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THE GEOLOGY OF THE PERSIAN GULF-GULF OF OMAN REGION: A SYNTHESIS

David A. Ross and Elazar Uchupi

Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

Robert S. White

Bullard Laboratories, University of Cambridge, United Kingdom

Abstract. During the Mesozoic most of the Arabian Peninsula, Persian Gulf, south-western Iran, and eastern Iraq constituted the Arabian platform. Deformation of the Musandam Peninsula in the Late Cretaceous and mid-Tertiary by compression (subduction) from the east and southwest, collision of the Arabian platform and Eurasian plate along the Zagros Crush zone during the Oligocene or early Miocene, and emplacement of the Zagros Mountains by gravitational sliding during the Neogene and Pleistocene have reduced the platform area to the Persian Gulf. Other factors that contributed to the reduction of the Arabian platform include the uplift of the Arabian Peninsula during the opening of the Red Sea in the Tertiary, tectonism of the Infracambrian Hormuz salt, upwarp of the platform sediment cover by basement uplift and/or salt tectonics, and a 600- to 400-m drop in sea level since the Cretaceous. At present, tectonism in the region is restricted to the northern edge of the Gulf of Oman where the Arabian plate is subducting the Eurasian plate from the south and to the Zagros Crush zone where the Arabian and Eurasian plates are colliding with one another.

Introduction

The Persian Gulf (gulf or Arabian Gulf) is a remnant of a former platform (Arabian platform) that during the Mesozoic and Paleogene formed most of Arabia, the gulf, eastern Iraq, and most of southwestern Iran. The geologic setting of this former platform is unique in that tectonism along its present northeast border ranges from continental collision to subduction. Such a setting makes it possible to see the changes in tectonic processes that take place along a plate boundary as it evolves from a subduction to a collision margin. The gulf itself lies atop the platform along a low between the deformed and undeformed platform segments. It lies between the Zagros Mountains to the northeast and the Arabian Peninsula (Arabian Shield) to the west, south, and southeast. It is connected to the Gulf of Oman by the narrow Strait of Hormuz which is constricted by the Musandam Peninsula (Figure 1). At the head of the Persian Gulf are the Tigris and Euphrates rivers which carry the drainage for all of Iraq and parts of Syria, Turkey, and Iran.

During R/V Atlantis II cruise 93 we made 15 multichannel seismic, gravity, magnetic, 3.5- and 12-kHz profiles, recorded nine sonobuoys, and obtained 17 gravity and piston cores in the Persian Gulf region. During the same cruise we also investigated the adjacent Gulf of Oman and obtained two multichannel and four single-channel seismic, gravity, magnetic, and 3.5- and 12-kHz profiles, and several sediment cores. The Gulf of Oman results were described by White and Ross [1979], and the significance of the sediment cores from both areas was reported by Stoffers and Ross [1979]. In this paper we present results from the Persian Gulf, some additional comments on the Gulf of Oman, and a brief synopsis of the geology of the Arabian platform.

Multichannel profiles were obtained with a six-channel streamer (each channel containing 150 elements) with a group spacing of 100 m, and a sound source consisting of 300-, 120-, 80-, and 40-cubic-inch (4.9, 2.0, 1.3 and 0.7 L.) airguns operating at 1850 psi (12.5 MN/m^2) and fired every 30 s. Trigger delays were used for individual guns so that the primary output pressure pulse of each gun was added in phase, whereas the bubble pulses were largely cancelled. A Hewlett-Packard X-Y plotter was used to monitor one channel in real time. Signals from all channels were digitally recorded and, after the cruise, were stacked and deconvolved. Wide-angle seismic profiles were obtained using expendable AN/SSQ41A sonobuoys with an 18-m hydrophone depth. Signals transmitted by the buoys were displayed on an X-Y plotter and recorded on magnetic tape using a four-channel FM tape recorder. These tapes were replayed later using a variety of different filter settings to enhance either the direct water wave from which the range was calculated (high-pass filtered) or the refraction (low-pass filtered). The sound velocity in water was determined from the water salinity and temperature [Wilson, 1960] using expendable bathythermographs (XBTs).

