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Title: MARINE SEISMIC REFRACTION STUDY BETWEEN CAPE SIMPSON AND PRUDHOE BAY, ALASKA

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A marine seismic refraction study, conducted in August 1976 by personnel from Oregon State University and the University of Connecticut between Cape Simpson and Prudhoe Bay, Alaska, provides data for analysis which yields a subsurface structural and geological cross-section of the area.

The results suggest that the structural homocline which dips to the east southeast on land extends to the offshore region as well. Correlation of geologic data from wells drilled on land with the refraction data permits tentative identification of geologic sequences on the basis of their seismic velocity. This study correlates 1.60 to 1.65 km/s layers to Quaternary sediments, 1.82 to 2.51 km/s layers to Tertiary strata, 2.91 to 3.40 km/s layers to Mesozoic formations to the east and 2.99 to 4.43 km/s to Early Mesozoic formations to the west. Velocities of 5.28 to 6.08 km/s are associated with probable argillite and phyllite of the Pre-Mississippian basement. At greater depths, refractors with velocities of 6.40 to 7.07 km/s are related to crystalline material which may be silicic or mafic. No seismic velocities typical of the upper mantle are present on the record sections, but a minimum depth

calculation places the Mohorovicic discontinuity deeper than 20 km. Although the observed crustal velocities are ambiguous towards theories of the origin of the Canada Basin and the tectonic history of the northern Alaska margin, they tend to favor the orocline-Rift theory of Carey (1955) over a subduction margin.

Marine Seismic Refraction Study Between Cape Simpson and Prudhoe Bay, Alaska

by

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TABLE OF CONTENTS

LIST OF FIGURES

MARINE SEISMIC REFRACTION STUDY BETWEEN CAPE SIMPSON AND PRUDHOE BAY, ALASKA

INTRODUCTION

One of the last regions on earth to be explored by modern man, the Arctic has only recently yielded to extensive scientific probing by earth scientists. As recently as 20 years ago, little was known about the form and composition of the Arctic Ocean Basin. Of particular interest to geologists and geophysicists, the thickness and composition of the earth's crust under the Arctic Ocean has been the subject of discussion for several decades. Ostenso (1962) begins his report of geophysical investigations in the area with the statement:

> The Arctic Ocean is the earth's least understood first order physiographic feature. Indeed, whether it may in the geological sense be properly considered an ocean is an open question.

Since this was written, scientific exploration of the Arctic Ocean Basin from drifting ice stations, aircraft, submarines and icebreakers, has advanced rapidly, but the tectonic history of the Arctic Ocean Basin, and in particular the Canada Basin and the topic of this study, the adjoining Beaufort Sea shelf, remains unclear. Reconstruction of the tectonic history of this margin requires information about the deep crustal structure of northern Alaska at the Beaufort Sea. Different authors attribute the formation of the margin to a wide variety of causes which include rifting because of counterclockwise rotation of Alaska (Carey, 1955; Rickwood, 1970; Tailleur and Brosgé, 1970), strike-slip faulting along a transform

fault (Tailleur, 1973; Herron et al., 1974; Yorath and Norris, 1975), or subduction within the Arctic Basin (Ostenso, 1972; Bally 1976). A great deal is known about the subsurface geology of the Arctic coastal plain from geologic, geophysical and drillhole data on land and considerable geophysical data are now available from published sources for the offshore area. Refraction data, however, is rare for the shelf area of northern Alaska. In 1976 a substantial amount of refraction data was successfully collected on the Alaskan shelf from an icebreaker. This thesis reports the analysis of thirteen refraction profiles obtained near the Beaufort Sea coastline of northern Alaska between Point Barrow and Prudhoe Bay. The text presents the data and its interpretation and relates the resulting structural and velocity sections to onshore geology.

ARCTIC BASIN PHYSIOGRAPHY

The relationship of the Arctic Basin to the study area is important to the significance of the seismic data. A brief description of the basic features present and theories for their formation provides a background for later reference.

The large number and rapid publication of Arctic studies in recent years results in confusion since individual features are given more than one name by different authors. The nomenclature of the major geomorphic features became further confused as new provinces were discovered and described. For simplicity, these features can be combined into three major geomorphic provinces as described by Herman (1974): continental margins, abyssal plains, and rises and ridges. Figure 1 shows the principal features and names of the Arctic Basin in the description which follows.

Continental Margins

The physiographic features which characterize the transition from continent to ocean are collectively referred to as the continental margin and consist of the continental shelf, slope, and rise.

Continental shelves are the shallow water submerged portions of continents and in the north polar region they underlie almost two-thirds of the Arctic Ocean. The continental shelves of Eurasia are anomalously shallow and wide. For example, the East Siberian shelf is 480-960 km in width and is similar to the Barents and Kara shelves which are 800-1440 km wide with water depths generally less

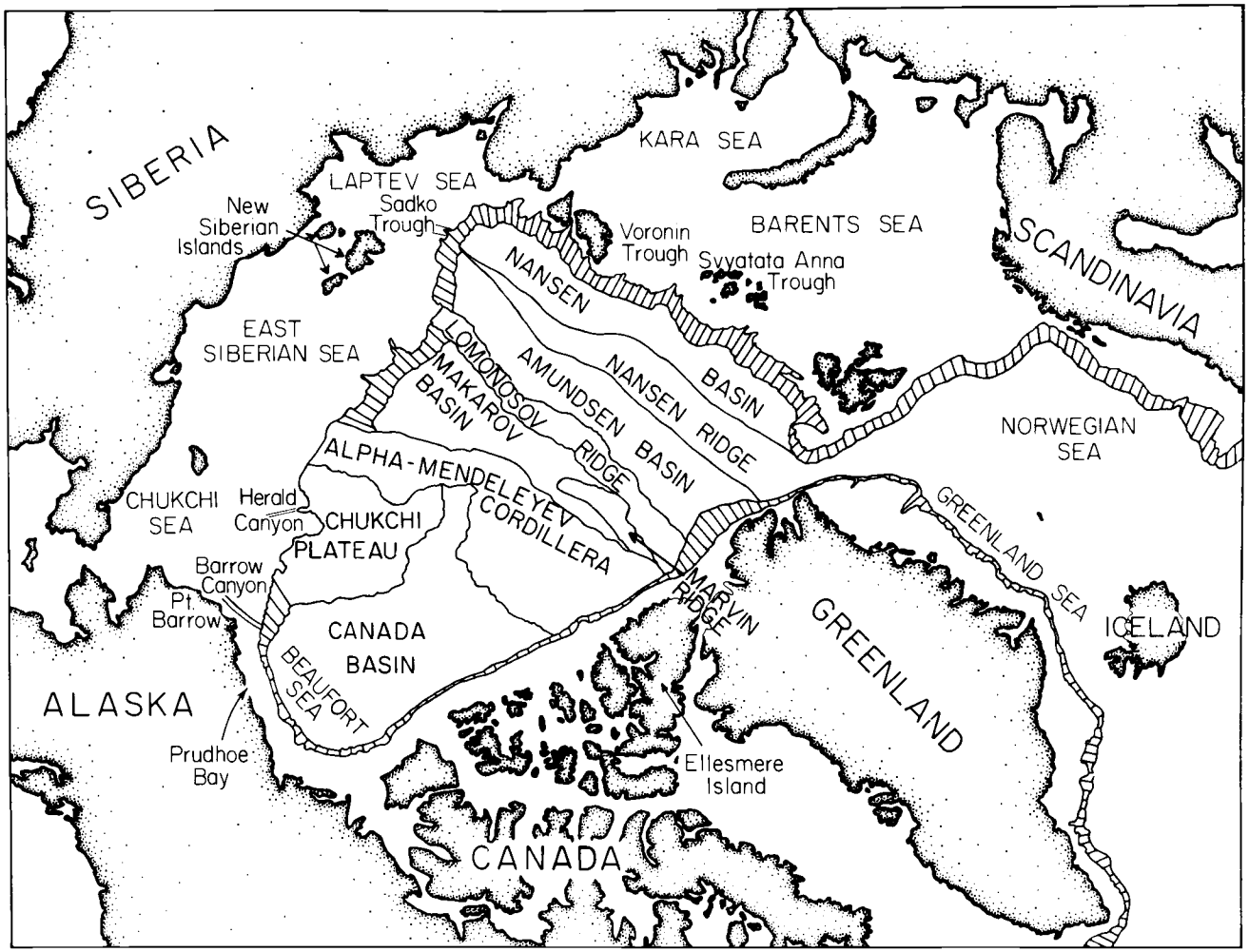


Figure 1. Physiographic map of the Arctic Ocean (from National Geographic Society, 1971).

than 100 m as shown in Figure 2. Geologic and geophysical information indicates that the shelves are formed by shales with thicknesses ranging from several meters to more than 18 km (Demenitskaya and Karasik, 1971). In contrast, the continental shelves off Greenland and northern Alaska are narrower and deeper being 32 to 80 km wide with water depths of approximately 200 m. These latter shelves are considered to be more like "normal" shelves and are underlain by a variety of sedimentary and metamorphic rocks.

The continental slope is the steeply dipping portion of the margin which lies seaward of the shelf. The main topographic features of Arctic continental slopes are canyons (Barrow Canyon, Herald Canyon) and submarine troughs (Sadko Trough, Voronin Trough, Svyatata Anna Trough) through which terrestrial and shallow water detritus is transported to the deep plains (Herman, 1974). The continental slopes begin at the usual depth of 200 m except off Greenland, where the break occurs at approximately 300 m (Ostenso, 1962). In general, the slope has gradients of 1 to 15° all around the Arctic Ocean. Off the coast of northeastern Alaska, the slope is the gentlest (1° to 2°) while it averages $4\frac{1}{2}^{\circ}$ offshore from the study area.

The continental rise lies at the base of the slope and has a very small gradient. In the Arctic Basin, rises are generally poorly developed features except in the Canada Basin (Fisher et al., 1958; Dietz and Shumway, 1961).

Abyssal Plains

Abyssal plains are smooth flat areas in deep water which cover

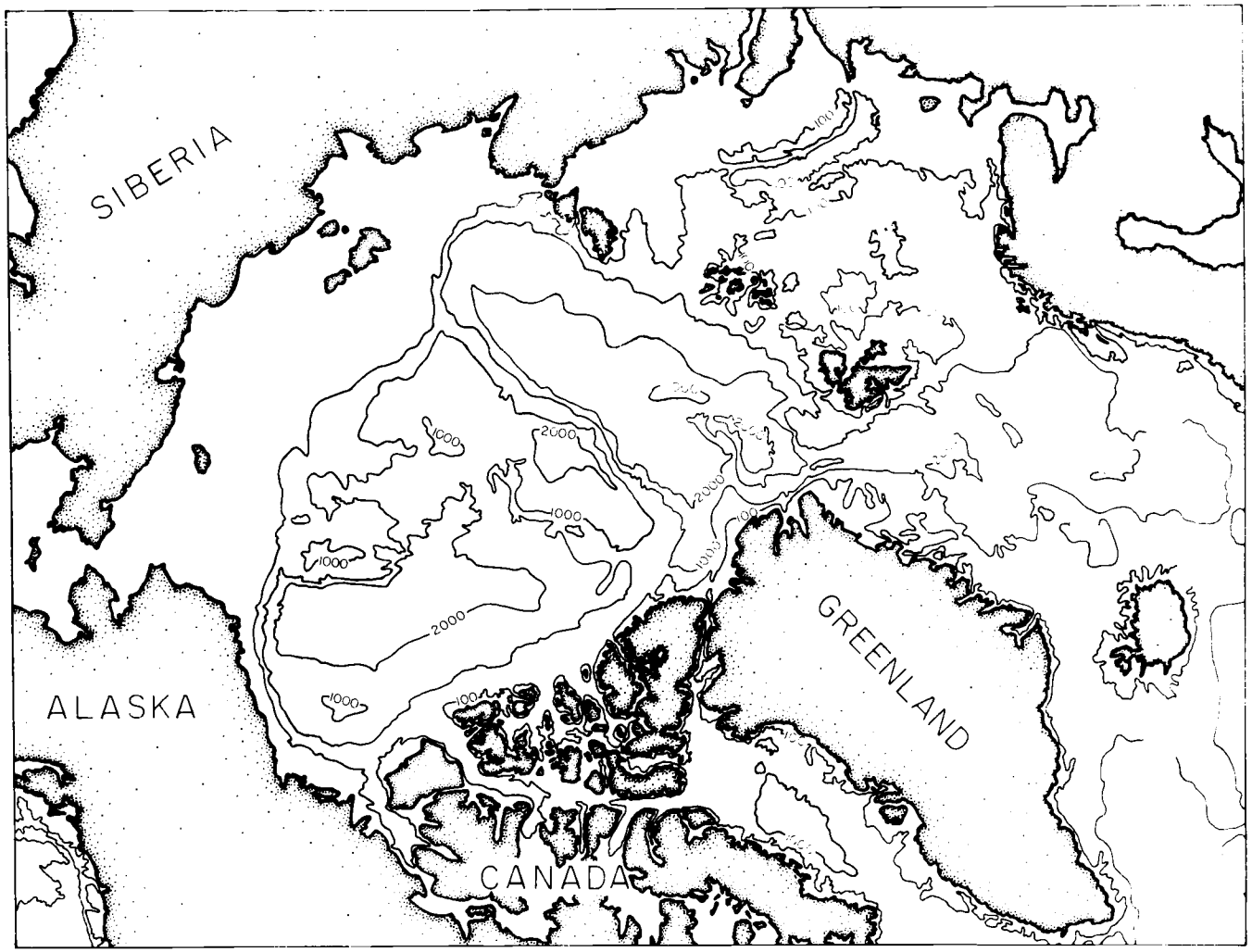


Figure 2. Bathymetric map of the Arctic Ocean (from National Geographic Society, 1971).

large areas. Three parallel ridges divide the Arctic Basin into four parallel elongate plains or basins. These four basins, in order across the pole from the North American to the Eurasian margin (Figure 1), are the Canada Basin, the Makarov Basin, the Amundsen Basin (also called Eurasian Basin, Fram Basin and Pole Abyssal Plain) and the Nansen Basin (also called Nautilus Basin, Barents Abyssal Plain and Sverdrup Plain).

The Canada Basin is surrounded by the steep continental slopes of Alaska and eastern Siberia and by the Alpha-Mendeleyev cordillera to the north. It has a rather smooth floor and a uniform depth of about 3800 m sloping gradually westward (Beal, 1969). Geophysical data suggest that the Canada Basin is underlain by a great thickness of sediments with a crystalline basement which slopes downward to the east (Ostenso and Wold, 1971).

The Lomonosov ridge, the Alpha-Mendeleyev cordillera and the Marvin ridge enclose the Makarov Basin. Its surface is featureless and remarkably flat at 400 m (Ostenso, 1962) but sediment thickness exceeds 2 km (Herman, 1974).

The Lomonosov ridge and the Nansen ridge (Arctic mid-oceanic ridge) delineate the Amundsen Basin which is also flat and featureless. It is about 4200 m deep and is underlain by from 2 to 3 km of sediments (Herman, 1974).

The Nansen Basin is the smallest and the deepest (it extends to a depth of 5180 m, Ostenso, 1962) of all four basins. The floor of the basin shows minor irregularities and is filled with sediment to a thickness of 2 or 3 km (Herman, 1974).

Rises and Ridges

Three subparallel submarine mountain ranges cross the Arctic Basin parallel to its minor axis and are, from the North American to the Eurasian margin, the Alpha Rise and Mendeleyev Ridge forming the Alpha-Mendeleyev cordillera, the Lomonosov Ridge and the Nansen Ridge (or Gakkel or Mid-Arctic Ridge) (Figure 1).

The Alpha-Mendeleyev cordillera is between 250 km and 1000 km wide and approximately 1800 km long. Broad plateaus join it to the Siberian and Canadian shelves (Herman, 1974). There is no evidence of seismic activity or volcanism associated with the Alpha-Mendeleyev cordillera. When first identified in 1957 this feature was thought to be of fault-block origin (Hunkins, 1961), but more recent analysis of the bathymetry (Beal, 1969; Hall, 1973) and aeromagnetic and gravity data (Vogt and Ostenso, 1970) have led to the suggestion that the cordillera is a fossil sea-floor spreading center. More recently, Herron et al. (1974) suggested that the Alpha-Mendeleyev cordillera is better interpreted as a fossil subduction zone. Its genesis is still unresolved.

The Lomonosov ridge is the narrowest (60 to 200 km) and highest (relief of about 300 m above the surrounding basin floor) of the ranges crossing the Arctic Basin. Significantly, it is both aseismic and magnetically quiet (Ostenso, 1962). It extends 1800 km from the continental shelf north of Ellesmere Island to the continental shelf north of the New Siberian Islands. Geophysical data suggest that the Lomonosov Ridge split from the former Eurasian continental margin

and was rafted to its present location by sea-floor spreading (Beal, 1969; Demenitskaya and Karasik, 1971; Vogt and Avery, 1974).

The seismically active Nansen Ridge extends from Greenland to the Laptev Sea and is about 2000 km long and 200 km wide. Individual peaks rise to 1500 m above the adjacent plains. Linear magnetic anomaly-patterns, typical of spreading ridges, are associated with the Nansen ridge (Rassokho et al, 1967) and it is considered to be an extension of the Mid-Atlantic Ridge (Heezen and Ewing, 1961) and a seismically active spreading center (Sykes, 1965).

THE ORIGIN OF THE ARCTIC BASIN

The numerous theories of the origin of the Arctic Basin are very diverse in spite of the common geological and geophysical evidence cited for support (Ostenso and Wold, 1971), but none is entirely satisfactory. For clarification, they are grouped into three main schools.

The Fixist theory originated in the USSR (Shatskiy, 1935; Belousov, 1955; Burkhanov, 1956; Hakkel', 1957; Saks, 1955, 1958) and in the USA (Eardley, 1948). They suggested that the Arctic Ocean is a foundered segment of a Paleozoic (or even Precambrian) continental platform. The theory proposed that geosynclinal sedimentation was followed by deformation and metamorphism, granite intrusion, and major uplifts within resulting fold belts. Such a history is similar to that of other ancient ocean basins which are believed to have closed in the past (Churkin, 1969). Though Eardley (1961) reversed his stand and favors an oceanic crust rather than a continental type for the Arctic Ocean, recent Russian geologists support the same theory (Pushcharovsky, 1960; Atlasov, 1964), but they describe the Arctic Ocean as a younger feature (Mesozoic or Cenozoic). More recently Meyerhoff (1970), who describes a Precambrian and Paleozoic evaporite distribution pattern in the Arctic, and Kiselev and Demenitskaya (1974) also support the foundered platform theory.

The Mobilistic theory describes the Arctic Ocean as a relatively young feature formed by continental drift through a process of rifting and sea-floor spreading during the Mesozoic and Cenozoic eras.

There are different and even sometimes incompatible theories within the Mobilists. Carey (1955) proposed an early idea where the Arctic Ocean originated from a crack which formed as North America spread apart from Eurasia by widening of the breach, like opening a pair of scissors, with the axis near the Alaska Range. Counterclockwise rotation of Alaska apart from the Arctic Canada would result in a bending of the western American orogenic belts, formation of the Alaskan orocline, and the opening of a huge triangular-shaped tension rift basin (the Arctic Ocean). Named the Orocline-Rift theory, it is still supported today (Tailleur and Brosgé, 1970; Rickwood, 1970; Freeland and Dietz, 1973). Tailleur (1973) proposed a variation of Carey's theory which favors a primitive north-south spreading axis bisecting the Beaufort Sea such that the Canada Basin was formed by rifting and sea-floor spreading in the Mesozoic.

