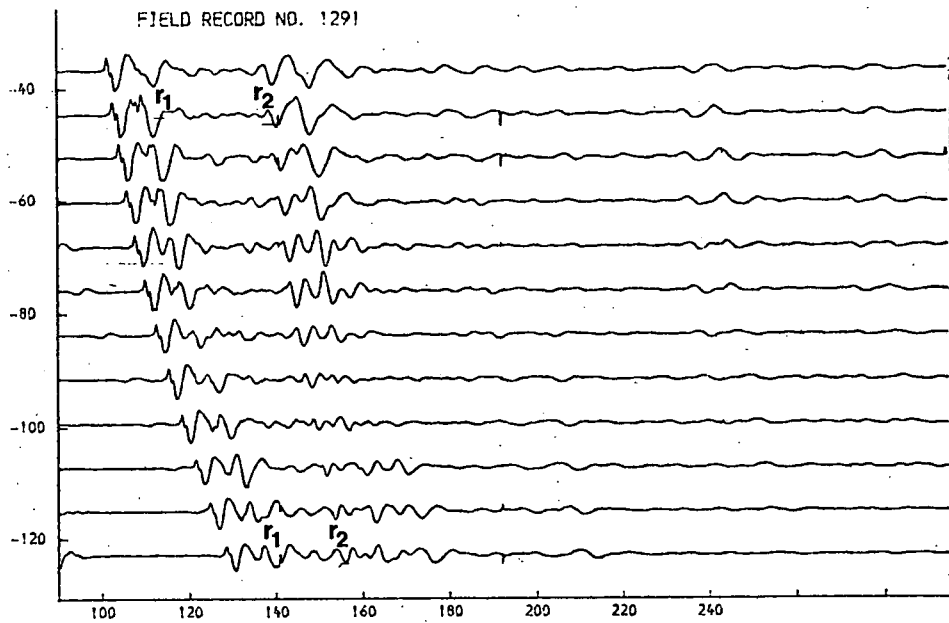


Fig. 9.132.

Field record obtained within 1km. of that shown in fig. 9.16.

those which have been examined more thoroughly. A further field record from the neighbourhood of the records nos. 1281 and 1289 appears in fig. 9.132. It shows similar characteristics to no. 1289 (fig. 9.16) in the appearance of the bottom reflexion and the subbottom reflectors r1 and r2. The nature of the reflector r1 matches that found in the models shown in figs. 9.124a and b in that its amplitude passes through a minimum at intermediate offset distance, an effect which appears to be associated with the presence of a velocity gradient at the top of layer 2. Such a minimum is less clear in the field record no. 1289 of fig. 9.16. This may suggest some lateral variation in a velocity gradient near the top of layer 2.

Field seismogrammes from the same neighbourhood as no. 496 (fig. 9.76) are shown in fig. 9.133. Amplitude information is not completely preserved in these records because of saturation, which is apparent particularly in the bottom events of record no. 513. (Along with irregular bottom and subbottom structure, saturation is one of the commoner reasons for rejecting much of the extensive (fig. 9.2) data set for the purposes of an initial attempt at detailed modelling.) The records of fig. 9.133 do appear to preserve enough information to suggest that the immediate subbottom should have much in common with what was found for record no. 496. in

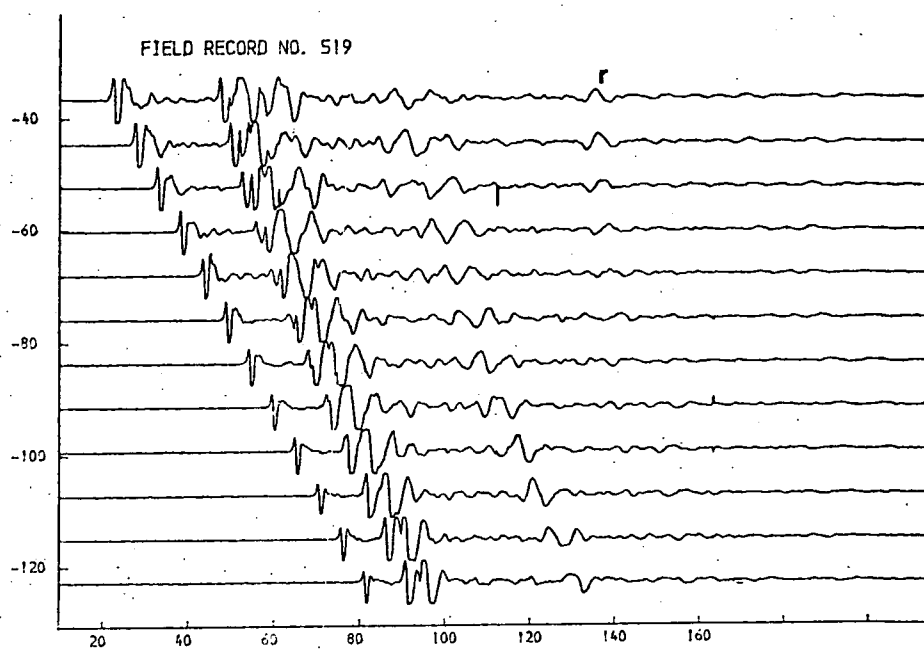
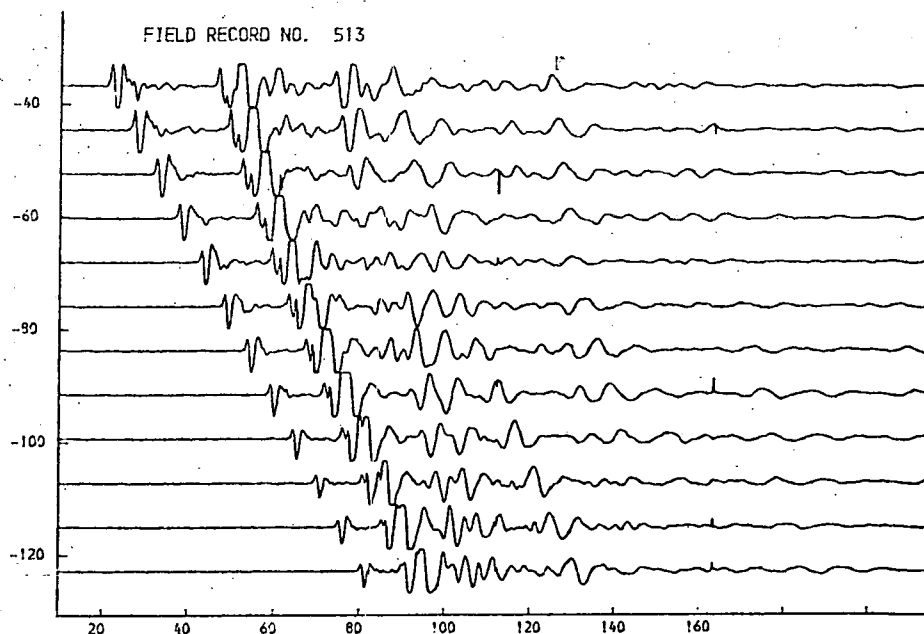


Fig. 9.133.

Field seismogrammes nos. 513 and 519 from the Ukalerk region of line 3 (see fig. 9.2), a few km. east of the record no. 496.

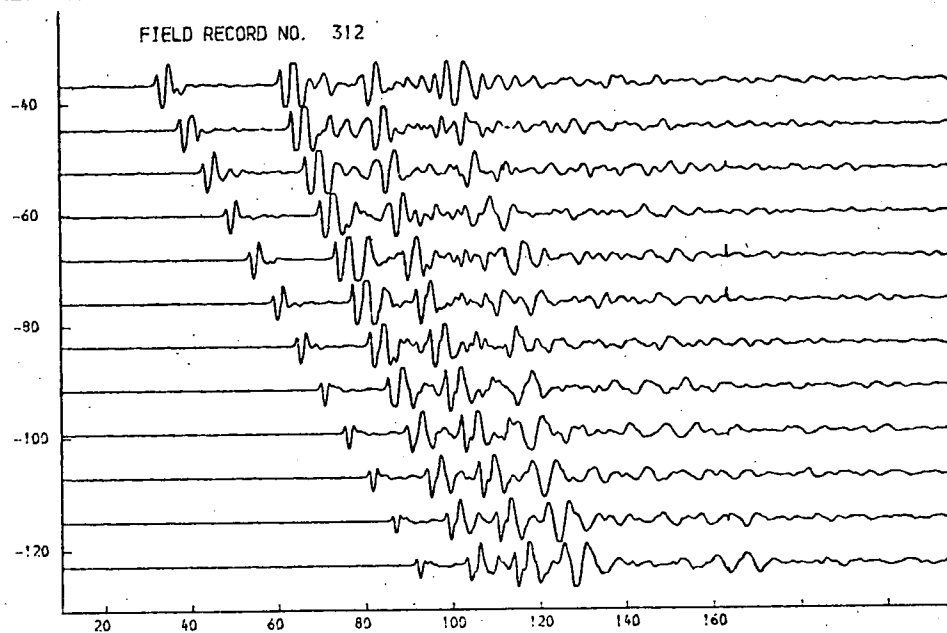
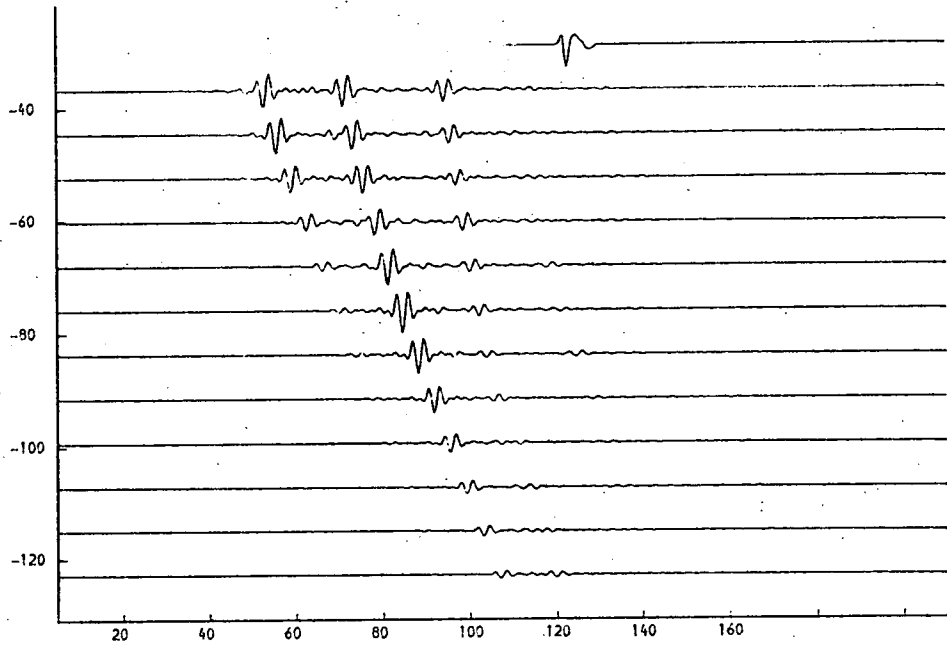
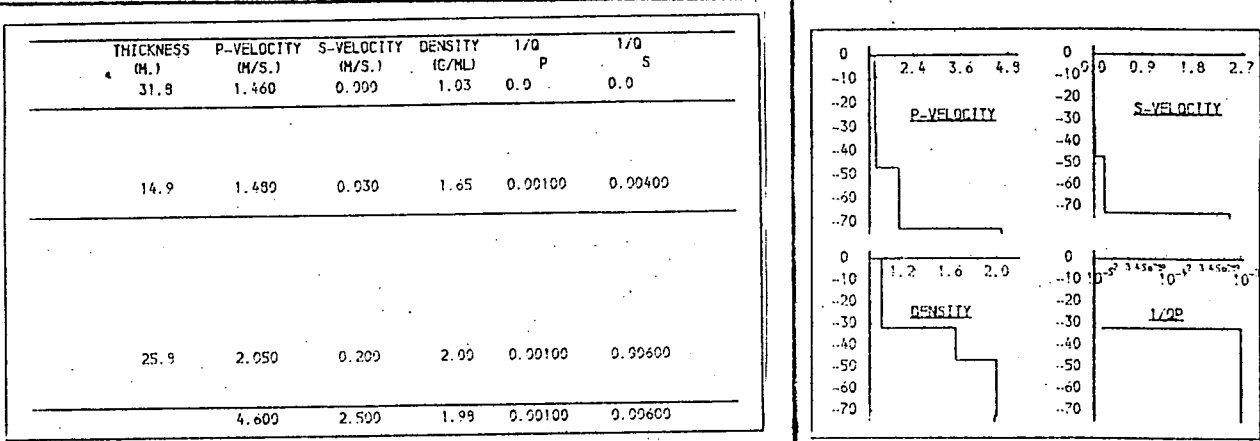


Fig. 9.134.

Field seismogramme no. 312 from an area between Amauligak and Kogyak (see fig. 9.2), together with a simple model.

particular, the appearance of high amplitudes in near-bottom events at high offset in a similar water depth, suggests that a simple layered model of the type shown in fig. 9.77a could be expected to be inappropriate for the records of fig. 9.133. These field records define a deeper reflector "r'" similar to that found in no. 496, but have, above this reflector, a complex series of arrivals whose explanation would require further modelling. One of these could be associated with a more abrupt transition to a deep low-velocity layer such as can be seen in fig. 9.121b in the event which appears in that figure at ca. 100msec. of travel time. The field record no. 312 from line 2, between the Amauligak and Kogyak areas (see fig. 9.2) is shown in fig. 9.134, together with a model based on simple layers. Again it seems that velocity gradients might be required to produce an appropriate model.

9.10 DISCUSSION

While it would appear that the most important point to emerge from the seismogramme modelling work is a p-velocity structure which is not suggested by more simple-minded approaches to data analysis, this should not obscure the fact that the modelling process can also be seen to provide a method of estimating shear wave velocity

and p- and s-attenuation. Previous attempts at measuring p-attenuation have relied on the "sediment wedge" technique (Li and Taylor Smith, 1969; Dodds, 1980) which requires cooperative geology, whereas the synthetic seismogramme approach can yield a value for Q_p when layer thickness is uniform. The values of attenuation which are found for the layer 2 (on the basis of a simple layered model) for the seismogrammes no. 1281 and 1289 can be compared with values summarised by Hamilton (1976, 1980). The suggested value (see table 9.1) for $1/Q_s$ of 0.03 falls within the range of 0.03 to 0.2 given by Hamilton as typical for water-saturated materials "of a type which might be found as marine sediments" (Hamilton had difficulty in finding extensive information on shear-wave attenuation in real marine sediments). The associated value for $1/Q_p$ of 0.001 is unusually low for a sand (Hamilton, 1976 gives a range of 0.023 to 0.04 for sands and 0.002 to 0.009 for silts and clays), and approaches the low value found by Frisk et al. (1981) in clays in the deep ocean (see section 9.3).

The low value might be taken as diagnostic of ice-bonding, although a rare experimental determination of p-attenuation in a water-saturated sand which had been frozen in the laboratory, Antsyferov et al. (1964), gave a value of $1/Q_p=0.014$. This laboratory determination (see section 4.2, fig. 4.13) was made at higher frequencies

than our field observations, but this frequency difference (little more than one order of magnitude) cannot be held to account for the difference in $1/Q_p$ if the lossy elastic model (see section 8.1) on whose acceptance for seabed sediments the field observation depends, is to be upheld. Other possible reasons for the difference are:

(i) Disparate grain-size distribution. Variations in the fraction of clay-size particles from 0 to 10% of a predominantly sandy sediment were shown by King (1967) to have a strong influence on p-attenuation.

(ii) The formation and growth of any type of crystal is known to depend on the length of time during which it is formed. Although the time scale on which an ice crystal will grow in the material of layer 2 in the locality of seismogrammes 1281 and 1289 is unknown, the discussion in section 4.4 on freeze-thaw cycles suggests that this time would be very much greater than that which would elapse during the freezing of a laboratory sample.

(iii) Incomplete ice bonding. The velocities suggested by the seismogramme modelling procedure, even in the presence of possible velocity gradients in the layer 2, are below the values expected for fully ice-bonded materials (see section 4.2, fig. 4.12), suggesting that ice bonding in this layer is only partial.

Shear wave velocity is a parameter which has been identified by Strachan (1983) as correlating with the

resistance to liquefaction of sands, a phenomenon which is of importance particularly in the design of the foundations of structures subject to cyclic loading such as can be induced by wave motion (Nataraja and Gill, 1983). The shear wave velocities determined in layers 2 and 3 of the seismogrammes 1281 and 1289 are higher than those measured by Strachan in water-saturated sands but lower than the field measurement of Hunter (1974), (see fig. 4.17) in land permafrost. The values tend to confirm the suggestion of partial ice-bonding which comes from the p-velocity measurements. In other sedimentary environments, it may be that s-velocity could be measured sufficiently accurately to be of use in such correlations as that of Strachan (1983), but the low precision achieved in the results of tables 9.1 and 9.2 suggest that the method may be of restricted worth for the determination of s-velocity. It should be noted, however, that no other method of measuring s-velocity from a vessel underway has been used successfully, and the values given here are probably the first found in-situ without station work.

THE SIGNIFICANCE OF RESULTS TO THE UNDERSTANDING OF
BEAUFORT SEA SEDIMENTS.

The p-velocity profile which is required for the modelling of the field record no. 496 contains a velocity

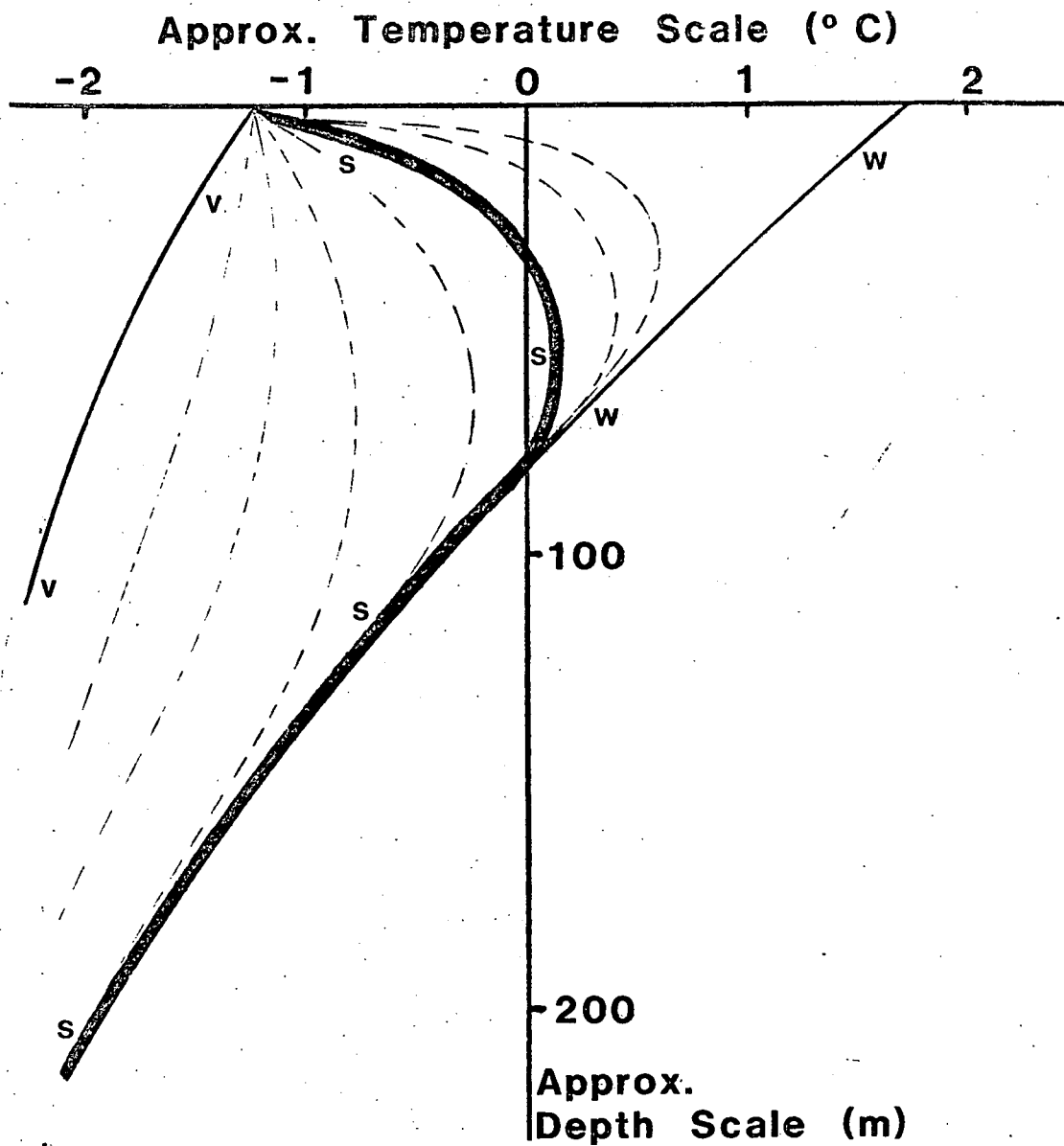


Fig. 9.135.

A possible temperature profile (heavy black line) which is suggested by the results of modelling the seismogramme no. 496.

variation with depth which has not previously been suggested for near-bottom sediments in the Beaufort Sea. The profile (see section 9.7) suggests significant, if partial, ice-bonding in the first 20m. below the seabed. The thin low velocity layer (if present) nearer the seabed could result from a thin lens of clay or mud within an otherwise uniform coarser sediment. At greater depth, the p-velocity decreases to the extent that ice-bonding should be absent. The arguments given in section 9.7 in the discussion associated with figs. 9.121 to 9.123 strongly suggest that gasification occurs in a zone above the reflector "r" of no. 496 (fig. 9.76). While it is nearly always possible to postulate the occurrence of some form of gas-producing mechanism within a sediment (see section 3.2, table 3.1), there may be another way in which gas could appear in this instance. If the reflector "r" represents the surface of permafrost which formed originally under subaerial conditions, then the frozen material could contain air, which would have little effect on its acoustic properties while material remained frozen, (see section 4.2, fig. 4.12), but would tend to depress the p-velocity (see section 3.5) when it thawed. (Gases other than air might also be frozen in.)

A possible explanation for such a profile is given in fig. 9.135 where the velocity variations are interpreted in terms of temperature. (This is done with the

reservations expressed in section 4.0 on the subject of equating sub-zero temperature with ice-bonding.) There is general acceptance of a thermal history of the Beaufort Sea shelf area which would suggest that the time-varying geothermal profile is similar to that proposed by Molochuskin (1973) for the Laptev Sea (see fig. 4.25, section 4.4), but some variation from this general scheme is possible.

The variation of velocity from high to low and back to high again (as is suggested by the interpretation of the seismogrammes presented) within what is probably a relatively uniform sedimentary unit suggests a temperature profile of the form shown by the line labelled "s" in fig. 9.135. Without the supposition of unlikely sources of heat within the sediment column, such a temperature profile cannot be in equilibrium, but may come about when a change in the bottom-water temperature alters the thermal gradient near the seabed. A change from the profile "v" to "w" would be expected following a drop in temperature of the water. The transition from one gradient to the other would have a time constant smaller than that associated with the temperature change resulting from marine incursion (fig. 4.25), but it could still take several hundred years for a transition of the sort shown in fig. 9.135 (see eqn. 4.3, section 4.4 and Hunter et al., 1976). A profile such as that of "s" in fig. 9.135

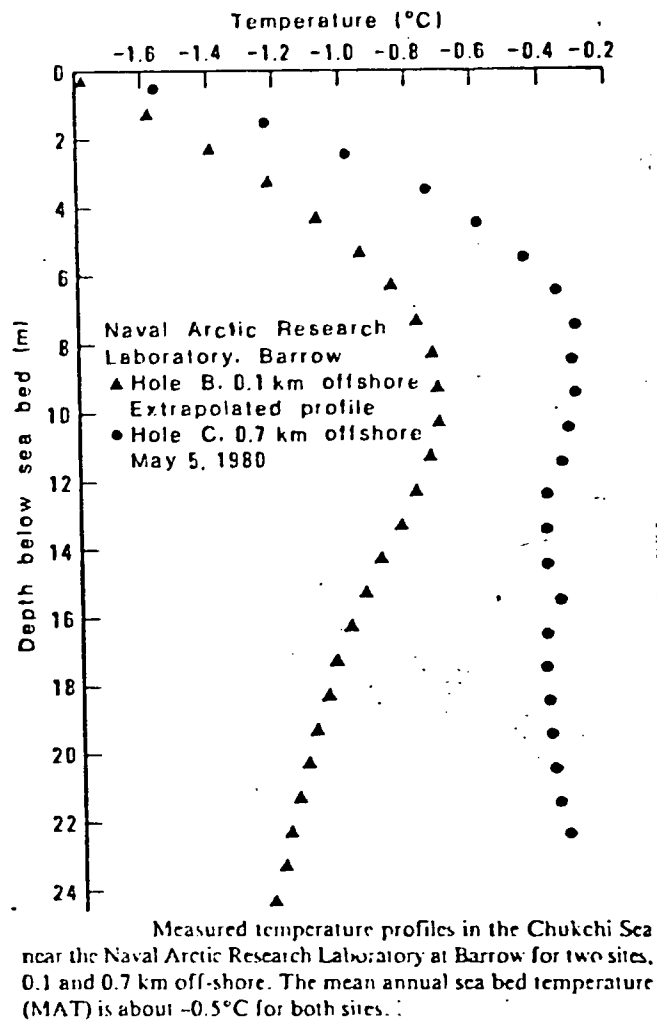


Fig. 9.136

Temperature profiles given by Osterkamp and Harrison (1982) for an area near Barrow, Alaska.

allows an unfrozen zone between two frozen zones. The temperature variations need not be exactly as shown in the figure, and the argument holds for smaller fluctuations in seabottom temperature than for the curves "w" and "v" as illustrated.

Although independent observations are not available for the specific locations discussed, Osterkamp and Harrison (1982) have suggested temperature profiles similar in character to that shown in fig. 9.135 (see fig. 9.136). Their work is based on borehole temperature measurements in the Prudhoe Bay area of the Beaufort Sea and in the Chukchi Sea, and is confined to locations within a few km. of the shoreline, but extends to water depths of at least 10m. (In close-inshore areas, sea-ice freezes to the seabed, but this does not normally occur in water over 2m. deep.) Harrison and Osterkamp (1982) note that seasonal variations in temperature are confined to the topmost 2m. of sediment, so that temperature variation at deeper levels is a result of longer-term trends. On the basis of resistivity measurements, Osterkamp and Harrison suggest that although permafrost (based on definition by temperature) occurs throughout their drillhole (fig. 9.136), ice-bonding is probably absent from most of it, presumably because of saline pore-water. The possibility of a velocity profile of the type suggested by the seismogramme modelling and an associated thermal

profile as described above provides an alternative explanation for the formation of underwater pingos (figs. 4.26, 4.27). Shearer (see fig. 4.26) associates these with lakes which have been submerged and turned into pingos in the same way as on land by the incursion of permafrost (see section 4.1) into a zone of water-saturated sediment beneath the old lake basin. If, alternatively, a thermal gradient in the seabed sediments were at one time such as to allow degradation of the frozen zone from the top (this requires a higher seabottom temperature than at present) a reservoir of fresh porewater could evolve, possibly protected from massive incursion by salty porewater by a thin cover of holocene clays. Subsequent lowering of the temperature of the bottom water would allow re-freezing to proceed from the seabed downwards, while (due to thermal inertia) thawing continued at greater depth to degrade the original permafrost. It is suggested that this is the situation in the areas where the seismogrammes have been analysed. Eventually degradation of the original permafrost from the top will stop and the unfrozen zone will be subjected to a squeezing action such as is likely to produce pingos.

9.11 SUMMARY.

The seismogramme technique described in this chapter provides a way of finding out more about sediment properties than is possible by the methods given in chapter 7. The modelling process suggests a p-velocity structure in Beaufort Sea sediments which is compatible with earlier interpretations (see section 4.4) but which introduces variations within the top 50m. of a character not previously recognised in offshore areas.

10. CONCLUSIONS AND DISCUSSION.

The acquisition of a greatly improved understanding of the properties and movements of seabed sediments remains one of the major scientific and practical problems of today (K.C. Dunham, personal communication, 1983). Specific problems and applications have been discussed in chapters 1,3,4 and 5. Other parts of the foregoing work have tried to identify ways in which methods of conducting acoustic surveys from vessels underway can be improved. This work was pursued in two ways: firstly by making minor modifications and additions to existing profiling equipment to allow the making of acoustic velocity measurements and secondly by constructing and operating new equipment to allow the gathering of multichannel data suitable for analytical treatment by methods established in the allied field of seismology. The results from the former method (see chapter 7) are as important as from the latter approach in that they are immediately applicable to the improvement of current survey practice (which has been described as "desultory" by Cratchley et al., 1982). The latter approach (see chapters 8 and 9) requires the considerable re-orientation of accepted field operations and data-reduction methods before its use could become widespread, but it is clearly necessary (see section 9.7) if erroneous conclusions about velocity structure deduced from simpler methods are to be avoided. Much recent research work on sediments has been based on deep-tow vertical-incidence reflexion profiling (e.g. Hill, 1982;

Spiess and Anderson, 1982; Piper and Normark, 1983). The results presented in this work reinforce the opinion of McKay and McKay (1982) that such a technique is a superficial way of analysing the acoustic response of seabed materials. The results also show (section 9.8) that a deep-tow deployment would be essential for the realisation of a geometrical disposition of source and receivers for the gathering of multichannel data optimal for analysis by the synthetic seismogramme method to reveal the velocity structure of the subbottom of the specific geographical areas where the observations discussed were made.

ECONOMIC INFLUENCES

Economic forces in this, as in most fields, strongly influence the application of potentially useful techniques. The cost of gathering conventional deep-tow vertical incidence profiles is a relatively small part of a survey operation mounted on a sea-going vessel of 30-50m. length with custom-designed navigation shore-stations and other geophysical and ancillary equipment such as sidescan sonar, echo sounders and a multitrace digital reflexion system for defining deeper structure. The adaptation of a 12-channel receiver of the type described in section 2.5 to a deep-tow deployment by

such means as suggested by McKay (1983) and the provision of appropriate recording facilities should add appreciably less than 10% to the cost of such a survey operation. A requirement to add the same capability to a small "day-boat" operation (which would not include deep-penetration seismic equipment, and could rely on inshore navigation systems) would have a much more significant influence on the cost. The data-processing costs associated with the modelling technique could be prohibitive for some purposes. Section 6 of appendix 4 gives some idea of the time of computation required for each seismogramme model. In terms of the accounting systems of the Northumbrian Universities Computing System, the total cost of the seismogramme models included with this work is ca. £10,000, a figure which would be doubled by considering those models which were constructed but have been discarded. While such a sum would be easily justifiable for a complete interpretation of the set of seismogrammes shown in the position plot of fig. 9.2, it would be hard to justify as reasonable for results from just three locations. In addition to the cost of computation, the time required by the person doing the modelling is considerable and would, in the commercial environment, double the cost stated above. These considerations suggest that the modelling technique, however powerful, may not be taken up readily by the marine geotechnical engineer until data processing methods

can be streamlined. The general decline in computing costs and increase in processing speed which is continuing may soon take care of part of this problem. The question of interpreter costs might be overcome in two ways: firstly by the development of quicker ways of entering and changing models, such as by the use of a "light-pen" to alter the graphical display of a parameter profile; secondly, given significantly greater computational power, by the derivation of suitable objective functions to allow the optimisation of a seismogramme model without significant human intervention.

DISCUSSION.

Until the synthetic seismogramme modelling technique can be put on an economic footing which is reasonable for most applications of sediment investigation, the methods given in chapter 7 can provide a convenient, if rudimentary aid to the characterising of seabed materials. The principal refinement required is in the field equipment available for the gathering of data (sections 2.3, 2.4, Appendix 1). Proper equipment would allow better definition of the velocity structure in such areas as those discussed in sections 7.4 and 7.5 and if provision was made to record the waveforms in a manner suitable for the recovery of field seismogrammes as shown

in chapter 9, would allow the archiving of data suitable for treatment by the more advanced modelling techniques when time and money permitted.

The results given in chapter 9 show that the modelling technique can yield, from suitable field records, parameters of interest in predicting geotechnical properties (chapter 1 and D. Taylor Smith, 1983), although it may be necessary to gain more experience in specifying the equipment deployment which will place acoustic source and receivers in the best position for minimising the error on such parameters (see e.g. section 9.10 on s-velocity value). The Beaufort Sea area may be regarded as offering rather extreme examples of velocity variations in sediments. While this may be so, it must be remembered that areas of the seafloor subject to the phenomenon of inundated permafrost are extensive (section 4.0), and that, as detailed sediment investigation is carried further into deeper water, off the Continental Shelf, free gas and gas hydrates (chapter 3) may produce equally large deviations from what might be expected. The U.S. Navy are developing equipment for gathering data in such an environment by means of an acoustic source and multichannel receiver towed near the seabed (Fagot, 1983). This is being done to help define sea-bottom types for modelling acoustic propagation in the water column (chapter 5) and is more free of economic

constraints than most projects. The interpretative methods discussed in this thesis will be appropriate for dealing with its field results when those start to appear.

The most striking and unexpected result to emerge from the seismogramme modelling work (chapter 9) is the recovery at the site no. 496 (section 9.7) in the Beaufort Sea (fig. 9.2) of a p-velocity profile which is quite different from the profile which emerges by treating the same data by travel-time methods (chapters 6 and 7). Although these simpler data reduction techniques do give indications of whether partial ice-bonding is present in the upper sediment layers (see section 9.1), a reconstructed seismogramme based on the parameters which they suggest can in some instances bear little resemblance to the field seismogramme from which the parameters were derived. This observation suggests that, although travel-time methods can be seen, by the results given in chapter 7 relating to areas of the Scotian Shelf, Great Lakes, Beaufort Sea and various inshore regions of Maritime Canada to offer significant improvement in the understanding of seabed materials over what can be gleaned from the vertical incidence profiler, still generally accepted as an appropriate research tool, it is not sufficient to rely on travel-time methods alone, even for the derivation of p-velocities. Such a finding has already become widely accepted in investigations into deep

rock structure, including structure in sedimentary rocks. The work presented here suggests that this acceptance must be extended to surficial sediments. Just as earlier seismological applications of seismogramme modelling have led to a new understanding of lithospheric p-velocity profiles, it appears that, unless there is an unusual peculiarity about the area of the Beaufort Sea where previously unsuspected zones of ice-bonding and gasification are revealed (section 9.7) there may be wide scope for the application of this method towards increasing general knowledge about sediment properties and processes.

Many details concerning the application of these modelling techniques to surficial sediments remain to be considered further (theoretical considerations have been mentioned in chapter 8, while some practical considerations in data gathering were raised in section 9.8), but the worth of pursuing investigations in this general manner appears to have been established by the work presented here. In relation to furthering the understanding of the surficial geology of the Beaufort Sea, the modelling work has uncovered in an offshore area, a velocity structure which suggests a subbottom temperature profile similar in character to what has been logged directly in holes drilled much closer to the shoreline. This in turn allows an alternative explanation

for the formation of the seabed mounds which are well-known topographic features in that region (see section 9.10).

SUMMARY

In the acoustic investigation of surficial seabed materials, an approach which makes use of a more detailed observation of the reflected sound field offers advantages over the currently accepted vertical incidence reflexion methods. Methods of varying complexity have been given for deducing acoustic properties from such observations. Application of the simpler methods (chapter 7) has contributed to the understanding of seabed materials in various parts of Canada while the more complex (chapter 9) suggest, in Beaufort Sea sediments, a distribution of partial ice-bonding not previously reported and not recoverable from the application of the simpler methods to the same field data. Since the required observational techniques do not increase survey time beyond what is needed by the conventional approach, it may be hoped that the considerations outlined in this thesis will significantly improve current practice in seabed investigations whether they are carried out to satisfy the aims of the sedimentologist, the civil engineer or the underwater acoustician.

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APPENDIX 1.

Further Equipment Adaptation.

This appendix gives details of the construction of the equipment used in making the observations of wide angle reflexions discussed in section 2.4.

A previous attempt at making wide angle reflexion profiles with an array of hydrophones is that of W.J. Porter et al. (1974) of the Raytheon Corporation. They constructed a 225 ft. (68.6m) long eleven channel array in a 1.25 inch diameter oil-filled sleeve. The hydrophones in this array were arranged so that their positions made for an even distribution of angular measurements based on the average geometries of water depth and subbottom layer thickness which they expected to encounter. They led all eleven channels back to be recorded on board ship. While not confined to operating in the very surface of the water, this array was not intended to operate at depths below about 10 metres.

In contrast with this, the hydrophone array described here was intended for use with the deep tow seismic profiling system which has been developed during the 1970's at the Nova Scotia Research Foundation (Bidgood, 1973). As such it had to be capable of submergence to depths of several hundred metres, preferably to 350m. The number of channels which could be brought back to the surface was limited. In the main 7-conductor signal cable connecting the underwater vehicle to the ship, only 3 conductors were available for a multi-hydrophone array. Accordingly, some kind of multiplexing arrangement was required. Various possibilities were considered, such as frequency

modulation for the simultaneous transmission of several receiver outputs over the one conductor, each signal being contained in a carrier bandwidth of 10KHz or thereabouts, or else the storage, delay and re-transmission in sequence of as many signals as could fit into the period between successive firings of the acoustic source. Both of these ideas involved an appreciable amount of electronic design and packaging into pressure cases which could fit into the underwater vehicle pod. This seemed likely to present problems, of which the worst might well be the interference from the high voltage cable at each firing of the sparker acting as the acoustic source. Even with the appreciable shielding of this pulsed high voltage line and the low impedance terminations of the signal wires large transients appear on these conductors. It is improbable that logic or memory chips could readily be made to behave themselves in such an environment.

A reversion to the technology of an earlier era in the form of the electro-mechanical relay seemed like a good way around some of these potential problems. If relays were to be used, no attempt would be made to take to the ship more than one channel of information from the hydrophone array, but that one channel would in turn be used to convey the signal from different hydrophones in sequence, so that different angles of reflexion from the seabed could be observed on successive firings of the

acoustic source. If the firing rate were matched properly to the speed of the boat, then the reflexions could be observed at the different angles from the same part of the seafloor if this was considered desirable in some circumstance, such as when a rapid change in topography or sedimentation pattern occurred.