Seafloor and subbottom dips were measured from a precision depth recorder and from the multichannel seismic reflection profiles. Since most dips are small (less than 2°), true velocities can be accurately determined from the sonobuoy profiles, even though they are unreversed. Due to the shallow depth of the Persian Gulf (less than 100 m), variable angle reflections recorded by the buoys were

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submerged in the water multiples and could not be used to determine interval velocities. Although refractions also were followed by a fan of multiples, we were able to use travel times of the first arrivals to calculate the refraction velocities.

Morphology

The Persian Gulf is 800 km long and from 185 to 115 km wide, has a median depth of about 35 m, and an area of 226,000 km^2 (Figure 1; see also Seibold and Vollbrecht [1969]). The main morphologic features are of tectonic origin with climate and rock types playing only secondary roles. At the head of the gulf is a flat deltaic surface formed by the floodplains of the Tigris, Euphrates, and Kara rivers. The Iranian coast is irregular in outline and backed by the northwest-southeast trending Zagros Mountains. These highs represent huge folds, the product of the Miocene to Plio-Pleistocene Zagros orogeny. Topographically the Zagros Mountains can be divided into three zones: a 50-km-wide coastal foreland consisting of elongate, symmetrical folds; a 225-km-wide belt, northeast of the coastal foreland where the sediments are deformed into larger and tighter folds than those in the foreland; and a 50-km-wide region of overthrusts (Zagros Crush zone) where the rocks are extensively deformed and the sedimentary sequence down to the Paleozoic is exposed [Koop and Stoneley, 1982]. Folding of the Zagros during the Miocene to Plio-Pleistocene may have been discontinuous, becoming younger to the southwest. Locally some of the coastal hills are cut by terraces and flanked on the seaward side by sand flats and salt marshes. A similar type of coast north of the Strait of Hormuz is fronted by a narrow belt of mangrove swamps [Emery, 1956].

In contrast the coastal region on the west side of the gulf is fairly subdued. From Kuwait to the Gulf of Bahrain, the coast is dominated by low sandy islands, beaches cut by shallow channels, and extensive salt swamps. The Gulf of Bahrain itself is shallow with many coral reefs forming shallow indentations. East of Qatar to longitude 53°30'E the coast is irregular, consisting of low cliffs backed by sand dunes [Emery, 1956]. From longitude 53°30'E to the Musandam Peninsula the coast is low, sandy, and characterized by many small islands, shallow inlets, and salt flats. A late Pleistocene (Wisconsin) marine deposit of calcarenite and quartz sand with dune sand interbeds forms cliffs along the shoreline. Its elevation ranges from slightly below sea level to as much as 150 m above sea level [Chapman, 1978b]. A bench of Holocene age (4000 years old) consisting of shells and serpulite reefs parallels the shore from Kuwait to Qatar and the southern shore of the Persian Gulf [Chapman, 1978b; Felber et al., 1978]. The elevation of these relict features reflects the recent tectonic instability of the region.

Salt flats or sabkhahs are common along the Arabian Peninsula coastal region, and are found many kilometers inland and as high as 150 m above sea level [Chapman, 1978b]. Other coastal features include low rolling plains where bushes and grass trap sand in hummocks, a wide belt of drifting sand and sand dunes, and gravel plains which represent deltas emplaced during the Pleistocene pluvials. Landward of the coastal region is the 80- to 250-km-wide, 250- to 400-m-high As Summan Plateau which is bordered on its eastern side by an 80- to 120-m-high scarp [Chapman, 1978b; Felber et al., 1978]. The indentation of this scarp by ancient stream valleys and the wave-cut terraces on nearby inselbergs is believed to be due to late Pliocene or earliest Pleistocene marine erosion when sea level was 150 m higher than now. Felber et al. state that the inland sabkhahs reaching elevations of 150 m above sea level were formed during the subsequent regression.

The Oman Mountains consist of nappes of ophiolites and Paleozoic and Mesozoic sediments, and slightly deformed Mesozoic rocks on the flanks of the mountains that dip gently westward and are almost entirely covered by flat-lying Tertiary sediments [Chapman, 1978a]. The northern tip of the Musandam Peninsula (northern extension of the Oman Mountains) is dominated by islands, and deep, steep-sided embayments. The coast east of the Oman Mountains, south of the Gulf of Oman, is an alluvial plain that descends gradually from the mountains to the west. The eastern Iran (Makran) coast, northeast of the gulf, is dominated by east-west trending mountains.

The islands in the Persian Gulf have several origins. Those in front of the Tigris-Euphrates Delta are of depositional origin representing large areas of muddy sediment [Emery, 1956]. Off Kuwait, Bahrain, and Qatar the numerous islands are formed by coral growth. The islands off the United Arab Emirates are composed of unconsolidated Holocene sand with cores of Pleistocene limestone of eolian and shallow water origins enclosing a series of tidal embayments. Extensive oolite formation occurs in the tidal deltas between the islands, and the carbonate buildups fronting the islands include fringing and barrier-type reefs [Evans, 1966] (for a detailed discussion of sedimentation in the Persian Gulf see Purser [1973]). Other islands in the Persian Gulf are of structural origin. Bahrain, for example, is a north-south anticline similar to the Qatar Peninsula. These north-south or northeast-southwest low dipping arch trends (Arabian "trends") characterize the Arabian coast and the adjacent shallow seafloor [Kassler, 1973]. The island chain off Iran is part of the Zagros fold belt. Most of the small round islands in the eastern half of the gulf represent salt diapirs.

Southeast of the present-day delta at the head of the Persian Gulf is a series of northwest trending ridges at intervals of 2 to 4 km having a maximum relief of 10 m. Seibold and Vollbrecht [1969], following Off [1963], stated that these features in shallower water are active mixed forms resulting from tidal current erosion and accumulation. The deeper

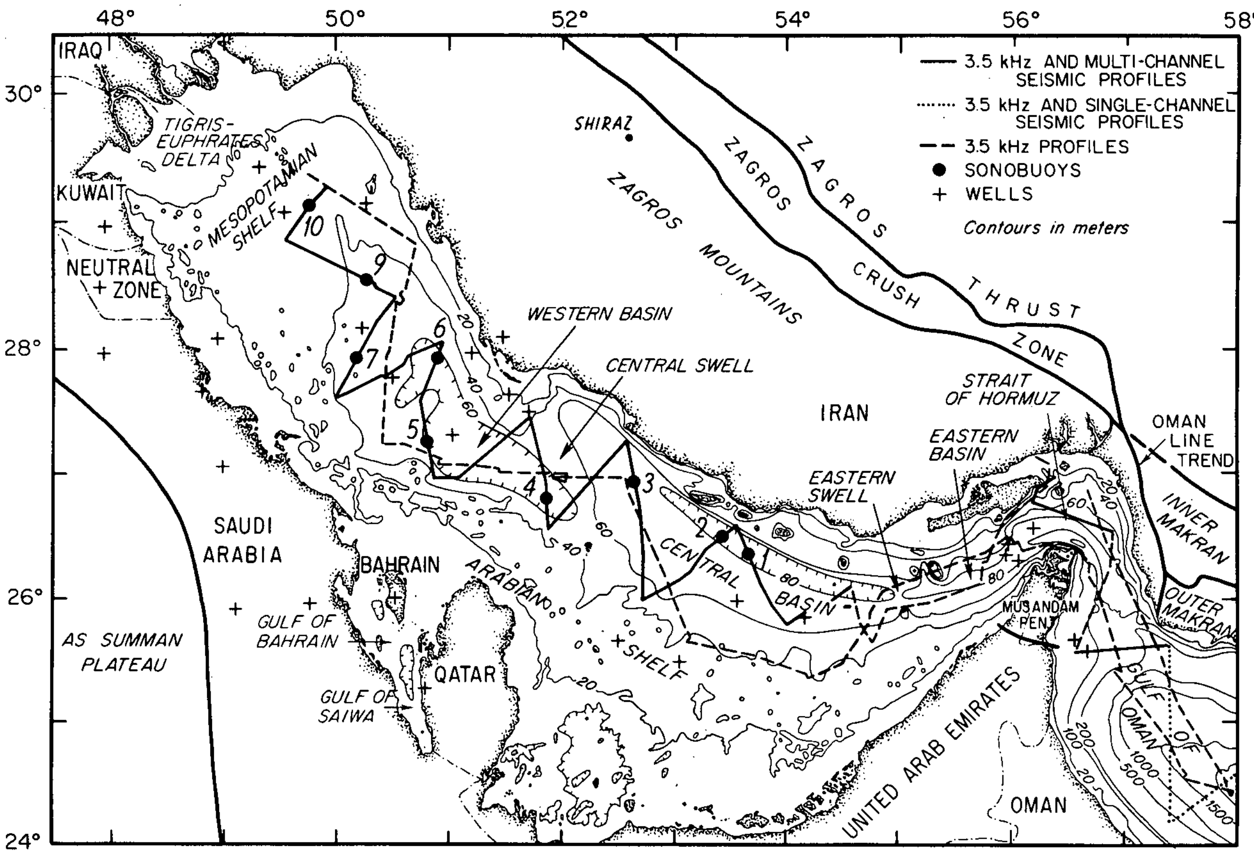

Fig. 1. Bathymetry of the Persian and Oman Gulfs. Compiled from data collected during R/V Atlantis II cruise 93, Seibold and Vollbrecht [1969] GEBCO sheet 5.05 [Laughton, 1975].

forms, which are partially covered with sediment, were felt to be relict and formed in the same manner when sea level was lower. Sarnthein [1972] with reference to studies of recent sediments in the southern margin of the Persian Gulf [Evans et al., 1969], however, believed that the ridges are drowned beach-dune ridges. Such an origin would explain the mixed character of grain sizes and components in the micrites as a mixture of relict eolian and beach carbonate sediment. Other ridges in the Central Basin of the gulf are 300 to 400 m apart and have relief of about 5 m. They are confined to depths mainly under 60 m, are erosional in origin, and were carved during the Holocene (see below).

From the head of the Persian Gulf, where up to 30-m-thick lobes of clastic sediment of the present-day delta prograde up to 100 km from the river mouths [Kassler, 1973], to the crest of the continental slope in the Gulf of Oman the seafloor of the gulf slopes gently seaward with a gradient of 10 m in 100 km (Sarnthein, 1972). At a water depth of 110-120 m at the shelf's edge the declivity of the seafloor steepens abruptly. The base of the continental slope is at approximately 1000 m depth beyond which the continental rise descends slowly eastward (Figure 1). The slope and rise off eastern Iran (Makran) is characterized by east-west trending ridges which exhibit relief of hundreds of meters. The slope on the western side of the Gulf of Oman is much more subdued, displaying broad structural terraces and fault-bound benches.

The seafloor of the Persian Gulf is asymmetrical in cross-section, the deepest water being close to the Iranian shore, with the Arabian side much gentler than the Iranian side. The curvolinear depression along the Iranian coast with its chain of islands along the top of the steeper eastern side lies on the boundary between the undeformed and the folded platform segments (Figure 1). The depression caused by downwarping of the lithosphere in response to the load emplaced by the Zagros Mountains to the north is divided into a series of basins by the northwest trending Central Swell and the north trending Eastern Swell.

Whereas the shelf off Iran is narrow and bound on its seaward side by a chain of islands, the shelf off Arabia is much wider, gentler, and has a complex topography of numerous flat-topped banks and shoals. Some of these highs are due to salt diapirism, and others display Arabian and/or Zagros structural trends. Still others may be erosional relicts related to Quaternary events.

The northern plunge of the Musandam Peninsula (northern extension of the Oman Mountains) constricts the mouth of the Persian Gulf. Its existence, since the late Tertiary (see below), has had a considerable influence on sedimentation in the Persian Gulf. Entrenched on the crest of this structural

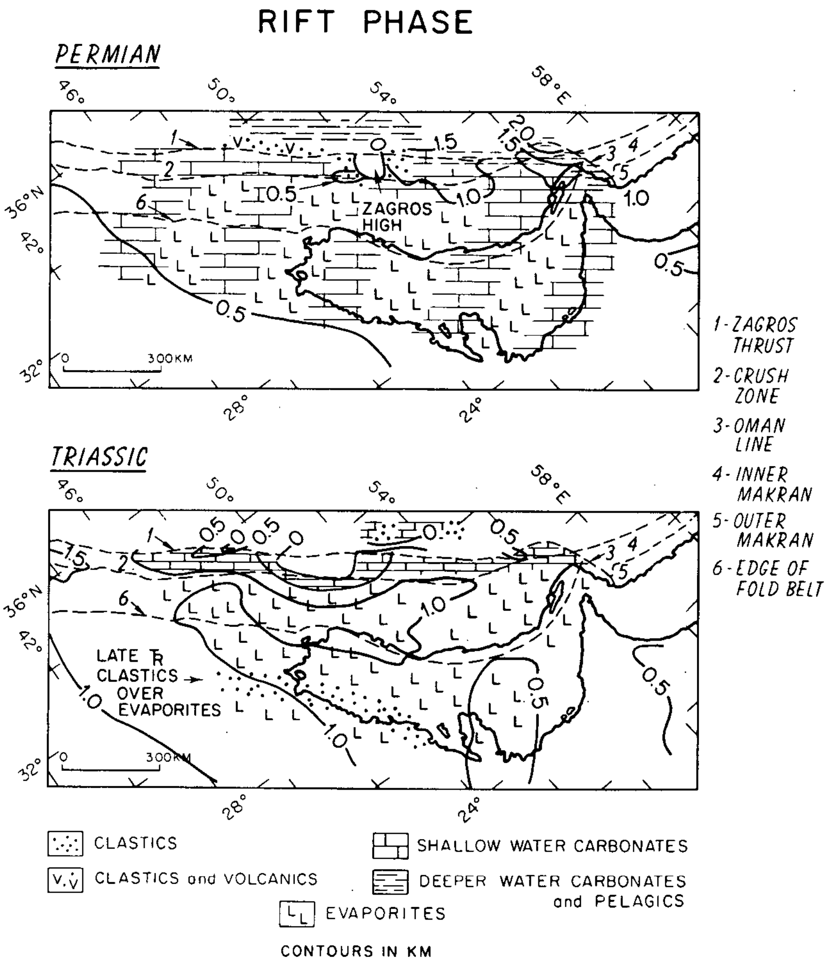

Fig. 2. Isopach and lithofacies maps of the rift sequence in the Persian Gulf region. Compiled and redrawn from maps from Koop and Stoneley [1982].

high in the Strait of Hormuz is a narrow depression where a maximum depth of 214 m was recorded during the cruise.

Depositional Patterns

Pre-Quaternary

Sediments in the Persian Gulf region can be divided into five depositional sequences: (1) a prerift unit of pre-Permian age, (2) a rift unit of Permian and Triassic age, (3) a drift unit of Jurassic and Early Cretaceous age, (4) a convergent unit of Late Cretaceous to Eocene age (Figures 2-4; see also Koop and Stoneley [1982]), and (5) a collision unit of Oligocene to Holocene age. The pre-Permian prerift unit which is over 3 km thick in the Zagros Mountains of Iran and was deposited on a broad shelf (Arabian platform) includes the Infracambrian Hormuz salt, clastics, and Cambrian, Devonian, and Carboniferous carbonates [Murris, 1980]. The distribution of these sediments is affected by unconformities of pre-Late Cambrian, pre-Devonian, Early Carboniferous age and the Late Carboniferous Hercynian unconformity. The sequence above the Hercynian unconformity is made up of prerift Late Carboniferous to Early Permian glacial deposits (only in Oman and Saudia Arabia; see McClure [1980]) and rift Permian basal sands, Permian clastic-volcanic sediments northeast of the Zagros Crush zone, and local red beds derived from the rifted terrain in the vicinity of the zone [Koop and Stoneley, 1982]. During the middle and Late Permian the region was blanketed by shelf carbonates and evaporites with reef carbonates along the northeast edge of the Arabian platform. Rift Triassic sedimentation also was dominated by carbonates and evaporites. A basin-wide unconformity, the breakup unconformity, of Late Triassic to Early Jurassic age marks the onset of seafloor spreading.

The Jurassic and Lower Cretaceous drift sequence is composed of Lower-Middle Jurassic shallow and deeper water carbonates, and pelagic marls, Upper Jurassic carbonates and evaporites, Lower Cretaceous (Berriasian to Aptian) shallow water carbonates, deeper water carbonate pelagic marl and clastics north of the emerging Arabian Shield (Arabian Peninsula), and a widespread Aptian carbonate transgressive unit. This trangression was terminated by a mid-Aptian regression that resulted in some erosion and a cycle of clastic deposition [Harris et al., 1984].

The onset of subduction near the end of the

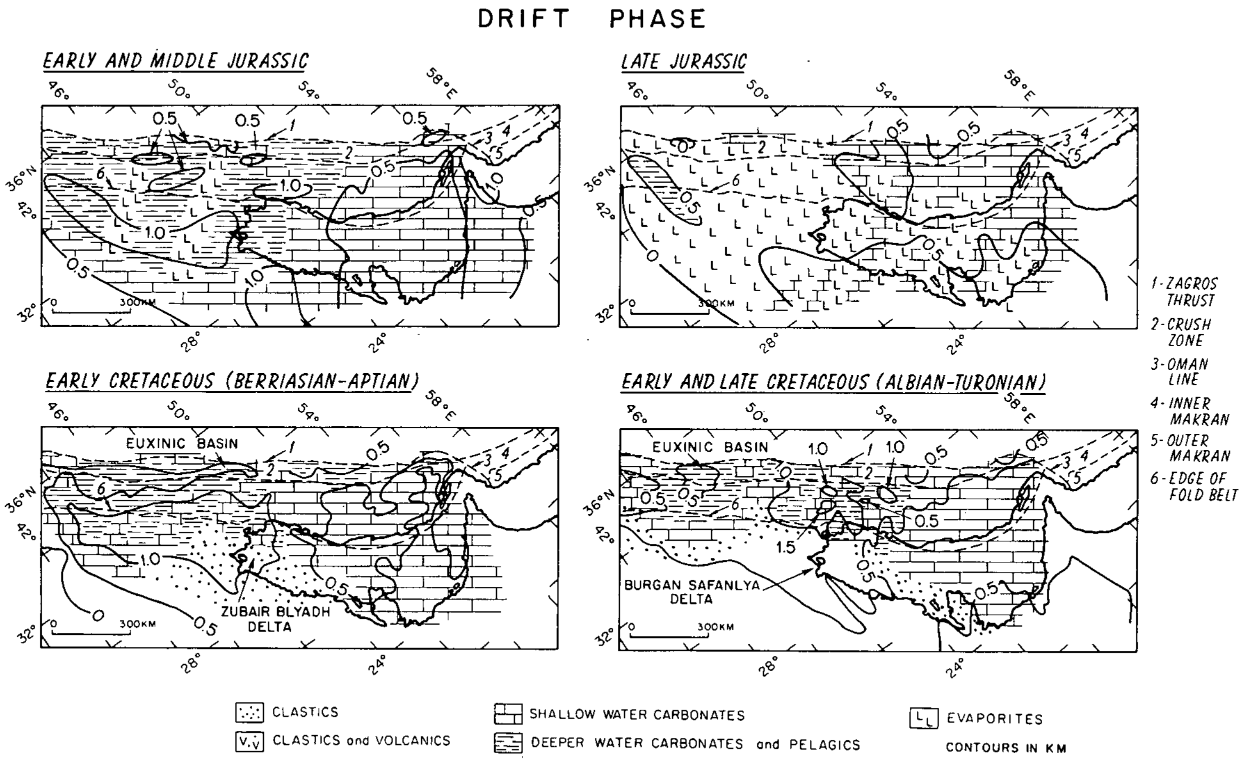

Fig. 3. Isopach and lithofacies maps of the drift sequence in the Persian Gulf region. Compiled and redrawn from maps from Koop and Stoneley [1982].

Turonian is marked by tectonism and erosion along north-south linear "Arabian" trends, and localized salt intrusions [Koop and Stoneley, 1982]. The subduction sequence in the Zagros Crush zone ranges from oceanic sediments and ophiolites probably emplaced during the Campanian, clastics (flysch) derived from oceanic rocks and deposited in linear foredeeps along the subduction front, deeper water shales south of the front, and platform carbonates along the edge of the Arabian Shield. As a result of uplift in the Late Cretaceous, most of the elongate basin south of the crush zone underwent erosion. The Paleocene-Eocene sediment facies ranges from clastics and shallow water carbonates along the crush zone to pelagic marls in the basin south of the zone, and platform carbonates and evaporites along the northeast edge of the Arabian Shield. Toward the end of the Eocene, shallow water carbonates blanketed the flysch along the crush zone. This together with the Maestrichtian to early Tertiary oceanic melange along the zone indicates that the Tethys southern seaway (High Zagros Alpine Ocean) still existed at that time.

In the Musandam Peninsula following a major emergence and period of nondeposition in the Turonian during which all of the Lower Cretaceous sediments on the outer carbonate platform were eroded, the platform collapsed and was overthrust by Tethyan oceanic basin sediments and ophiolites [Searle et al., 1983]. After the destruction of the margin, shallow marine upper Maestrichtian limestones were deposited unconformably over the allochthonous rocks. Resubsidence of the northern portion of the Oman foredeep continued from the Paleocene to the late Oligocene (Pabdeh foredeep facies).

During the early Oligocene, sedimentation was restricted to a shallow elongate basin south of the Zagros Crush zone with marl deposition in the center of the basin, neritic carbonates along the basin edges and deltaic sedimentation in Kuwait. In the early Miocene, shallow water carbonates and evaporites prograded over the deeper water marls and deltaic clastics. According to Koop and Stoneley [1982], closure of the Tethys was probably complete by the Oligocene or early Miocene. During closure, basement shortening and thickening along the crush zone by thrusting occurred by reactivating normal faults [Jackson and Fitch, 1981] that had developed during the rifting phase in the Permian and Triassic. As a result of uplift along the crush zone, overlying sediments were decoupled from the basement by décollement within the Infracambrian Hormuz salt and the overlying lower Miocene evaporites. Sliding above décollements in response to basement shortening was responsible not only for the "whaleback" folds of southwestern Iran, but also for similar folds in the United Arab Emirates and those west of the Musandam Peninsula [Searle et al., 1983]. The Miocene-Plio-Pleistocene sediments deposited during the formation of the Zagros folds (Zagros orogeny) consist of Miocene evaporites and carbonates that were followed by Miocene-Pliocene molasse red clastics as the depocenter shifted to its present position in the Persian Gulf.

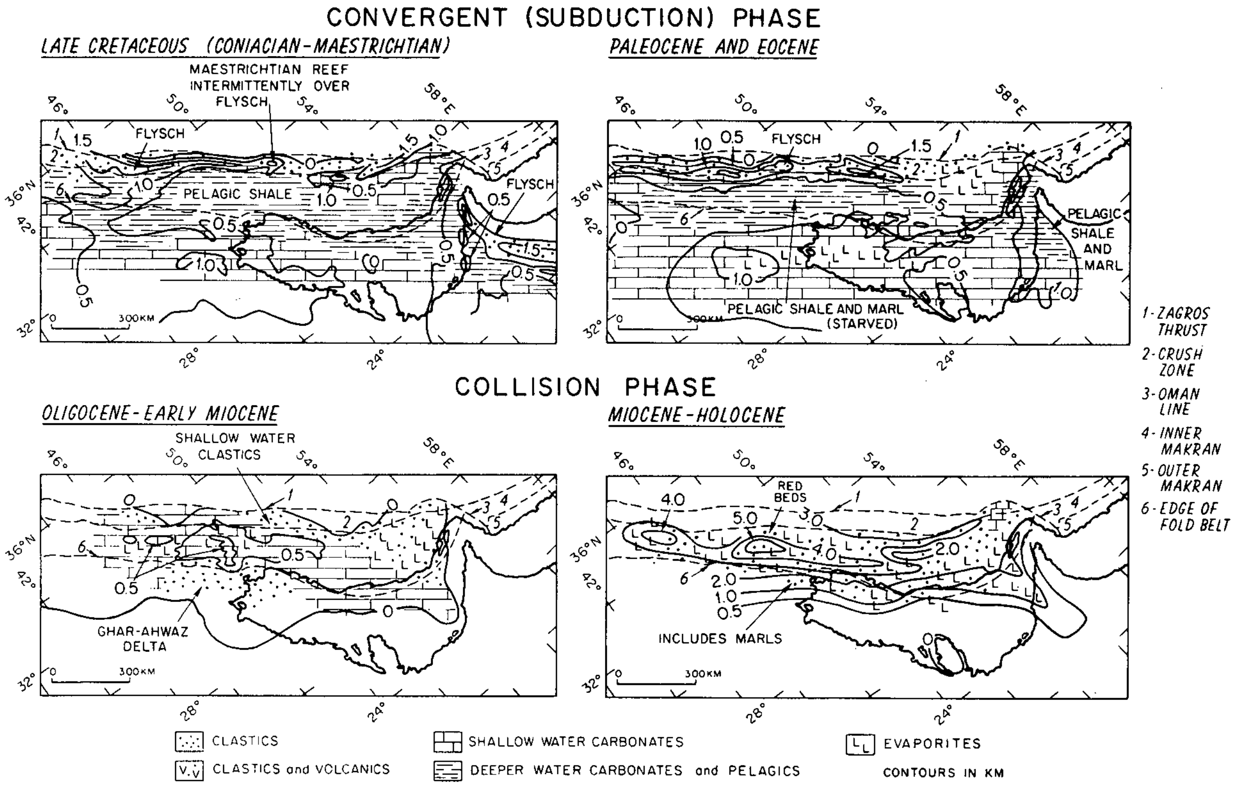

Fig. 4. Isopach and lithofacies maps of the convergent and collision sequences in the Persian Gulf region. Compiled and redrawn from maps from Koop and Stoneley [1982].

Quaternary

The Persian Gulf attained essentially its present form by Plio-Pleistocene time with the uplift of the Musandam Peninsula and the emplacement of the Zagros Mountains. The Strait of Hormuz also is a recent geologic feature formed by the northward tilting of the Musandam Peninsula as indicated by the ria-type shoreline north and east of the peninsula [Searle et al., 1983]. During the Wisconsin regression, continental conditions prevailed over the Persian Gulf as the sea retreated to the Strait of Hormuz. It was during this regression that carbonate-cemented sand dunes east of Qatar and beach-dune ridges at the head of the gulf were deposited. The gulf formed a large river valley across which the waters of the Tigris and Euphrates rivers emptied directly into the Gulf of Oman. The main trunk river was located along the contact between the Zagros folds and the Arabian platform. Debouching into the main stream was a complex system of distributaries from both the Arabian and Iranian sides of the gulf. The streams off Iran were much shorter and had steep gradients, whereas the Arabian ones flowed at a low grade as they crossed the gentle Arabian shelf. The Arabian drainage system was influenced by localized uplift of salt diapirs that forced the streams laterally, leaving the highs as erosional remnants [Kassler, 1973]. The buried, irregular topography shown on the 3.5-kHz profiles represents the fluvial surface eroded during the Wisconsin regression (Figures 5 and 6). Associated with this erosional surface is a grayish-brown detrital silt 30,000 to 12,000 years old [Stoffers and Ross, 1979].

In the course of the Holocene or Flandrian transgression that began about 11,000 years ago with the sea reaching the northern end of the Persian Gulf 8500 years ago, the Wisconsin regressive surface was blanketed unconformably by an acoustically transparent unit (Figures 5 and 6). Cores described by Stoffers and Ross [1979] indicate that the unit consists of light gray carbonate mud deposited 12,000 to 6000 years ago, and olive gray marl deposited during the last 6000 years. On the Arabian side most of the accumulation of carbonate muds occurred in topographic lows reducing the relief of the Wisconsin regressive surface [Kassler, 1973]. The fine-grained sediments tend to thicken toward the Iranian side where the carbonate muds become mixed with clastic sediments [Pilkey and Noble, 1966; