SEISMIC REFRACTION MEASUREMENTS IN BRANSFIELD STRAIT

By M. J. G. Cox

ABSTRACT. Five seismic refraction lines were surveyed in the north-east part of Bransfield Strait during January 1962. A description of the equipment used and the survey technique is given, together with a description of the corrections applied to the data before the time-distance graphs were plotted. The cross-sections calculated from the corrected time-distance graphs are discussed. The area is thought to be part of a transition zone between continental and oceanic structures. At the north-east extremity of the area of survey a high-velocity rock (about 7·0 km./sec.) occurs at a depth of approximately 0·5 km. This is correlated with the dunite-serpentine which occurs on Gibbs Island 8 km. from the end of the line.

FIVE seismic refraction lines, each about 15 miles* (27·8 km.) in length were surveyed in Bransfield Strait during January 1962, as shown in Fig. 1. Two ships were used for the survey, H.M.S. *Protector* acting as the firing ship and R.R.S. *Shackleton* as the receiving ship. The survey was carried out to determine the broad structure of the Earth's crust in this locality, and to act as a control for the interpretation of the gravity and magnetic data, obtained during and following the 1959–60 season (Griffiths and others, 1964).

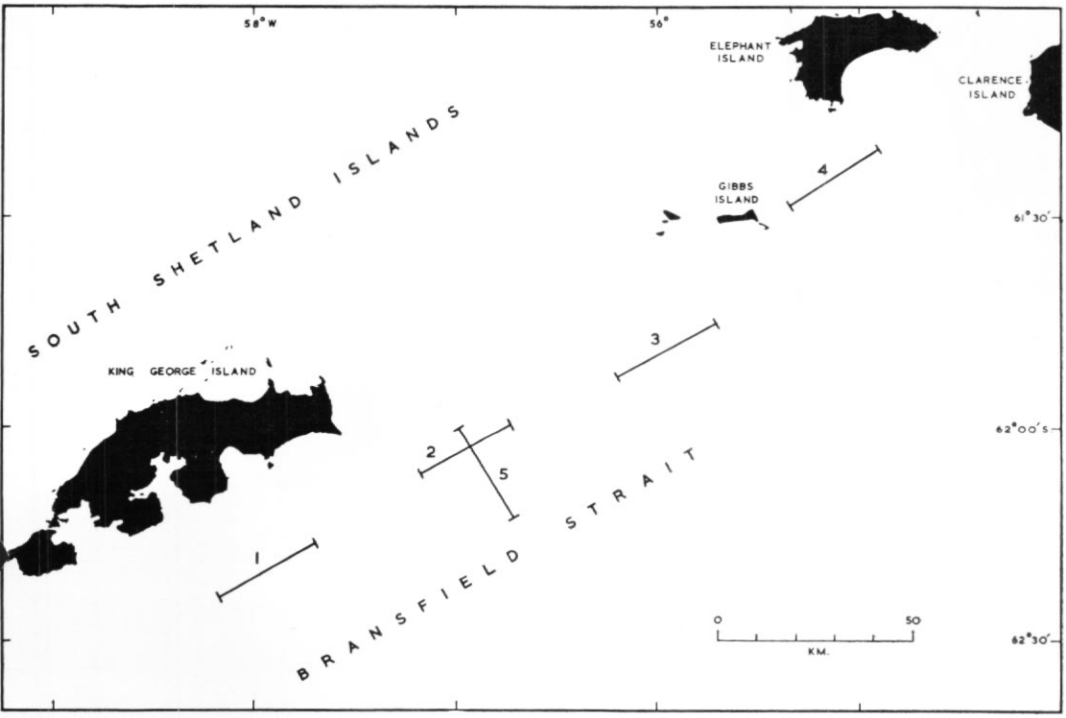

Fig. 1. Map of the north-east part of Bransfield Strait showing the seismic refraction lines surveyed in the 1961–62 season.

The main geological facts relevant to the present survey are summarized below. The geology of King George Island has been described by Hawkes (1961), where the main feature is a central core composed of Jurassic volcanics and rocks of the Andean Intrusive Suite, flanked to the north and to the south by Tertiary and recent volcanic rocks. Fragments of gneiss and quartzite of the Basement Complex have been observed, but none of these have

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* Throughout this report the "mile" unit refers to the "nautical mile", i.e. 1·852 km.

been *in situ*. The major fault pattern lies along the length of the island with the upthrow side to the north. The rocks of Gibbs Island and Elephant Island have been described by Tilley (1930) and Tyrrell (1945). Elephant Island and the western half of Gibbs Island are composed almost entirely of schists. Dunite-serpentine is widespread on the eastern half of Gibbs Island and appears to have been intruded parallel to the foliation planes of the schist (Tyrrell, 1945).

Previous geophysical work in the area has been described by Griffiths and others (1964). The relevant features of the Bouguer anomaly map are a sharp elongated high of about +130 mgal centred over the South Shetland Islands, a single value of +182 mgal on Gibbs Island, values of +154 mgal close to the south coast and +106 mgal on the north coast of Elephant Island. The high on Gibbs Island must be associated with the high-density dunite-serpentine which has a density of about 3·3 g./cm.^{3} (Adams, 1951, p. 74). The main magnetic feature of the area is an elongated high of about 800 gamma situated to the north of the South Shetland Islands and a wider shallower trough in Bransfield Strait (Griffiths and others, 1964, p. 38).