Another useful property of relays is that many of them have been rendered obsolete and as such are, at the time of writing available at little or no cost, (a matter of considerable importance when no funding is available for a project). The relay used for switching from one hydrophone to another was one taken from an ex-military wireless receiver (R-962/ARR-52), where it had operated a stepped tuning mechanism. With it, 17 inputs could be gated sequentially to one output, and the first inclination was to make a hydrophone array with some 17 elements, each with its pre-amplifier. These elements would be spliced into a 0.7cm. diameter seismic signal cable having 26 twisted wire pairs. This idea came up against a number of difficulties. There was the problem of acquiring 17 hydrophone elements and the tedium of constructing 17 preamplifiers, but, perhaps more important was the problem of feeding this 0.7cm. cable with its many conductors through the bulkhead of a pressure case. All such connexions in the basic N.S.R.F. Deep Tow System are made with underwater connectors, the use of packing glands

for feeding through into pressure cases having long since been abandoned as a practice fraught with dangers of flooding. No suitable connector for such a large number of conductors seemed to be readily available. It appeared that it might be better to use a smaller number of hydrophones and to rely on well-tried methods of entering pressure cases by way of available connectors. This placed an upper limit of 6 on the number of hydrophones which could be accommodated. In the event, an array of 5 hydrophones was constructed, the hydrophones being spaced at 40ft (12.2m) intervals. (Some provision was made for the addition of a sixth, but this has not been done). Such an array would provide reflexions at a variety of angles, depending on the height at which the deep tow vehicle was flying off the bottom. However, as a way of providing a quasi continuous profile, a means was contrived of generating a greater number of angles. This was done by arranging to have 8 sparkers as acoustic sources instead of just one. The eight would be spaced along a 35ft (10.7m) array at 5ft (1.52m) intervals. By a suitable sequence of switching between the sparkers and hydrophones, a total of $8 \times 5 = 40$ different reflexion angles could be observed. Apart from yielding far greater redundancy in the effective number of equations available for solution to give velocities in the subbottom layers, this quasi continuous profile would minimise the risk of the incorrect identification of an arrival, which is a

IN

OUT

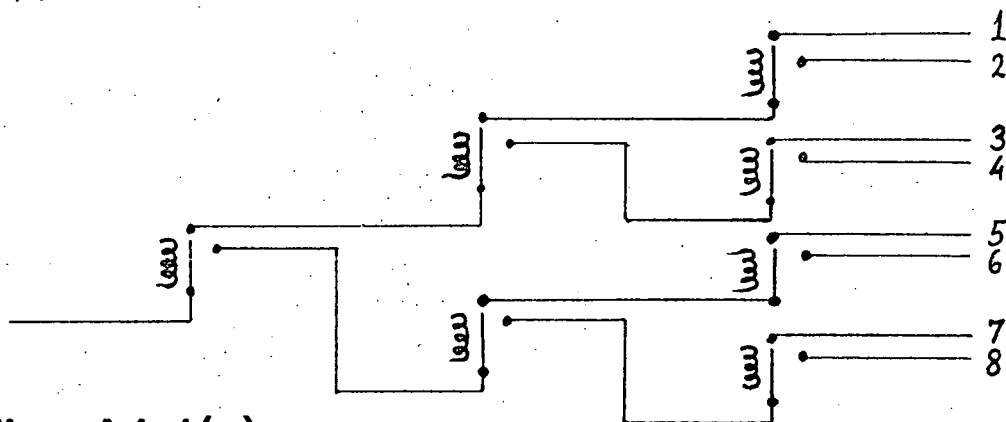


Fig. A1.1(a)

Switching relays in series. Seven are required to gate a single input to any of eight outputs. Only the paths for the high voltage are shown.

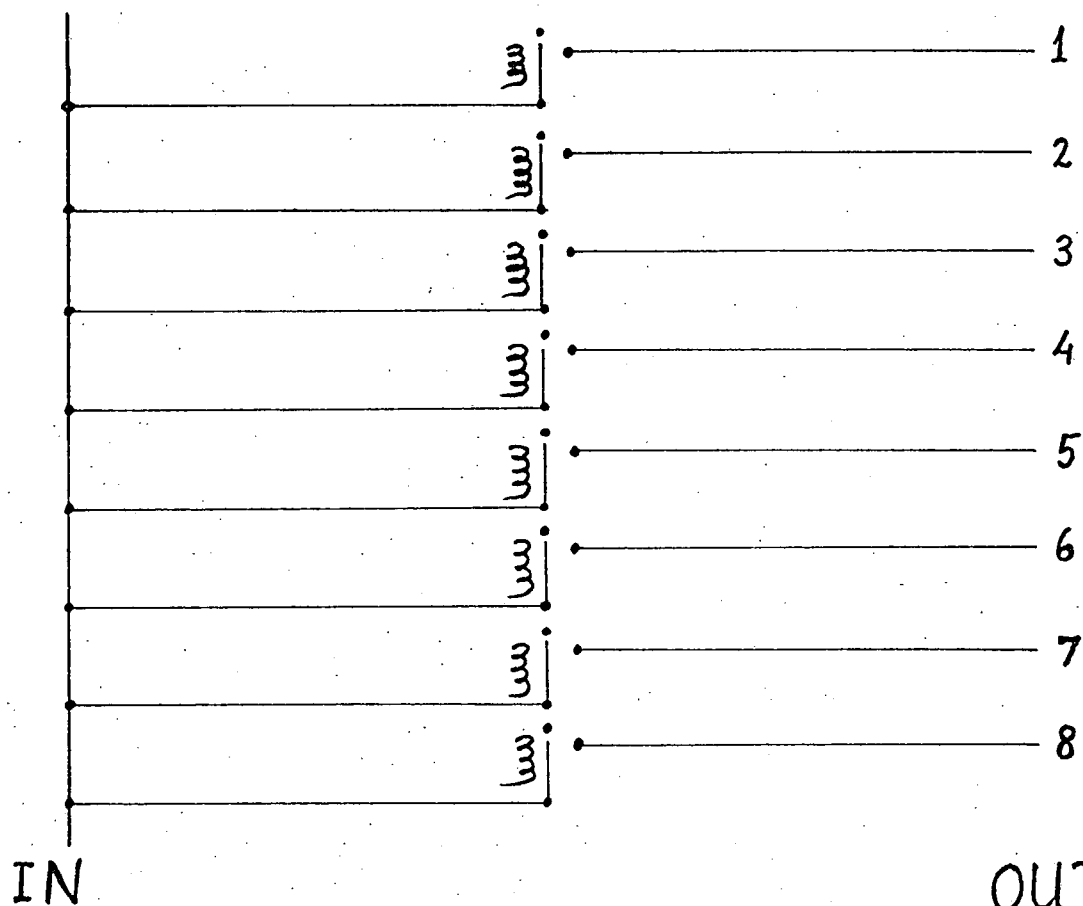


Fig. A1.1(b)

Switching relays in parallel. Eight are required to gate a single input to any of eight outputs. Again no low voltage control circuitry is shown

danger with discontinuous profiles, and might also allow a useful plot of the variation of reflexion amplitude with reflexion angle to be constructed. The construction of 8 suitably matched sparkers posed no particular problem, but the construction of a remotely controlled switch to divert the high voltage to each sparker in sequence was something more of a departure from any prior experience.

DESIGN OF HIGH VOLTAGE SWITCH MECHANISM.

As a first step towards the construction of a high voltage switch mechanism, the behaviour was investigated of a surplus 10amp three-pole two-way uncased relay. This relay was connected to the customary Teledyne high voltage source, and used to gate its output to one or other of two typical sparker heads. When the relay was left in air, no gating was accomplished because a spark could easily jump the 1.3mm. gap between the contacts on its way to the "switched off" sparker head, but when the relay was immersed in transformer oil, it worked perfectly well as a gate. Encouraged by this observation, a multiple gate was designed. There appeared to be two basic ways of doing this. A number of relays could be stacked in series as shown in fig. A1.1(a), or else connected in parallel (fig. A1.1(b)). It will be seen that the series method uses one relay fewer, but this advantage was thought to be

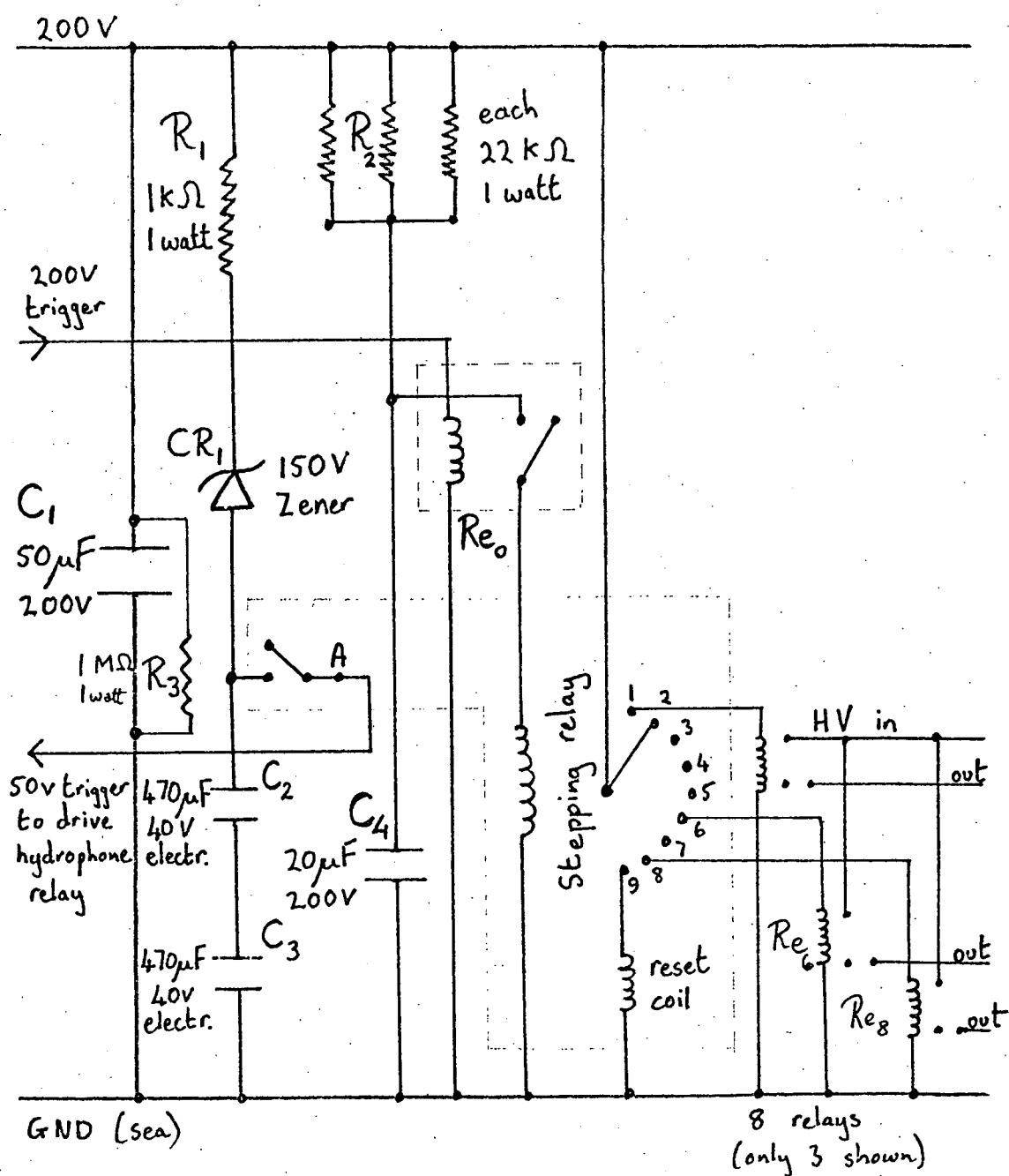


Fig. A1.2

The submersible high voltage switch mechanism. The switch at A is actuated momentarily by the stepping relay when it resets to position 1.

more than offset by the more complicated arrangements which would have to be made for gating the relays in the correct sequence. With the parallel configuration, all that would be required would be to turn each relay on and off in sequence, something which could be done with a stepping relay like that used for the hydrophone switching. In practice, a slightly different stepping relay was used, one which could reset to its first position after any predetermined number of steps, up to 18. The eventual circuit for the high voltage switch mechanism is shown in fig. A1.2. The basic low voltage supply to the circuit is 200v D.C. Surges on this supply are smoothed by the capacitor C1. This capacitor stores a potentially hazardous charge, so that to minimize risks to operators handling or dismantling the switch, the bleed resistor, R3, is required. It will discharge the capacitor in about a minute, appreciably less time than would normally be required for shutting down high voltage supplies, retrieving equipment from the sea and disconnecting. The 200v D.C. supply is used to actuate the relays Rel to Re8 when it is gated to them in turn by the stepping relay. The stepping relay is itself activated by the discharge through it of the capacitor C4 by way of relay Re0, which is driven by an input 200v trigger pulse. The reason for this method of driving the stepping relay is that it is essentially a low impedance device (ca. 100ohm) whereas the relays Re0 to Re8 are of

much higher (10 kilohm) impedance. The 200v trigger must be expected to have to be fed down a cable of up to 30ohm resistance before reaching the switch. The capacitor discharge method of triggering the stepping relay leads to a good quick stepping action, while preventing overheating by minimising the time for which current flows. The same technique is applied to the generation of the output trigger used to drive the hydrophone stepping switch, housed separately from, but in close proximity to the high voltage switch. This is done by the discharge of capacitors C2 and C3 through the switch A, which is actuated by a cam on the resetting of the stepping relay to position 1. It was necessary to supply a lower voltage to the hydrophone relay. This was derived by inserting the zener diode CR1 in the charging circuit of C2 and C3.

The period between successive triggerings of the switch has to be sufficient to allow capacitor C4 to charge up through R2. The time constant of the circuit shown is 0.14sec., which is adequate for triggering at intervals of 0.5sec., but inadequate for 0.25sec. triggering. The time constant could be reduced by lowering R2, but it has been found that 0.5 or 0.75sec. triggering is usually what is wanted, a sparker being fired either 2 or 3 times at 0.25sec. intervals before switching to the next.

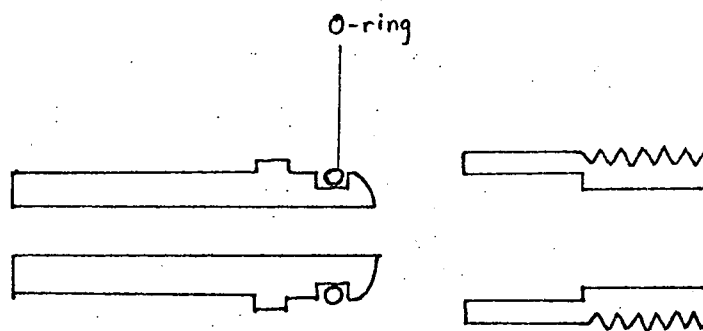


Fig. A1.3(a)

The fittings used as a basis for the construction of the high voltage connector.
(approximately life size)

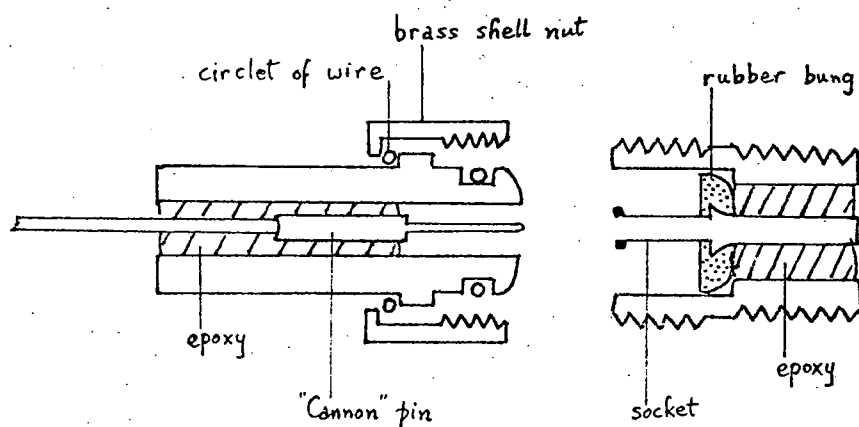


Fig. A1.3(b)

The completed connector pair.

Although the overall size of the high voltage switch described above is only 29cm. long and 9.5cm. in diameter, it was too large to fit into any underwater housing which was readily available. Further, there remained the problem of how to lead nine connexions (one input, eight output) for the high voltage cables through the wall of such a housing. The standard high voltage connectors used in N.S.R.F. equipment are large and expensive. It was necessary to make something smaller and cheaper, and to forego some of the robustness of the large connectors. This was achieved by modifying an idea of F. Smith (personal communication) in making use of C.P.V.C. plastic plumbing fittings. One such fitting is readily available with an O-ring seal already on it, and it is an ideal starting point for making an underwater connector. The basic male and female plumbing fittings are illustrated in fig. A1.3(a).

The parts added to these to make the connector pair are:

- (i) An electrical pin and socket pair as used in the conventional high voltage connector.
- (ii) A shell nut taken from a brass union (again from the plumber's).
- (iii) A small rubber bung (see fig. A1.3(b)).

The only machining required is the cutting of a thread in one of the plastic fittings to mate with the brass shell

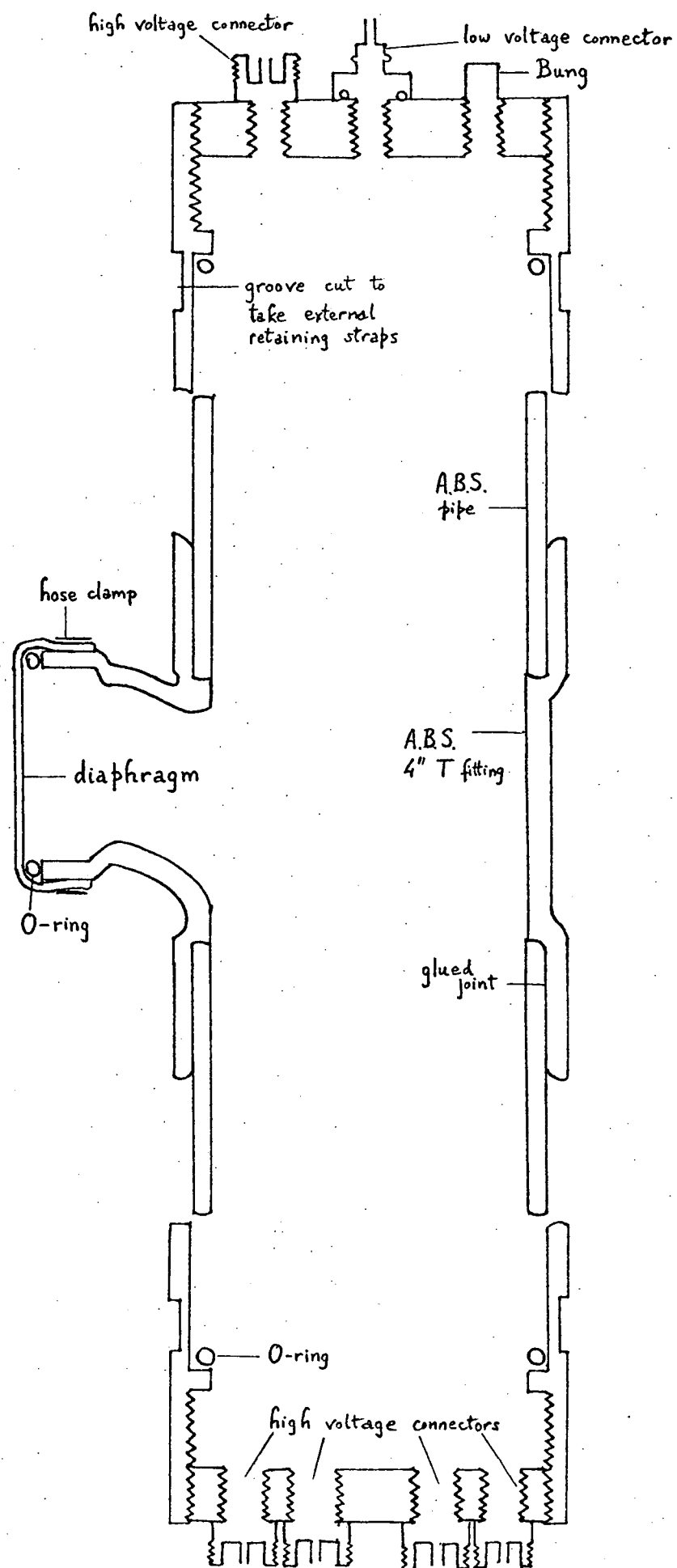


Fig. A1.4

The underwater housing for the H.V. switch.

nut. The connectors are completed by potting the electrical pin and socket into place with epoxy glue. The resulting connector has as water seals, both the original O-ring seal of the fitting and a face seal formed by the bearing down of the foremost edge of the male connector on the rubber bung. Such connectors have been tested to depths of 200m. without leakage.

The switch housing is itself shown in fig. A1.4. It is constructed of 10cm. A.B.S. sewer pipe and fittings, the principal components being a T-junction (5cm. side arm) and two pipe-thread to glue-joint adapters. The adapters were carefully selected to make sure that they had an interior lip suitable for the seating of an O-ring. Two short lengths of pipe, lathe-cut to ensure a true and smooth end surface for mating on the O-rings, were glued into the central T-junction. Special flat end caps were machined from 1.2cm. thick white nylon, threaded to fit the ABS couplings and tapped for the various connectors required. The case was to be sealed by pushing in the end pieces to seat against the O-rings. Pressure was maintained by an external arrangement of hose clamps. The case was designed to be oil-filled, and a diaphragm on the 5cm. open end of the T-junction would take up volume differences due to variations in temperature and pressure, so that the pressure inside the case remained the same as that outside it. The same oil would act as the dielectric

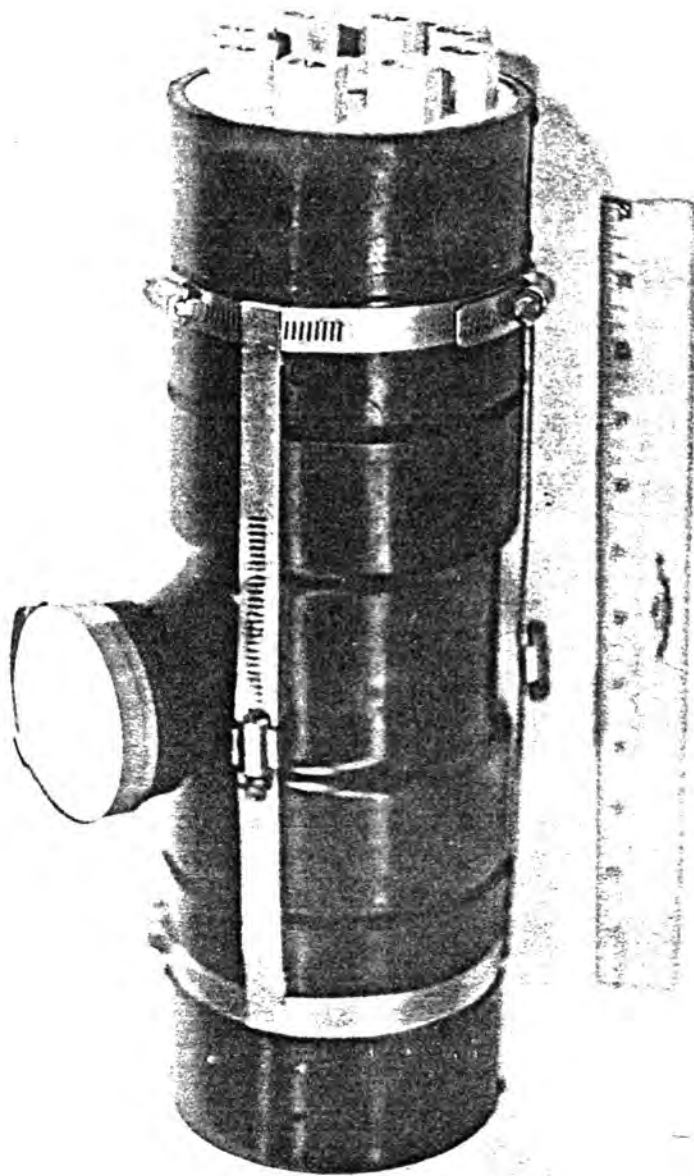


Fig. A1.4(b)

The underwater housing for the high voltage switch, showing the end-cap with the eight outputs to the sparker array, the pressure-equalising diaphragm on the side arm, and the external arrangement of hose clamps which holds the case together. (The scale at the side of the case is 1 foot long.)

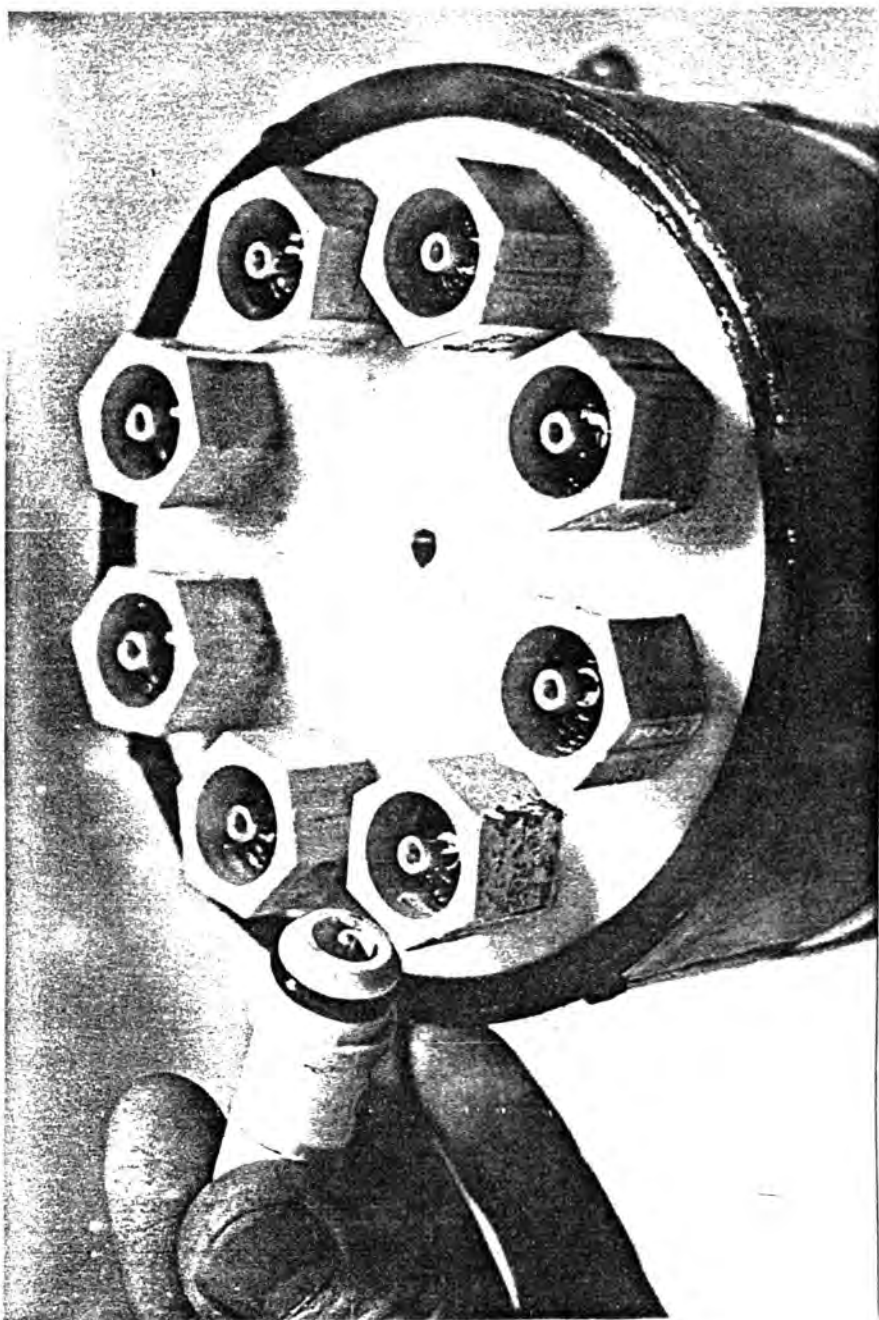
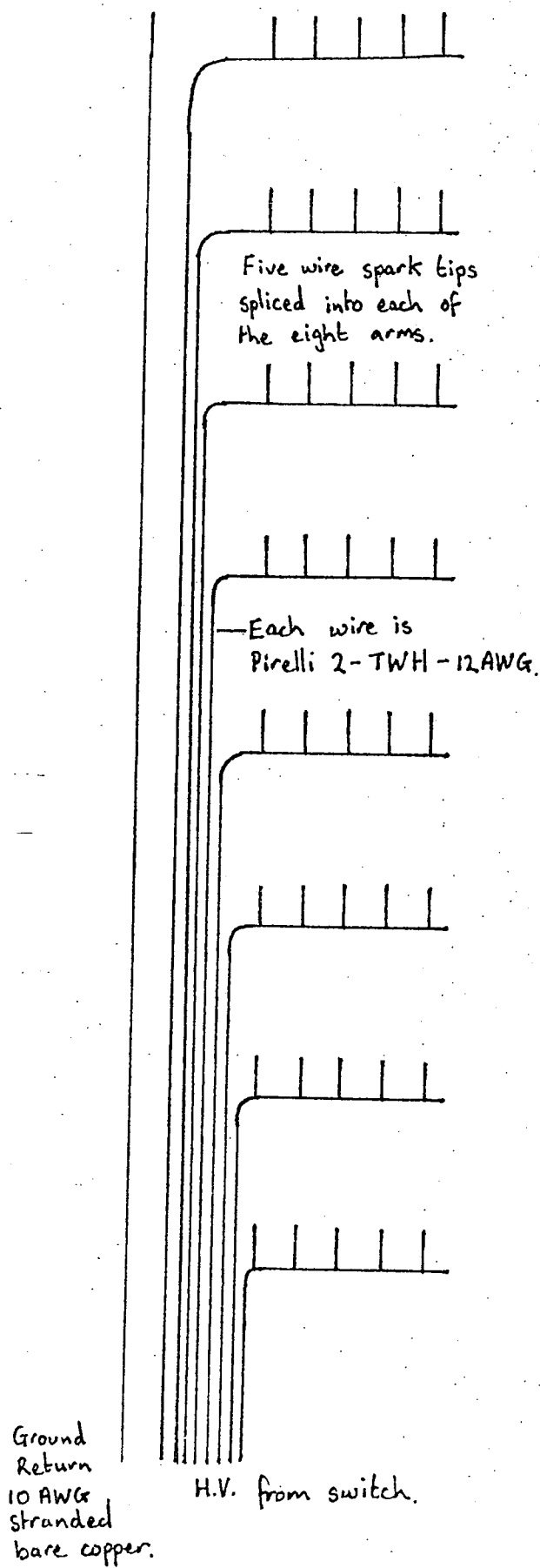


Fig. A1.4(c)

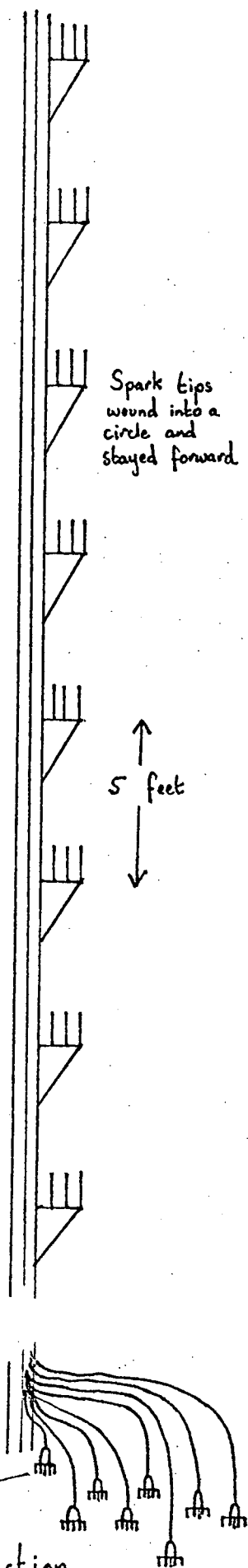
A close-up view of the end-cap of the underwater housing to show the high voltage connectors. The threads cut for the retaining shell nut (see fig.A1.3(b)) were a feature added after this photograph was taken.



Electrical Layout

Fig. A1.5

The 8-sparker array.



Mechanical Construction

to allow the relays to operate properly, without arcing. The sparker array which the switch would feed was constructed of 12-gauge house wire, supplied by Teledyne in their multitip sparkers (see fig. A1.5).

The hydrophone array was constructed as shown in fig. A1.6. In many ways, this has proved to be the least satisfactory part of the equipment, probably because its construction did not involve any novel ideas, and in consequence, received less attention. Its principal shortcoming is the appreciable discrepancy in the levels of output from the 5 hydrophones. It appears that this is a result of considerable scatter in the characteristics of the field effect transistors used in the pre-amplifiers. Unfortunately, the mechanical construction of the array does not lend itself to easy dismantling, and attempts to balance the outputs have till now been confined to adjustments which can be made at the input to the hydrophone switch (see fig. A1.7).

FIELD OPERATIONS.

The first, and until now, the only full trials of the system were done from a small boat in Halifax Harbour. With the boat stationary, the sparker array and H.V. switch were lowered in the water to test the operation of

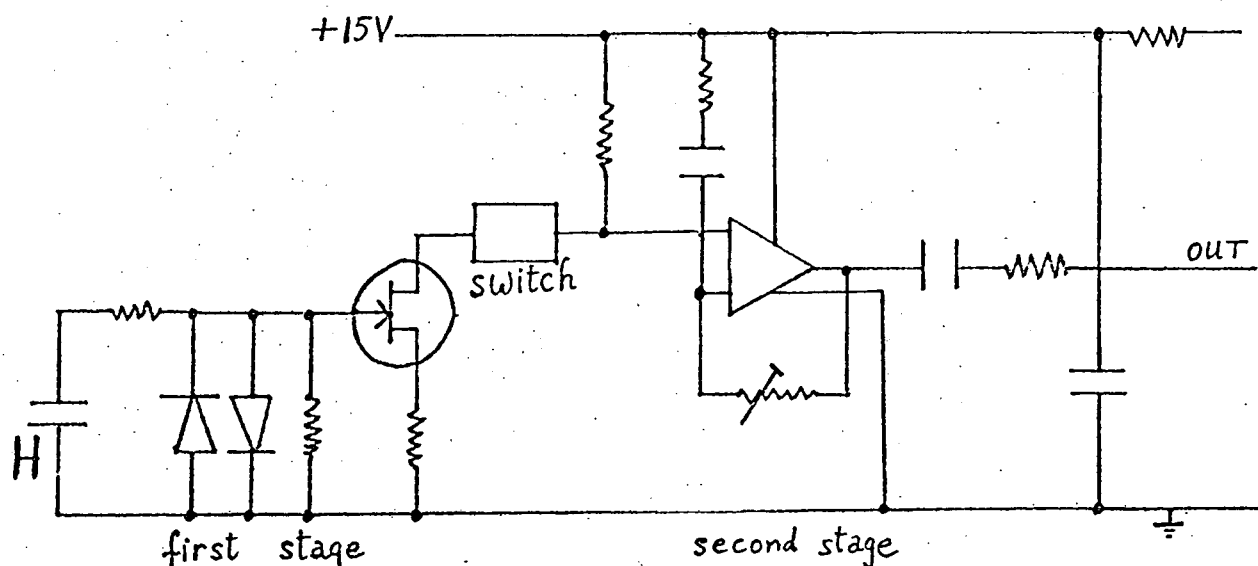


Fig. A1.7(a) The pre-amplification applied to the signal of the hydrophone, H, is accomplished in two stages by field effect transistor and operational amplifier. The 5 hydrophones, each with its own first-stage pre-amplifier were fed via the switch to a single second-stage preamplifier. (Circuits after Prentiss, 1973 and Bidgood, 1975).

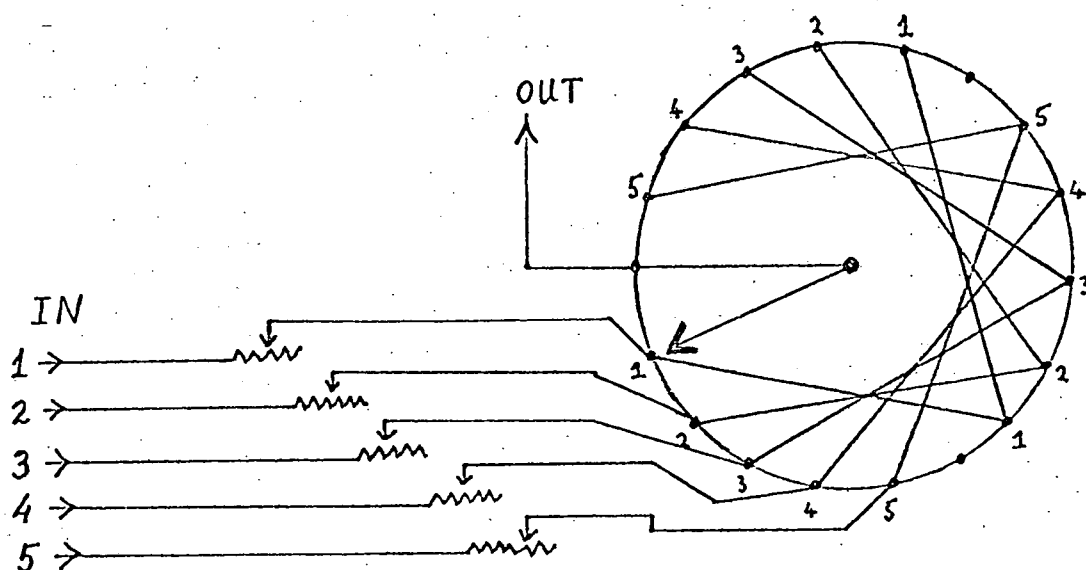
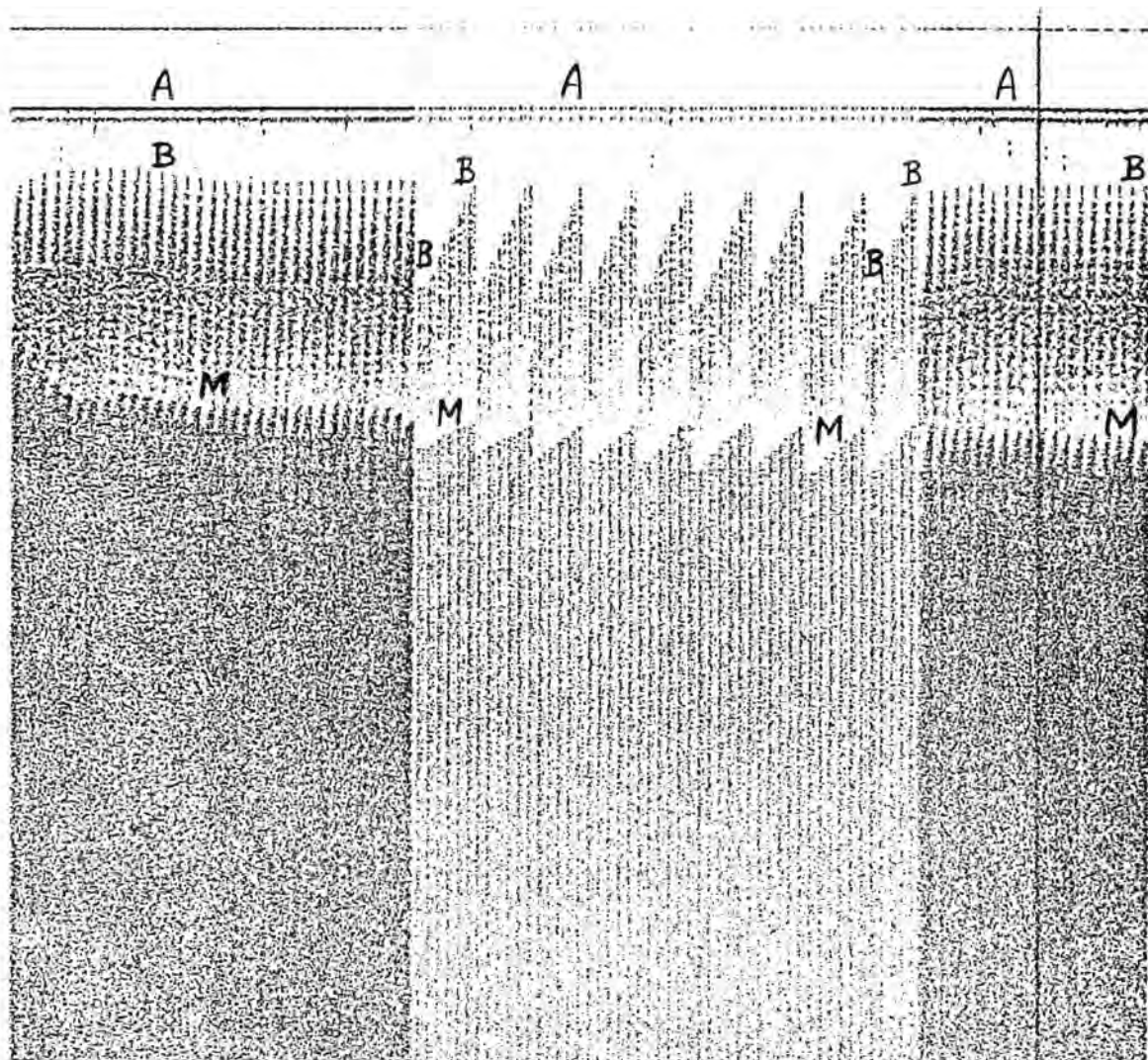


Fig. A1.7(b) The hydrophone switch assembly, showing the location of variable resistors inserted in an attempt to balance the hydrophone outputs.



increased recorder paper speed

Fig. A1.8

Output of switched sparker array as received by a surface hydrophone. Boat stationary.

Trigger event	A
Direct arrival	B
Bottom reflected arrival	M.

the switch, while the output of the sparkers was monitored by a hydrophone close to the surface, near the boat. It will have been realised that the high voltage switch is not really designed to turn a flow of current on and off. It is rather a series of gates which, when opened in turn while no current is flowing, will re-direct the current flow when it starts again. Accordingly it is necessary to arrange to syncopate the triggering of the high voltage source with the stepping of the switch. Suitable triggers were devised, the one as a pulse output directly by the graphic recorder which displayed the hydrophone record (an EPC 4600), the other by passing this pulse through delay circuits in a N.S.R.F. surface profiling electronics console. The result of firing the array in this way is shown in fig. A1.8.

When it had been established that the switch would operate satisfactorily, it was hauled back on board, because no suitable underwater vehicle could be made available on the small boat to allow the switch housing to be towed in the water. The switch was left on deck while the spark array remained in the water, towing just below the surface. The switched hydrophone was deployed such that its leading hydrophone took up a position just astern of the hindmost sparker. Profiles of the sort shown in fig. A1.9(a) were obtained over an area of the seabed shown in fig. A1.9(b). Unfortunately, a recorder failure

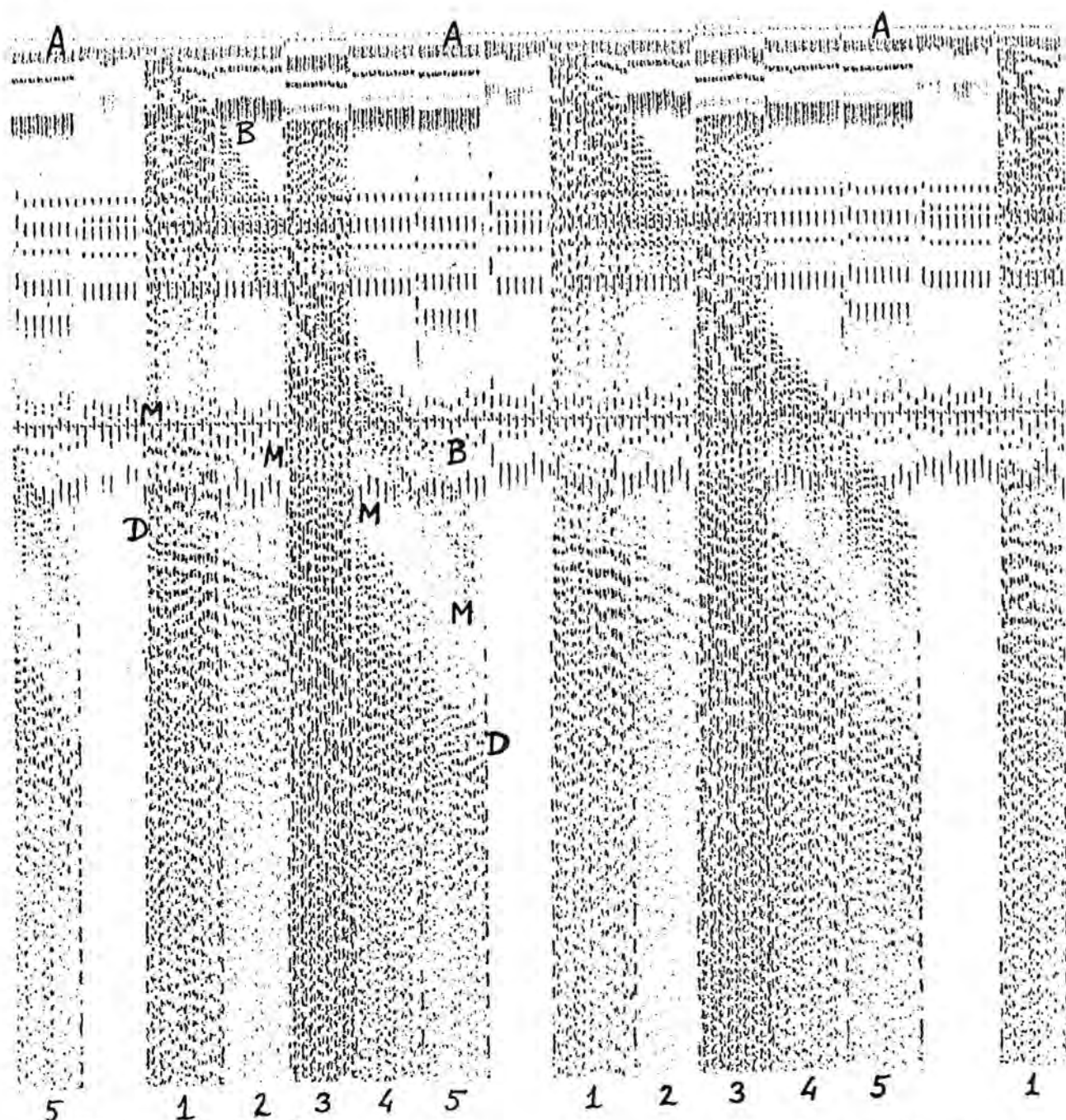


Fig. A1.9(a) Quasi-continuous wide angle reflection profiles made in Halifax Harbour by the switching sparker and hydrophone.

Trigger Event	A	Direct Arrival	B
Bottom reflection	M	Subbottom reflection	D

The "banding" on the record is a result of imbalance in the hydrophone outputs (hydrophones labelled 1 to 5). On the sixth band, no hydrophone is connected, leaving a blank which serves to separate the profiles one from another.

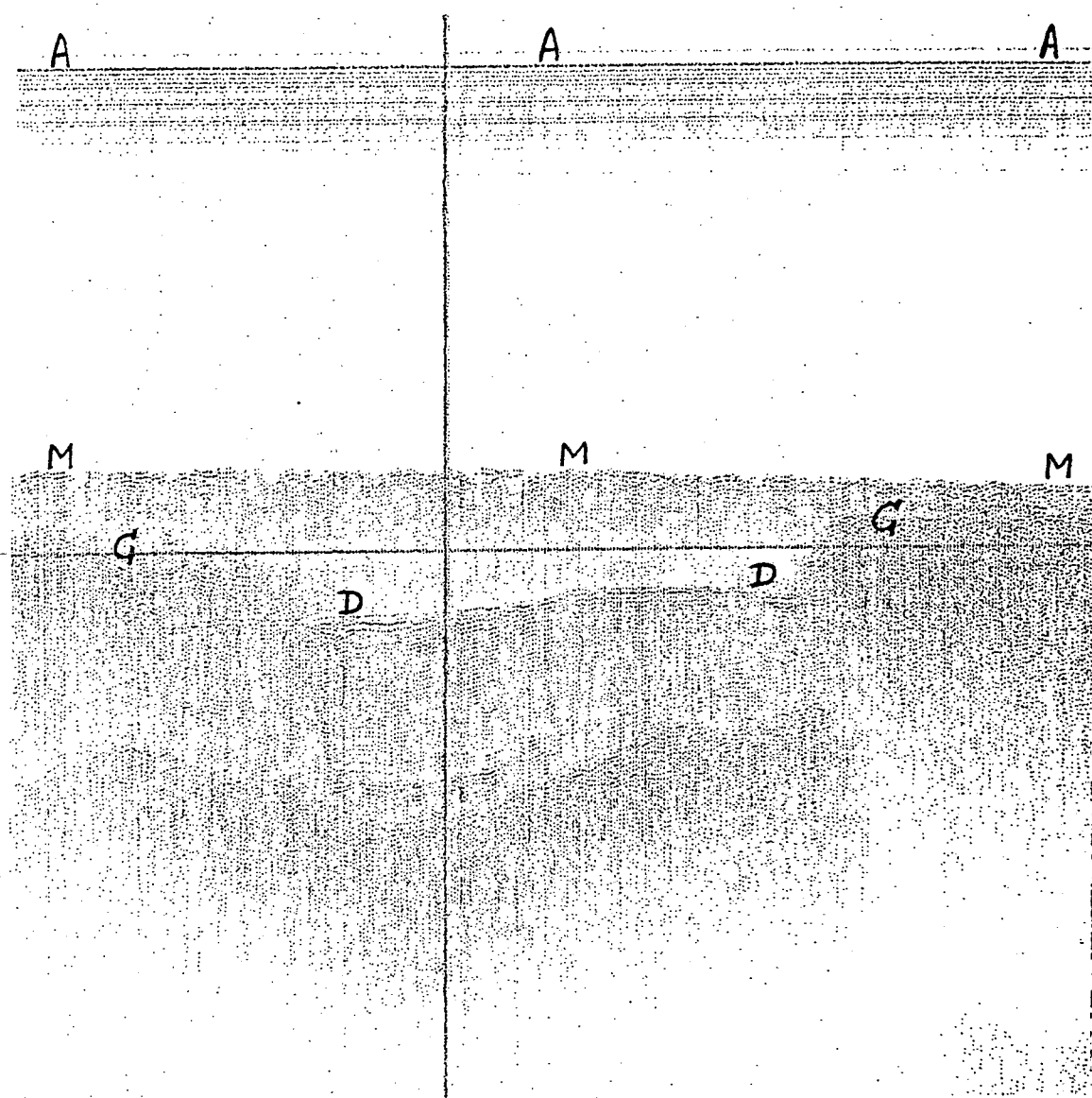


Fig. A1.9(b)

Vertical incidence boomer profile made over the same area as fig.A1.9(a).

Trigger	A
Seabed	M
Subbottom	D
Gas in sediments	G.

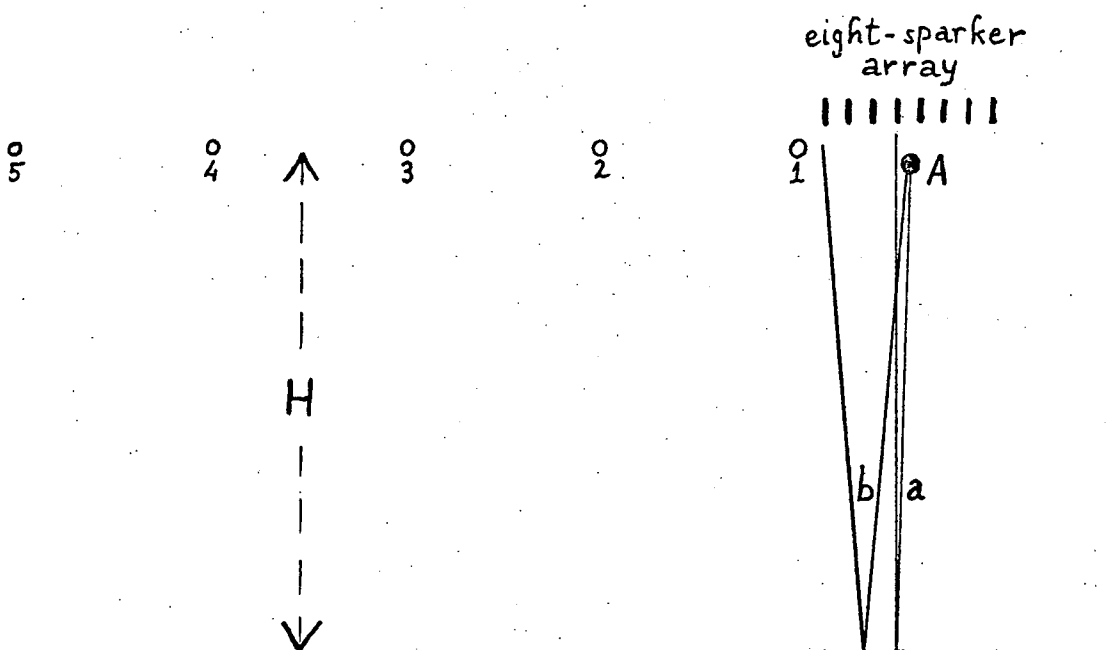


Fig. A1.10

Illustration of the potential problems in making a vertical incidence profile with a switching sparker array. The hydrophones numbered 1 to 5 are the switching hydrophones for wide angle work. The hydrophone at "A" is the one used for vertical incidence reflections. The table shows the expected distortions in the recording of the seabed.

Clearance off Bottom H (feet)	Time Difference Between Paths a and b (msec)	Resultant Fluctuation in Displayed Value of H (feet)
150	0.2	1.0
100	0.3	1.5
50	0.6	3.0

at the crucial moment prevented the making of a vertical incidence profile simultaneously with the switched profile, so that there is no indication of what degradation occurs in a vertical incidence profile when a switched sparker array replaces the more usual single sparker. Fig. A1.10 gives a brief outline of this question and shows that there should be little or no "graininess" in a vertical incidence profile made in this way, provided the spark array is 35 metres or more clear of the seabed.

An interpretation of the profile of fig. A1.9a gives a velocity in the sediment layer which is indistinguishable from that in the seawater (reflector assumed horizontal, in the absence of a vertical incidence profile).

APPENDIX 2.

Programme for analysis of headwave data.

A programme for the analysis of headwave arrival times
observed with equipment towed near the seabed.

A programme was written for dealing with the headwave travel-time data gathered in the Nerlerk area of the Beaufort Sea (see section 7.3). This programme is based on relationships given in McKay and McKay (1982), (see equations no. 1 and 4 of that paper).

Errors $\delta(p)$, where $\delta(p)$ is the error on parameter p , are treated as follows:

(i) Errors arising from the use of eqn. 1 of McKay and McKay, in determining the towing depth "h" of the trailing hydrophone (see fig. 5 of McKay and McKay).

$$\delta h = [(b+c)/2a]\delta(b+c) + (d/2a)\delta d + [(b+c)^2 - d^2]/4a^2 \delta a$$

(ii) Errors arising in the formation of terms for the quadratic equation formed from equation no. 4 of McKay and McKay by the substitution $t = \tan(\theta/2)$.

For the equation $pt^2 + rt + s = 0$,

$$p = -[(h_2 + h_1) + v_1 T]$$

$$r = 2\sqrt{[d^2 - (h_2 - h_1)^2]}$$

$$s = [(h_2 + h_1) - v_1 T]$$

from which δp , δr , δs can be derived as:

$$\delta p = \delta h_1 (\partial p / \partial h_1) + \delta h_2 (\partial p / \partial h_2) + v_1 \delta T (\partial p / \partial T)$$

$$\delta r = \delta d (\partial r / \partial d) + \delta h_1 (\partial r / \partial h_1) + \delta h_2 (\partial r / \partial h_2)$$

$$\delta s = \delta p$$

where for each parameter error and for each partial derivative, the modulus of its value is implied.

(iii) Errors arising in the solution of the quadratic equation in the standard form:

$$\delta t = \delta p (\partial t / \partial p) + \delta r (\partial t / \partial r) + \delta s (\partial t / \partial s)$$

(iv) The error arising in the arctangent solution:

$$\theta = 2 \tan^{-1}(t)$$

The above error analysis is incorporated in the programme which follows. The programme was written in "BASIC" for an Apple II+ microcomputer and is interactive in nature. It prompts for input and, at each prompt, it expects the observed value of the travel time requested, together with an estimate of the error in measurement of the travel time. Values for the constants of the specific geometrical deployment used at the Nerlerk site are incorporated in the programme, but at the beginning of a run, interactive requests are made to reset these, if required, to accommodate other deployments. Provision is made to calculate and display the velocity in the lower layer on the basis of each of two values for the water velocity, the one being input as the best available estimate of the velocity expected in the bottom water, the other being calculated from the horizontal offset of the trailing hydrophone and the measured travel time of the direct acoustic arrival.

The programme has not been run on machines other than the Apple II+ and its close relatives the Apple II and IIe. It probably would require alteration to run on other machines.

DLIST

```

100 REM      PROG TO DEAL WITH HEADWAVE DATA
105 REM
106 REM
110 REM      *** VARIABLES ***
115 REM
120 REM      SCAL      NO. OF SCALE DIVS PER MSEC
130 REM      VELO      ASSUMED WATER VELOCITY
140 REM      XFAC      VELO/SCAL
150 REM      TIDIR > DIRECT
160 REM      TDEP > BOTTOM U.I. >
170 REM      THEAD > HEADWAVE >
180 REM      SUI > SURFACE U.I.>
190 REM      SWA > SURFACE W.A.>    ALL ARR. TIMES IN SCALE DIVS
195 REM
200 REM      UIHO > HORIZ. U.I.>
210 REM      UUIO > VERTICAL U.I> SOURCE-RECEIVER SEPRN. (METRES)
220 REM      WAHO > HORIZ. W.A.>
225 REM
230 REM      H1 > SOURCE ABOVE SEABED
240 REM      H2 > REMOTE HYD. ABOVE SEABED
250 REM      XSUI > SOURCE BELOW SEA SURFACE
260 REM      XH > REMOTE. HYD. BELOW SEA SURF.    (ALL IN METRES)
265 REM
270 REM      XA >
280 REM      XB >    QUADRATIC TERMS
290 REM      XC >
295 REM
300 REM      XT      TAN (0.5*THETA)
310 REM      TEET2    0.5 * THETA
320 REM      XTNET    THETA - THE 'CRITICAL' ANGLE
330 REM      RUELO2    VELO IN LOWER LAYER WITH UPPER AS 1.0
340 REM      WUELO     WATER VELO DET. FROM MEASUREMENT
350 REM      UUILO2D    VELO IN LOWER LAYER WITH UPPER AS WUELO
360 REM      U2LO2A    VELO IN LOWER LAYER WITH UPPER AS VELO
365 REM
370 REM
375 DIM XLN$(20),FX$(10)
380 REM      ***   DEFINE SCALE FACTORS   ***
381 D1$ = "": REM  CNTRL-D CHAR
382 D$ = CHR$(4)
383 RITTER$ = "PR#4"
384 PRINT D$:"PR#0"
390 INPUT " ENTER SCALE FACTORS (Y), OR DEFAULT (N) ":Q0$
392 IF Q0$ = "N" THEN 420
394 INPUT "INPUT SCAL(0.3908) ":SCAL
396 INPUT "INPUT VELO (1.460) ":VELO
398 INPUT "INPUT UIHO (2.14) ":UIHO
400 INPUT "INPUT WAHO (44.65) ":WAHO
402 INPUT "INPUT UUIO (0.76) ":UUIO
404 GOTO 490
405 REM
410 REM      ***   DEFINE GEOMETRY OF TOW   ***
420 UIHO = 2.14
430 WAHO = 44.65
440 UUIO = 0.76
442 SCAL = 0.3908

```

```

444 VELO = 1.460
445 REM
450 REM     NOTE THAT ALTHOUGH THE MAIN CALCULATION DOES NOT NEED
460 REM     AN ACCURATE KNOWLEDGE OF THE WATER VELOCITY,
470 REM     A PROPER APPLICATION OF THE ABOVE TOW-GEOMETRY PARAMETERS
480 REM     DOES DEPEND ON A GOOD VALUE FOR THE WATER VELOCITY.
485 REM
490 REM     ***** ONERR GOTO 500
494 REM
500 PRINT D$;"PR#0"
501 INPUT "INPUT LINE NO.:";XLN$
502 I% = - 1
510 REM
520 REM
530 REM
540 INPUT "INPUT FIX NO.:";FX$
550 REM
560 INPUT "INPUT VERTICAL INC. DEPTH AND ERROR ";TDEP,D2
565 INPUT "INPUT DIRECT ARRIVAL AND ERROR ";TIDIR,D1
570 INPUT "INPUT REFRACTED TIME AND ERROR ";THEAD,D3
580 INPUT "INPUT U.I. SURFACE REFLECTN AND ERROR ";SUI,D4
585 INPUT "INPUT W.A. SURFACE REFLECTN AND ERROR ";SMA,D5
590 REM
600 REM
610 PRINT D$;"PR#4"
620 REM     HOME
621 FOR IJ = 1 TO 5 STEP 1
622 PRINT ""
623 NEXT IJ
630 PRINT SPC( 20);"HEADWAVE ANALYSIS"
631 PRINT ""
632 PRINT ""
635 PRINT SPC( 5);XLN$; TAB( 60);FX$
636 PRINT ""
650 PRINT "DIRECT    U.I. DEPTH  HEADWAVE    U.I. SURFACE  W.A. SURFACE"
670 PRINT TIDIR; TAB( 12);TDEP; TAB( 24);THEAD; TAB( 36);SUI; TAB( 50);SMA
680 PRINT D1; TAB( 12);D2; TAB( 24);D3; TAB( 36);D4; TAB( 50);D5
700 REM
710 REM
750 REM     *** DEFINE GEOMETRY ***
800 REM
810 REM
820 REM     ** CONVERT INPUT DATA TO METRES **
830 XFAC = VELO / SCAL
840 TIDIR = TIDIR / SCAL
845 XIDIR = TIDIR * VELO
846 F7XIDIR = D1 * XFAC
850 XDEP = TDEP * XFAC
851 X9 = SQR (XDEP * XDEP - UIHO * UIHO)
852 H1 = (X9 - UIHO) / 2.0
853 E1H1 = 0.5 * ((XDEP * F6) / X9 + (D6 * UIHO) / X9 + D7)
860 H1 = ( SQR (XDEP * XDEP - UIHO * UIHO) - UIHO) / 2.0
870 REM     NOTE THAT THIS EXPRESSION FOR 'H1' IS EXACT.
880 REM     THERE IS NO APPROXIMATION
890 REM     'H1' IS THE VERTICAL DISTANCE FROM REFLECTOR TO SOURCE
900 REM
910 THEAD = THEAD / SCAL
911 D3 = D3 / SCAL
920 REM