The Alpha-Mendeleyev cordillera is a major unexplained feature of the Arctic Ocean and numerous speculations about its genesis have appeared. Beal (1969), Vogt and Ostenso (1970), and lately Hall (1973) support the idea of a fossil spreading center to account for the Alpha-Mendeleyev cordillera. According to this theory, the Arctic Ocean formed from a Paleogene spreading axis at the location of the Alpha-Mendeleyev cordillera. The spreading axis became inactive 40 m.y. ago when the locus of spreading shifted from the Alpha-Mendeleyev cordillera to the Nansen ridge, a presently active spreading center. The theory suggests that this new axis of Arctic spreading intersected the European continental margin and rifted off a section of continental crust that was translated northward to form

the present Lomonosov Ridge (Ostenso and Wold, 1971).

In contrast to this idea, Herron et al. (1974), favor the Alpha-Mendeleyev cordillera as a fossil subduction zone and incipient island arc. During Paleozoic time, the Kolymski plate (now part of Siberia) drifted south towards the Canadian Arctic archipelago. The Kolymski block remained part of North America until Jurassic time when it broke away and drifted towards Siberia creating the opening of the American basin by sea-floor spreading. It collided with Siberia in Early Cretaceous time and later subduction occurred at the location of the Alpha-Mendeleyev cordillera during Upper Cretaceous to Paleogene time to accomodate compression created by the opening of the North Atlantic Ocean. The Eurasia basin opened by sea-floor spreading during the Cenozoic where the Lomonosov ridge was the seaward edge of the Eurasian continental margin prior to Cenozoic.

Another theory, suggested by Yorath and Norris (1975), supports the existence of a Mesozoic spreading axis along the Canadian continental margin.

King et al. (1966), Rassokho et al. (1967), Demenitskaya and Hunkins (1971) form a third school known as fixist-mobilists who divide the Arctic Ocean into the Eurasia and America basins, and combine continental drift and continental subsidence to account for the formation of the Arctic Ocean. The Eurasia basin is a result of spreading processes and the Amerasia basin formed by geosynclinal subsidence.

Churkin (1973) gave a slightly different explanation by proposing

that a proto-Amerasian basin in the Early Paleozoic or even Pre-Cambrian underwent different openings and closings by geosynclinal subsidence followed by continental margin tectonics of sea-floor spreading type.

THE ORIGIN OF THE BEAUFORT SEA

The origin of the Canada Basin, of which the Beaufort Sea occupies the southeastern part, is of prime importance for the evaluation of the petroleum potential of the vast continental shelf northwest of Alaska. Many uncertainties remain concerning the nature of the Canada Basin, its origin and age.

The Canada Basin is now generally believed to be oceanic (Churkin, 1969; Wold et al., 1970; Hunkins, 1963; Tailleur, 1973; Hall, 1973; Vogt and Avery, 1974), so that the Canada Basin is a real, although small, ocean deep. Its age depends strongly on its interpreted origin, which at various times has been attributed to rifting due to rotation of Alaska, as the result of sea-floor spreading at different ridge locations, and as the result of subduction within the Arctic Basin as described earlier. More detailed information regarding these theories and the geologic evidence for them is given in the descriptions which follow.

The theory of rifting by rotation of Alaska away from Canada (Rickwood, 1970; Tailleur and Brosgé, 1970; Tailleur, 1973) supports a post-Triassic age for the Beaufort Sea, based mainly on paleogeographic and palinspatic fits between Arctic Alaska and Arctic Canada.

Sea-floor spreading as the origin of the Canada Basin has been mentioned very often, but different loci of spreading have been advocated: spreading at the mid-oceanic ridge (Nansen ridge) explains the Beaufort continental margin as a rifted margin, formed in pre-Cretaceous time (Vogt and Avery, 1974); Ostenso (1972)

supported this hypothesis but mentioned that some of the original Canada Basin crust may have subsequently been subducted under the Beaufort Sea continental margin. Inter-arc type of sea-floor spreading that occurs behind island arcs (Karig, 1971) would give a Carboniferous age for the Beaufort Sea (Vogt and Avery, 1974) while spreading at the Alpha-Mendeleyev cordillera during the Paleozoic leads to a Permo-Carboniferous or older oceanic crust beneath the Beaufort Sea (Hall, 1973). A Mesozoic spreading axis along the Canadian continental margin is favored by Yorath and Norris (1975) and suggests a Jurassic to Early Cretaceous time for the origin of the Beaufort Sea, wherein the passive continental margin exists as a transform fault. Another hypothesis which favors the Beaufort Sea continental margin as a transform fault suggests that a north-south sea-floor spreading axis bisected the Beaufort Sea at right angles to the present margin during the Jurassic and Middle Cretaceous (Tailleur, 1973). Herron et al., (1974) proposed a different variation which starts with the closure of a proto-Amerasian basin during the Early Paleozoic time and continued with a Jurassic opening of the Amerasian basin together with a spreading ridge along the Canadian continental margin and a transform fault along the Alaskan shelf edge.

Churkin (1969) noted that the southern part of the Canada Basin (which includes the Beaufort Sea) is rimmed by an Early Paleozoic geosynclinal belt and he later explained (Churkin, 1970; Churkin, 1972; Churkin, 1973) the history of the Canada Basin as an early development of geosynclines along the margins of the modern Canada Basin followed by deformation, metamorphism, granitic intrusion, and major uplift

and clastic-wedge sedimentation. He concluded that a proto-Canada Basin with continental margin tectonics existed in the Early Paleozoic (Cambrian to Middle Devonian).

Bally (1976) favored a subduction zone on the North Slope of Alaska and theorized on the tectonic history of the Beaufort Sea. If the basement of the continental shelf is a Paleozoic fold belt, there is a hint that Late Paleozoic-Triassic block faulting was associated with the initiation of the subsidence of the Beaufort shelf. Major block faulting may have occurred during the Mid-Jurassic and Early Cretaceous probably due to the rise of the Brooks Range (Rickwood, 1970). Since the Upper Cretaceous sediments lie unconformably over the Lower Cretaceous sequence, this unconformity indicates that much of the block faulting ceased during the early Upper Cretaceous and was followed by subsidence of older grabens and the overlying sediments accompanied by growth faulting and associated structure. Almost all the strike-slip faulting reported occurred also in Upper Cretaceous-Tertiary so that subduction of continental crust during Mesozoic-Tertiary is suggested on the North Slope of Alaska and is evident by extensive decollement structures in the Romanozoff Mountains and adjacent foothills.

Even though no final choice can be made among these theories, all of which cite geologic and geophysical evidence, a minimum Late Cretaceous age for the origin of the Beaufort Sea can be deduced from geological observations. Undisturbed sediments indicate that there has been no motion between the floor of the Canada Basin and Alaska since at least Early Cretaceous time (Rickwood, 1970; Tailleur,

1973). Furthermore, Collins and Robinson (1967) found that a submarine canyon leading into the Beaufort Sea east of Point Barrow had been cut to a depth of more than 1370 m and filled with sediments during the Late Cretaceous.

NORTHERN ALASKA STRUCTURE AND BATHYMETRY

Against this framework of mega-tectonics, it is appropriate to focus attention to the study area on the continental margin of northern Alaska. Geologically mapped from outcrop and well data on land, the main structural features are well-known. They consist of three major features which trend in an east-west direction (Figure 3). The Brooks Range on the south is a major topographic and structural feature which separates northern Alaska from the rest of the state and forms a major watershed. This north-facing slope drains into the Arctic Ocean and is the basis for the regional name North Slope. The parallel and strongly asymmetrical Colville geosyncline has its axis near its southern margin close to the mountain front. From thicknesses exceeding 9 km (Morgridge and Smith, 1972), the Lower Cretaceous to Tertiary sediments in the Colville geosyncline thin to the north towards a structural high known as the Barrow High or Barrow Arch. The Barrow Arch is a broad Lower Paleozoic basement arch which is oriented roughly parallel to the coastline and plunges to the southeast. The south flank of the arch has gentle dips of generally less than 2° while the north flank is somewhat steeper (Morgridge and Smith, 1972).

The Beaufort Sea has one of the narrowest continental shelves found anywhere in the Arctic (Figure 3) since it is only 70 km wide between the Colville and MacKenzie river deltas (Wold et al., 1970). Another peculiarity of the continental shelf in this area is its shallowness since it is generally less than 64 m deep (Carsola et al.,

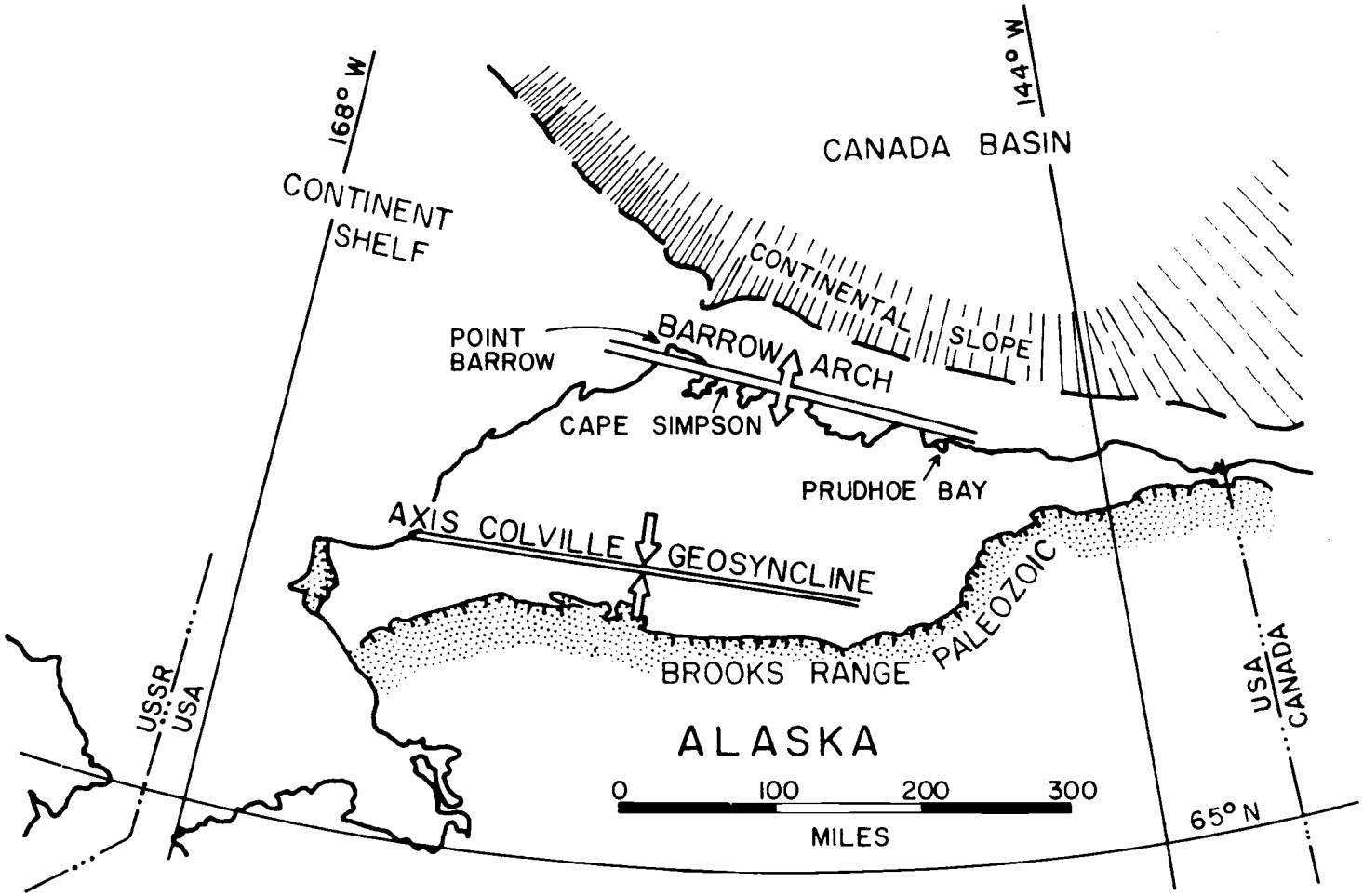


Figure 3. General structural framework of northern Alaska (from Rickwood, 1970).

1961) although the depth averages 20 m in the study area.

The continental slope is steep, linear and strongly sculptured by slumps and leveed channels off the western and central Beaufort shelf (Grantz et al., 1975). The gradient of the slope ranges from $1\frac{1}{2}^{\circ}$ to 5° (Ostenso, 1962) and the steepest part is just west of 141°W longitude (where the study area is located) with an average gradient of $4\frac{1}{2}^{\circ}$ (Figure 4).

It is worth noting that the Beaufort seacoast may receive from the main rivers and numerous streams emptying into the Beaufort Sea on the order of 30 million tons of solid material per year. This is the principal explanation for the high rate of sedimentation in this area (Wold et al., 1970).

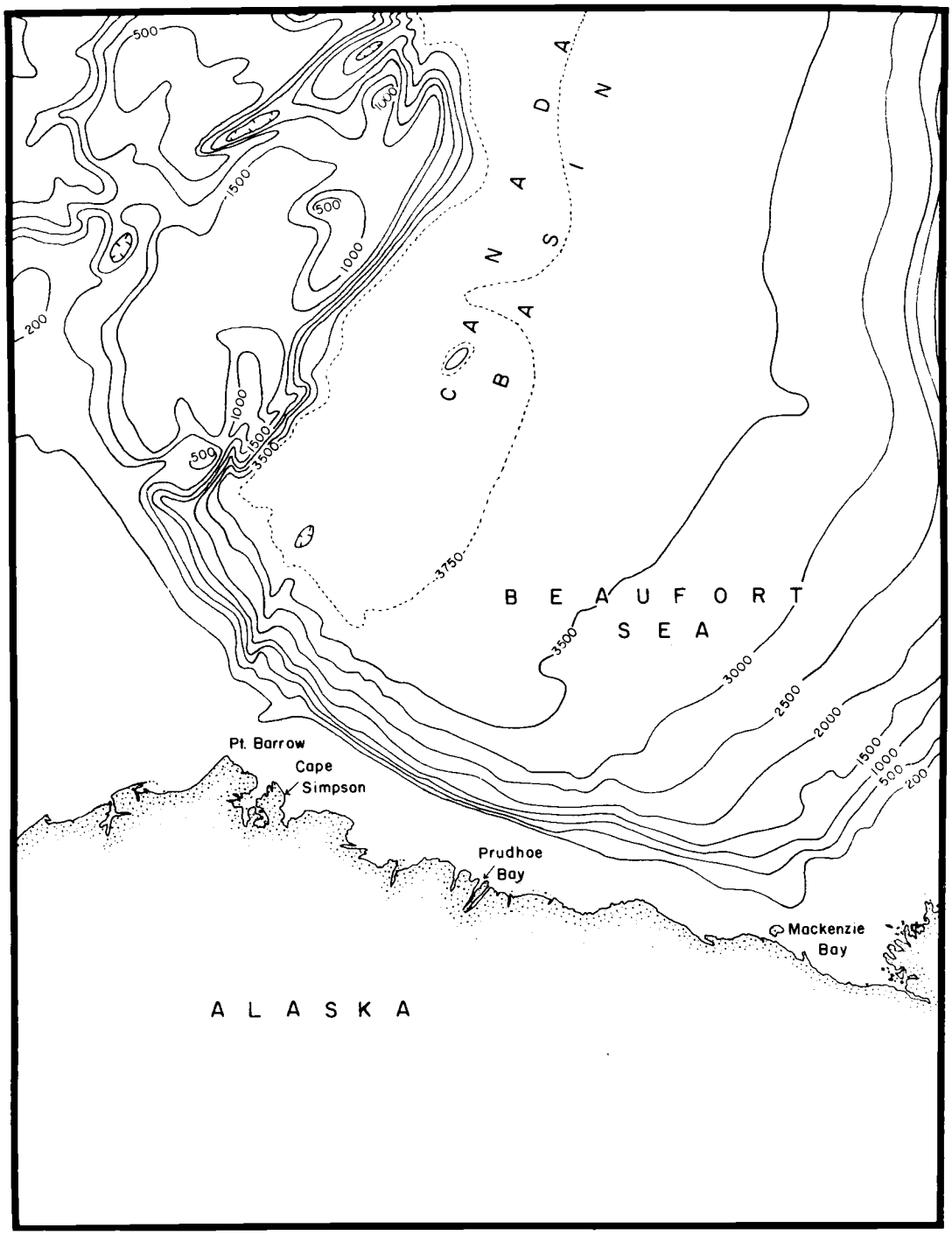


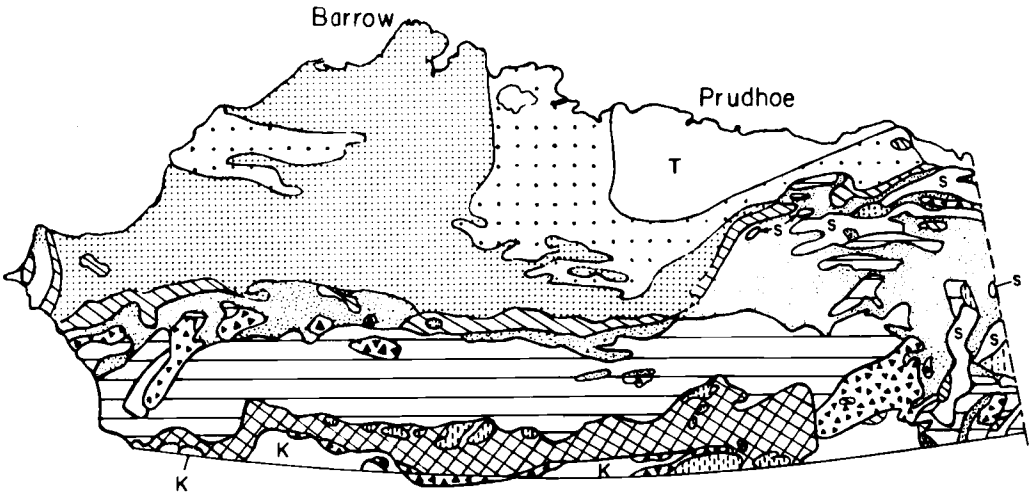
Figure 4. Bathymetric map of the Beaufort Sea, Alaska (from Tectonic map of North America, U.S.G.S., 1969).

GEOLOGY OF THE BARROW ARCH AREA

The continental shelf of northern Alaska is a sedimentary basin containing mainly Tertiary and Cretaceous material. Geophysical evidence indicates that the sediments dip eastward (Ostenso, 1962; Woolson et al., 1962; Hunkins, 1966; Rickwood, 1970; Ostenso and Wold, 1971; Morgridge and Smith, 1972). Most of the offshore geological structure has been deduced from test wells onshore, interpretation of seismic reflection and offshore extension of the land geological structure (Figure 5). Morgridge and Smith (1972) describe a geological cross section of the Barrow Arch area based on well data. The cross-section shown on Figure 6 extends from west to east through the South Barrow gas field, the Simpson wells, the Topogoruk no. 1 dry hole, the Fish Creek test well, the Colville no. 1 dry hole and the Prudhoe Bay discovery well and extends farther southeast along the arch and northeast into the Beaufort Sea. The main sedimentary sequences named according to their geological age are:

1) Pre-Mississippian Sequence
2) Carboniferous Sequence
3) Permian and Lower Triassic Sequence
4) Triassic and Jurassic Sequence
5) Lower Cretaceous Sequence
6) Upper Cretaceous, Tertiary and Quaternary Sequence

These sequences are described in some detail in the discussion which follows and Figure 7 summarizes the stratigraphic section.



EXPLANATION

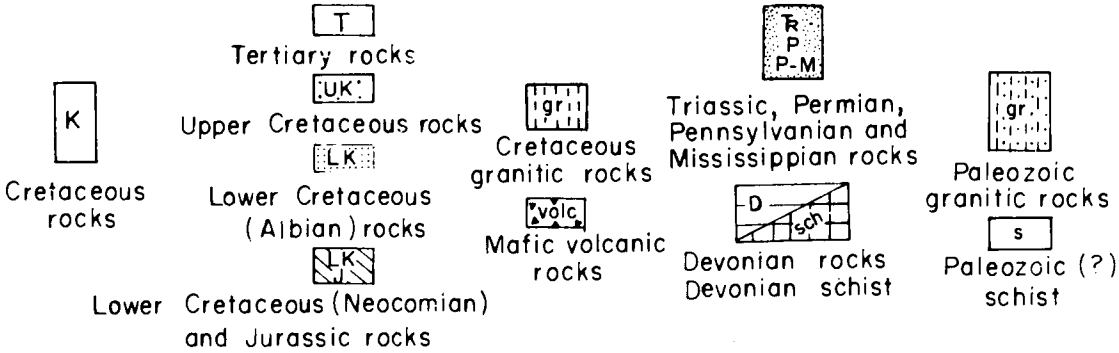


Figure 5. Geologic map of northern Alaska (from Brosgé and Tailleur, 1970).

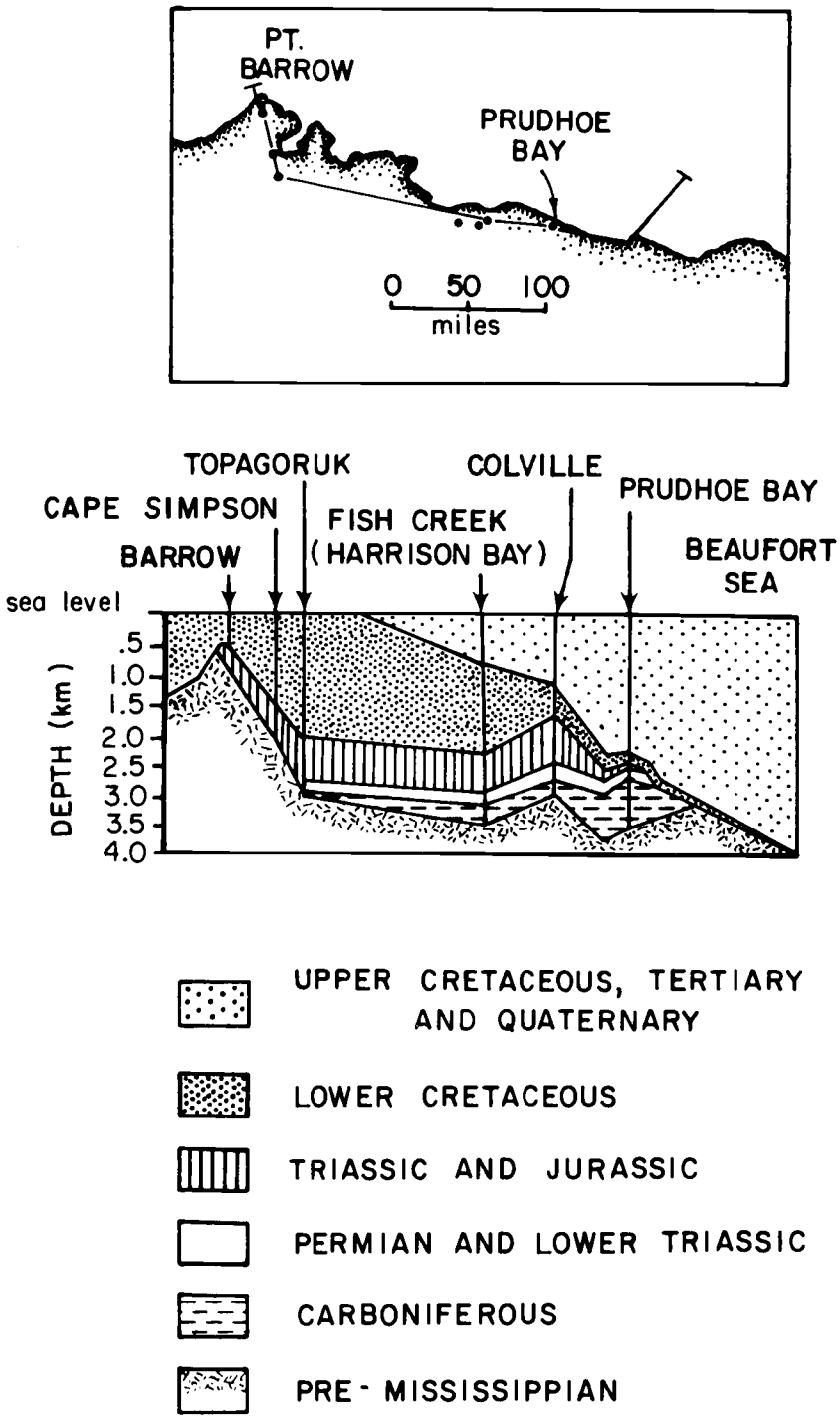


Figure 6. Present time cross-section of the Barrow Arch area (from Morgridge and Smith, 1972).

GEOLOGIC AGE				STRATIGRAPHIC NAME	
ERA	PERIOD		EPOCH		
CENO-ZOIC	QUATERNARY		PLEISTOCENE	GUBIK FORMATION	
	TERTIARY		PALEOCENE to MIOCENE or PLIOCENE	SAGAVANIRKTOK FORMATION	
MESOZOIC	CRETA-CEOUS	UPPER	CAMPANIAN	COLVILLE GROUP	SCHRADER BLUFF FORMATION
			CONIACIAN and TURONIAN		SEABEE FORMATION
			CENOMANIAN	NANUSHUK GROUP	GRANDSTAND FORMATION
		LOWER	ALBIAN	TOROK or TOPAGORUK FORMATION	
				OUMALIK FORMATION	
			NEOCOMIAN	PEBBLE SHALE	
	JURASSIC			KINGAK FORMATION	
	TRIASSIC			SHUBLIK FORMATION	
PALEOZOIC	PERMIAN			SADLEROCHIT FORMATION	
	CARBON-IFEROUS	PENNSYLVANIAN		LISBURNE GROUP	
		MISSISSIPPIAN		KAYAK FORMATION	
	PRE-MISSISSIPPIAN			no name	

unconformity

No indication of thickness

Figure 7. Stratigraphic ages of northern Alaska.

Pre-Mississippian Sequence

Extensive geological and geophysical investigations of the Arctic Slope showed eastward-dipping basement rocks to be at a depth of 0.76 km under Barrow (Dana, 1951; Woollard et al., 1960; Morgridge and Smith, 1972; Gryc, 1970), 2 km at Cape Simpson (Robinson, 1959), 3.2 km at Topagoruk (Gryc, 1970) and more than 3 km under Prudhoe Bay (Rickwood, 1970; Brosgé and Tailleur, 1970; Morgridge and Smith, 1972). These rocks were identified as argillite and phyllite which transmitted seismic compressional waves at 5.2 km/s (Ostenso and Wold, 1971). This altered and highly distributed sequence is believed to be of different ages at different places: Precambrian in northern Alaska (Payne et al., 1951), pre-Upper Devonian at Prudhoe Bay (Rickwood, 1970; Morgridge and Smith, 1972), Middle or Lower Devonian at Topagoruk no. 1, test well (Gryc, 1970), Early Paleozoic argillites at Point Barrow (Brosgé and Tailleur, 1970), and Late Precambrian phyllites at Cape Simpson (Brosgé and Tailleur, 1970). As the uncertainty remains, these rocks will be called Pre-Mississippian since Mississippian rocks constitute the oldest well-dated layer.

In the vicinity of Point Barrow and Cape Simpson, P-wave velocities of 6.4 km/s were recorded in what was believed to be granite underlying the argillite-phyllite sequence (Ostenso and Wold, 1971). This crystalline basement has been found west of Point Barrow (Johnson and Chiburis, 1977) and east of Prudhoe Bay in the Canadian Shield (Hobson, 1962; Overton, 1970). Ostenso (1962) interprets a gravity high as due to a ridge of high density

non-magnetic material, probably granite, which parallels the continental shelf. Its crest would be approximately under the 200 m isobath.

Carboniferous Sequence

Pennsylvanian-Mississippian continental and shallow-marine clastic sediments and shallow-water shelf carbonate beds overlie unconformably the Pre-Mississippian sequence (Morgridge and Smith, 1972). This sequence is missing at the Barrow and Simpson wells where it has certainly been removed by erosion.

Marine clastic rocks of the Kayak Formation primarily composed of shale represent the base of the Carboniferous sequence. Some red rock beds are present in the Colville delta (Brosgé and Tailleur, 1970) and mudstones with minor sandstones having a thickness of at least 1.8 km are present at Prudhoe Bay (Rickwood, 1970).

The carbonates of the Lisburne Group overlie the Kayak Formation, and range in age from Mississippian to Permian. At Prudhoe Bay, the Lisburne Group is represented by more than 1.2 km of Pennsylvanian limestone (Rickwood, 1970). It is missing in Topagoruk well no. 1, near Barrow, but is presumed from seismic reflections to be present farther west (Brosgé and Tailleur, 1970).

Permian and Lower Triassic Sequence

The sandy and conglomeratic Sadlerochit Formation rests disconformably upon the Lisburne carbonates. It is absent over the Barrow High (certainly eroded) and the upper part of the formation

is eroded at Topagoruk and Prudhoe. The Sadlerochit clastic sediments were derived from erosion of the Mississippian-Pennsylvanian-Permian Lisburne carbonate unit and older quartz and chert-rich rocks which were exposed in uplifts north of the present Arctic coastline (Morgridge and Smith, 1972). This sequence is truncated by the Upper Triassic Formation near Barrow (Brosgé and Tailleur, 1970) and is composed of 280 m of shales and sands at Prudhoe Bay (Rickwood, 1970) truncated by the Cretaceous formation (Brosgé and Tailleur, 1970).

Triassic and Jurassic Sequence

A thin layer (60 to 80 m) of Lower Triassic Shublik Formation, consisting of chert and fossiliferous limestone in the western regions, and dark shale and phosphatic limestone in the northeast, extends over the entire North Slope overlying the Sadlerochit Formation (Brosgé and Tailleur, 1970). It is represented by limestone and sandstone at Point Barrow and by argillaceous siltstone, silty claystone and sandstone at Cape Simpson (Collins, 1961; Robinson, 1959). A uniform blanket of Upper Triassic limestone, chert and shale of the Shublik Formation marked the last of a long persistent depositional pattern. It locally contains thin sands, glauconite or oolite around the Barrow area where it laps onto basement (Brosgé and Tailleur, 1970).

The Jurassic thick dark marine shale of the Kingak Formation forms a thick layer: more than 610 m at Colville and Topagoruk (Morgridge and Smith, 1972) and 550 m at Prudhoe Bay (Rickwood, 1970).

Thin sandstones and siltstones at both the base and the top of the Jurassic section at Point Barrow and Cape Simpson have been derived from a northern source area (Morgridge and Smith, 1972).

Lower Cretaceous Sequence

A thin layer of dark shale, called the Pebble Shale, overlies the Kingak Formation (Brosgé and Tailleur, 1970; Collins, 1961; Robinson, 1959). It generally underlies Albian rocks and unconformably overlies Jurassic rocks, so it is regarded as Neocomian in age (Brosgé and Tailleur, 1970) and may extend to Upper Jurassic (Robinson, 1959; Detterman, 1973). At Prudhoe Bay, all the pre-Cretaceous formations were truncated and sealed by this Early Cretaceous shale and a hydrocarbon trap was formed (Morgridge and Smith, 1972).

The Pebble Shale is overlain unconformably by Albian clay shale of the Oumalik Formation (Collins, 1961; Robinson, 1959). The thickness of this formation varies from 0 m to 210 m near Point Barrow (Collins, 1961) and from 56 m to 520 m in different test wells at Cape Simpson (Robinson, 1959).

The Torok or Topagoruk Formation is also of Albian age and lies above the Oumalik Formation. This thick layer (550 m at Barrow, Collins, 1961; 250 m to 790 m at Cape Simpson, Robinson, 1959; more than 1 km at Fish Creek, Robinson and Collins, 1959) is composed of marine clay shale, siltstone and sandstone (Collins, 1961; Robinson, 1959; Robinson and Collins, 1959; Rickwood, 1970).

Above the marine Albian are the non-marine and marine

subgraywackes of the Nanushuk Group. It is Middle Albian to Cenomanian in age and grades downward and northward into the Torok Formation (Brosgé and Tailleur, 1970). The Nanushuk Group is represented by the Grandstand Formation on the North Slope of Alaska and 300 m of clay shale, sandstone and siltstone are indicated at Cape Simpson by Robinson (1959), and about 100 m of sandstone and shale of the Grandstand Formation at Point Barrow by Collins (1961). This formation is missing in many wells, probably due to erosion.

Upper Cretaceous, Tertiary and Quaternary Sequence

The youngest rocks are poorly consolidated Upper Cretaceous and Tertiary marine and non-marine deposits which form a thick layer dipping east and north and derived from the Brooks Range in the south, which continued to rise during the Cretaceous (Rickwood, 1970). They are about 100 m thick at Point Barrow (Collins, 1961), 800 m at Cape Simpson (Robinson, 1959), around 700 m at Fish Creek (Robinson and Collins, 1959), and 2.44 km at Prudhoe Bay (Rickwood, 1970; Morgridge and Smith, 1972). These rocks make up the Quaternary Gubik Formation, the Tertiary Sagavanirktok Formation and the Upper Cretaceous Colville Group, which lies partly unconformably on the Nanushuk Group (Brosgé and Tailleur, 1970).

The sedimentary rocks of the Colville Group are generally divided into two formations. At the base of the Colville Group, the Coniacian and Turonian clays and shales of the Seabee Formation lie on the rocks of the Nanushuk Group with some local unconformities at Cape Simpson and Point Barrow (Brosgé and Tailleur, 1970). The

Seabee Formation varies from 0 m to 380 m thickness at Cape Simpson (Robinson, 1959) to about 200 m thickness at Fish Creek (Robinson and Collins, 1959). At the top of the Colville Group, the Campanian Schrader Bluff Formation is composed of 0 to 440 m of marine clay shale at Cape Simpson (Robinson, 1959), of about 450 m of marine clay shale with interbedded siltstone at Fish Creek (Robinson and Collins, 1959) and of 610 m of clays, shales and coals in the lower part of the formation, and sandstones and siltstones in the upper part at Prudhoe Bay (Rickwood, 1970).

The Colville Group is overlain by the marine sands, silts and clays of the Sagavanirktok Formation believed to be Early Tertiary to Miocene or Pliocene age (Brosgé and Tailleur, 1970). Rickwood (1970) reported it to be .9 to 1.5 km thick at Prudhoe Bay and it may be as much as 2.13 km thick along the east part of the coastal plain (Brosgé and Tailleur, 1970). On the west part of the coastal plain no Tertiary strata have been found (Robinson, 1959; Robinson and Collins, 1959; Collins, 1961).

In northern Alaska the coastal plain is masked by gravel and sand of the Quaternary Gubik Formation which lie on nearly undeformed Upper Cretaceous rocks in the west (Point Barrow, Cape Simpson, Fish Creek) and, in other places, on Tertiary sediments (Prudhoe Bay) (Churkin, 1973).

PERMAFROST

Permafrost is defined as earth material within which the temperature has been below $0^{\circ}C$ for several years without regard to the percentage of moisture present. The permafrost layer is not physically homogeneous and the velocity of compressional waves through it ranges from 2.0 to 4.0 km/s (Woolson et al., 1962; Hobson, 1962; Lachenbruch, 1970; Hunter and Hobson, 1974). The effect of permafrost where it has been measured is to increase the velocity of the conpressional wave, that is, to produce a high-velocity layer, thus complicating seismic exploration work in the Arctic compared to other areas. A low-velocity layer (sea water and unfrozen sediments) overlying a high-velocity layer (permafrost) constitutes a model amenable to interpretation by the seismic refraction method.

Permafrost, referred to as remnant subsea permafrost, is present under the coastal waters of the Beaufort Sea (in the study area) in the form of isolated lenses of thicknesses ranging from 30 m to 600 m (Lachenbruch, 1970; Lewellen, 1974; Judge, 1974; Hunter and Hobson, 1974) with greater thicknesses close to the present northern shorelines (Lachenbruch, 1970; Judge 1974). The determination of the thickness of the permafrost is problematic and, since the base of the permafrost is a seismic interface between high and low-velocity materials, seismic refraction methods are not suitable for this particular case.

In this study, it was not possible to distinguish between permafrost and an unfrozen sediment layer and hence sediments with

velocities of 1.82-2.51 km/s could be permafrost or sediment. The records may be unsuitable for observing the presence of permafrost because of the large shot spacing (about .7 km). For a better determination of permafrost the use of a high-resolution shallow-penetration boomer with a close shot spacing and a multielement streamer might be effective. This technique would use the nearest hydrophone for reflection results and more distant hydrophones for refraction. Permafrost detected by seismic refraction work would need to be verified by drilling and sampling.

REFRACTION MEASUREMENTS BETWEEN CAPE SIMPSON AND PRUDHOE BAY, ALASKA

Location

Personnel from Oregon State University and the University of Connecticut conducted marine seismic refraction studies in the western Beaufort Sea to obtain structural and velocity information on the continental margin. In contrast to most previous refraction studies in the region which were made from stations located on the ice, these data were obtained using standard marine seismic techniques during the Arctic summer. Thirteen profiles were run between Cape Simpson and Prudhoe Bay in August 1976 from the USCG icebreaker Burton Island with helicopter support. The profiles, ranging in length from 23 to 75 km, were roughly parallel to the coastline in about 20 m of water. The reason for the apparently random direction of individual profiles shown in Figure 8 was heavy ice conditions. Because of these navigation difficulties only three lines were reversed. These lines are Lines 10-11 and 12-13, Lines 22A-23 and 24B-25, and Lines 24A-25 and 22B-23, where the numbers indicate the end points of the lines. The seven remaining lines (Lines 8A-9A, 8B-9B, 16-17, 18-19, 20-21, 26A-27A, 26B-27B) were interpreted as single-ended lines.

The profiles lie in a region of low gravity gradient between moderate gravity highs on land and a series of substantial gravity highs at the edge of the upper slope (Figure 9).