INSTRUMENTATION AND SURVEY TECHNIQUE

The equipment used for the survey was a simplified version of that commonly used in marine refraction work (Officer and others, 1959). The recording equipment was installed in R.R.S. *Shackleton*, with the hydrophones suspended away from the ship. Four hydrophones were suspended from a positively buoyant cable at intervals of 200 ft. (61 m.). The hydrophones possessed slight positive buoyancy and were arranged to float at a depth of about 150 ft. (46 m.) below surface level, supporting a small loop of the negatively buoyant hydrophone lead. A fifth hydrophone was suspended on a separate cable close to the ship and, in addition, a geophone was attached to the bottom of the ship. Each detector output was amplified using a transistor amplifier, designed by H. A. D. Cameron in the Department of Geology, University of Birmingham, and recorded on a conventional galvanometer camera. Five of the amplifying channels were designed to have a low frequency response (about 5–100 c./sec. with a flat response in the range 5–20 c./sec.) to detect the ground wave, which varied in frequency from 8 to 15 c./sec. The sixth channel had a higher frequency response (about 20–300 c./sec.) and was used, in conjunction with the separate hydrophone, for recording the water-wave arrivals. It was found, however, that all the channels recorded the water-wave adequately. One of the seismic records is shown in Fig. 2.

In order to reduce the background noise, all non-essential ship's machinery was stopped during recording. Drift of the ship produced noise due to the dragging of the hydrophones through the water near surface level. To minimize this noise about 100 ft. (30·5 m.) of the buoyant hydrophone cable and the separate hydrophone were brought inboard and then released approximately 20 sec. before firing the shot, thus allowing all hydrophones time to sink to their correct depth and minimizing their mechanical coupling to the ship. On several occasions, when the drift rate was exceptionally high, the cables were released 10 to 15 sec. before firing the shot.

Charges, varying in weight from 50 to 200 lb. (22·6 to 91 kg.), were made up from naval demolition charges of T.N.T. and fired electrically from H.M.S. *Protector*. A firing cable was streamed behind the ship initially with both the charge and the firing mechanism on the ship's after deck. The charge was dropped into the sea 30 sec. before firing so that it had sunk to a depth of about 200 ft. (61 m.) approximately 500 ft. (152 m.) astern of the ship when detonated. When the largest charges were used, for the longer-range shots, a longer loop of cable was streamed behind the ship and the charge was released 40 sec. before firing the shot. The shot instant was relayed from H.M.S. *Protector* to R.R.S. *Shackleton* by radio and the depth of water below each shot point was recorded on the ship's echo sounder.

Small moored buoys carrying radar targets were used on a number of the lines to facilitate the calculation of the position and drift of the receiving ship during the course of a firing sequence. On the occasions when buoys were not used the position of the ship was fixed using radar bearings and distances from neighbouring islands. During a firing sequence the receiving ship was re-positioned when its distance from the end of the line became greater than about 0·5 miles (0·93 km.). On a calm day no re-positioning was necessary.

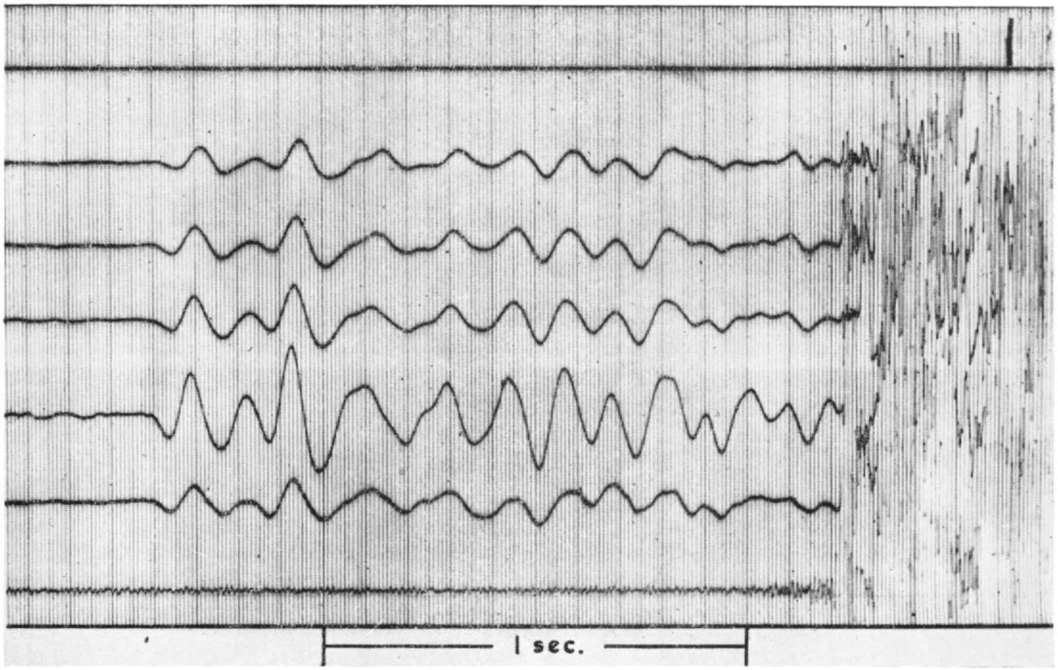

Fig. 2. An example of a seismic record showing the ground-wave and water-wave arrivals.

Wherever possible the distance between the two ships was calculated from the water-wave arrival times. Since on several occasions the shot instant was not recorded, the radar distance between the two ships was noted for each shot. When no shot instant was recorded, its position on the record was calculated from the radar distance and the water-wave arrival. The records were developed immediately after recording and an approximate running travel-time graph was plotted. Changes in charge size and the firing schedule were then made as directed by the quality of the records and the travel-time plot. A typical line involved firing shots at ranges of 8, 6, 4, 5, 7, 10, 12, and 15 miles (14·8, 11·1, 7·4, 9·3, 13·0, 18·5, 22·2, and 27·8 km.). In this way there was a distance of at least 2 miles (3·7 km.) and a minimum time interval of 12 min. between each shot fired; the firing ship could proceed with constant speed and direction, apart from the 180° turn necessary between the 4 and 5 mile (7·4 and 9·3 km.) shots. The variations in the sea-bottom topography along the line were recorded on R.R.S. *Shackleton*'s echo sounder whilst steaming from one end of the line to the other.

INTERPRETATION

Before the time-distance graph is plotted the refraction arrival times must be corrected for variations brought about by differences in the depth of water beneath the different shot points using either a constant depth or a sloping interface as a reference datum (Officer and others, 1959, p. 29). The depth corrections applied to the original data were all within the range +0·5 to −0·5 sec. In reading the record emphasis is always placed on the beginning of the first arrival. Later arrivals, due to the refraction of the seismic waves at another interface, are often difficult to pick since they may overlap the wave-trains associated with earlier arrivals.

Explosions at moderate depths in water have one or more pulses associated with them; these are caused by oscillations of the gas bubble generated by the explosion. They have the appearance of later shots on the record and, in some cases, may produce arrivals of amplitude comparable to that of the original explosion. In this case the first arrival seen on the record may have been produced by a bubble pulse rather than by the original explosion. The time interval between the instant of detonation and the first bubble depends on the charge size, the type of explosive used and the depth at which detonation takes place. The average time interval for line 1 was 0·33 sec.

The fitting of the refraction lines was carried out using a least squares analysis where there were sufficient points on the line to justify the use of this method. The reverse times, or travel times for the complete line in the two opposing directions, are theoretically the same by the principle of reversibility providing the arrivals have been obtained from the same layer. Small discrepancies in the reverse times can easily be accounted for by the lines shot in the two directions not being coincident. From the time-distance graph the up-dip and down-dip velocities and the two intercept times (where the distance is zero) can be read off and used to calculate the true velocity of the layer and its depth. This calculation assumes a constant velocity and dip over the segment of the interface from which the arrivals have been refracted (Heiland, 1946; Jakosky, 1960).

Apart from line 4, no evidence was obtained from first refraction arrivals of any unconsolidated or semi-consolidated sedimentary velocities, so that all the data for these layers have been obtained from second arrival times. If only the first arrivals had been considered, the thicknesses and dips of the crustal layers would have been somewhat less. The time-distance graphs together with the corresponding calculated cross-sections for lines 1 to 5 are shown in Figs. 3–7. The main features of each line are discussed below.

Line 1

The time-distance graph for line 1 is shown in Fig. 3. No topographic corrections were applied to the original data as the total variation in depth along the line was less than 10 fathoms (18·3 m.). The first layer, with a true velocity of 1·9 km./sec., is well defined and the intercept times result in a depth comparable with the sea depth. The second layer, which has an apparent velocity of 2·7 km./sec. when shooting towards the south-west, was not detected on the line shot towards the north-east, but there is additional evidence for its presence since a similar velocity was measured on lines 2, 3 and 5. In the interpretation it was assumed that this layer had a horizontal upper surface. The next layer obtained from the time-distance graph has a true velocity of 3·3 km./sec. This is found to be a very thin layer, about 200 m. thick, but the second arrivals from it are very marked. The crustal layers, with true velocities of 5·0 and 7·1 km./sec., are well defined and the reverse times for each layer are in good agreement with each other.

Line 2

Topographic corrections were applied for a level base line at a depth of 500 fathoms (0·92 km.). The corrected time-distance graph for this line is shown in Fig. 4. The top two layers are fairly well defined, although the reverse times for the lower of these two layers differ appreciably. The next layer, with an apparent velocity of 4·3 km./sec., was only detected on the line shot towards the north-east, but there is further evidence for this layer from line 5, which was shot perpendicular to line 2 (Fig. 1). The next layer, with a true velocity of 6·1 km./sec., is well defined. The apparent velocity of 8·6 km./sec. obtained on the line shot towards the south-west could be an up-dip velocity for either the 6·1 km./sec. layer or for a deeper layer. As a true velocity of 7·1 km./sec. was obtained on line 1, the apparent velocity of 8·6 km./sec. was thought to be better interpreted as an up-dip velocity of this layer.

Line 3

Topographic corrections were applied for a horizontal base line at a depth of 700 fathoms (1·28 km.). The corrected time-distance graph for this line is shown in Fig. 5. The top two layers, with apparent velocities of 2·0 and 2·7 km./sec., are poorly defined and were only detected shooting towards the north-east and south-west respectively, but their existence has been included in the interpretation, where their true velocities have been assumed to be identical with their apparent velocities. The crustal layers with true velocities of 4·2 and 5·9 km./sec. are well defined and the reverse times agree well for each layer.

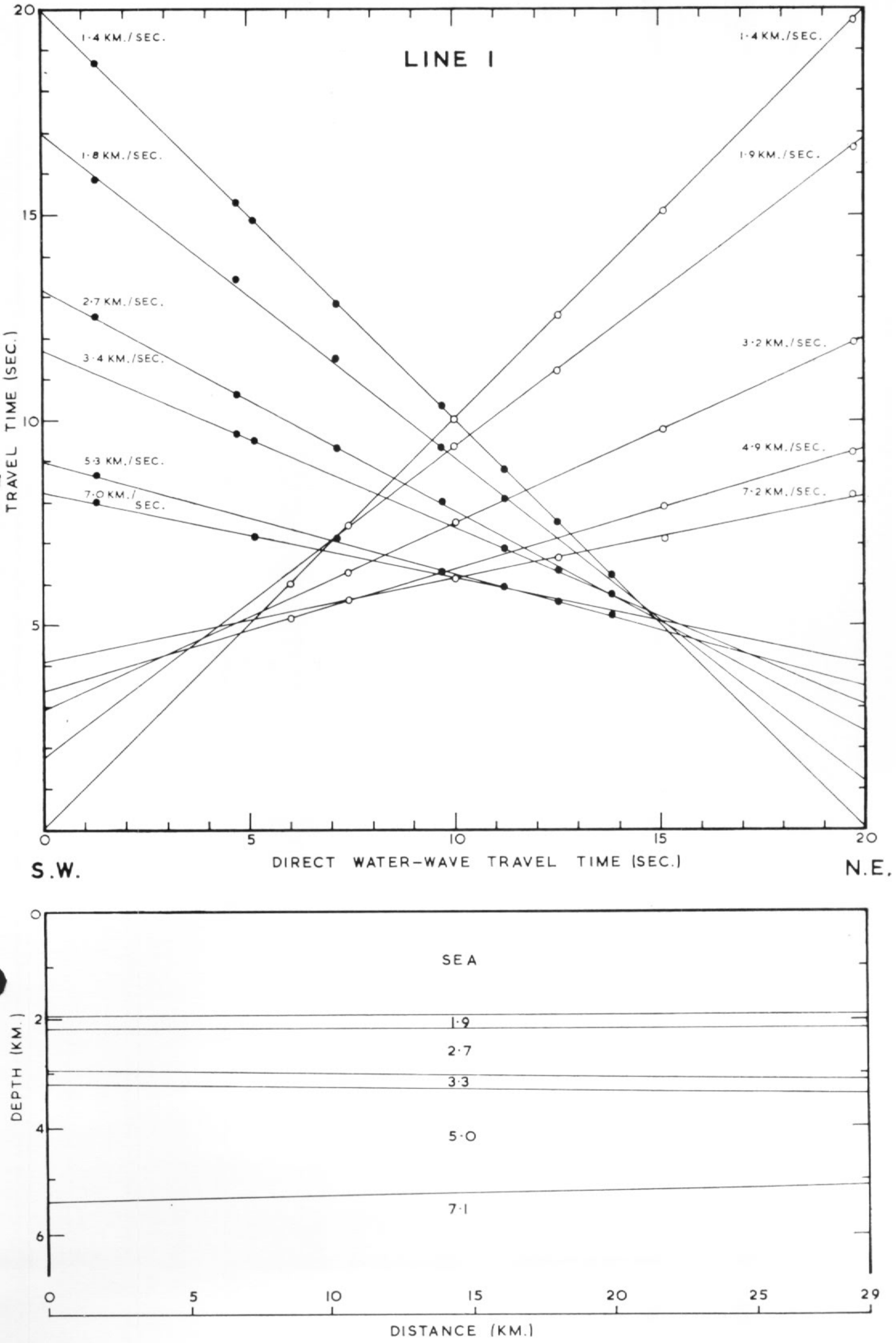

Fig. 3. Time-distance graph and calculated cross-section for line 1.

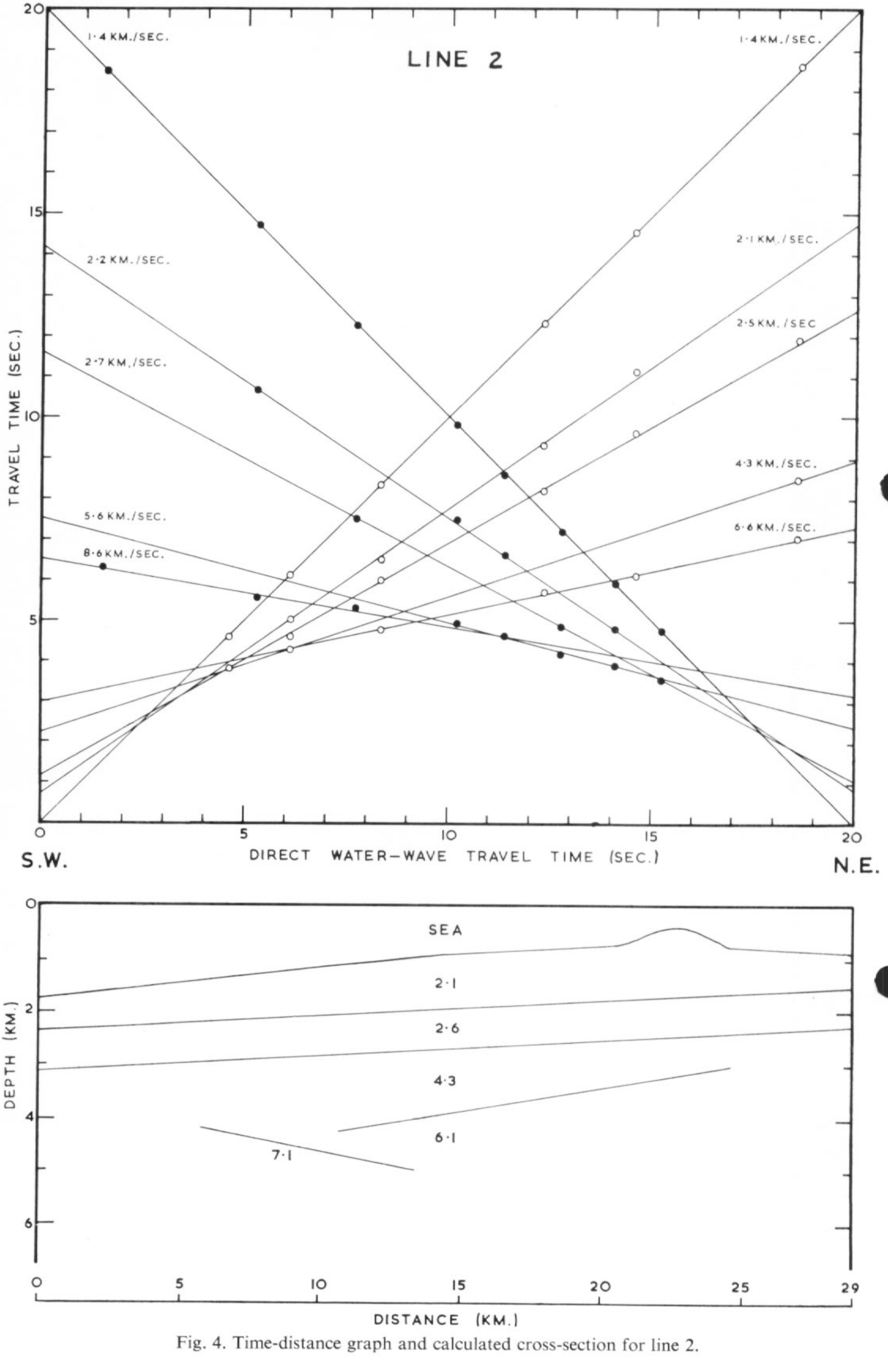

Fig. 4. Time-distance graph and calculated cross-section for line 2.

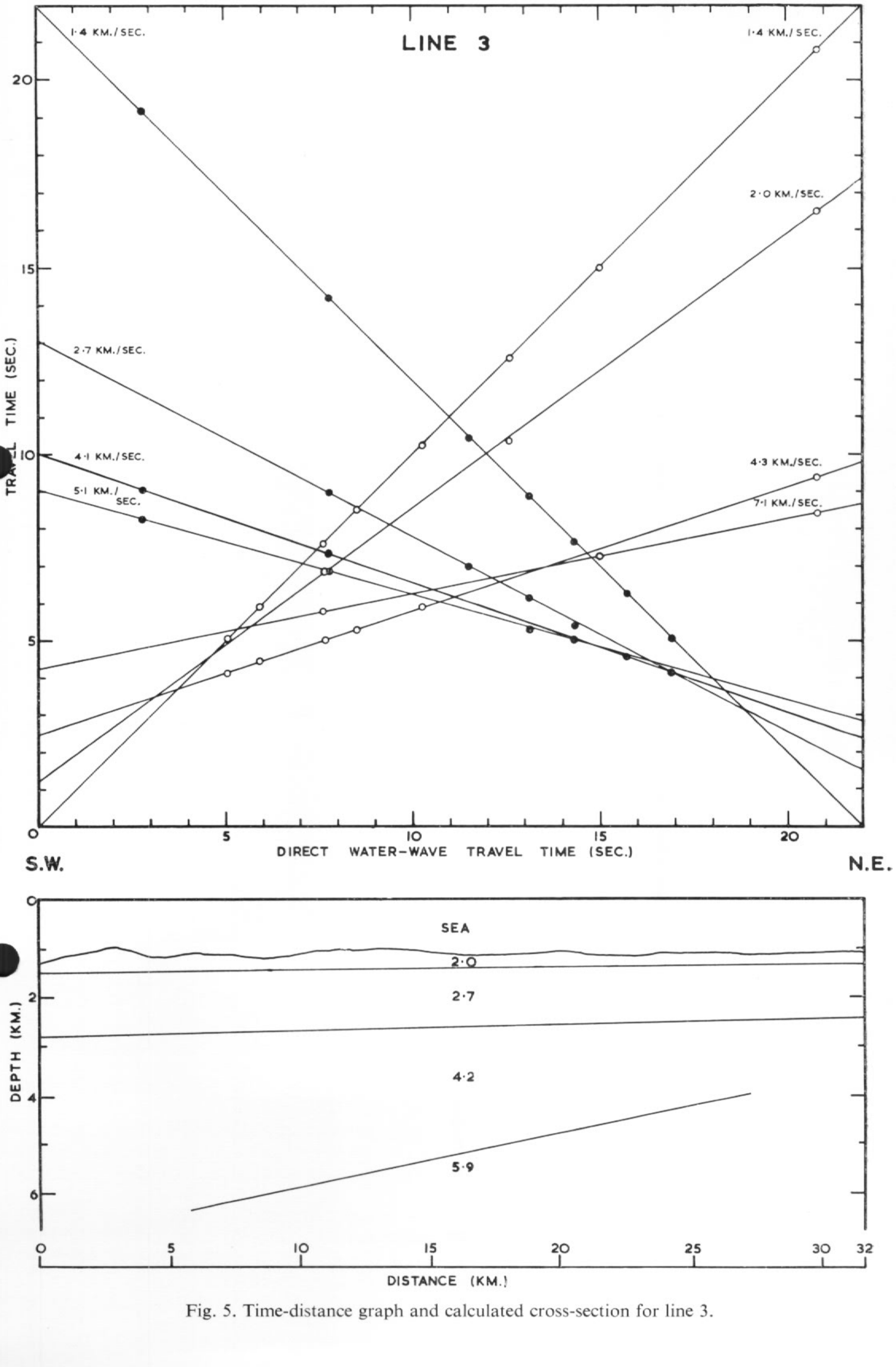

Fig. 5. Time-distance graph and calculated cross-section for line 3.

Line 4

Topographic corrections were applied for a horizontal base line at a depth of 400 fathoms (0·73 km.). The corrected time-distance graph for this line is shown in Fig. 6. The first point on the line shot towards the north-east is of considerable interest as the seismic wave has travelled a horizontal distance of 4·04 miles (7·5 km.) in 2·00 sec. After the appropriate deduction has been made for the time taken to travel through the water, the wave must have had a velocity of 7·0 km./sec. in the rock immediately below the sea bottom. If sediment is present in this area, the velocity would have to be higher than 7·0 km./sec. Alternatively, the velocity could reasonably be as low as 6·5 km./sec. if the shot had been fired above a steep slope on the sea bottom. In this case the distance travelled by the seismic wave in water not only depends on the depth of water beneath the shot and the velocity of the material constituting the sea bottom but also on the slope of the sea bottom. A true velocity of 7·0±0·5 km./sec. was therefore used for this. A sediment velocity of 2·0 km./sec. was assumed for the top layer of the north-east part of the line. Underlying this assumed layer there is a well-defined layer of true velocity 3·0 km./sec. The apparent velocity of 4·0 km./sec. on the line shot to the north-east almost certainly represents a down-dip velocity of the 7·0 km./sec. layer, which implies a dip of 23°. The apparent velocity of 5·3 km./sec. could then be interpreted as being due to a decrease in the dip of this interface to 9°. On the reversed line the arrivals are very weak at the longer ranges, but there is a suggestion of an apparent velocity of 22·0 km./sec. If this is combined with the 4·0 km./sec. velocity, a layer of true velocity 6·4 km./sec. dipping at 20° is obtained. This dip then decreases to 6° at the north-east end of the line.

Line 5

Topographic corrections were applied for a horizontal base line at a depth of 500 fathoms (0·92 km.). The corrected time-distance graph for this line is shown in Fig. 7. The true velocity of 2·1 km./sec. for the top layer is well determined. On the line shot towards the south-east an apparent velocity of 2·7 km./sec. was obtained. It was assumed that this was also the true velocity for the layer. The next layer, with a true velocity of 4·3 km./sec., is well defined with good agreement of the reverse times. The deepest layer obtained on this line has poor agreement of the reverse times. However, if the two apparent velocities represent a constantly dipping interface, then a true velocity of 6·4 km./sec. is found for this layer. This compares with a velocity of 6·1 km./sec. on line 2, which was shot perpendicular to line 5 (Fig. 1).

CONCLUSIONS

The seismic results have been summarized in the structure cross-section shown in Fig. 8. The formations with velocities of about 2·0 and 2·7 km./sec., and the thin layer with a velocity of 3·3 km./sec. obtained on line 1, are interpreted as constituting sedimentary layers. The increase in velocity with depth is associated with consolidation (Ewing and Nafe, 1963). These sedimentary layers extend down to a depth of about 3 km. The velocity of about 4·2 km./sec. on lines 2 and 3 falls within the range for volcanic rocks and for consolidated sediments. Because of the abundance of volcanic rocks in the area, this layer may be best correlated with the volcanic rocks of King George Island. The 5·0 km./sec. layer on line 1 is suggested as being igneous or metamorphic. Later work in this area, however, suggests that there are two layers with respective velocities of about 4·5 and 5·5 km./sec. rather than the single velocity of 5·0 km./sec. (personal communication from W. A. Ashcroft). The velocity of 6·0 km./sec. is characteristic of the principal refracting layer of the continents (Gutenberg, 1955). The high-velocity layer observed on line 1 at a depth of about 5 km. is probably a rock which is intermediate in composition between that which is primarily crustal and that which is primarily mantle.

The above structure shows marked differences when compared with a typical oceanic structure which is composed of about 4·5 km. of water, 0·5 km. of sediments or layer 1, 1·75 km. of layer 2 (velocity about 5·1 km./sec.), 4·75 km. of layer 3 (velocity about 6·7 km./sec.) overlying the mantle or layer 4 (velocity about 8·1 km./sec.) (Raitt, 1963). The

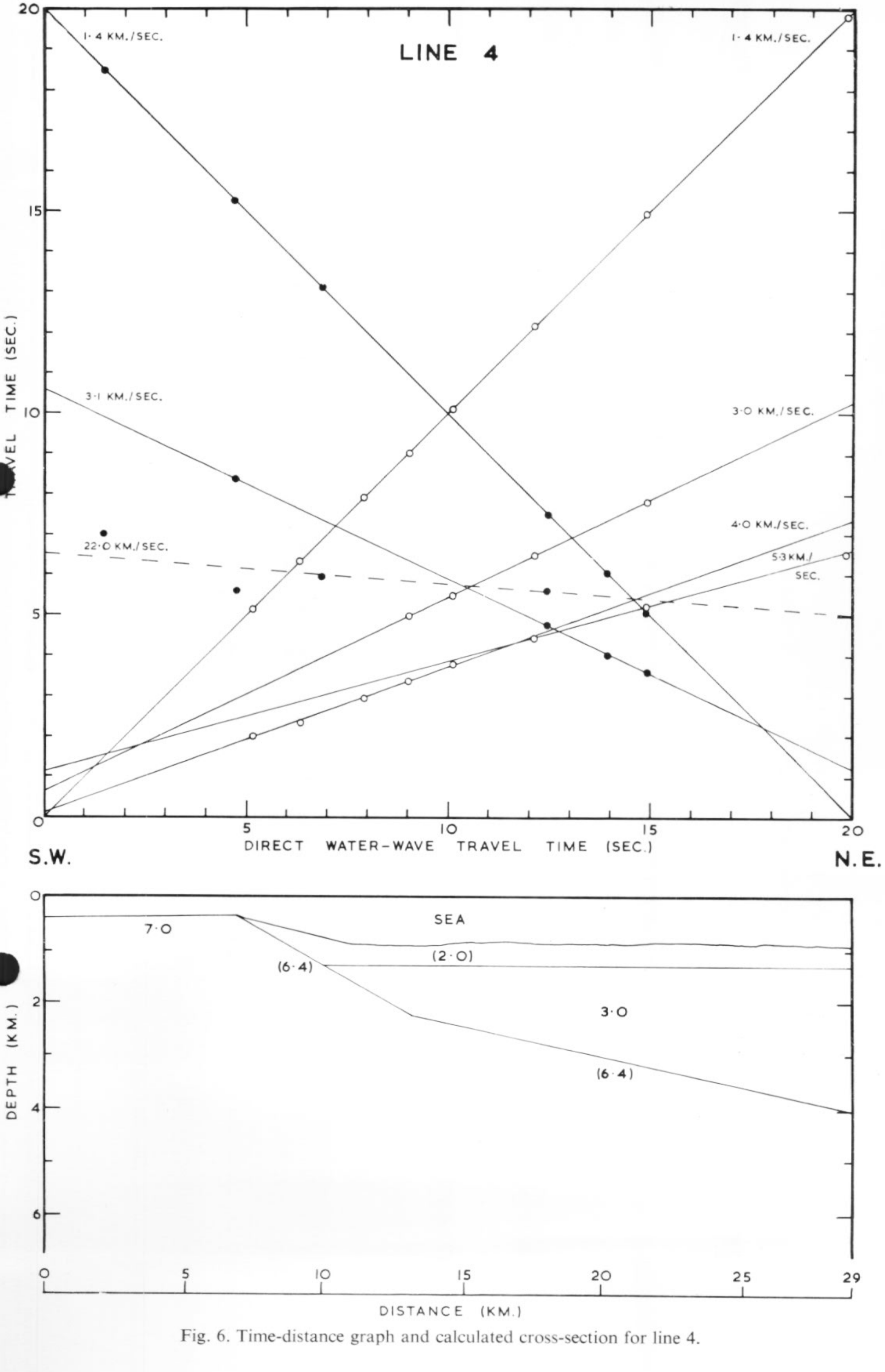

Fig. 6. Time-distance graph and calculated cross-section for line 4.

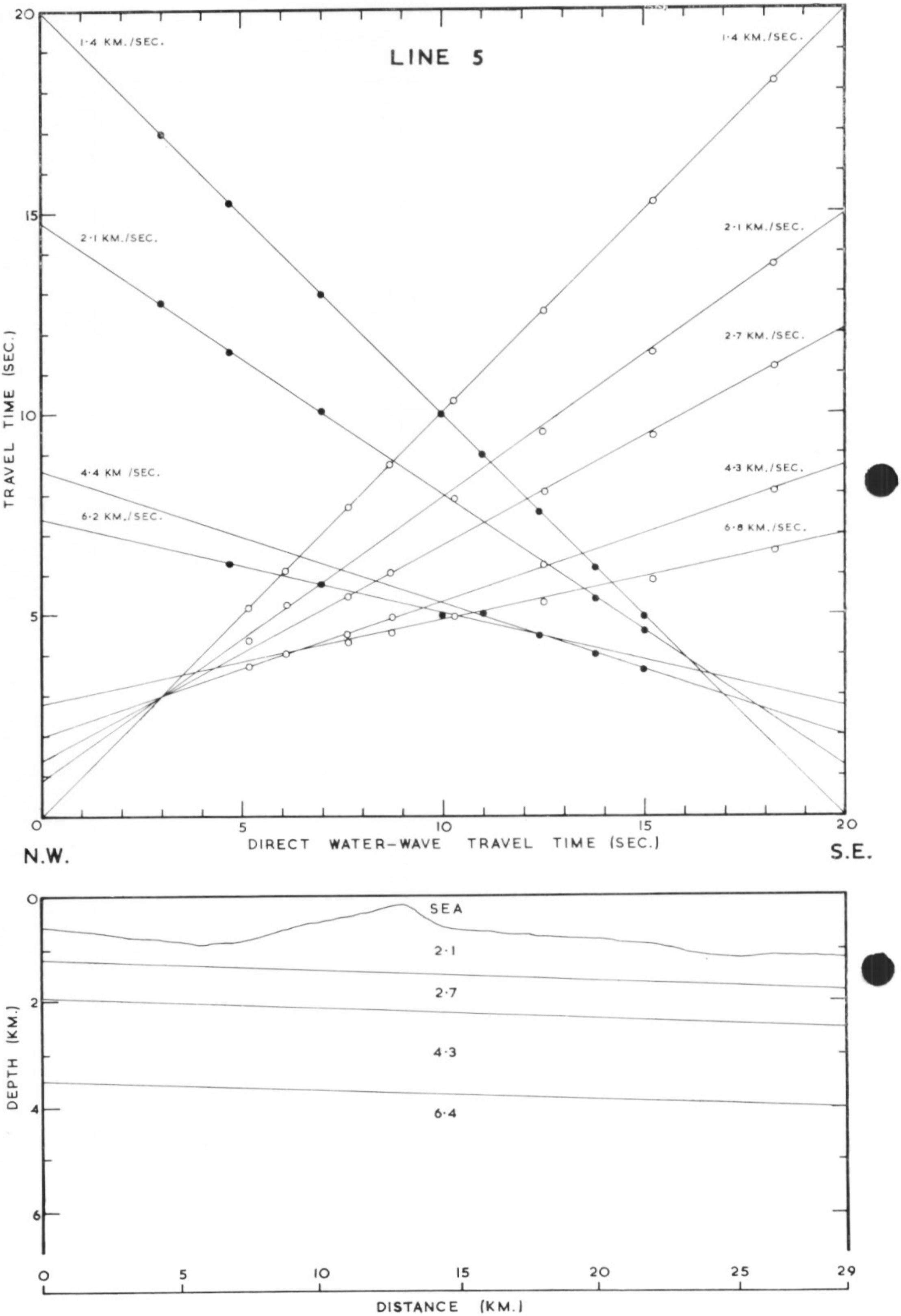

Fig. 7. Time-distance graph and calculated cross-section for line 5.

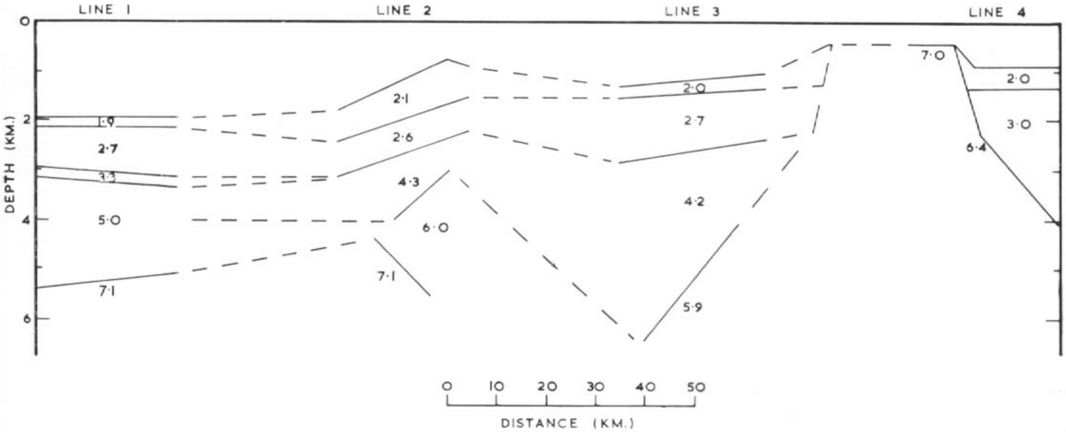

Fig. 8. Structural cross-section deduced from lines 1 to 4. The vertical to horizontal exaggeration is 10 : 1.

layer 3 velocity increases to about 7·0 km./sec. near the continents (Drake, Ewing and Sutton, 1959).

The north-east part of Bransfield Strait thus exhibits velocities which are more typically continental and, at the same time, thicknesses of layers below the sea bottom which are more typically oceanic. It is therefore suggested that this area is part of a transition zone between continental and oceanic structures.

The high-velocity layer obtained on line 4 is almost certainly due to the dunite-serpentine which occurs on Gibbs Island (Tyrrell, 1945) close to the south-western extremity of the line. From line 3, the south-western part of line 4 and the gravity data, very approximate limits can be placed for the boundaries of the dunite-serpentine. It is recommended that detailed sea magnetic work should be carried out in the immediate vicinity of Gibbs Island in order to locate the boundaries of the dunite-serpentine with greater accuracy.

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