```

```

930 REM ** PROCEED TO FIND VERT POSH OF REMOTE HYDRO.
940 XSUI = (SUI * XFAC + UUIO) / 2.0
941 E3XSUI = (D4SUI * XFAC + D7UUIO) / 2.0
950 XH = ((XFAC * SWA) ^ 2 - X1DIR ^ 2) / (4.0 * XSUI)
951 E4XH = (XFAC * SWA * XFAC * D5SWA) / (2.0 * XSUI) + (X1DIR * F7X1DIR) / (2.0 * XSUI) + (XH * E3XSUI) / XSUI
960 H2 = XSUI + H1 - XH
961 E2H2 = E3XSUI + E1H1 + E4XH
962 IF H2 < 0.0 THEN 1150
963 H3 = 10.0 * H1
964 H3 = 10.0 * H1
966 IF H2 > H3 THEN 1160
970 REM 'H2' IS THE DISTANCE ABOVE REFLECTOR FOR
980 REM REMOTE HYDRO. ASSUMING FLAT HORIZON
990 REM NOW SET UP QUADRATIC TERMS
1000 REM
1010 XA = H1 + H2 + VELO * THEAD
1011 E5XA = E1H1 + E2H2 + VELO * D3THEAD
1015 PRINT "X1DIR,H1,H2 "X1DIR,H1,H2
1019 Y1 = SQR (X1DIR ^ 2 - (H2 - H1) ^ 2)
1020 XB = - 2.0 * (SQR (X1DIR ^ 2 - (H2 - H1) ^ 2))
1021 E6XB = (2.0 * X1DIR * F7X1DIR) / Y1 + 2.0 * (H2 - H1) * (E1H1 + E1H2)
1030 XC = VELO * THEAD - (H1 + H2)
1031 E7XC = E5XA
1033 Z3 = SQR (XB * XB - 4.0 * XA * XC)
1035 XQ = (- XB - SQR (XB * XB - 4.0 * XA * XC)) / (2.0 * XA)
1036 PRINT " XA,XB,XC "XA,XB,XC
1037 PRINT " XQ "XQ
1038 F9XQ = ABS (E5XA * (- XB / (2.0 * XA * XA) + XC / (XA * Z3) + Z3 / (2.0 * XA * XA))) + E6XB * (1.0 + XB / Z3) / (2.0 * XA) + E7XC * (1.0 / Z3)
1040 TEET2 = ATN (XQ)
1041 E9TEET2 = (1.0 / (1.0 + XQ ^ 2)) * F9XQ
1050 XTHET = 2.0 * TEET2
1051 F1XTHET = 2.0 * E9TEET2
1055 PRINT "TEET2,XTHET "TEET2,XTHET
1060 RUELO2 = 1.0 / SIN (XTHET)
1061 F2RUELO2 = (1.0 / (SIN (XTHET) * TAN (XTHET))) * F1XTHET
1070 UUELO = WAHO / T1DIR
1080 U1LO2 = UUELO * RUELO2
1085 F4U1LO2 = UUELO * F2RUELO2
1090 U2LO2 = VELO * RUELO2
1095 F5U2LO2 = VELO * F2RUELO2
1100 PRINT " RELATIVE REFRACTED VELOCITY "RUELO2; SPCK 5;F2RUELO
1110 PRINT " WATER VELOCITY DETERMINED "UUELO
1120 PRINT " ASSUMED WATER VELOCITY "VELO
1130 PRINT " REFRACTOR ON BASIS OF UUELO "U1LO2; SPCK 5;F4U1LO2
1140 PRINT " REFRACTOR ON BASIS OF VELO "U2LO2; SPCK 5;F5U2LO2
1142 PRINT D1$:"OPEN RESULTS,L512"
1144 PRINT D1$:"READ RESULTS,R0"
1145 INPUT M
1146 PRINT D1$
1147 PRINT " M (READ FROM FILE) = "M
1148 M = M + 1
1149 PRINT "WRITE RESULTS,R":M
1150 PRINT D1$:"WRITE RESULTS,R":M; PRINT M
1151 PRINT XLN$: PRINT FX$: PRINT SCAL: PRINT VELO: PRINT UUIO: PRINT WAHO
1152 PRINT UUIO: PRINT TDEP: PRINT D2: PRINT T1DIR: PRINT D1: PRINT THEAD: PRINT D3
1153 PRINT SUI: PRINT D4: PRINT SWA: PRINT D5: PRINT RUELO2: PRINT F2RUELO2: PRINT UUELO
1154 PRINT D1$
1155 PRINT D1$:"WRITE RESULTS,R0"
1156 PRINT M

```

```
1158 PRINT D1$:"CLOSE RESULTS"
1159 PRINT "  M (AFTER CLOSE) = " ; M
1160 GOTO 1170
1162 PRINT "H2 < 0.0      " ; H2
1164 GOTO 1170
1166 PRINT "  H2 TOO BIG  " ; H2
1170 I% = I% + 1
1171 IF I% > 0 THEN 500
1173 H2 = H1
1174 PRINT "REPEAT WITHOUT H2 CORRECTION  "
1175 GOTO 1000
1200 GOTO 100
1250 END
```


APPENDIX 3.

Programme for analysis of W.A. reflexion data

A Fortran programme for carrying out the data reduction of wide angle reflexion profiles made with source and receiver deep in the water column.

FORTTRAN PROGRAMME FOR WORKING OUT LAYER VELOCITIES FROM
REFLEXION TRAVEL TIME DATA.

The subroutine LSQ fits a straight line to an array of N points X,Y. It returns the line's gradient A+SIGA and intercept B+SIGB, and number (IDEV) of the data point farthest from the line (by a distance DEVM).

The subroutine REORD rejects the IDEVth data point from the N-dimensional arrays X and Y, and re-orders these arrays as N-1 dimensional arrays.

The main programme calculates hydrophone positions and reduces travel times to a common horizontal datum before using subroutine LSQ to fit a straight line to $T^2 - X^2$ data to arrive at an average velocity by the method of Green. The velocity in the lower layer is then found from this average velocity together with the velocity in the water column.

The velocity in the water column is assumed constant and its value is found from the difference in arrival time of the trailing hydrophones of known separation.

INPUT.

CARD IMAGES 1,2, AND 3.

Alphanumeric header which is reprinted on the output without change.

CARD IMAGE 4. I10,2F10.0,I10

- (i) Number of interfaces (including the seabed) (<16).
- (ii) Number of scale divisions per millisecond for input travel time data on cards 6 and onwards.
- (iii) Distance of vertical incidence hydrophone above acoustic source (in feet).
- (iv) Method of specifying vertical movement of deep-tow fish (if any) over the horizontal distance relevant to this wide-angle calculation.
 - = zero or blank - no movement.
 - = 1 - vertical displacement between first and last switching position is given as YDIP and read in on the card following travel time cards.
 - = 2 - vertical height of fish above (horizontal) seabed is given for each switching position on card following travel time cards.

CARD IMAGE 5. 1615

- (i) Number of travel times identifying seabed.
- (ii) Number of travel times identifying 1st subbottom.
- (iii) Number of travel times identifying 2nd subbottom.
- etc.
- (n+1) Number of travel times identifying nth subbottom.

CARD IMAGE 6. 8F10.0

- (i) Travel time of direct wave to vertical incidence hydrophone.
- (ii) Travel time of direct wave to the 2nd hydrophone.
- etc.
- (n) Travel time of direct wave to the nth hydrophone.

CARD IMAGE 7. 8F10.0

- (i) Travel time of bottom reflexion to vertical incidence hydrophone.
- (ii) Travel time of bottom reflexion to the 2nd hydrophone.
- etc.
- (n) Travel time of bottom reflexion to the nth hydrophone.

CARD IMAGE (7+N)+1. 8F10.0

Value or values for YDIP as explained in card 4(iv) above. If no vertical movement of the tow fish is considered, then this card image will be absent from the input.

In addition to the above, a dipping reflector can be specified by giving the amount of the dip in the last field (71 - 80) of the travel time card for the layer in question. This field should be blank (or zero) for horizontal layers. A number punched in this field should be the amount of reflector rise as seen on the vertical incidence record between the start and end of the switching cycle.

Note that the "travel times" referred to in cards 6 to (7+n)+1 above are to be given in the arbitrary scale divisions. These values are converted to true travel times by the programme, provided that the scale factor on card 4(ii) is properly specified.

```

1      SUBROUTINE REORD(X,Y,IDEV,N)
2      C
3      C
4      C      PURPOSE IS TO REMOVE ONE ELEMENT 'IDEV' FROM THE ARRAYS X & Y
5      C      AND TO SHUFFLE ALONG ALL HIGHER NUMBERED ELEMENTS TO FILL UP
6      C      THUS LEAVING ARRAYS WITH ONE ELEMENT LESS THAN WHEN THE SUBROUTINE
7      C      WAS ENTERED
8      C      THIS SUBROUTINE IS HELPFUL AS PART OF THE PROCESS OF REJECTING
9      C      'BAD POINTS WHILE CURVES ARE BEING FITTED TO DATA
10     C
11     C
12     DIMENSION X(N),Y(N)
13     DO 100 I=1,N
14     IF(I - IDEV)100,2,2
15     2 IF(I - N)3,4,4
16     4 RETURN
17     3 X(I)=X(I + 1)
18     Y(I)=Y(I + 1)
19     100 CONTINUE
20     RETURN
21     END
22     SUBROUTINE LSQ(X,Y,A,B,SIGA,SIGB,N,IDEV,DEVM)
23     C
24     C      THIS IS A LEAST SQUARES FITTING ROUTINE
25     C      IT USES BASIC LEAST SQUARES TECHNIQUES
26     C      'A' IS THE LINE'S GRADIENT
27     C      'B' IS THE INTERCEPT
28     C      'SIGA' & 'SIGB' ARE THE STANDARD DEVIATIONS ON A & B
29     C      'IDEV' IDENTIFIES THE MOST DEVIANT VALUE OF 'Y'
30     C
31     DIMENSION X(N),Y(N)
32     SX=0.0
33     SY=0.0
34     SX2=0.0
35     SXY=0.0
36     DO 100 J=1,N
37     SX=SX + X(J)
38     SX2=SX2 + X(J)*X(J)
39     SY=SY + Y(J)
40     SXY=SXY + X(J)*Y(J)
41     100 CONTINUE
42     XN=FLOAT(N)
43     Q=XN*SX2 - SX*SX
44     A=(XN*SXY - SX*SY)/Q
45     B=(SY*SX2 - SX*SXY)/Q
46     SD2=0.0
47     DEVM=0.0
48     DO 200 J=1,N
49     DEV=(Y(J) - A*X(J) - B)**2
50     SD2=SD2 + DEV

```

```

51      IF (DEV - DEVM) 3,3,4
52      4 DEVM=DEV
53      IDEV=J
54      3 CONTINUE
55      200 CONTINUE
56      Q1=(XN - 2.0)*Q
57      IF (Q1) 5,6,5
58      6 SIGA=0.0
59      SIGB=0.0
60      RETURN
61      5 CONTINUE
62      SIGA=SQRT(XN*SD2/Q1)
63      SIGB=SQRT(SX2*SD2/Q1)
64      RETURN
65      END
66      PROGRAM PNAME (INPUT,OUTPUT,TAPE5=INPUT,TAPE6=OUTPUT)
67      C      PROGRAMME FOR THE REDUCTION OF WIDE ANGLE REFLECTION PROFILES
68      C      OBTAINED FROM DEEP TOW SEISMIC EQUIPMENT ALASDAIR G. MC KAY 1979
69      C      THE METHOD IS THE 'T-SQUARED : X-SQUARED' METHOD AS DESCRIBED E.G.
70      C      IN MAYNARD ET AL., 1973, MODIFIED TO TAKE ACCOUNT OF THE VARYING
71      C      LEVELS TAKEN UP BY THE MID-WATER DEPLOYMENT OF AN ARRAY
72      C      OF HYDROPHONES
73      C
74      C
75      C      NPT(J)    NO. OF POINTS(NO. HYDROPHONES + V. INC.) IN LAYER 'J'
76      C      NINT     NO. OF INTERFACES(INCLUDING BOTTOM) LESS THAN 16
77      C      DPMS     SCALE FACTOR: NO. OF DIVISIONS PER MSEC.
78      C      COROF    OFFSET CORRECTION FOR DISTANCE BETWEEN VERT. INC.
79      C      HYDROPHONE AND SOUND SOURCE (IN FEET .... VERTICAL)
80      C
81      C
82      DIMENSION AMNH(10),HY2(10),R(100),D(8),BC(8),CD(8),TTOT2(8),X2(8),
83      IVAV(10),ERRAV(10),AYY(10),TEXT(20),NPT(16),Y(6)
84      DATA NMAX,HYSEP/6,160.0/
85      MR=5
86      MI=6
87      IKID=1
88      90 CONTINUE
89      WRITE(MI,98)
90      98 FORMAT(1H1)
91      C      READ AND WRITE IDENTIFYING TEXT
92      DO 100 I=1,3
93      READ(MR,1)(TEXT(J),J=1,20)
94      1 FORMAT(20A4)
95      11 FORMAT(1H ,20A4)
96      100 WRITE(MI,11)(TEXT(J),J=1,20)
97      2 FORMAT(8F10.3)
98      3 FORMAT(16I5)
99      READ(MR,5)NINT,DPMS,COROF,IDDIP
100     5 FORMAT(I10,2F10.0,I10)

```

```

101      READ(MR,3) (NPT(J),J=1,NINT)
102      C
103      C      READ IN DIRECT WAVE DATA
104      C
105      READ(MR,2) (D(J),J=1,NMAX)
106      C
107      C      READ IN BOTTOM AND SUBBOTTOM-REFLECTED DATA
108      C
109      DO 200 K=1,NINT
110      J1=(K - 1)*8 + 1
111      J2=J1 + 7
112      READ(MR, 2) (R(J),J=J1,J2)
113      200 CONTINUE
114      C
115      C      PRINT OUT DATA
116      C
117      WRITE(MI,6)
118      6 FORMAT(///," NUMBER OF INTERFACES      SCALE FACTOR      OFFSET")
119      WRITE(MI,7)NINT,DPMS,COROF
120      7 FORMAT(9X,I2,12X,F12.3,6X,F5.1)
121      WRITE(MI,8)
122      8 FORMAT(/," NUMBER OF POINTS PER INTERFACE")
123      WRITE(MI,9) (J,J=1,NINT)
124      9 FORMAT(8I10)
125      WRITE(MI,9) (NPT(J),J=1,NINT)
126      WRITE(MI,4)
127      4 FORMAT(/," DIRECT AND REFLECTED TRAVEL TIMES")
128      WRITE(MI,2) (D(J),J=1,NMAX)
129      DO 707 K=1,NINT
130      J1=(K - 1)*8 + 1
131      J2=J1 + NPT(K) - 1
132      WRITE(MI,2) (R(J),J=J1,J2)
133      707 CONTINUE
134      C
135      C      CALCULATE VELOCITY FROM POINTS NO. 6 & 2 OF DIRECT WAVE DATA
136      C      TAKING THESE HYDROPHONES AS BEING 160 FEET APART
137      WVEL=0.3048*HYSEP*DPMS/(D(6) - D(2))
138      C
139      C      VELOCITY CALCULATED IS IN M./MSEC.      PRINT THIS.
140      C
141      WRITE(MI,95)WVEL
142      95 FORMAT(" ACOUSTIC VELOCITY IN WATER CALCULATED FROM HYDROPHONES NO
143      1. 2 AND 6      ",F10.4," KM./SEC.",//)
144      COROF = COROF*0.3048*DPMS/WVEL
145      AY1=(R(1) - COROF)/2.0
146      IDDIP=IDDIP + 1
147      GO TO (51,52,53),IDDIP
148      51 DO 59 K=1,NMAX
149      Y(K)=AY1
150      59 CONTINUE

```



```

151      GO TO 60
152 52 READ (MR,2) YDIP
153      DO 58 K=1,NMAX
154      Y(K)=AY1 + YDIP*(FLOAT(K-1)/FLOAT(NMAX-1))
155 58 CONTINUE
156      GO TO 60
157 53 READ (MR,2) (Y(K),K=1,NMAX)
158      DO 75 K=1,NMAX
159      Y(K)=(Y(K) - COROF)/2.0
160 75 CONTINUE
161 60 CONTINUE
162      J=1
163      NPT11=NPT(1)
164      DO 500 K=2,NPT11
165      HY2(K) = ((R(K))**2 - (D(K))**2)/(4.0*Y(K))
166      AMNH(K) = Y(K) - HY2(K)
167      APLH=AY1 + HY2(K)
168      THETA= ASIN(AMNH(K)/D(K))
169      BC(K) = (AMNH(K)/APLH)*SQRT(R(K)**2 - APLH**2)
170      CD(K) = D(K)*COS(THETA) - BC(K)
171      AC = SQRT(AMNH(K)**2 + BC(K)**2)
172      TTOT2(K) = (R(K) + AC)**2
173      X2(K) = (CD(K) + 2.0*BC(K))**2
174 500 CONTINUE
175      TTOT2(1)=4.0*AY1**2
176      X2(1)=0.0
177      CALL LSQ(X2,TTOT2,AA,BB,SIGA,SIGB,NPT(1),IDEV,DEVM)
178      WRITE(MI,21)
179 21 FORMAT(" SEABED REFLECTION",/,7X,"GRADIENT",9X,"INTERCEPT",13X,
180 1"GREATEST DEVIANT")
181      WRITE(MI,22)AA,SIGA,BB,SIGB,IDEV,DEVM
182 22 FORMAT(2X,F7.4," : ",F7.5,5X,F8.2," : ",F7.4,5X,I10,3X,F7.2)
183      NPTJJ=NPT(J)
184      WRITE(MI,23) (TTOT2(I),I=1,NPTJJ)
185 23 FORMAT(" SQUARED TRAVEL TIMES",8F10.2)
186      WRITE(MI,24) (X2(I),I=1,NPTJJ)
187 24 FORMAT(" SQUARED DISTANCES ",8F10.2)
188 C
189 C      PROCEED TO CALCULATIONS FOR LOWER INTERFACES
190 C
191      AYY(1)=AY1
192      VAV(1)=WVEL
193      ERRAV(1)=0.0
194 C      THE ABOVE STATEMENT SETS THE ERROR ON THE WATER VELOCITY TO 0.0
195 C      IT MAY BE NECESSARY TO CHANGE THIS SOMETIME
196 C
197      DO 400 J=2,NINT
198      J1 = (J - 1)*8 + 1
199 C
200 C

```

```

201 C      INSERT THIS SECTION IN AN ATTEMPT TO ACCOUNT FOR THE DIPPING
202 C      LAYER PROBLEM.      THIS METHOD IS RATHER CRUDE
203      IF(R(J1 + 7))81,82,81
204 81 DIPA=ATAN((R(J1 + 7)*VAV(J - 1))/(DPMS*HYSEP*0.3048))
205      COSTH=COS(DIPA)
206      GO TO 83
207 82 COSTH=1.0
208 83 CONTINUE
209 C
210 C
211      AY=(R(J1) - COROF)/2.0
212      AYY(J)=AY
213      NPTJ3=NPT(J)
214      DO 600 K=2,NPTJ3
215      HY=HY2(K) + (AY - Y(K))
216      APLH=AY + HY
217      BCPCD=BC(K) + CD(K)
218      X3 =(2.0*BCPCD*AY)/APLH
219      DK=SQRT((X3-BCPCD)**2 + AMNH(K)**2)
220      K1=J1 + K - 1
221      TTOT2(K)=(R(K1) + DK)**2
222      X2(K)=X3*X3
223 600 CONTINUE
224      TTOT2(1)=4.0*AY**2
225      X2(1)=0.0
226      WRITE(MI,31)J
227 31 FORMAT(///," REFLECTION FROM INTERFACE NO.",I3)
228      NPTJ1=NPT(J)
229      WRITE(MI,23) (TTOT2(I),I=1,NPTJ1)
230      WRITE(MI,24) (X2(I),I=1,NPTJ1)
231      WRITE(MI,41)
232 41 FORMAT(6X,"GRADIENT",14X,"INTERCEPT",14X,"GREATEST DEVIANT",1X,
233 1"AVERAGE VELOCITY   INTERVAL VELOCITY AVERAGE TO THIS REFLECTOR"
234 2)
235      NPTJ2=NPT(J)
236      DO 800 K=2,NPTJ2
237      KQ=NPT(J) - K + 2
238      CALL LSQ(X2,TTOT2,AA,BB,SIGA,SIGB,KQ      ,IDEV,DEVM)
239 C
240 C      CALCULATE AVERAGE VELOCITY, INTERVAL VELOCITY AND ERRORS
241 C
242      VAV(J)=SQRT(WVEL*WVEL*COSTH**2/AA)
243      VAVM=SQRT(WVEL*WVEL/(AA+SIGA))
244      ERR=ABS(VAVM - VAV(J))
245      IF(ERR - 0.000001)55,55,54
246 54 ERRAV(J)=ERR
247 55 CONTINUE
248      VSAV=(AYY(J)*VAV(J) - AY1*WVEL)/(AYY(J) - AY1)
249      ESAVM=ABS(VSAV - (AYY(J)*VAVM - AY1*WVEL)/(AYY(J) - AY1))
250      VINT=(AYY(J)*VAV(J) - AYY(J-1)*VAV(J-1))/(AYY(J) - AYY(J-1))

```

```

251 C      VINIM=(Ayy(J)*VAVM-Ayy(J-1)*(VAV(J-1)-ERRAV(J-1)))/
252      VINIM=(Ayy(J)*VAVM-Ayy(J-1)*(VAV(J-1)+ERRAV(J-1)))/
253      1(Ayy(J) - Ayy(J-1))
254      ERRIN=ABS (VINIM - VINT)
255      WRITE(MI,42)AA,SIGA,BB,SIGB,IDEV,DEVM,VAV(J),ERRAV(J),VINT,ERRIN,
256      1VSAV,ESAVM
257      42 FORMAT(2X,F7.4," : ",F7.4,5X,F8.4," : ",F7.4,5X,I10,3X,F7.4,2X,
258      1F6.3," : ",F6.3,5X,F6.3," : ",F6.3,5X,F6.3," : ",F6.3)
259      CALL REORD(X2,TTOT2,IDEV,KQ)
260      800 CONTINUE
261      400 CONTINUE
262      WRITE(MI,33)
263      33 FORMAT("/" TRAVEL TIMES      DEPTH BELOW FISH (M)      DEPTH BELOW SE
264      *ABED (M) ")
265      DEPO=((R(1)/DPMS)*WVEL)/2.0
266      DO 920 K33=1,NINT
267      J33=(K33 - 1)*8 + 1
268      T1=R(J33)/DPMS
269      DEPl=(T1*VAV(K33))/2.0
270      ERDEP=DEPl*ERRAV(K33)/VAV(K33)
271      DEP2=DEPl - DEPO
272      WRITE(MI,35)T1,DEPl,ERDEP,DEP2
273      35 FORMAT(3X,F10.3,5X,F10.3," : ",F10.3,5X,F10.3)
274      920 CONTINUE
275      GO TO (90,99),IKID
276      99 CALL EXIT
277      END

```

nd of file

APPENDIX 4

Methods of handling 12-channel digital data.

Methods of handling the 12-channel digital field data and programmes associated with the construction of synthetic seismogrammes.

1. TRANSFER OF SEISMOGRAMME FILES FROM APPLE TO CIFER.

The transfer of the 12-channel digital seismogrammes from the 5" floppy discs written in the "Normanton" format of the Geological Survey of Canada (G.S.C.) system for archiving high resolution seismic data to the storage media of the Northumbrian Universities Multiple Access Computing Network (NUMAC) was a cumbersome procedure, and required the use of an intermediate machine - a "Cifer" microcomputer. Before transfer of data to the Cifer could begin, the files had to be transferred within the Apple II machine from the "Appledos" to the "CP/M" operating system, and, before that could be done, translation of the compact "Normanton" format into the normal "DOS" file format had to be arranged. This involved the adaptation of an Apple "Basic" programme of the G.S.C. designed to read numbers individually from the files of the Normanton discs. With considerable help from R. Jackson, the programme which follows was written to reconstruct files into a suitable format for transfer to the CP/M operating system of the Apple II. The programme is interactive and prompts for the insertion of source and target discs. The programme runs under "DOS" and is invoked by the command LOAD TRANSFER.

J LIST

```

100 TR = 0: PNT = 0: V = 0: SF = 0: S
    K = 0: GI = 0: HC = 0: LC = 0: T
    D = 0: TS = 0: DIM BN(12)
102 AD = 46989: POKE AD,166: POKE
    AD + 1,222: POKE AD + 2,70: POKE
    AD + 3,2: POKE AD + 4,167: REM
    LDX ERRCOD JMP ERR PRINT
105 ONERR GOTO 100
110 GOTO 370
120 REM *****
130 REM * TR = TRACE *
140 REM * 1<=TR<=12 *
150 REM * PNT= POINT IN TRACE *
160 REM * 1<=PNT<=1024 *
170 REM * V = READING AT PNT *
180 REM * 0<= V <= 1024 *
190 REM * BIASED AT 512 *
200 REM * SF = SHOT OFFSET *
210 REM * GI = GROUP INTERVAL *
220 REM * SK = # OF STACKS *
230 REM * HC = HIGH CUT *
240 REM * LC = LOW CUT *
250 REM * TD = TIME DELAY *
260 REM * TS = TIME SCALE *
270 REM * DR$= DIRECTION *
280 REM * HE$= HEADER *
290 REM * D$ CNTL-D *
300 REM * SL = INPUT SLOT *
320 REM * DV = INPUT DRIVE *
340 REM * A$ = GENERAL ANSWER *
360 REM *****
370 D$ = CHR$ (13) + CHR$ (4)
380 TEXT
390 DV = 1: SL = 6: POKE 251,DV: POKE
    252,SL
790 HOME : VTAB 12
800 PRINT "INSERT NIM/TO DISK IN
    " : INVERSE : PRINT "DRIVE"
    : NORMAL : PRINT " : DV: "
    : PRINT " : : INVERSE : PRINT
    "SLOT": NORMAL : PRINT " :
    SL: PRINT : PRINT "AND PRESS
    " : INVERSE : PRINT "RETURN
    " : NORMAL
810 GET A$: IF ASC (A$) < > 13
    GOTO 810
815 POKE 43624,DV: POKE 43626,SL
    : REM $AA68,$AA6A
820 HOME
830 ONERR GOTO 2000
840 CALL 768
845 IF PEEK (206) < > 160 GOTO
    900
850 HOME : VTAB 10: PRINT "NO FI
    LE ENTERED - PRESS ANY KEY
    : GOTO 840 GOTO 790

```

```

900 CALL 16911: REM $402F
905 ONERR GOTO 2100
910 GOTO 980
920 REM *****
930 REM * THIS CALL HAS MOVED *
940 REM * THE PARAMETERS FROM *
950 REM * $5000 TO THE BASIC *
960 REM * VARIABLE LOCATION *
970 REM *****
980 DR$ = CHR$ ( PEEK (20495)): RE
    DIRECTION AT $500F
990 HOME : PRINT
1000 INVERSE
1010 FOR I = 0 TO 7: HE$(I) = CHR$
    ( PEEK (20486 + I)): PRINT H
    E$(I): NEXT : NORMAL : REM
    HEADER
1011 REM *** NEW PROGRAM SECTIO
    N ***
1012 SPEED= 255
1013 ONERR GOTO 2200: PRINT
1014 PRINT "GIVE NAME OF OUTPUT
    FILE"
1016 INPUT F$
1041 PRINT D$:"OPEN ";F$:" ", D2,
    S6"
1042 PRINT D$:"WRITE ";F$: FOR I
    = 0 TO 7: PRINT HE$(I): NEXT
    I: PRINT : PRINT CHR$ (10):
1043 PRINT SF;" ";GI;" ";SK;" ";
    HC;" ";LC;" ";TD;" ";TS;" ":
    PRINT CHR$ (10)
1050 FOR TR = 1 TO 12
1051 CNT = 0
1060 FOR PNT = 1 TO 1024
1100 CALL 16384
1110 PRINT V;" ";
1111 CNT = CNT + 1
1112 IF CNT < 10 GOTO 1116
1113 CNT = 0
1114 PRINT : PRINT CHR$ (10):
1116 REM
1120 NEXT PNT
1130 PRINT : PRINT CHR$ (10):
1140 NEXT TR
1150 PRINT D$:"CLOSE ";F$
1900 END
2000 PRINT "ERROR 1": END
2100 PRINT "ERROR 2": END
2200 PRINT D$:"CLOSE ";F$
2210 PRINT "ERROR 3": END
3POKE 33,33

```

Once a seismogramme has been written into a regular "DOS" file on a new disc, it is moved again by use of the interactive utility programme APDOS which re-writes the file to another disc in a form suitable for further treatment under the CP/M operating system. The utility programme APDOS itself runs under the CP/M system. Of eight seismogrammes on a Normanton-format disc, only two can be held on one disc in this translated format.

Files are moved to a Cifer microcomputer by linking it with the Apple and running the appropriate utility "export" and "import" programmes, with both Apple and Cifer running under CP/M with the prompt "A>", the commands:

E filename (Apple)

I B:filename (Cifer)

will transfer the seismogramme to the disc in slot "B" of the Cifer. Such a disc will hold 4 seismogrammes. The total running time of Apple and Cifer for the complete transfer of one seismogramme is about 30 minutes.

2. TRANSFER FROM CIFER TO M.T.S.

After a disc had been written up in the Cifer system, it could be moved into the NUMAC network by another suitably connected Cifer microcomputer, by the use of

utility programmes. The necessary programmes were:

- (i) CLEARWAY which allows routing to the
 "NUNET" network of NUMAC, on
 receiving the command "N".
- (ii) TRANSFER which initiates connexion to
 the NUMAC system.
- (iii) CL18:TRANSFER which runs under the M.T.S.
 system of NUMAC and initiates
 the transfer of files.

In addition to the above utilities, there was sometimes a need for the configuration routine "&SDEF" if the two Cifer machines involved in the overall process of file transfer were running under different versions of their operating systems.

Transfer to the M.T.S. system could take a varying amount of time depending on how busy the system was, but this would usually be from 30 to 60 minutes per seismogramme.

REFORMATTING OF FILES IN THE M.T.S. SYSTEM.

Files transferred to the M.T.S. system were written and stored in the "free-format" style with numbers separated by commas. This was expanded into a fixed format

compatible with the output of the synthetic seismogramme programme, so that the same plotting programme could be used to handle both types of record. The re-formatting programme is given on the next page:

THIS PROGRAMME RE-FORMATS FILES SENT TO MTS IN FREE
 FORMAT. THE INPUT IS AS SET BY THE APPLE-CIFER-MTS
 TRANSFER PROGRAMMES. THE OUTPUT IS IN THE FORMAT :

16I7

BOTH INPUT AND OUTPUT FILES ARE HEADED BY THREE LINES OF
 TEXT/PARAMETERS OF WHICH THE LAST LINE IS BLANK

THE NUMBERS ARE INPUT AS INTEGERS (0 - 1024) and are
 output as INTEGERS IN THE RANGE
 (-9999 - +9999)

```

    DIMENSION IA(1024,12),J(10),JJ(4),AA(20)
    DO 350 IH=1,3
      READ(5,22) (AA(IU),IU=1,20)
      WRITE(6,22) (AA(IU),IU=1,20)
350  CONTINUE
    22 FORMAT(20A4)
      DO 200 K=1,12
        DO 100 I=1,103
          10  FORMAT(10I8)
            READ (5,10,END=999) (J(IC),IC=1,10)
999  CONTINUE
            DO 600 IQ=1,10
              2  FORMAT(I15)
                IR=10*(I-1) + IQ
                IA(IR,K)=IFIX(9999.0*(FLOAT(J(IQ)-512))/512.)
360  CONTINUE
370  CONTINUE
380  CONTINUE
390  CONTINUE
    C  NOW WRITE OUT ON '6' IN FIXED FORMAT
      DO 300 K=1,12
        DO 400 II=1,64
          IMX=II*16
          IMN=IMX-15
          WRITE(6,1) (IA(I,K),I=IMN,IMX)
460  1  FORMAT(16I7)
470  400 CONTINUE
480  300 CONTINUE
490  CALL EXIT
500  END

```

End of file

4. SEISMOGRAMME PLOTTING PROGRAMME.

The Fuchs-Mueller synthetic seismogramme programme had been adapted previously to use NUMAC plotting facilities, but there were some drawbacks with the output format for the present application. A new output programme was written. It incorporates the plots of the models and parameters as seen in the many seismogramme figures included with this work. The programme makes extensive use of calls to the "GHOST" graphical subroutine package. These utility plotting subroutines produce a plot-file which is relatively independent of the hard copy devices used to produce plots, so that the programme should be portable to systems which support the use of "GHOST". The input files are:

FILE NO. 1.

The source wavelet which had been used in the seismogramme computation. FORMAT (10F7.0). This file is terminated by "-9999."

FILE NO. 2.

Information about the seismogramme model as output on file no. 4 by the programme SYNSEI. This is essentially a repeat of the information input to SYNSEI for the construction of the synthetic seismogramme and is used by the plotting programme to produce plots of variations in the parameters modelled.

FILE NO. 5.

The seismogramme as output by SYNSEI (the Fuchs-Mueller programme), or else the field seismogramme output of the re-formatting programme already described.

FILE NO. 7.

Literal string data used by the plotting programme for titling purposes.

There is only one output file from this programme. It is the file containing the plotting instructions as produced by the "GHOST" package. This file can be turned into a hard-copy plot by using it as input for the NUMAC utility programme "*MTSPLOT".

Seismogramme plotting programme

INPUT FILES: 1 Source wavelet for seismogramme computation
 2 Information output on '4' by SEIS
 5 Seismogramme with header as output on '3' by SEIS
 7 String data (same for all plots) held in PLTIN(0)

OUTPUT FILE 9 Plot-file as produced by GHOST graphics package

DIMENSION XS(100),AA(20,3),B(20),X(1024),Y(1024),A(30),DEEP(30)
 DIMENSION DENTIF(10),D(30), BB(30),RHO(30),QAI(30),QBI(30),
 * NREV(30),XLAY(2),YLAY(2),YI(30),YI1(30)
 DATA MR,XMIN,XMAX,YMIN,YMAX /5,0.125926,1.196296,0.07799,.8030/
 DATA CHARBL, DMIN,DMAX/4H , -130.7,-21.5/
 DATA RMIN,RMAX,SMIN,SMAX/0.13,0.90,0.845,0.960/
 DIMENSION PVELO(3),SVELO(3),DENS(3),QP(3),QS(3)
 DIMENSION BSTR1(20),BSTR2(20)

READ(7,1001)(BSTR1(J),J=1,20)
 READ(7,1001)(BSTR2(J),J=1,20)

1001 FORMAT(20A4)

READ(7,1002)(PVELO(J),J=1,3)
 READ(7,1002)(SVELO(J),J=1,3)
 READ(7,1002)(DENS(J),J=1,3)
 READ(7,1002)(QP(J),J=1,3)
 READ(7,1002)(QS(J),J=1,3)

1002 FORMAT(3A4)

YSTEP=0.051995
 YINV=1.0
 DELT=204.8
 YOFF=1.666666667*YSTEP

10 CALL PAPER(1)

DO 100 I=1,2
 READ(MR,1)(AA(J,I),J=1,20)

1 FORMAT(20A4)

100 CONTINUE

READ(MR,2)NPTS
 IF(NPTS.NE.0) GO TO 808
 NPTS=1024
 YINV=-1.0

2 FORMAT(I10)

808 CALL PSPACE(XMIN,XMAX,YMIN,YMAX)

READ(2,403)MDECKS,MDECKR,NOL

403 FORMAT(40X,3I5)

INO= NOL

DO 404 IKK=1,INO

404 READ(2,1)(B(J),J=1,20)

READ(2,405)TMIN

TMIN = TMIN*1000.0

405 FORMAT(54X,F7.3)

TMAX=TMIN + DELT

```

51      REWIND 2
52      CALL MAP (TMIN,TMAX,DMIN,DMAX)
53      CALL AXES
54          CALL CTRMAG(15)
55          CALL CTRORI (0.0)
56      C      CALL THICK(3)
57          DO 101 I=1,20
58          B(I)=AA(I,1)
59      101  CONTINUE
60          CALL PLOTCS(TMIN+10.0,DMAX-6.5,B,80)
61      C      CALL THICK(0)
62          CALL MAP(0.0,0.0,0.0,0.0)
63          DO 300 I=1,NPTS
64          X(I)=FLOAT(I)
65      300  CONTINUE
66          DO 200 I=1,12
67          READ(MR,3,END=200) (Y(K),K=1,NPTS)
68          DO 601 KQ=1,NPTS
69          Y(KQ) = Y(KQ)*YINV
70      601  CONTINUE
71      3    FORMAT(16F7.0)
72          I1=I-2
73          YMAX=-(YSTEP*FLOAT(I1)+ YOFF) + 0.762597
74          YMIN=YMAX - YSTEP
75          CALL PSPACE(XMIN,XMAX,YMIN,YMAX)
76          XPTSS=FLOAT(NPTS)
77          CALL MAP(0.0,XPTSS,-9999.0,9999.0)
78          CALL PTPLOT(X,Y,1,NPTS,-2)
79      200  CONTINUE
80          IF (NPTS.EQ.1024) GO TO 999
81          ENDDATA=-9999.
82          DO 2510 J=1,256
83      2510  Y(J) =0.0
84          NAL= 32
85          BALMAX=0.
86          DO 2540 J=NAL,1015,10
87          JDC=J+9
88          READ(1,216) (XS(II),II=1,10)
89      216  FORMAT(10F7.0)
90          DO 2520 I=J,JDC
91          IJK=I-J+1
92          IF (XS(IJK)-ENDDATA) 2515,2547,2515
93      2515  Y(I) = XS(IJK)
94      2520  CONTINUE
95      2540  CONTINUE
96      2547  I=I-1
97          DO 2548 J=1,I
98          SMAX=ABS(Y(J))
99          IF (SMAX.GT.BALMAX) BALMAX=SMAX
100     2548  CONTINUE

```

```

101      DO 2549 J=1,I
102      FAC = 9999.0/BAIMAX
103 2549 Y(J) = Y(J)*FAC
104      NOUT=NPTS/2
105      XPTSS=FLOAT(NOUT)
106      YMAX=2.0*YSTEP - YOFF + 0.762597
107      YMIN=YMAX - YSTEP
108      DXX=(XMAX - XMIN)*0.50 + XMIN
109      CALL PSPACE(DXX,XMAX,YMIN,YMAX)
110      CALL MAP(0.0,XPTSS,-9999.0,+9999.0)
111      DO 303 I=1,NOUT
112      X(I) = FLOAT(I)
113 303 CONTINUE
114      CALL PTPLOT(X,Y,1,NOUT,-2)
115 C
116 C ***** PLOT OF MODEL *****
117 C
118      READ(2,121) (DENTIF(K),K=1,10),MDECKS,MDECKR,NOL
119 121 FORMAT(10A4,3I5)
120      DO 122 IK=1,MDECKS
121      READ(2,123)DEEP(IK)
122 123 FORMAT(F9.2)
123 122 CONTINUE
124      IQ=MDECKS + 1
125      NO2 = NOL - MDECKS
126      DO 124 IK=1,NO2
127 124 READ(2,125)D(IK),A(IK),BB(IK),RHO(IK),QAI(IK),QBI(IK)
128 125 FORMAT(F9.1,3X,2(F9.3,3X),F8.2,2X,2(F9.5,3X))
129 C
130 C
131      THICK=0.0
132      NOL = NO2 - 1
133      DO 130 IK=1,NOL
134      THICK=THICK + D(IK)
135      YI1(IK) = -THICK
136 130 CONTINUE
137 C
138 C Work out interface positions for plotting
139 C
140      DO 131 IK= 1,NOL
141 131 YI(IK) = 1.0 - (-YI1(IK)/THICK)
142 C
143 C Now proceed with plot.
144      CALL FRAME
145 C
146      CALL PSPACE(0.02,0.93,0.01,0.50)
147      CALL BORDER
148      CALL CTRORI(0.0)
149      CALL CTRMAG(12)
150      CALL PSPACE(0.05,0.90,0.40,0.480)

```

```

151      CALL MAP(0.0,1.0,-0.1,1.0)
152      CALL PLOTCS(0.10,.80,BSTR1,80)
153      CALL PLOTCS(0.10,.50,BSTR2,80)
154      XLAY(1) = 0.0
155      XLAY(2) = 1.0
156      YLAY(1) = 1.0
157      YLAY(2) = 1.0
158      CALL PTPLOT(XLAY,YLAY,1,2,-2)
159      C      YLAY(1) = 0.0
160      C      YLAY(2) = 0.0
161      C      CALL PTPLOT(XLAY,YLAY,1,2,-2)
162      C
163      C      Rewind 2, and read in layer parameters for plot.
164      C      plot parameters for water layer before call to new PSPACE.
165      REWIND 2
166      NQ=MDECKS + 2
167      C
168      C      The effect of the loop to 140 is to read through the file '2'
169      C      until the appropriate layer parameters are reached.
170      C
171      DO 140 IK=1,NQ
172      140  READ(2,1) (B(J),J=1,20)
173      READ(2,1) (B(J),J=1,20)
174      CALL PLOTCS(0.1,0.17,B,80)
175      C
176      C      Ready for new PSPACE and mapping to plot seabed layers.
177      C
178      CALL PSPACE(0.05,0.90,0.020,0.40)
179      FACT= (THICK - (-YI1(2)))/THICK
180      CALL MAP(0.0,1.0,-0.04,FACT)
181      C
182      C      As the water layer has already been done loop starts at IK=2.
183      DO 133 IK = 2, NO1
184      READ(2,1) (B(J),J=1,20)
185      YLAY(1) = YI(IK)
186      YLAY(2) = YI(IK)
187      YP= YI(IK) - 0.025
188      IF( IK.NE.NO1 ) YP=(YI(IK) + 3.0*YI(IK+1))/4.0
189      IF( IK.EQ.NO1 ) B(1)=CHARBL
190      IF( IK.EQ.NO1 ) B(2)=CHARBL
191      IF( IK.EQ.NO1 ) B(3)=CHARBL
192      CALL PLOTCS(0.10,YP,B,80)
193      CALL PTPLOT(XLAY,YLAY,1,2,-2)
194      133  CONTINUE
195      C
196      C      Plot out other information relating to synthetic seismogramme
197      C
198      CALL PSPACE(0.98,1.00,0.05,0.98)
199      CALL MAP(0.0,1.0,0.0,1.0)
200      CALL CTRORI(1.0)

```



```

201      READ(2,1) (B(J),J=1,20)
202      CALL PLOTCS(0.80,0.02,B,80)
203      C
204      C      Move above to plot parameter variations
205      C
206      C      CALL FRAME
207      CALL CTRORI(0.0)
208      CALL CTRMAG(12)
209      C
210      C      Plot p-velocity, s-velocity, density, 1/Qp, 1/Qs each
211      C      in its own small block
212      C
213      CALL PSPACE(0.02,0.56,0.53,1.01)
214      CALL BORDER
215      Y1=-1.03*THICK
216      CALL PSPACE(0.09,0.27,0.77,0.99)
217      VPMIN=0.9*AMIN(A,NO2)
218      VPMAX=1.1*AMAX(A,NO2)
219      DX = FLOAT(IFIX(10.*((VPMAX-VPMIN)/3.0)))/10.0
220      C      This value for dx is derived to give reasonable axis annotation.
221      CALL MAP(VPMIN,VPMAX,Y1,0.05*THICK)
222      CALL AXESSI(DX,10.0)
223      CALL POSITN(VPMIN,0.0)
224      CALL JOIN(A(1),0.0)
225      DO 606 K=1 , NO1
226          CALL JOIN(A(K),Y1(K))
227          CALL JOIN(A(K+1),Y1(K))
228      606  CONTINUE
229      CALL JOIN(A(NO2),Y1)
230      CALL UNDLIN(1)
231      XPL = VPMIN + (VPMAX - VPMIN)*0.3
232      YPL = -THICK*0.35
233      CALL PLOTCS(XPL,YPL,PVELO,10)
234      CALL UNDLIN(0)
235      C
236      C      Plot s-velocity
237      C
238      CALL PSPACE(0.36,0.54,0.79,0.99)
239      VPMIN=0.9*AMIN(BB,NO2)
240      VPMAX=1.1*AMAX(BB,NO2)
241      DX=FLOAT(IFIX(10.*((VPMAX-VPMIN)/3.0)))/10.0
242      CALL MAP(VPMIN,VPMAX,Y1,0.05*THICK)
243      C      CALL CTRSIZ(0.035)
244      CALL AXESSI(DX,10.0)
245      CALL POSITN(VPMIN,0.0)
246      CALL JOIN(BB(1),0.0)
247      DO 616 K=1 , NO1
248          CALL JOIN(BB(K),Y1(K))
249          CALL JOIN(BB(K+1),Y1(K))
250      616  CONTINUE

```

```

251      CALL JOIN(BB(NO2),Y1)
252      CALL UNDLIN(1)
253      XPL = VPMIN + (VPMAX - VPMIN)*0.3
254      YPL = -THICK*0.35
255      CALL PLOTCS(XPL,YPL,SVELO,10)
256      CALL UNDLIN(0)
257  C
258  C      Plot density
259  C
260      CALL PSPACE(0.09,0.27,0.55,0.75)
261      VPMIN=0.9*AMIN(RHO,NO2)
262      VPMAX=1.1*AMAX(RHO,NO2)
263      DX=FLOAT(IFIX(10.*((VPMAX-VPMIN)/3.0)))/10.0
264      CALL MAP(VPMIN,VPMAX,Y1,0.05*THICK)
265      CALL AXESSI(DX,10.0)
266      CALL POSITN(VPMIN,0.0)
267      CALL JOIN(RHO(1),0.0)
268      DO 626 K=1 , NO1
269          CALL JOIN(RHO(K),Y1(K))
270          CALL JOIN(RHO(K+1),Y1(K))
271  626  CONTINUE
272      CALL JOIN(RHO(NO2),Y1)
273      CALL UNDLIN(1)
274      XPL = VPMIN + (VPMAX - VPMIN)*0.3
275      YPL = -THICK*0.35
276      CALL PLOTCS(XPL,YPL,DENS,7)
277      CALL UNDLIN(0)
278  C
279  C      Move to lower position in frame and plot 1/Qp, 1/Qs
280  C
281      CALL PSPACE(0.36,0.54,0.55,0.75)
282      VPMIN=0.9*AMIN(QAI,NO2)
283      IF (VPMIN-0.00001) 2777,2778,2778
284  2777 VPMIN=0.00001
285  2778 CONTINUE
286      VPMAX=1.1*AMAX(QAI,NO2)
287      CALL MAPXL(VPMIN,VPMAX,Y1,0.05*THICK)
288      VPMIN=1.01*VPMIN
289  C      Insert loop to ensure no Qp is too small.
290      DO 1123 KQ=1, NO1
291          IF (QAI(KQ)-0.00001) 3777,3888,3888
292  3777 QAI(KQ)=VPMIN
293  3888 CONTINUE
294  1123 CONTINUE
295      CALL AXEXLI(10.0)
296      CALL POSITN(VPMIN,0.0)
297      CALL JOIN(VPMIN,Y1)
298      CALL JOIN(VPMIN,0.0)
299      CALL JOIN(VPMIN,Y1(1))
300      CALL JOIN(QAI(2),Y1(1))

```

```

301      DO 636 K= 2 , NO1
302          CALL JOIN(QAI(K),YI1(K))
303          CALL JOIN(QAI(K+1),YI1(K))
304      636  CONTINUE
305          CALL JOIN(QAI(NO2),Y1)
306          CALL UNDLIN(1)
307          CALL MAP(0.0,1.0,0.0,1.0)
308          XPL = 0.4
309          YPL = 0.62
310          CALL PLOTCS(XPL,YPL,QP,4)
311          CALL UNDLIN(0)
312      C
313      C
314      999  CALL GREND
315          STOP
316          END
317

```

End of file

5. GRADIENT PROGRAMME.

As a help in preparing an input file for the running of the systhetic seismogramme programme "SYNSEI", a programme was written to take a basic layered model and to divide one or more layers into sub-layers for the modelling of gradients. This programme is run interactively from a NUMAC terminal. The file to be input should be a listing of the basic layered-model file, because the gradient programme identifies the line to be set as a gradient by its line number. The programme prompts for:

- LINENO. The number of the line which is to be set
 as a gradient (Integer I10).
- NODIV. The number of subdivisions into which the
 layer is to be divided (Integer I10).
- PMMPS. The number of m/sec. between the minimum
 and maximum velocity required. This may
 be either positive or negative (Real F10.0).
- I1 Set=1 if s-velocity is to be varied
 proportionally with the p-velocity.
 Otherwise set=0 or blank. (Integer I5)
- I2 Set=1 if density is to be varied
 proportionally with the p-velocity.
 Otherwise set=0 or blank. (Integer I5)
- I3 Set=1 if the attenuation is to be varied
 proportionally with the p-velocity.
 Otherwise set=0 or blank. (Integer I5)

The output file is in the appropriate format for input to "SYNSEI", but will often require further editing when gradients are desired which are more complex than the simple linear gradient output by this programme.

```

1      C      File to introduce gradients into specified layers
2      C      The file to be altered is read from MR (=5)
3      C      The file written into is MI (=6)
4      C      The alteration parameters are read from the keyboard MT (=1)
5      C      Prompting and formatting is given by the screen MY (=2)
6      C      Specify : Line number of file to be given gradient
7      C      Number of layers into which it must be divided
8      C      Number of metres per sec. for half-range of
9      C      variation
10     C      I1,I2,I3 : set = 1 for similar gradt. in
11     C      Svel,Rho,Qp (Qs is not altered)
12     C
13     C
14     DIMENSION TEX(40)
15     DATA MR,MI,MT,MY/5,6,1,2/
16     1      WRITE(MY,2)
17     2      FORMAT(' LINENO. | NODIV. | PMMPS | I1 | I2 | I3 |')
18     READ(MT,3)LIN,NODIV,PMMPS,I1,I2,I3
19     PMMPS=PMMPS/1000.0
20     3      FORMAT(2I10,F10.5,3I5)
21     GO TO 91
22     150     WRITE(MY,151)
23     151     FORMAT(' LINENO. |')
24     READ(MT,3)LIN
25     91      WRITE(MY,61)LIN
26     61      FORMAT(' LINE NUMBER BEING SET AS GRADIENT ',I5)
27     IF (LINE-LIN+1) 20,21,20
28     20      READ(MR,11,END=999,ERR=999)LINE,(TEX(J),J=1,40)
29     WRITE(MY,98)LINE
30     98      FORMAT(' line read = ',I6)
31     WRITE(MI,12) (TEX(J),J=1,40)
32     11      FORMAT(I7,5X,40A4)
33     12      FORMAT(40A4)
34     IF (LINE-LIN+1) 20,21,20
35     21      READ(MR,14)LINE,THK,PVEL,SVEL,RHO,NREV,QP,QS
36     WRITE(MT,19)THK,PVEL,SVEL,RHO,NREV,QP,QS
37     14      FORMAT(I7,5X,4F10.0,I10,2F10.0)
38     19      FORMAT(' LINE ALTERING ',I2X,4F10.5,I10,2F10.6)
39     VMIN=PVEL-PMPS
40     VEL=VMIN
41     TST=(THK/PVEL)/FLOAT(NODIV)
42     SVELN=SVEL
43     RHON=RHO
44     QPN=QP
45     DPMPS=2.0*PMMPS/FLOAT(NODIV)
46     VEL=VEL - 0.5*DPMPS
47     DO 222 K=1,NODIV
48     VEL=VEL + DPMPS

```

```

49      THKS=TST*VEL
50      IF (I1.EQ.1) SVELN=SVEL*(VEL/PVEL)
51      IF (I2.EQ.1) RHON= RHO*(VEL/PVEL)
52      IF (I3.EQ.1) QPN= QP*(VEL/PVEL)
53      QSN=QS
54      WRITE(MI,15)THKS,VEL,SVELN,RHON,NREV,QPN,QSN
55      15  FORMAT(4F10.5,I10,2F10.6)
56      222 CONTINUE
57      WRITE(MY,8)
58      8   FORMAT(' DO YOU WISH TO CHANGE MORE LINES ?',/,
59      */, 'TYPE "1" TO CHANGE ONLY LINENO. "2" FOR ALL PARAMS.')
60      READ(MT,9)IQ
61      9   FORMAT(I1)
62      IF (IQ-1)27,150,27
63      27  IF (IQ-2)20,1,20
64      999 STOP
65      END
66

```

End of file

6. STRUCTURE OF FILES FOR RUNNING THE SYNTHETIC
SEISMOGRAMME PROGRAMME.

An example is given of a "batch" job prepared for running in the "M.T.S." operating system on the IBM 4341/1 computer of the NUMAC network. Batch jobs prepared in this way can only start to run immediately on entry to the machine when the requested running time does not exceed 300sec. of C.P.U. time. Longer runs are "held" until the system judges itself sufficiently free of other tasks to handle them. There is consequently a need to restrict the running time of seismogramme construction if the modelling process is to be in any way "interactive". Even at very quiet periods, the fastest actual running time for a "300 sec." job is about 10 minutes. Entering it and plotting it on a screen may take a further 5 minutes. Consequently a 15 minute turnaround is the minimum time between thinking up a simple variation on a model and seeing the resultant seismogramme. Where precise comparison with a field record is required, the seismogramme must be plotted on paper. This occupies a further 15 minutes. Usually this time would be longer, at busy periods, up to 10 times as long. An example of the structure of a "batch job" follows:

Example of a batch job. Files -CSEISQ & -CSEIS2Q contain the Fuchs-Mueller modelling programme.

```

$SIG GPM2 T=300
$COP WIGGLE -WIG
$RUN -CSEISQ+-CSEIS2Q+*GHOST 1=-A 2=*DUMMY* 3=-L 4=-R 5=*SOURCE* 6=-B
7=-G 9=-P 10=-C 11=-X 12=-D 15=-E 99=-Pl
00041000000101116110000 1 AS 300 THIN HV LAY. P=1.50 ARC301 NO.496 2 2
0.00010 1.460-0.0100000 1.027 3 0.000001 0.000002
0.02660 1.460 0.0000001 1.027 3 0.000001 0.000002
0.00010 1.460-0.0100000 1.027 3 0.000001 0.000002
0.02660 1.46000 0.0000001 1.02700 3 0.000001 0.000002
0.00100 1.48000 0.010 1.60 3 0.001000 0.010000
0.00030 1.60250 0.010 2.00 3 0.001000 0.010000
0.00032 1.68750 0.100 2.00 3 0.001000 0.010000
0.00036 1.86250 0.250 2.00 3 0.001000 0.010000
0.00039 2.03749 0.400 2.00 3 0.001000 0.010000
0.01645 2.21249 0.500 2.00 3 0.001000 0.010000
0.00040 2.05249 0.400 2.00 3 0.001000 0.010000
0.00040 1.90249 0.270 2.00 3 0.001000 0.010000
0.00040 1.75249 0.100 2.00 3 0.001000 0.010000
0.00200 1.55000 0.020 2.00 3 0.001000 0.010000
0.00040 1.75249 0.150 2.00 3 0.001000 0.010000
0.00040 1.90249 0.300 2.00 3 0.001000 0.010000
0.00040 2.05249 0.600 2.00 3 0.001000 0.010000
0.00025 2.65249 1.100 2.00 3 0.001000 0.010000
0.00250 2.80000 1.200 2.00 3 0.001000 0.010000
0.00160 2.40000 0.9 2.00 3 0.001000 0.030000
0.00125 2.15000 0.5 2.00 3 0.001000 0.030000
0.00100 1.80000 0.25 2.00 3 0.001000 0.030000
0.04175 1.50000 0.001 2.00 3 0.001000 0.030000
0.50000 3.500 2.0000000 1.98 3 0.001000 0.030000 0.0

.03658 .12274 .007833
100000000. 0.010
0.600 3 30.00 2.00 200.00 250.00 500.00 500.00
0.00040000 512 0 1
00. 468. 859. 2577. 2655. -3437. -6561. -4687. -703. 1640.
2734. 2890. 2890. 1874. 1171. 1015. 390. -546. -1249. -1327.
-1406. -1171. -703. -234. -9999.
$COM
$COM
$RUN CPLT2N+-CSEIS2Q(329,352)+*GHOST 1=-WIG 2=-R 5=-L 7=PLT 9=-QLT1 SPRINT=*DUMMY*
$RUN *MTSPLOT SCARDS=-QLT1 SERCOM=*DUMMY* PAR=SITE=NEWCPLOT,SCALE=0.425
$SIG

```

Example of input to plotting programme (model)

```

AS 300 THIN HV LAY. P=1.50 ARC301 NO.496      2      2      24
0.1      1.460      0.0      1.03      0.00000      0.00000
26.6      1.460      0.000      1.03      0.00000      0.00000
0.1      1.460      0.0      1.03      0.00000      0.00000
26.6      1.460      0.000      1.03      0.00000      0.00000
1.0      1.480      0.010      1.60      0.00100      0.01000
0.3      1.602      0.010      2.00      0.00100      0.01000
0.3      1.688      0.100      2.00      0.00100      0.01000
0.4      1.863      0.250      2.00      0.00100      0.01000
0.4      2.037      0.400      2.00      0.00100      0.01000
16.4      2.212      0.500      2.00      0.00100      0.01000
0.4      2.052      0.400      2.00      0.00100      0.01000
0.4      1.902      0.270      2.00      0.00100      0.01000
0.4      1.752      0.100      2.00      0.00100      0.01000
2.0      1.550      0.020      2.00      0.00100      0.01000
0.4      1.752      0.150      2.00      0.00100      0.01000
0.4      1.902      0.300      2.00      0.00100      0.01000
0.4      2.052      0.600      2.00      0.00100      0.01000
0.3      2.652      1.100      2.00      0.00100      0.01000
2.5      2.800      1.200      2.00      0.00100      0.01000
1.6      2.400      0.900      2.00      0.00100      0.03000
1.2      2.150      0.500      2.00      0.00100      0.03000
1.0      1.800      0.250      2.00      0.00100      0.03000
41.7      1.500      0.001      2.00      0.00100      0.03000
500.0      3.500      2.000      1.98      0.00100      0.03000
0.60 3 30.0 2.0 200. 250. 500. 500. 0.1E+09 0.0100 0.00040000 512

```

String data as read by plotting programme

THICKNESS (M.)	P-VELOCITY (M/S.)	S-VELOCITY (M/S.)	DENSITY (G/ML)	1/Q P	1/Q S
P-VELOCITY					
S-VELOCITY					
DENSITY					
1/QP					
1/QS					

Source signal (terminated by -9999.)

```

00. 468. 859. 2577. 2655. -3437. -6561. -4687. -703. 1640.
2734. 2890. 2890. 1874. 1171. 1015. 390. -546. -1249. -1327.
-1406. -1171. -703. -234. -9999.

```