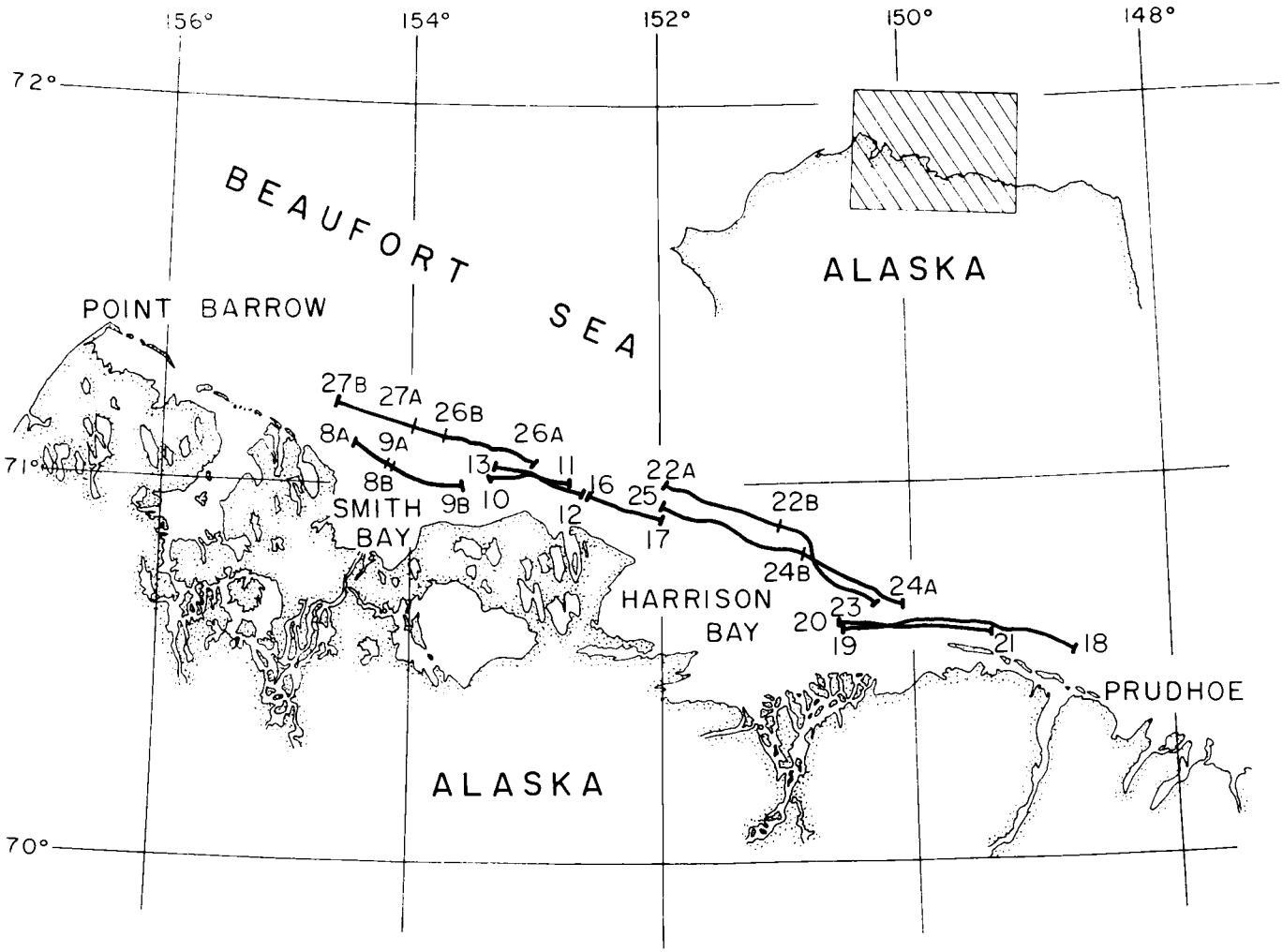


Figure 8. Location map showing the locations of seismic refraction profiles.

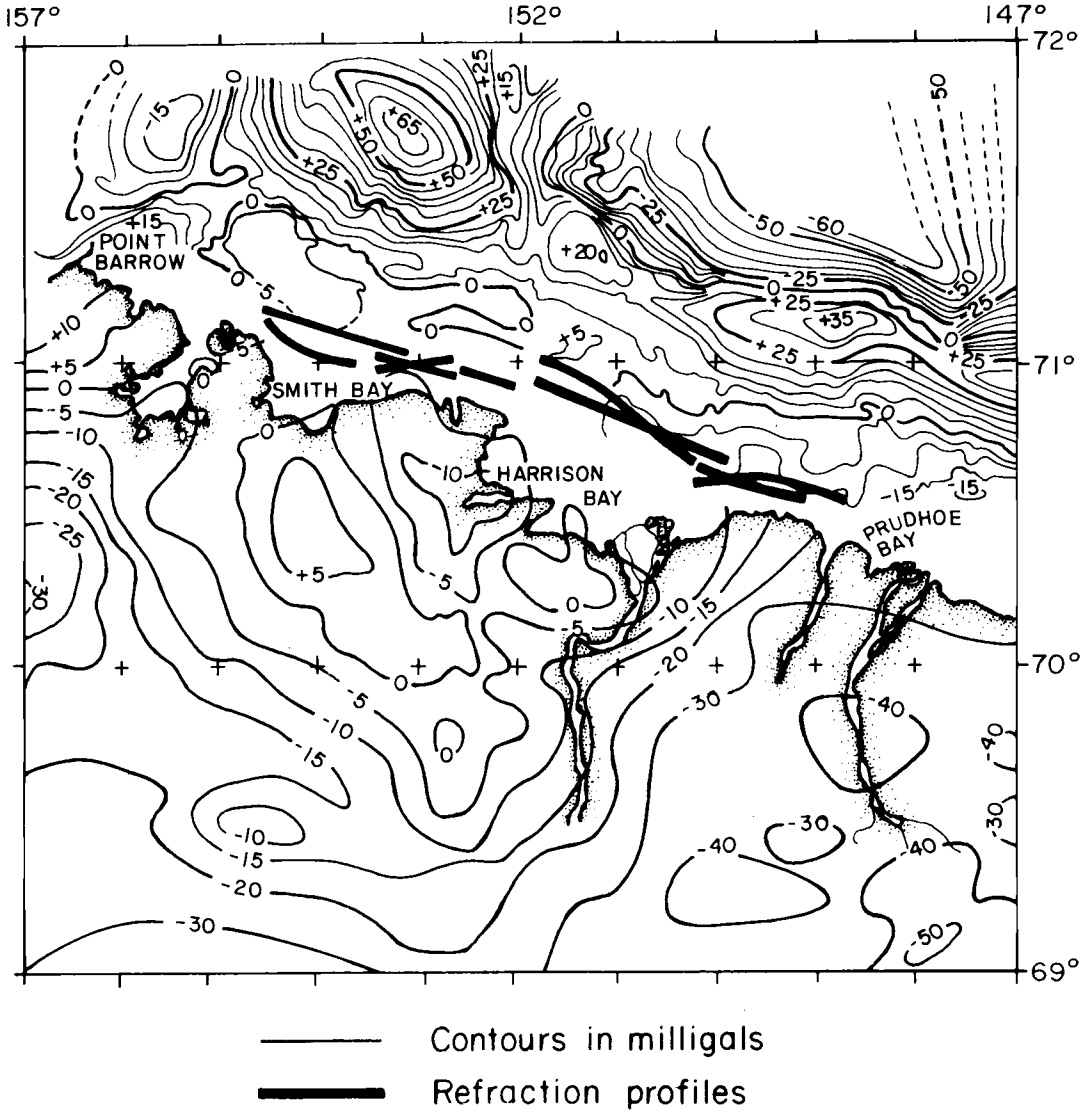


Figure 9. Gravity anomaly map of northern Alaska: offshore free-air gravity anomalies from Dehlinger (1978) and land Bouguer anomalies from Woolson et al. (1962) together with the locations of refraction profiles.

Techniques

A zone of ice-free or semi ice-free water exists along the northern continental shelf area out to a depth of approximately 2000 meters during one month of the year, and existence of this zone permitted operation of standard marine refraction techniques with sonobuoys and explosive charges.

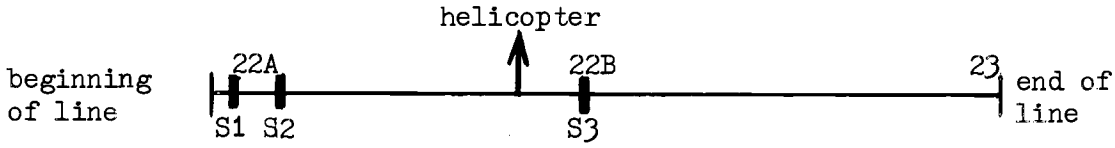
The lines were shot in the standard marine fashion by dropping charges in the water from the fantail of the icebreaker, using powder fuses and tilt table, while the ship was underway. Rotor-mounted Yagi antennas received the sonobuoy signals which were recorded on a 4-channel tape recorder. Two of the 4 channels were sonobuoy signals, one was a clock signal and one was a shot-break signal from a streamer 3 m long which was towed close behind the ship. When radio contact with the sonobuoys was lost, at a distance of about 30 km because of earth curvature, a portable recording unit consisting of a radio receiver, an amplifier, and strip-chart and tape recorders installed in a helicopter, monitored the sonobuoy. The helicopter remained within radio reception range of the sonobuoy.

Expendable naval sonobuoys of the type AN/SSQ 41A were modified for extended time operation by addition of dry-cell batteries which worked quite well in spite of the cold water. Explosive charges of less than 10 pounds were made up of nitro-carbonitrate (Nitromon) in 1 pound metal cans. Explosive charges between 30 and 660 pounds were made up of Tovex in 30 pound plastic bags.

Shots were detonated every three minutes at a ship speed of 10 kt which was slightly variable because of ice conditions. This

resulted in a shot spacing of about .7 km. During the helicopter operations, shots were detonated at intervals of from 5 to 15 minutes resulting in a shot spacing of 1.4 to 4.3 km.

A special sonobuoy deployment was used in our study: two sonobuoys (S1, S2) were dropped at the beginning of each line and for a long line an intermediate sonobuoy (S3) was deployed in the middle of the line. This resulted in a special line numbering with letter A for the first sonobuoys and letter B for the intermediate one (for example, Lines 22A-23 and 22B-23).



The satellite navigation equipment was inoperative during the time of the experiment, therefore all the navigation was by radar fixes to land points at 15 minute intervals. The ice coverage ranged from 0 to 8 octas during the course of the experiment and required frequent course changes and caution on the part of the shooter not to hit floating ice during the explosive drops.

The currents were not negligible and affected the sonobuoy drift. The direction and magnitude of the drift was estimated by combining water wave travel time and navigation. The in-line component of drift ranged from 0 to 1.65 m/s and was quite variable from line to line.

DATA ANALYSIS

A combination of manual and computer-aided manipulations shown in the flow chart of Figure 10 transform the raw data to record sections and velocity depth profiles.

Arrival times for ground and water waves picked on each seismogram and other information such as bathymetry, ship velocity and streamer length, constitute an input data file for program TIMCORM. The computer program TIMCORM computes corrected ground and water wave arrival times to a datum, making corrections for the shot instant due to separation of shot and streamer and surface and bottom corrections at receiver and shot. Using these results, the computer program REFPLTT produces a travel-time plot with the corrected arrival times of the ground waves at the distance calculated from corrected water wave arrival time. A first interpretation with seismic velocities is then made and the data examined for possible errors. Using the corrected output file obtained from program TIMCORM, the computer program REDPLOT2 produces a reduced time plot where arrival times reduced with a velocity of 5.00 km/s are plotted versus distance. A profile which includes some helicopter refraction data requires an additional routine. A data file is prepared from the ship navigation as an input to computer program UTMSTEVE and its two subroutine programs UTMGRIDB and UTM. These programs plot the navigation data on a Universal Transverse Mercator projection. Shot positions are added manually and arcs drawn using a compass at a radius corresponding to the water wave travel time. Sonobuoy drift

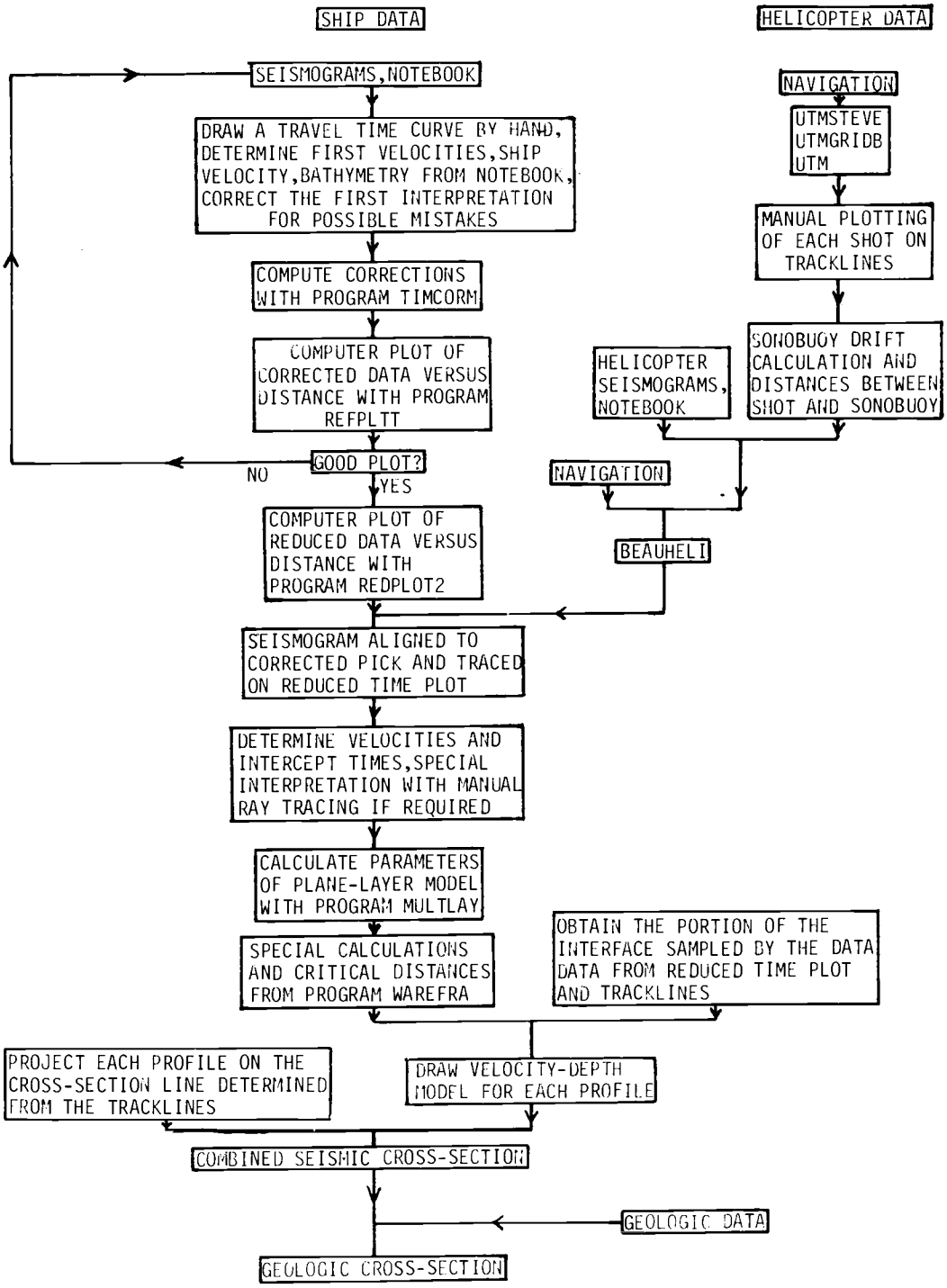


Figure 10. Flow chart of data analysis.

due to wind and currents appears as a non-coincidence of the arcs. A drift rate parallel to the trackline is assumed to apply to more distant shots where water wave arrival times may not be detected by the sonobuoy. The shot distance is then the distance between the shot location and the drift position of the sonobuoy at the time of the shot. The drift distances between the shot and the sonobuoy and the ground wave arrival times from the helicopter refraction data constitute an input file for program BEAUHELI. Computer program BEAUHELI computes time corrections between the ship and the helicopter clocks, the corrected distance between the shot and the sonobuoy, and the corrected travel time for the ground arrival. The strip chart record from the helicopter is then aligned to the corrected ground arrival time as plotted on the reduced time plot and traced.

Apparent velocities and intercept times are obtained from this record section and put into the computer program MULTLAY. The computer program MULTLAY computes a model composed of N dipping plane layers from reversed seismic refraction profile data using the formulation of Adachi (1954). The program requires two sets of apparent velocities and intercept times from single-ended refraction data. Complications such as non-reciprocal data or early intercept times require a manual ray tracing method which also uses the formulation of Adachi (1954). On each seismic model obtained from MULTLAY the portion of an interface drafted with heavier lines on Figure 11 through Figure 23 represents only that portion of the layer which gives rise to observable seismograms. The offset distances of the rays are obtained from the program WAREFRA. The computer program

WAREFRA is based also on the formulation of Adachi (1954) and requires the seismic model obtained from MULTLAY as an input data file. For the combined velocity-depth section (Figure 24), each profile was projected onto a straight line drawn through the tracklines. The final seismic cross-section along the chosen line includes all the individual seismic models and is interpreted as seismic layers of the same range of velocities. Integrating the velocity-depth section with surface geological information, regional tectonic interpretations, published well-data, and other interpretations of nearby geological and geophysical data, produces the final synthesis. The final result is a correlation of seismic layers with geologic formations (Figure 26).

In a technical report, Bee et al. (1979) give trackline maps, shot locations, radar navigation points, sonobuoy drifts, input data files, computer program listings and output data files with the structural information.

RESULTS

The profiles parallel the coast in a general east-west direction. Figures 11 to 23 show the thirteen profiles (Lines 8A-9A, 8B-9B, 10-11, 12-13, 16-17, 18-19, 20-21, 22A-23, 22B-23, 24A-25, 24B-25, 26A-27A, 26B-27B) plotted as reduced record sections where the reducing velocity is 5.0 km/s. The distance is in water wave travel time seconds where the water velocity was estimated at 1.44 km/s. All the velocities are in km/s. The seismic models determined from interpreted reduced sections are composed of plane horizontal or plane dipping layers. These are shown above the record section with a vertical exaggeration of 3:1. Heavy lines on the refractors indicate the interfaces responsible for observed arrivals and contain the horizontal offsets for upgoing rays. Only the reversed profiles resulted in true velocities, and on all the single-ended profiles apparent velocities are assumed to be true velocities. Arrivals from a thin upper sediment layer are not detected as first arrivals because of the wide shot spacing, but its presence is required to obtain the observed water depth (about 20 m). A velocity of 1.60 km/s on Lines 26A-27A and 26B-27B and 1.65 km/s on Lines 10-11 and 12-13 was obtained for this layer from the arrival on the first seismogram. For other lines, the velocity for this layer was assumed to be 1.60 km/s. All estimated velocities are in parentheses. With reference to the record sections presented in Figures 11 through 23, the following paragraphs describe the basis for the interpretation of each line and the velocity-depth section computed.

Line 8A-9A

Line 8A-9A was shot to the east close to Cape Simpson and is only 11 km long. The records presented in Figure 11 indicate a poorly-determined velocity of 2.23 km/s based on first and second arrivals and a well-determined velocity of 3.02 km/s based on strong first arrivals. A higher velocity of 4.68 km/s fits first arrivals at the end of the line. The seismic model has five horizontal layers, including the water and sediment velocities of 1.44 km/s and 1.60 km/s, with depths of .24 km for the 2.23 km/s refractor, .76 km for the 3.02 km/s refractor and 1.86 km for the 4.68 km/s refractor.

Line 8B-9B

Line 8B-9B presented in Figure 12 continues Line 8A-9A to the east and runs close to and parallel to the coast. The absence of data at the beginning of the line forced the inclusion of a velocity of 2.23 km/s based on the interpretation of line 8A-9A. Very strong first arrivals indicate a 2.83 km/s and a 4.24 km/s velocity. These are followed by a long segment with a velocity of 5.63 km/s based on first and second arrivals. Evidence for a higher velocity is reasonably strong on these records even though the line is only 24 km long and a 7.07 km/s velocity results from a number of strong arrivals near the end of the line. The corresponding velocity structure is a seven-layer model with all the layers being assumed horizontal and the apparent velocities assumed to be true velocities. The depths are 0.24 km for the 2.23 km/s refractor, 0.63 km for the

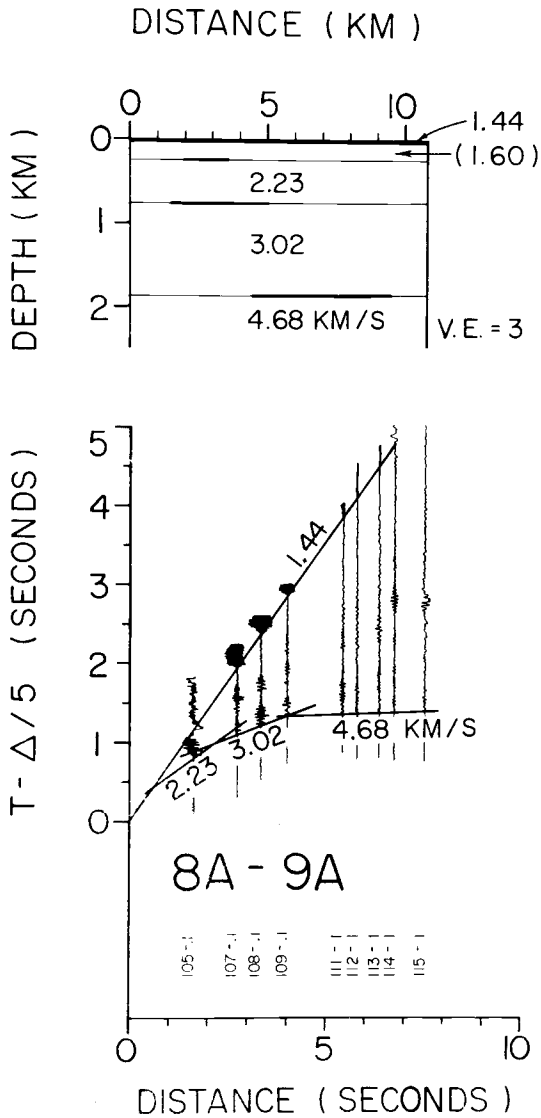


Figure 11. Line 8A-9A record section and velocity-depth model interpretation. Below each seismogram is listed the shot number and the charge weight in pounds.

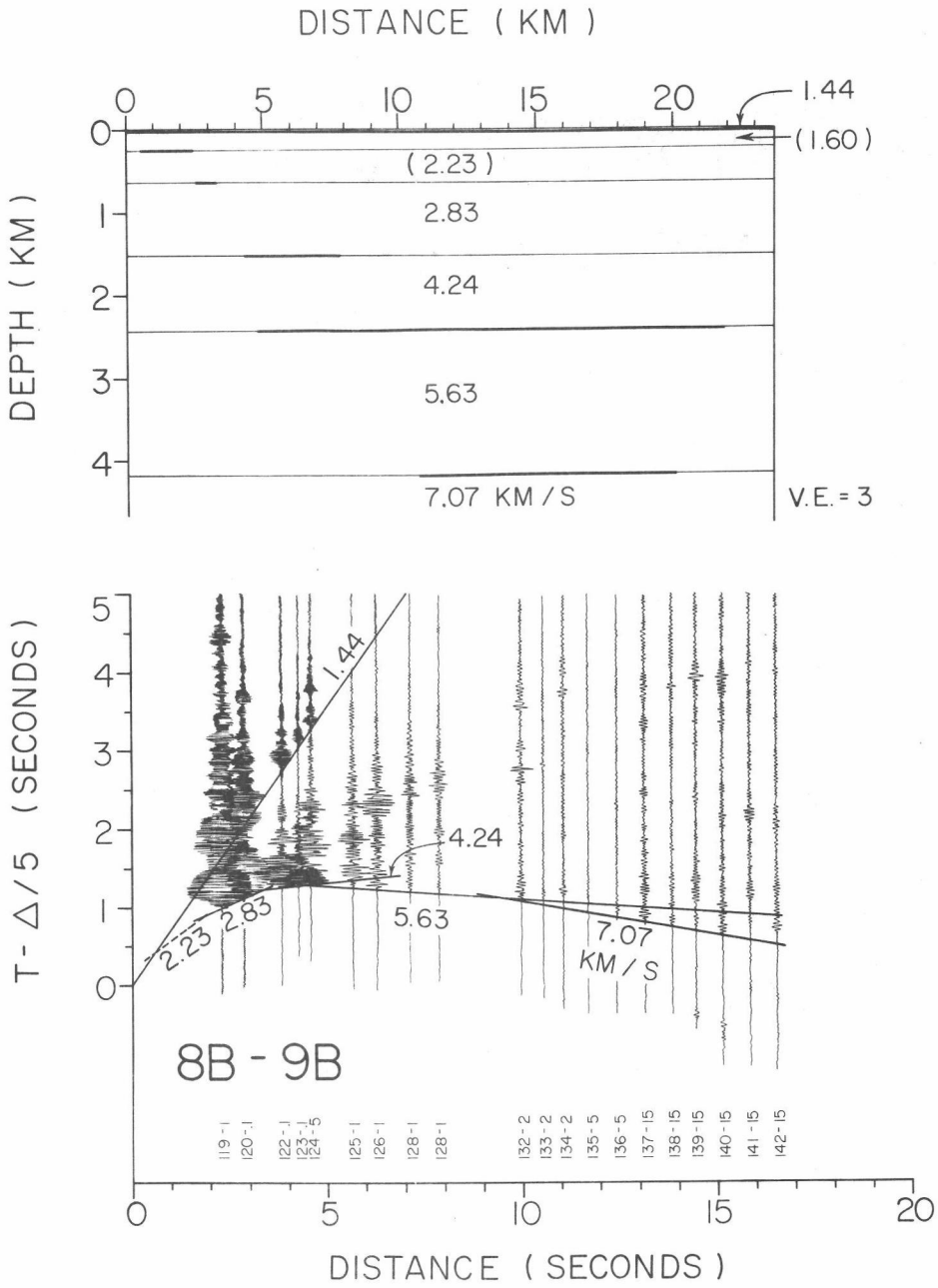


Figure 12. Line 8B-9B record section and velocity-depth model interpretation.

2.83 km/s refractor, 1.52 km for the 4.24 km/s refractor, 2.43 km for the 5.63 km/s refractor and 4.18 km for the 7.07 km/s refractor.

Lines 10-11 and 12-13

Situated between Cape Simpson and Cape Halkett, Line 10-11 (Figure 13) from west to east reverses Line 12-13 (Figure 14) from east to west. On both lines the upper layers are well-determined and easily reversed to obtain true velocities of 1.44 km/s, 1.65 km/s (apparent velocities of 1.65 km/s on both Line 10-11 and Line 12-13), 2.51 km/s (2.51 km/s on both lines), and 3.07 km/s (3.04 km/s on Line 10-11 and 3.09 km/s on Line 12-13). The deeper layers were more delicate to interpret and required manual ray tracing as described earlier. Very strong second arrivals indicate velocities of 4.06 km/s on Line 10-11 and 3.80 km/s on Line 12-13. Unfortunately, these two apparent velocities do not satisfy the reciprocity condition and both layer velocity and layer dip are assumed to change along the profile. The arrivals have been attributed to a 4.14 km/s velocity layer with a downdip of 1° on Line 10-11, and to a horizontal refractor with a velocity of 3.79 km/s on Line 12-13. For the deepest layer, the layer dip and velocity for the western end were adjusted to obtain the observed 6.91 km/s apparent velocity to the east on Line 10-11 and the 7.13 km/s apparent velocity to the west on Line 12-13. The result is a refractor with a true velocity of 7.02 km/s which dips up to the east with a 0.08° dip. Repeating this procedure at the eastern end, a 6.00 km/s apparent velocity on Line 12-13 and 6.91 km/s apparent velocity on Line 10-11 result in

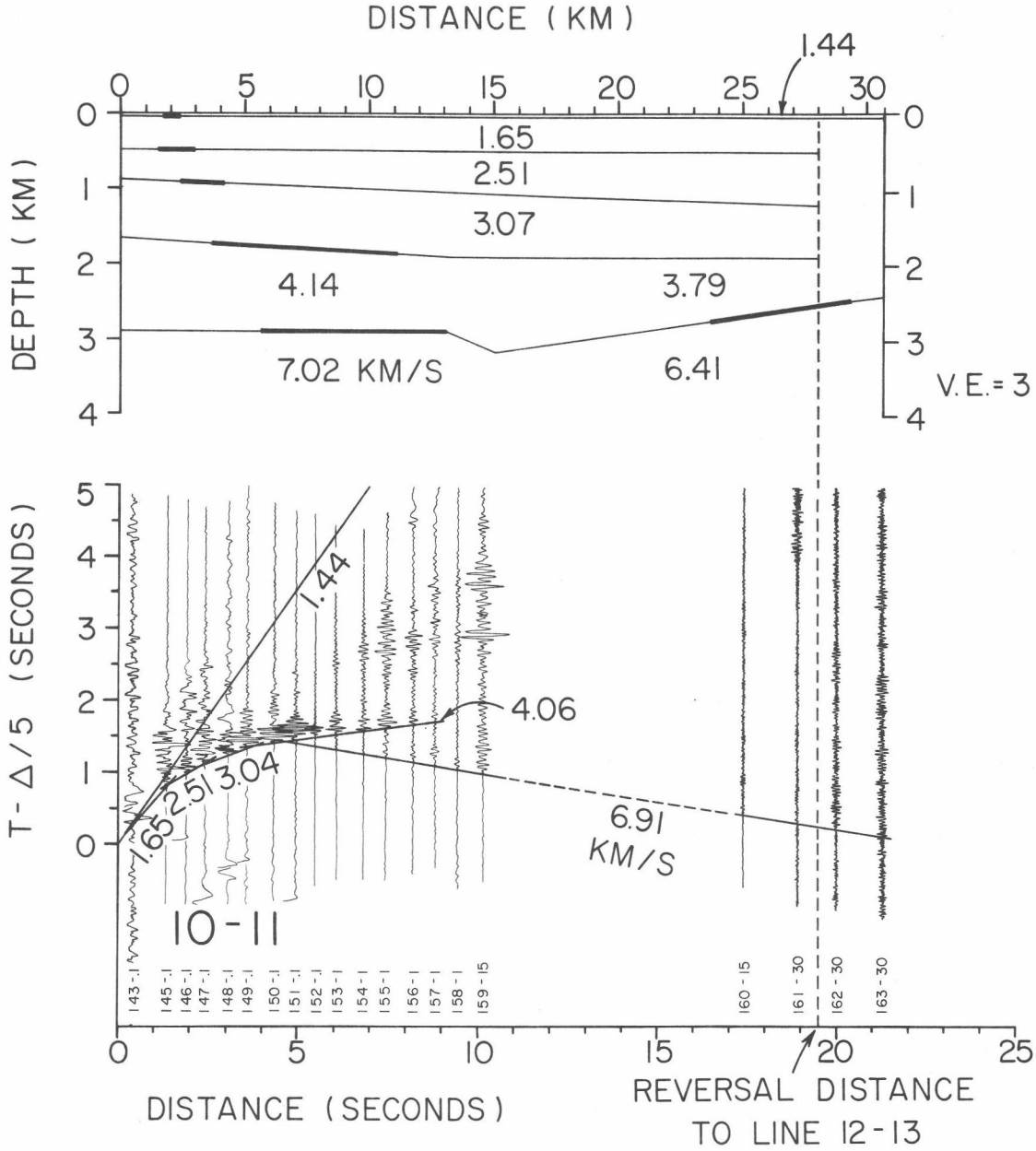


Figure 13. Line 10-11 record section and velocity-depth model interpretation.

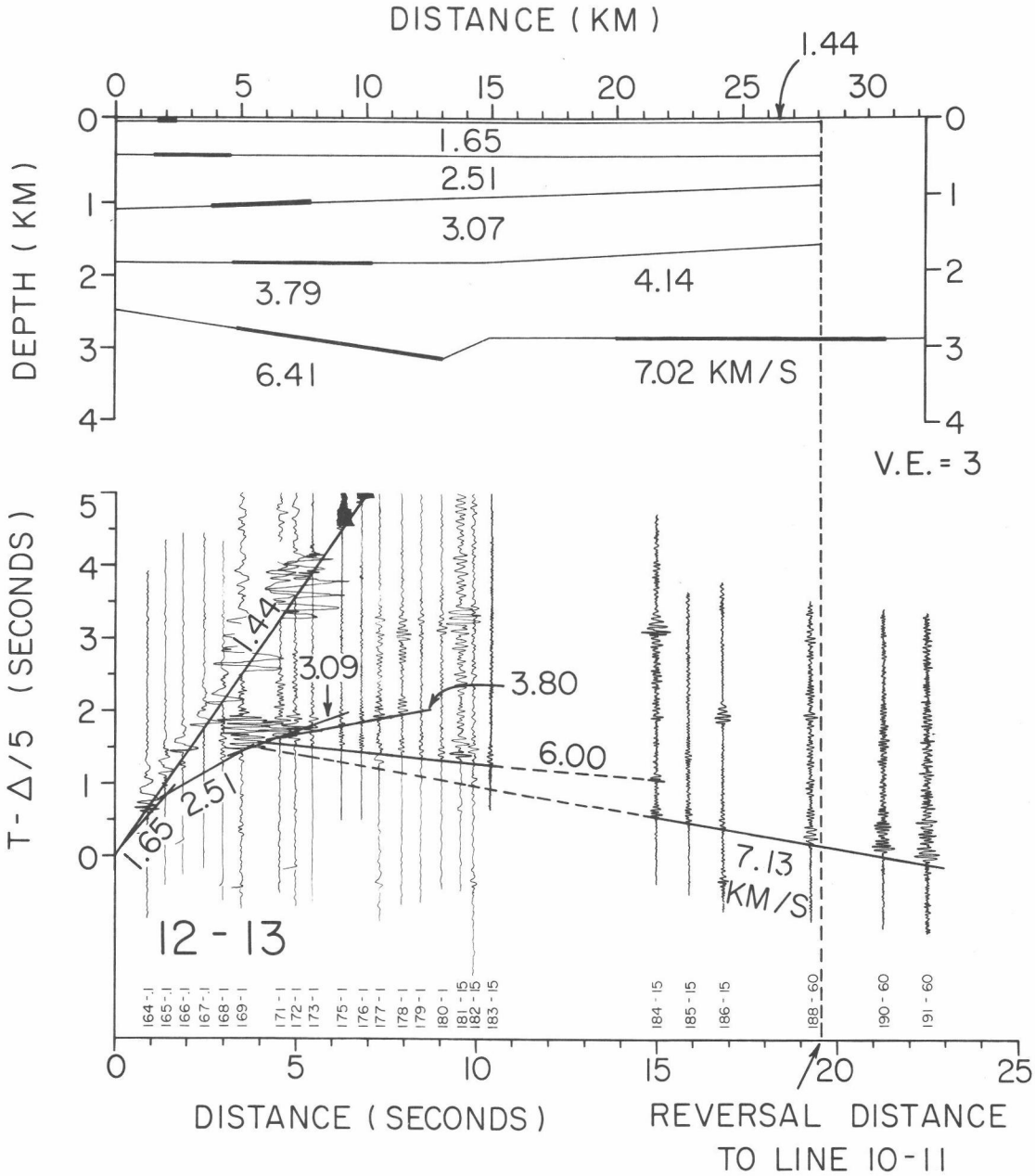


Figure 14. Line 12-13 record section and velocity-depth model interpretation.

a 6.41 km/s refractor with a large dip of 3.2° down to the west. This interpretation of the deep layers is not esthetically pleasing because the 6.91 km/s velocity arrivals seem to be continuous on Line 10-11 and hence should form a single continuous layer. However there appears to be no other easy way to explain the lack of 7.13 km/s arrivals in the middle of Line 12-13. The reversal travel time for the model agrees with the observed travel time to within .05 sec. The absence of data in the middle of these two lines is due to helicopter launches while the ship continued underway. The resulting seismic model has six layers including the water layer with a 1.44 km/s velocity, a 1.65 km/s layer at a 0.01 km depth, a 2.51 km/s layer at a 0.68 km depth, a 3.07 km/s refractor dipping down to the east from depths of 0.88 km to 1.13 km, a fifth layer with a 4.14 km/s velocity on the west side at approximately a 1.80 km depth and a 3.79 km/s velocity on the east side at a 1.89 km depth, and the deepest layer with a velocity of 7.02 km/s on the west side with a slight updip to the east at about a 2.90 km depth and a velocity of 6.41 km/s on the east side with a steep updip to the east from depths of 3.15 km to 2.53 km.

Line 16-17

Line 16-17 shown in Figure 15 lies close to Cape Halkett and extends for a distance of 30 km to the east. The records are slightly noisy and the only two velocities really well-determined are a 1.82 km/s velocity based on first arrivals and a 3.62 km/s velocity based on strong first and second arrivals. An intermediate

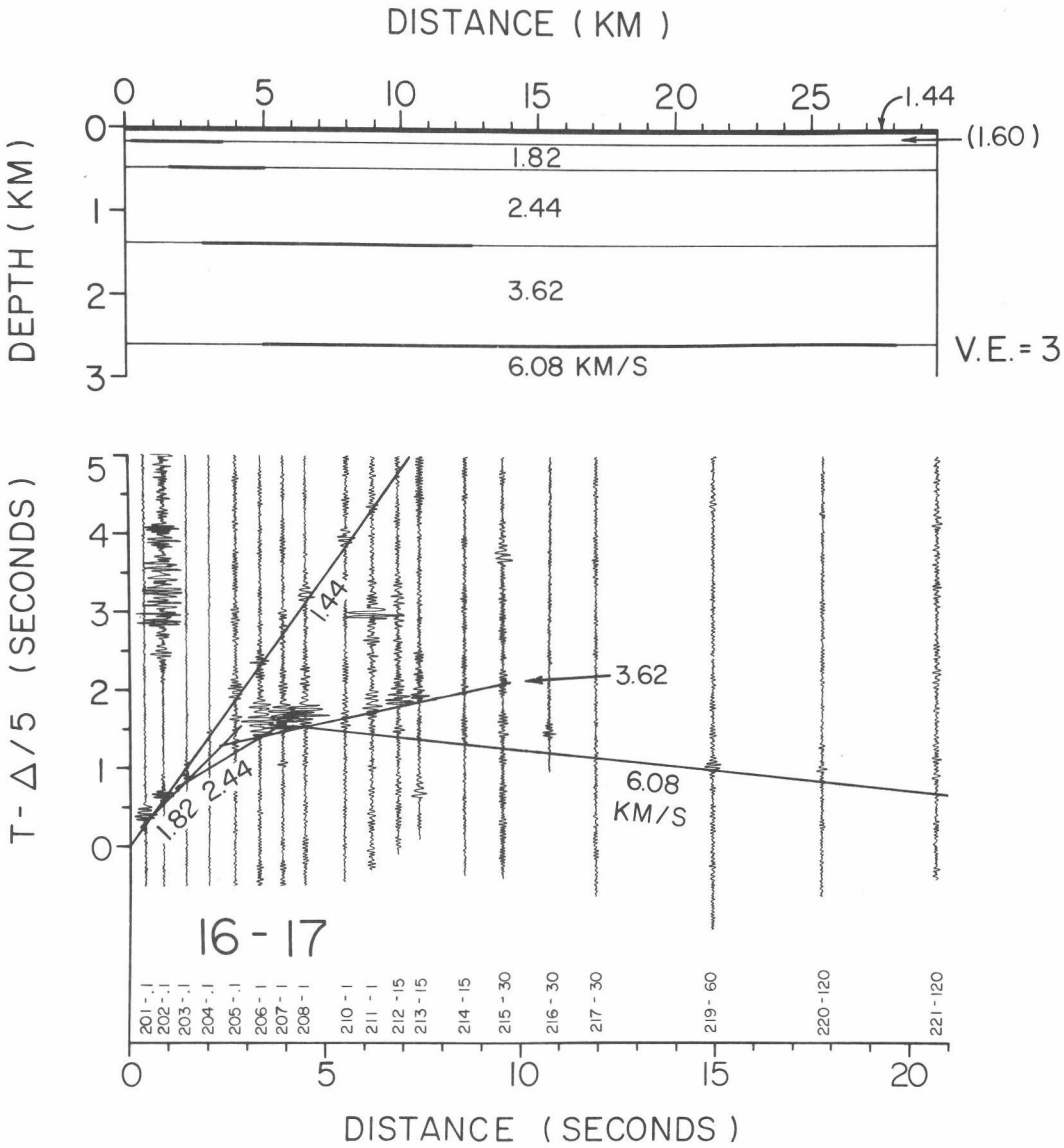


Figure 15. Line 16-17 record section and velocity-depth model interpretation.

2.44 km/s velocity and a higher 6.08 km/s velocity are suggested but the evidence is weaker. The velocity structure is composed of six horizontal layers, in addition to the 1.44 km/s and 1.60 km/s layers a 1.82 km/s interface at a 0.16 km depth, a 2.44 km/s at a .47 km depth, a 3.62 km/s at a 1.38 km depth and a 6.08 km/s at a depth of 2.60 km.

Line 18-19

Line 18-19 shown in Figure 16 is the easternmost line of the study and extends from close to Prudhoe Bay towards the west. The poor quality of the records of Line 20-21 did not permit a reversal with Line 18-19 as initially planned. On Line 18-19, seen as first arrivals close to the origin, are arrivals with an apparent velocity of 1.84 km/s followed by strong first arrivals which indicate apparent velocities of 2.32 km/s and 3.31 km/s. These are followed by a long segment with a velocity of 5.66 km/s. The corresponding seismic model has a length of 24 km and has six horizontal layers including 1.44 km/s and 1.60 km/s layers. At a depth of 0.20 km lies a layer with a velocity of 1.84 km/s, underlain by a layer with 2.32 km/s velocity at 0.63 km depth. The fifth and sixth layers with velocities of 3.31 km/s and 5.66 km/s are respectively at 1.60 km and 3.36 km depths.

Line 20-21

Line 20-21 shown in Figure 17 runs from west to east close to Oliktok Point. The records are very noisy, because of a substandard

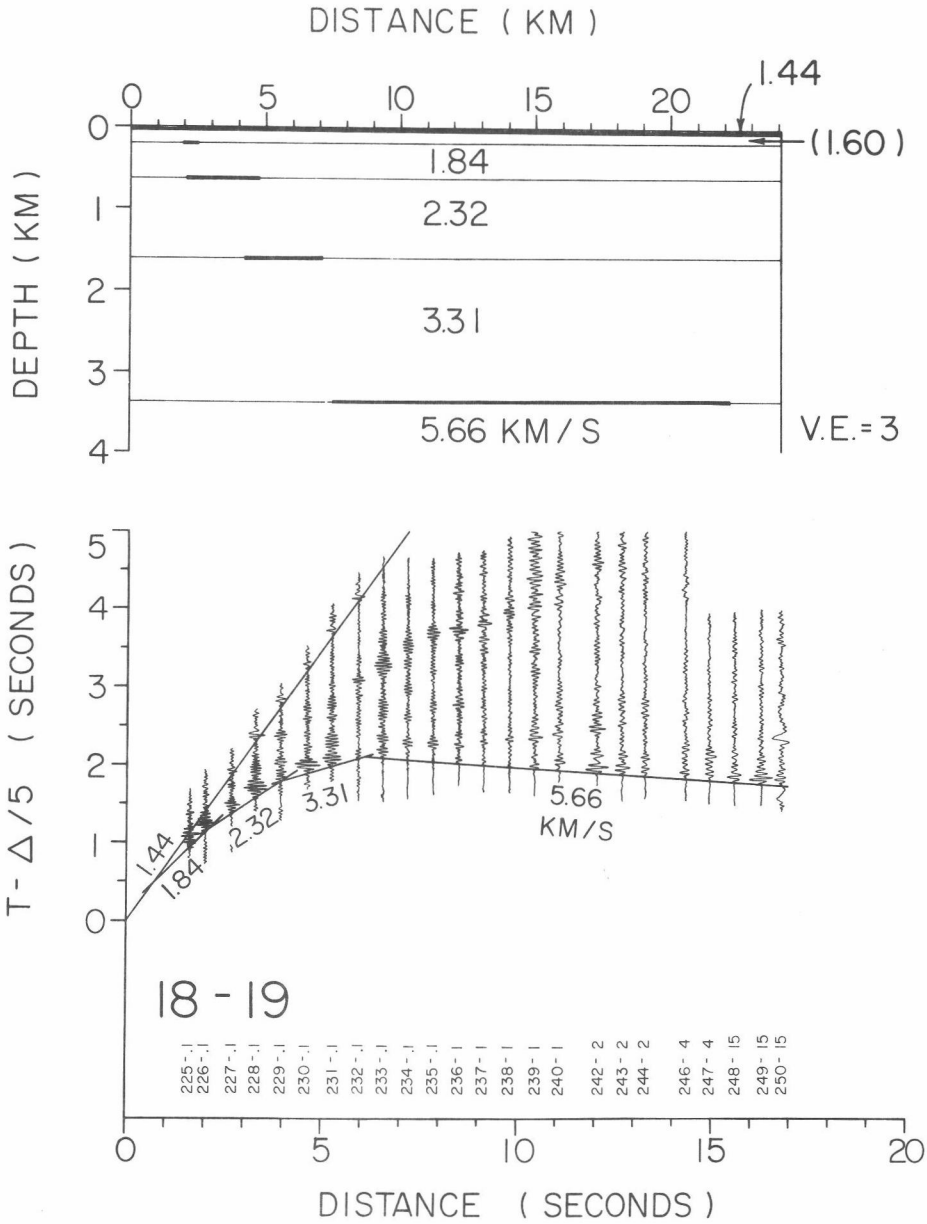


Figure 16. Line 18-19 record section and velocity-depth model interpretation. Second arrivals between 6 and 14 seconds do not correspond to any good refractor but may be sideswipe arrivals.

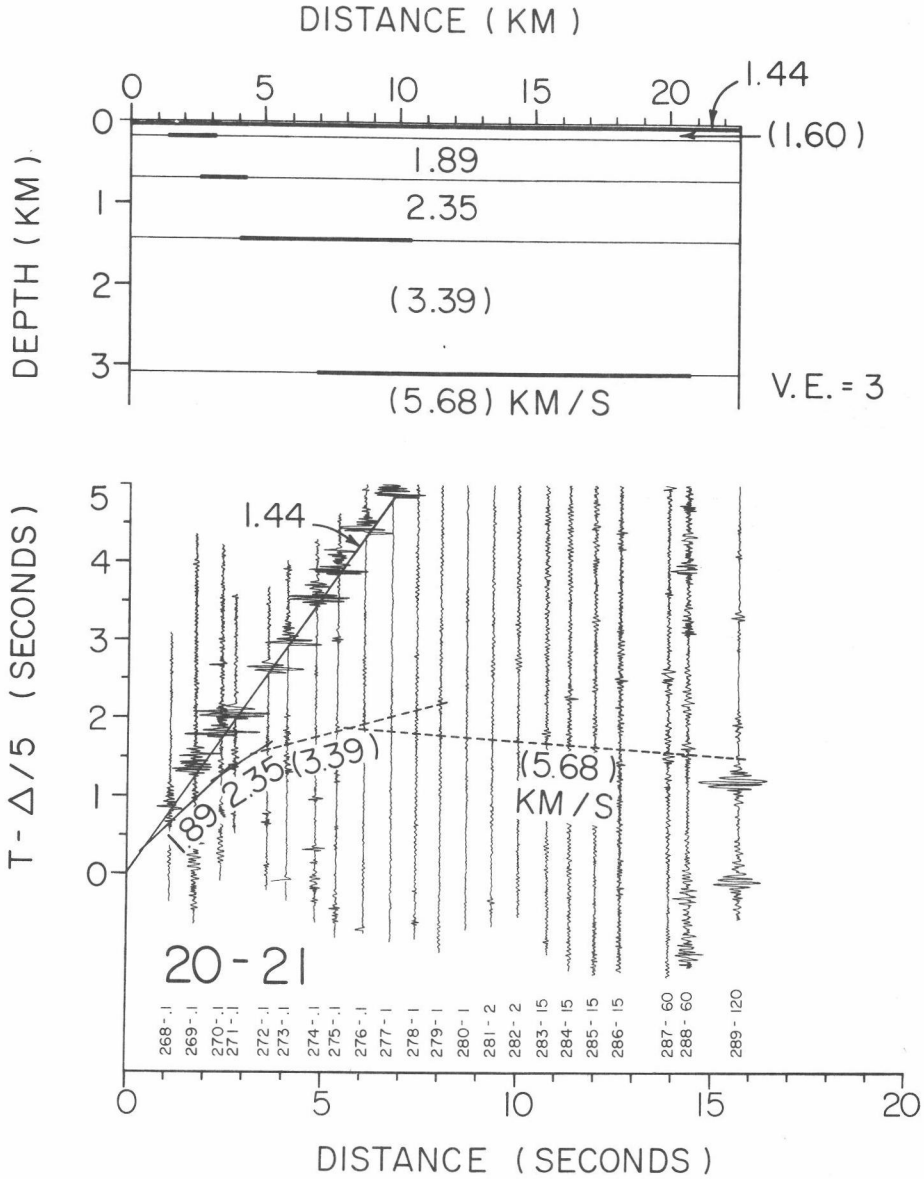


Figure 17. Line 20-21 record section and velocity-depth model interpretation.

sonobuoy and most velocities have been estimated by analogy to Line 18-19. The corresponding seismic model is 23 km long and has six layers with velocities of 1.44 km/s, 1.60 km/s, 1.89 km/s, 2.35 km/s, 3.39 km/s and 5.68 km/s and corresponding interface depths of 0.02 km, 0.18 km, 0.69 km, 1.43 km and 3.06 km respectively.

Lines 22A-23 and 24B-25

In Harrison Bay, Line 22A-23 (Figure 18) from west to east and Line 24B-25 (Figure 19) from east to west are reversed over a distance of 46 km. Line 22A-23 shows strong arrivals over its total length of 69 km and first seen are two first arrival apparent velocities of 1.83 km/s and 2.35 km/s. Apparent velocity arrivals with a velocity of 2.93 km/s are based mainly on strong second arrivals and the highest apparent velocity of 5.78 km/s can be seen almost over the entire remaining length of the line.

Line 24B-25 has less data but shows strong arrivals for the determination of 2.31 km/s, 2.88 km/s and 6.03 km/s apparent velocities. A 1.83 km/s velocity had to be estimated because no data exist close to the origin. The large amplitude and high frequency of the waves at the beginning of the line prevented the tracing of the entire seismograms, but the water waves visible later on individual seismograms permitted calculation of the shot distance for the reduced section. The corresponding velocity structure with true velocities and dipping layers is a six-layer model with layers of velocities 1.44 km/s and 1.60 km/s overlying a third layer of velocity 1.83 km/s with a 0.17° updip to the east from a 0.28 km to

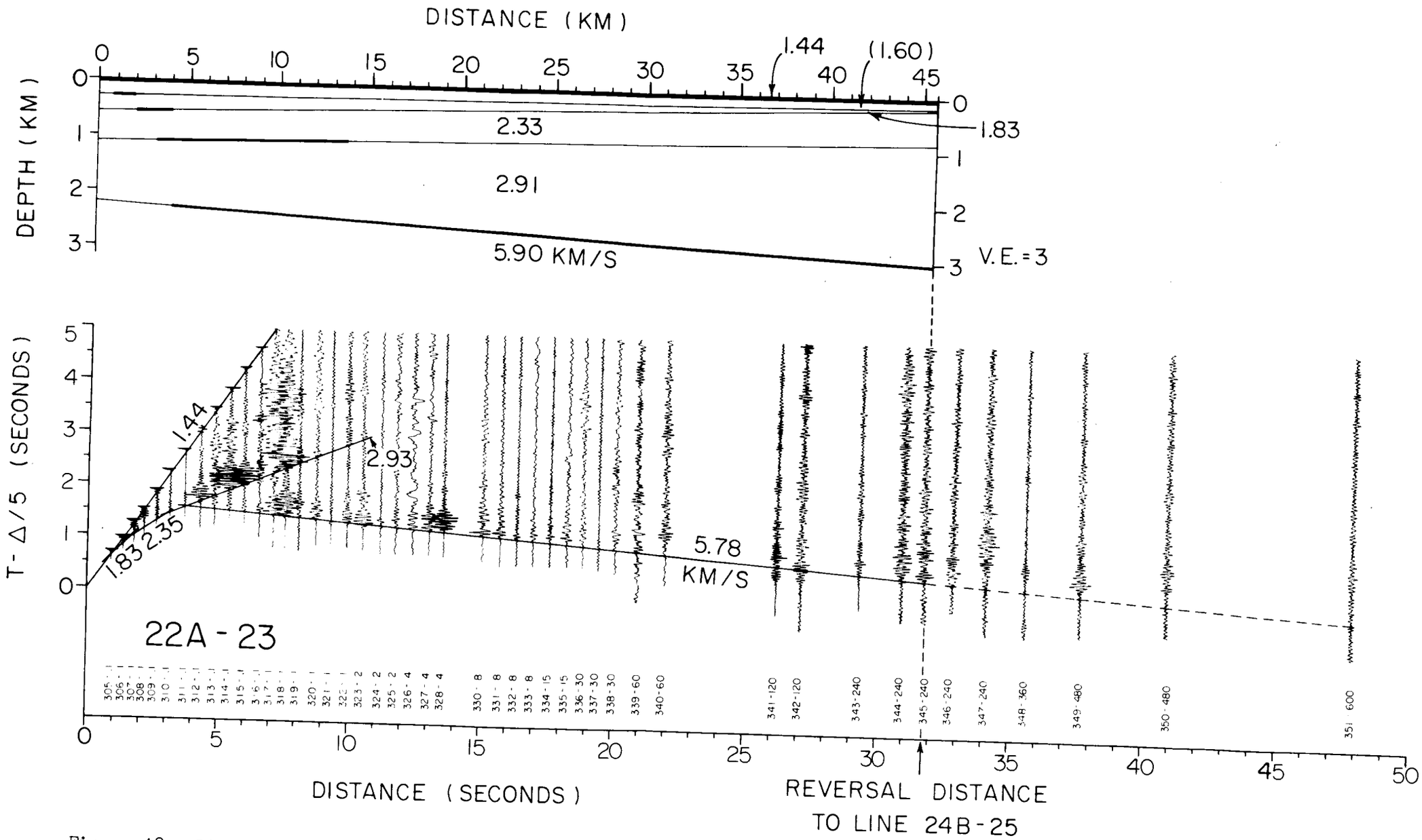


Figure 18. Line 22A-23 record section and velocity-depth model interpretation.

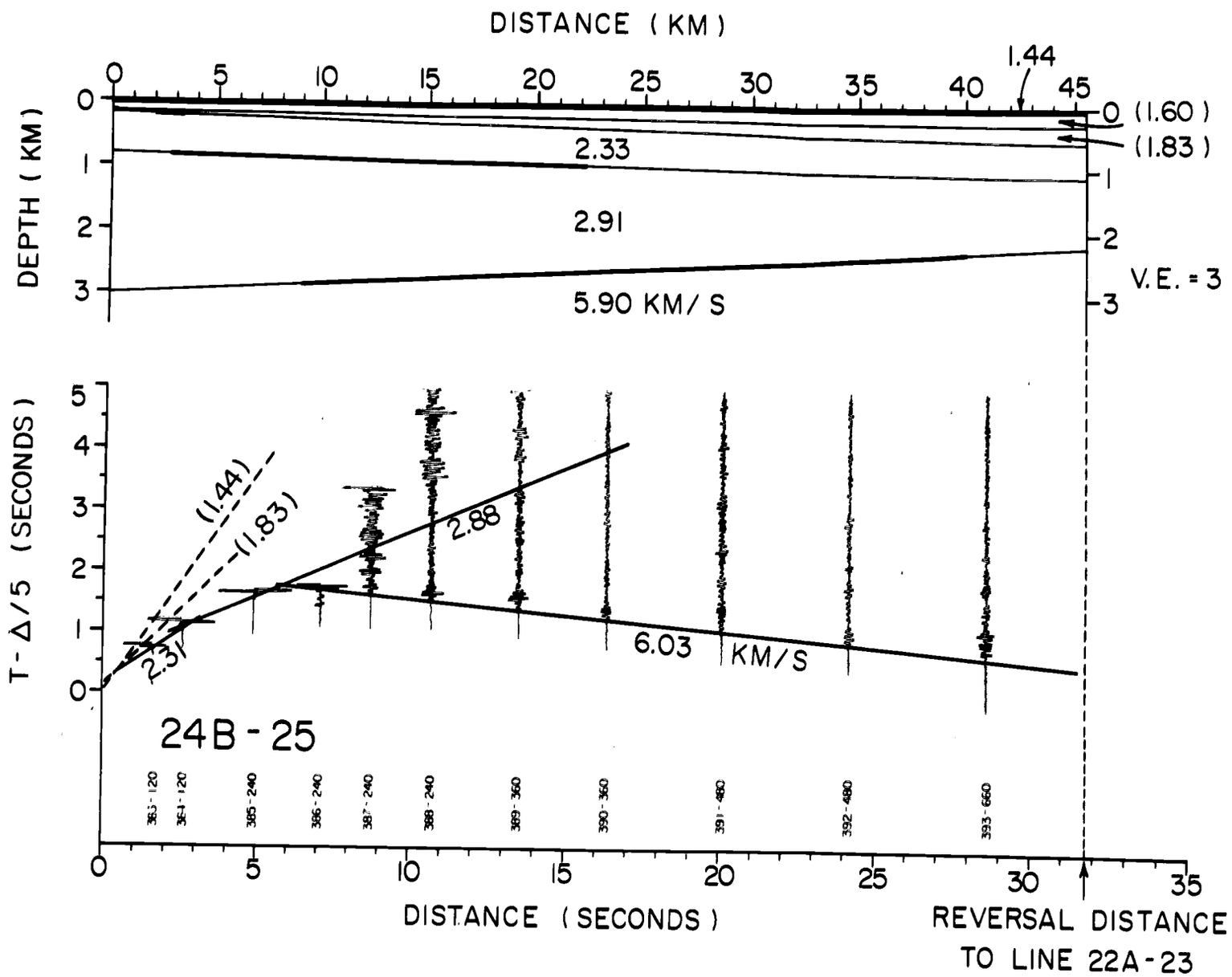


Figure 19. Line 24B-25 record section and velocity-depth model interpretation.

a 0.16 km depth. A fourth layer with a velocity of 2.33 km/s dips up to the east at a dip of 0.45° from a depth of 0.56 km at the west end to a depth of 0.19 km at the east end. Layer five has a true velocity of 2.91 km/s and also shows updip to the east of 0.32° from a depth of 1.10 km to 0.83 km. The deepest layer has a velocity of 5.90 km/s and the interface dips down to the east with a 1° dip from a depth of 2.20 km to 3.03 km.

Lines 24A-25 and 22B-23

Also situated in Harrison Bay, Line 24A-25 (Figure 20) from east to west and Line 22B-23 (Figure 21) from west to east reverse over a distance of 41 km. On Line 24A-25, first arrivals close to the origin and second arrivals at greater distances show a 1.94 km/s velocity and a 3.40 km/s velocity. Evidence for a higher velocity of 5.75 km is less convincing on these records because of noisy traces and a lack of strong clear arrivals. Line 24A-25 out to a distance of 75 km is the longest line of the study and helicopter records on the most distant portion show a 6.51 km/s velocity. The corresponding velocity on Line 22B-23 was not seen and hence the interpretation had to be treated with manual ray tracing. The strong first arrivals of Line 22B-23 show clearly 1.94 km/s, 3.39 km/s and 5.69 km/s velocities. In contrast to the other lines situated in Harrison Bay (Lines 22A-23 and 24B-25), the seismic model shows a lateral change in the velocity structure in its upper part. For example, the 2.33 km/s velocity is not seen and the 1.83 km/s and 2.91 km/s velocities increase respectively to 1.94 km/s and 3.40

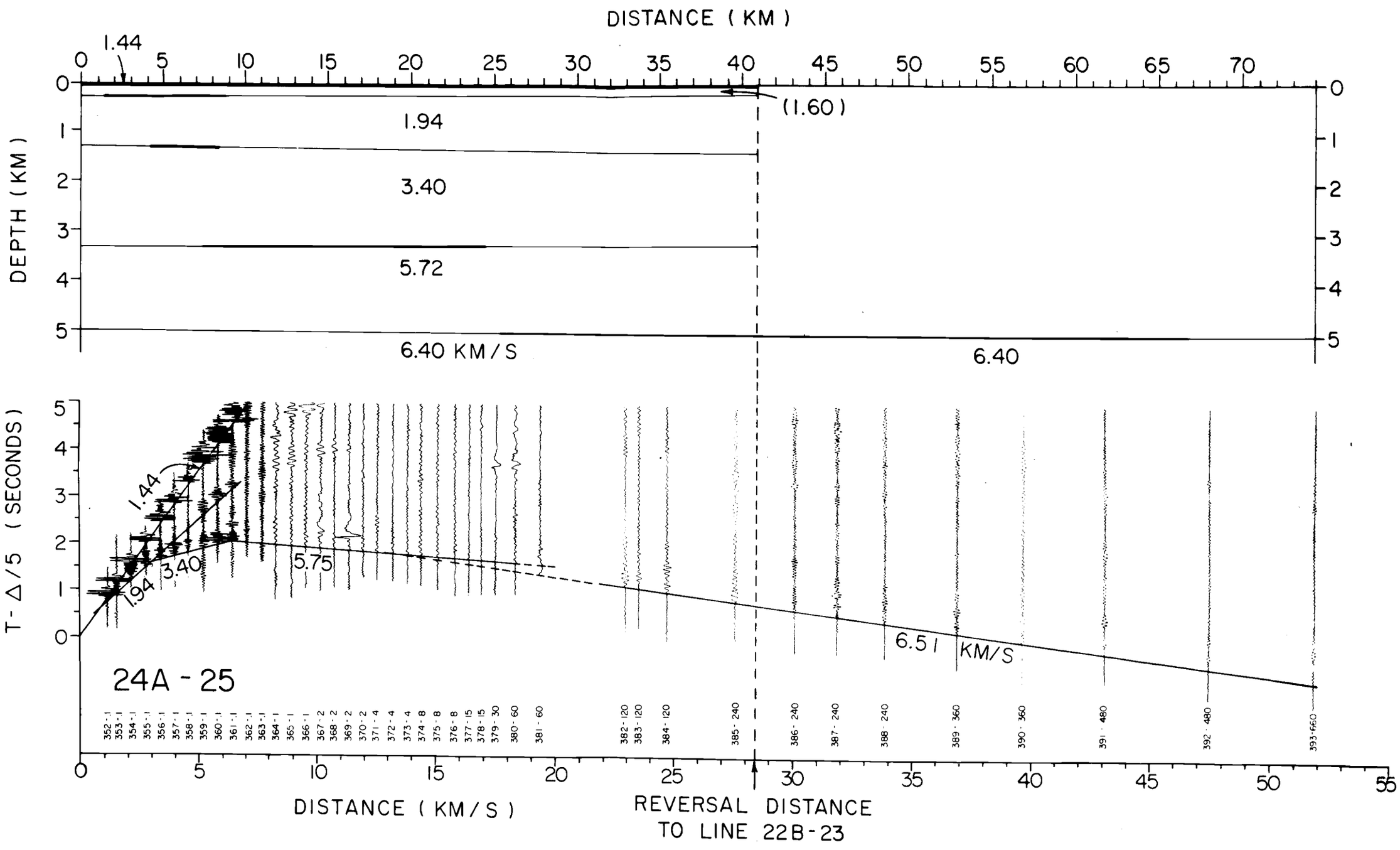


Figure 20. Line 24A-25 record section and velocity-depth model interpretation.

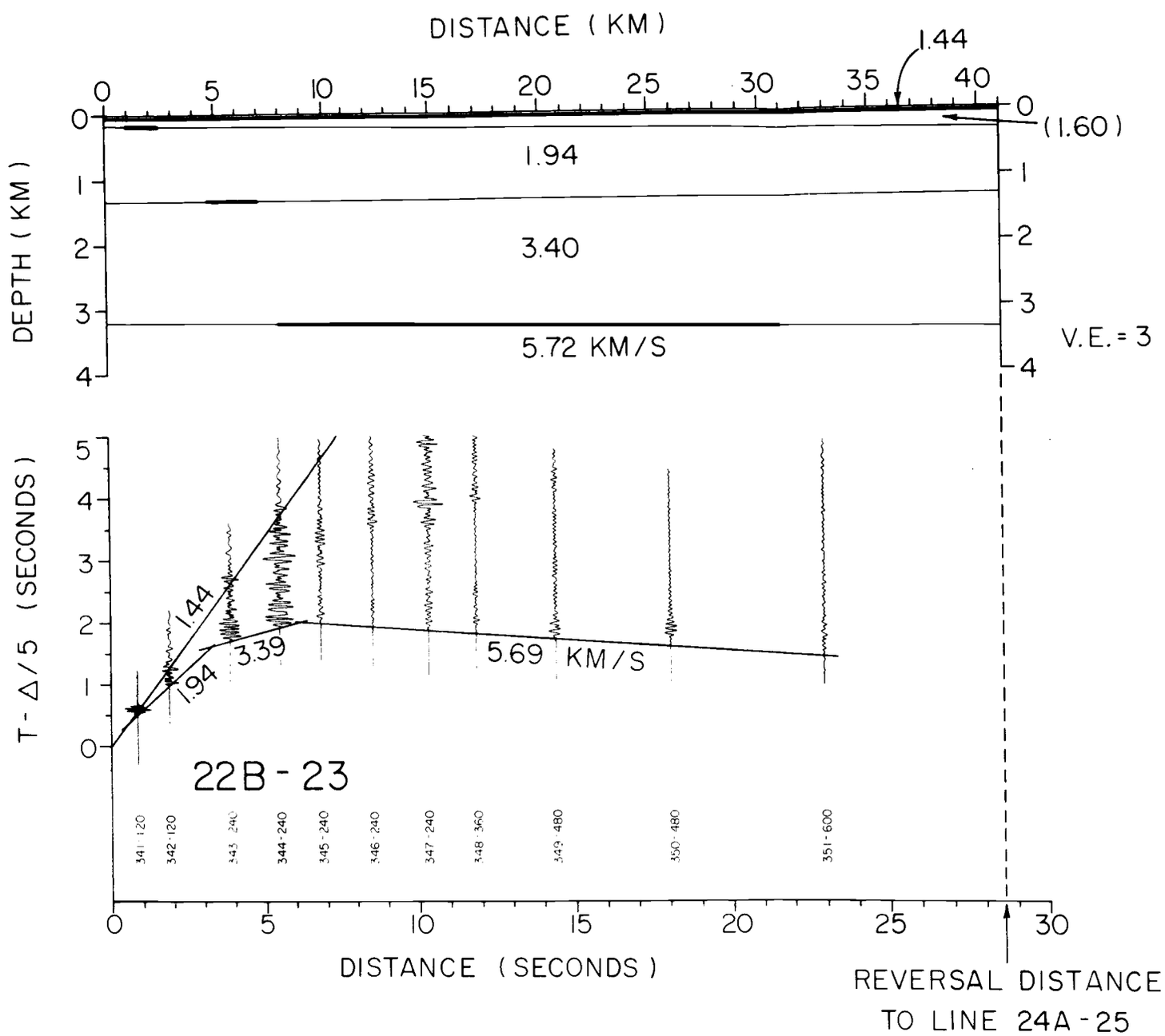


Figure 21. Line 22B-23 record section and velocity-depth model interpretation.

km/s. Although the 5.90 km/s apparent velocity is similar (5.72 km/s for model 22B-24A), there is an additional higher velocity of 6.40 km/s. The 1.94 km/s dips down to the east with a 0.17° dip from 0.16 km to 0.30 km and the 3.40 km/s layer dips up slightly at an angle of 0.03° to the east from a depth of about 1.32 km. The 5.72 km/s layer slopes down to the east with a 0.18° dip and from a depth of 3.19 km. The apparent velocity of 6.51 km/s had to be obtained from the two reversed models, where the downgoing ray travels through model 22B-24A and upgoing ray travels through model 22A-24B, by ray tracing. The best fit is a horizontal layer at a depth of 5 km with a true velocity of 6.40 km/s. The travel time at the reversal distance from the model agrees with the observed travel time to within .04 sec.

Line 26A-27A

Line 26A-27A shown in Figure 22 was profiled from east to west and is situated at the western end of the study area close to Cape Simpson but more offshore than Lines 8A-9A and 8B-9B. A 1.60 km/s velocity is weakly determined but is based on one first arrival and one second arrival. The subsequent first arrivals are strong and clear and they show apparent velocities of 2.13 km/s and 2.99 km/s. Strong second arrivals have a 4.14 km/s apparent velocity while 5.28 km/s apparent velocity arrivals are weak because of the noisy traces of the end of the line. The seismic model extends to a distance of 32 km and to a depth of about 3 km. The six layers consist of the water layer and a layer with a velocity of 1.60 km/s at 0.02 km,

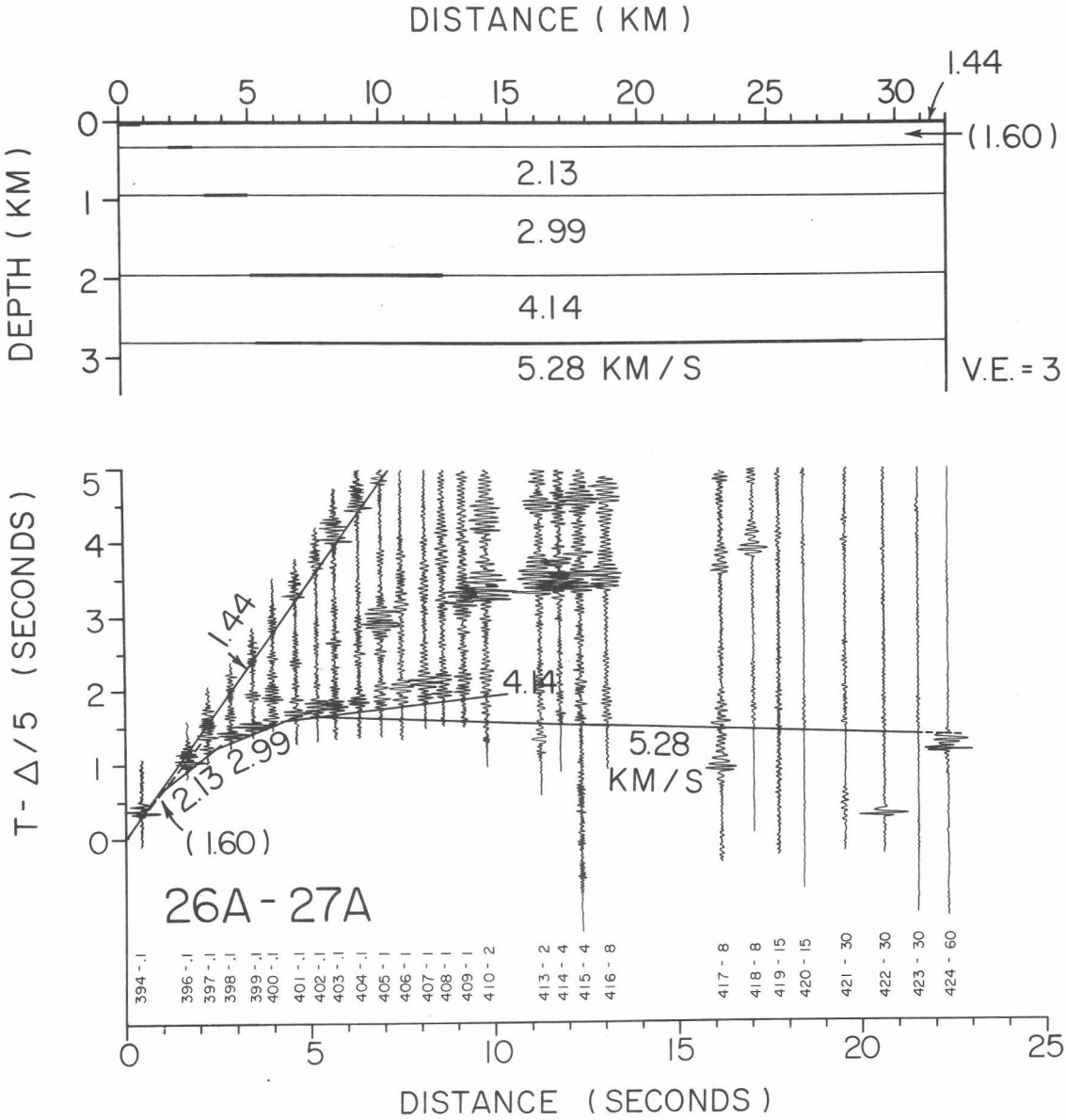


Figure 22. Line 26A-27A record section and velocity-depth model interpretation.

underlain by layers with velocities of 2.13 km/s and 2.99 km/s at depths of 0.32 and 0.96 km respectively. The deepest layers have velocities of 4.14 km/s and 5.28 km/s at depths of 1.95 km and 2.82 km respectively. The general ringing nature of the traces is typical of shelf profiles due to uniform layering, shallow water and the bubble pulse associated with the explosion.

Line 26B-27B

This line shown in Figure 23 is the western extension of Line 26A-27A and is 34 km in length. Seen as first arrivals close to the origin are waves with velocities of 1.60 km/s, 2.10 km/s and 3.43 km/s. These are followed by a segment with a velocity of 4.43 km/s based on strong first arrivals. Unfortunately, a higher velocity of 6.88 km/s is poorly determined by weak first arrivals. The velocity structure contains six layers with velocities of 1.44, 1.60, 2.10, 3.43, 4.43, and 6.88 km/s and corresponding interfaces at depths of 0.02, 0.44, 1.20, 2.16 and 5.40 km.

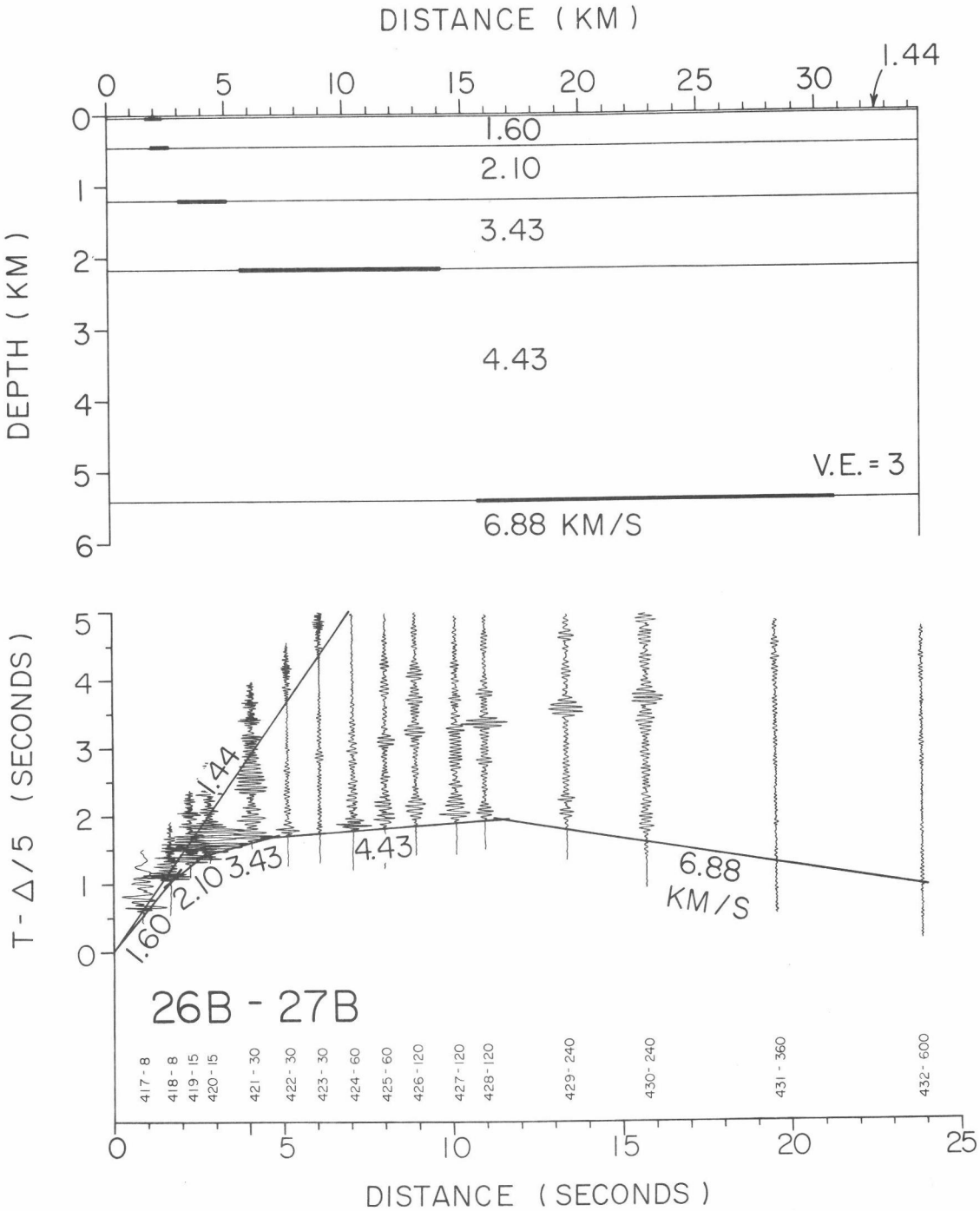


Figure 23. Line 26B-27B record section and velocity-depth model interpretation.

INTERPRETATION

The composite section shown in Figure 24 summarizes the subsurface velocity and depth information calculated from the refraction data. The heavy lines on the interfaces, corresponding to the observed arrivals as shown in Figures 11 through 23, were projected onto a composite section which passes through the profiles. The small arrows along each interface correspond to the appropriate refraction lines shown at the top of the figure. The refraction layers from adjacent lines have been correlated on the basis of velocity as indicated in Figure 24 by light lines. In general, the section thickens to the east.

Figure 25 is a section at the west end of the section of Figure 24 which shows velocity and depth changes between two near-shore lines and the continental slope.

The eastward and northward dip of the structure agrees with earlier geological and geophysical (but largely non-seismic) studies in the same area (Ostenso, 1962; Woolson et al., 1962; Hunkins, 1966; Rickwood, 1970; Ostenso and Wold, 1971; Morgridge and Smith, 1972).

To the east, beneath a layer at the sea bottom with an assumed velocity of 1.60 km/s to 1.65 km/s lies an additional layer with a velocity of 1.82 to 1.94 km/s. This layer is not seen to the west either because it is truncated or because it is too thin to be detected with this data. A uniform layer with velocities of 2.10-2.51 km/s overlies a 2.91-3.43 km/s velocity layer which thickens to the east. A layer seen only on the western part of the cross-section has

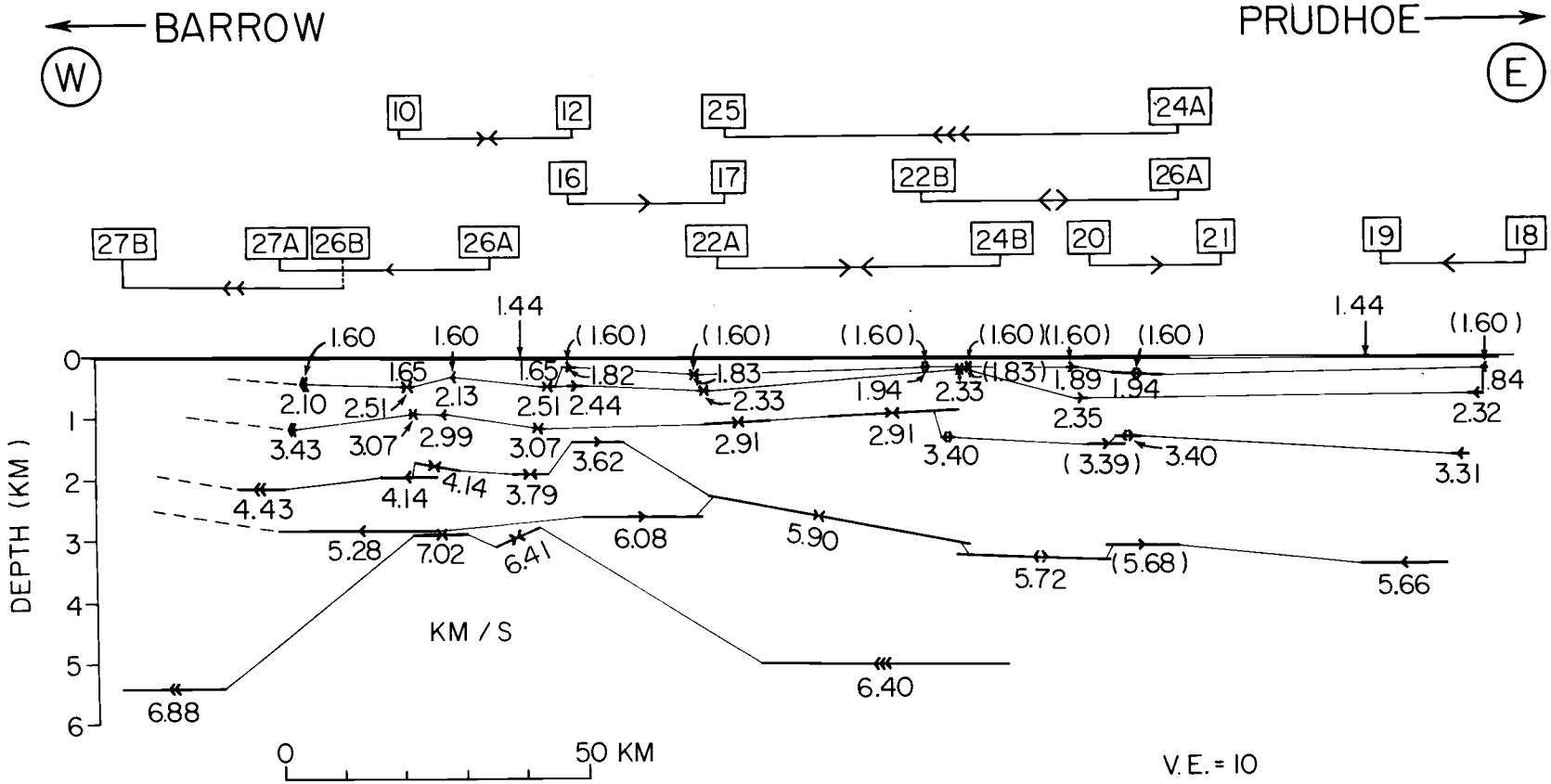


Figure 24. East-west velocity-depth section summarizing the subsurface velocity and depth information calculated from the refraction data.

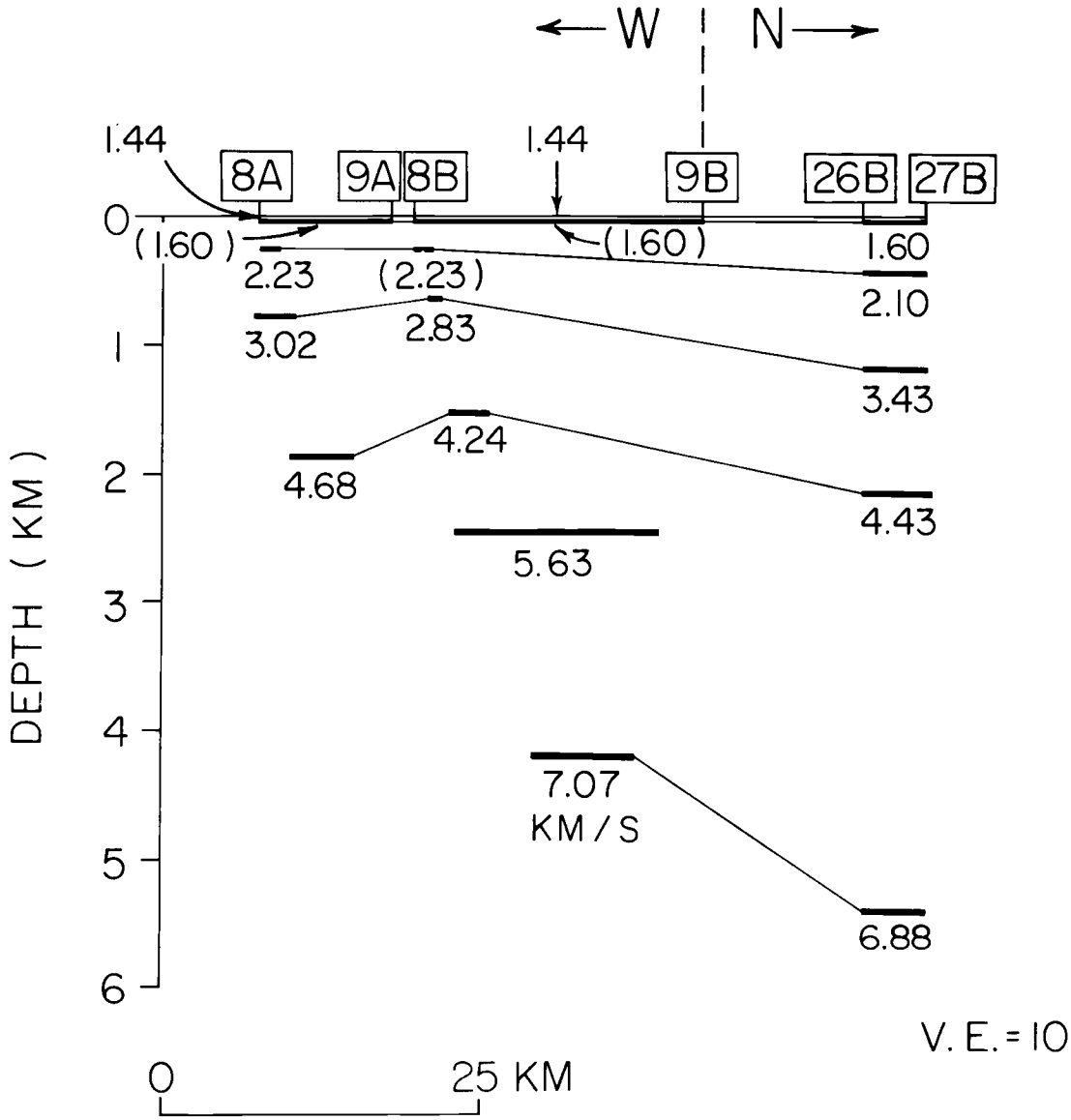


Figure 25. Velocity-depth section north of Smith Bay.

velocities ranging from 3.62 to 4.43 km/s. As it has a moderate thickness (.5 to 1.0 km), that it is not seen east of station 17 suggests that this layer may be truncated by erosion to the east. A high velocity basement (5.28-6.08 km/s) layer lies deeper to the east. A higher velocity layer of 6.40 to 7.02 km/s shallows beneath stations 10 to 12 but is not detected beyond station 24 to the east.

A correlation between the velocities measured in this study and the age of the formations as determined from test wells drilled on land and other previous studies is shown on a geologic cross-section in Figure 26. It appears that the general structural homocline to the ESE present onshore extends offshore also. The first layer with velocities of 1.60-1.65 km/s is probably sediments of the Gubik Formation deposited during the Quaternary. We interpret the second layer with velocities of 1.82-2.51 km/s as Tertiary in origin and as belonging very likely to the Sagavanirktok Formation. This agrees with the results of Grantz et al. (1975) from an airgun sonobuoy profile parallel to the western Beaufort coast 30 miles offshore where an observed 1.8-2.9 km/s velocity layer is correlated with the Tertiary section observed in the Prudhoe Bay wells. The base of the gently eastward-dipping Tertiary layer lies at depths of .8 to 1.3 km and overlies a thick Mesozoic layer which also dips to the east. The base of the Mesozoic layer lies at a depth of 2.4 km at the center of the cross-section and deepens to 3.2 km near Prudhoe Bay. Near Prudhoe Bay this layer is probably composed of primarily Upper Cretaceous sands and shales of the Colville and Nanushuk

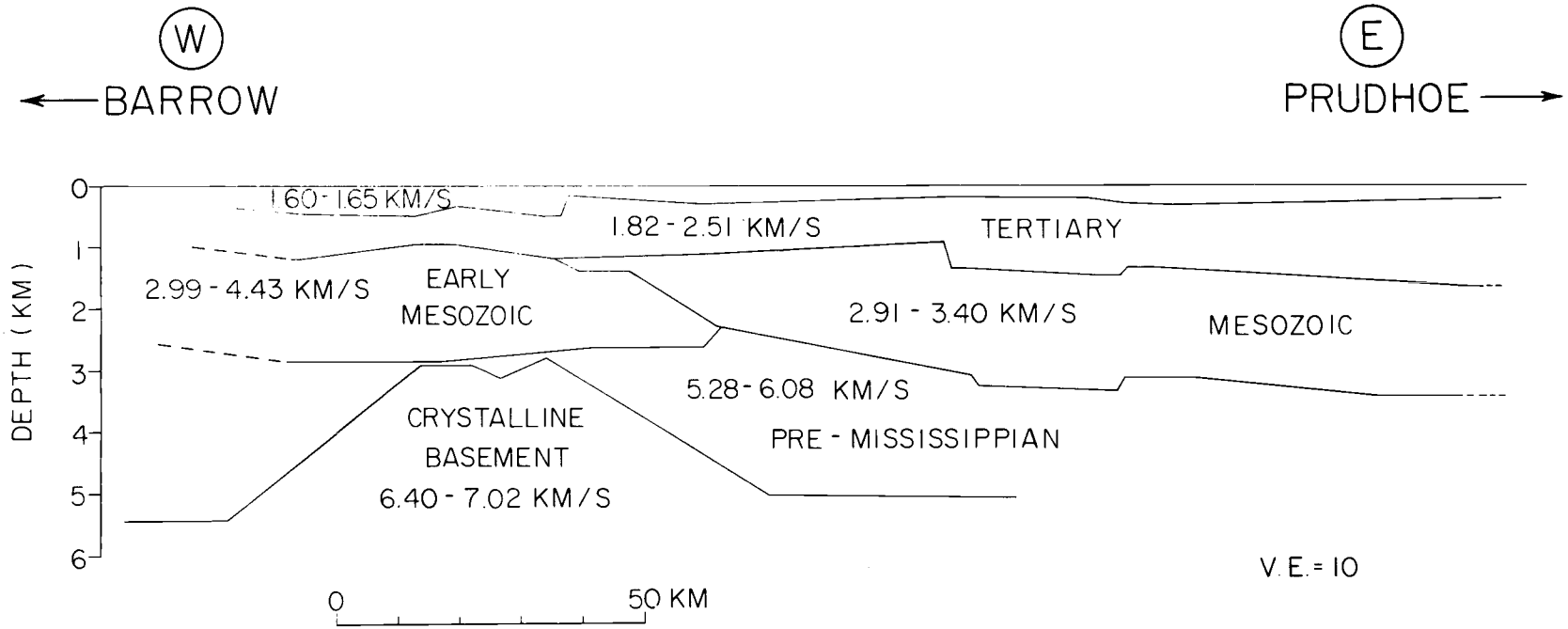


Figure 26. Velocity-depth-age cross-section summarizing the correlation between the seismic velocities and the geologic age of the formations.

Groups, and possibly also of Jurassic and Triassic sediments (Rickwood, 1970). The Lower Cretaceous is absent from wells on the east and has most probably been removed by erosion. The Upper Cretaceous layer thins to the west and is absent in wells at Point Barrow. The Upper Cretaceous layer is either absent or too thin to be seen on the western refraction profiles. Early Mesozoic formations to the west include mainly Lower Cretaceous sandstones and shales of the Oumalik and Torok Formations and also Jurassic shales of the Kingak Formation, reported to be 0.5 km thick along the coastline parallel to the cross-section (Rickwood, 1970; Morgridge and Smith, 1972), and Triassic sediments of the Shublik Formation. The velocities (2.91-4.43 km/s) and the depths of the Mesozoic layer are in good agreement with previous work done in this area, to the west in the Barrow area and to the east in Canada (Robinson, 1959; Robinson and Collins, 1959; Collins, 1961; Woolson et al., 1962; Sander and Overton, 1967; Rickwood, 1970; Hofer and Varga, 1972; Morgridge and Smith, 1972; Grantz et al., 1975; Johnson and Chiburis, 1977). The Permian and Carboniferous sediments of the Lisburne Group are likely to be present at the base of the Mesozoic and Early Mesozoic layers, but form probably a too thin layer to be detected with the refraction method. Informations from test wells indicate that the basement dips eastward and lies at a depth of 0.8 km at Barrow (Dana, 1951; Woollard et al., 1960; Morgridge and Smith, 1972; Gryc, 1970), 2 km at Cape Simpson (Robinson, 1959), about 3 km at Topagoruk (Gryc, 1970) and more than 3 km at Prudhoe Bay (Rickwood, 1970; Brosgé and Tailleur, 1970; Morgridge and Smith, 1972). Hence the

layer which dips eastward at about these depths is probably the Pre-Mississippian basement composed of argillite and phyllite with P-wave velocities ranging from 5.28 to 6.08 km/s. Ostenso and Wold (1971) reported in the vicinity of our study area a 5.2 km/s velocity layer to be the argillite and phyllite basement. Woolson et al. (1962) found at Point Barrow a refractor with velocities of 5.5 km/s to 6.1 km/s and concluded that it was probably the Pre-Mississippian basement. Further east in the Canadian Shield, velocities of 6.0 km/s have been interpreted as the Pre-Mississippian basement (Sander and Overton, 1965) and west of Point Barrow, Johnson and Chiburis (1977) correlated a 5.1 km/s velocity to argillite and slate. Between Smith Bay and Harrison Bay, the argillite and phyllite sequence overlies a crystalline basement with velocities between 6.40 and 7.02 km/s. The crystalline basement is tentatively identified as silicic in character by comparison with velocities observed by others between Barrow and Cape Simpson (Ostenso and Wold, 1971), in western Canada (Hobson, 1962) and west of Barrow (Johnson and Chiburis, 1977). Ostenso (1962) interprets a gravity high trending parallel to the coast as due to a ridge of high density non-magnetic material - probably granite. This high velocity basement could also represent oceanic layer 3 (Overton, 1970). Finally, since velocities higher than 6.40 km/s are not detected out to distances of 75 km and since the upper mantle velocity in the region has been measured at 8.2 km/s (Sander and Overton, 1965; Overton, 1970; Berry and Barr, 1971; Mair and Lyons, 1978) a minimum depth estimate for the Mohorovicic discontinuity along the line of profiles is 20 km.

This is in general agreement with estimates from gravity measurements by Woollard et al. (1960) and Wold et al. (1970).

SUMMARY AND DISCUSSION

A steep continental slope defines the Alaska margin which borders the Beaufort Sea. The data presented here suggests that the structural homocline which dips to the east southeast on land extends to the offshore region as well. Correlation of geologic data from wells drilled on land with the refraction data permits tentative identification of geologic sequences on the basis of their seismic velocity. This study correlates 1.60 to 1.65 km/s layers to Quaternary formation, 1.82 to 2.51 km/s layers to Tertiary formation, 2.91 to 3.40 km/s layers to Mesozoic formation to the east and 2.99 to 4.43 km/s layers to Lower Cretaceous, Triassic and Jurassic formations to the west. Mississippian, Pre-Mississippian and pre-Devonian formations, which in this region form the base of the economically attractive stratigraphic section, correlate with velocities of 5.28 to 6.08 km/s. Relatively high velocities of 6.40 to 7.02 km/s from refractors at the greatest depths are probably related to crystalline material which may be silicic or mafic. Although no seismic velocities typical of the upper mantle are present on the record sections, a minimum depth calculation places the Mohorovicic discontinuity deeper than 20 km.

The velocity-depth section north of Cape Simpson (Figure 25) suggests a 3° dip down to the north on the north side of the Barrow Arch in contrast to the south-dipping strata south of it. The section parallel to the coastline (Figure 24) indicates Pre-Mississippian basement which dips 0.3° to the ESE. The stratigraphic

section east of Cape Halkett is most likely composed primarily of Upper Cretaceous and Tertiary material and forms a wedge which thickens to the east from 1.5 km to 3.3 km. Lower Cretaceous material is believed present only on the west end of the profile as Upper Cretaceous is absent on the west end. Both layers have certainly been eroded. The sedimentary section thins rapidly to the west and off the west end of the profile, Pre-Mississippian basement is detected in wells at a depth of 760 m at Point Barrow.

The crystalline basement which rises to a minimum depth of 3 km at Line 10-12 (Figure 24) exhibits lower velocities to the east than to the west (6.4 km/s versus 7.0 km/s). It is uncertain whether this change represents the variability of seismic velocity within material of the same composition or whether there is a change in composition. The small variation of the gravity field along the line suggests normal variability of velocity without change in composition. The anticlinal form of the crystalline basement may be an artifact caused by the line of profiles crossing the Barrow Arch at a very oblique angle. Hence, the western end of the profiles (Line 26-27) lies north of the Barrow Arch and the eastern end (Line 18-19) lies on its southern flank.

The observed crustal velocities are rather ambiguous toward theories of the origin of the Canada Basin and the tectonic history of the northern Alaska margin. The regional trend is parallel rather than perpendicular to the coastline perhaps influenced more by tectonics in eastern Alaska than by the proximity of the continental margin. The velocities of the crystalline basement do not

permit its identification as silicic or mafic. However, there is a perhaps coincidental agreement between crustal velocities observed in this study and those observed in the Canadian Archipelago (Hobson, 1962; Hobson and Overton, 1967; Sander and Overton, 1967; Overton, 1970; Hofer and Varga, 1972). If this comparison is valid, then the Orocline-Rift theory of Carey (1955) provides a simple but adequate explanation of the seismic observations and implies that the crystalline basement is silicic in character.

The lack of island-arc structures or andesitic petrology argues against a convergent or subduction origin for the northern margin. The steep bathymetric expression of the slope suggests either a convergent of strike-slip origin unless crustal rebound after rifting has maintained the steep physiography. The crystalline basement could conceivably be a remnant of partially subducted oceanic crust but the absence of magnetic character along the margin argues against such a possibility.

Although the refraction results of this study are inconclusive in resolving the problem of the origin of the Canada Basin, they tend to favor the Orocline-Rift theory over a subduction margin. Further data are required to resolve the remaining uncertainties and with the current economic interest and exploration activities in the region, new and crucial information is probably forthcoming.

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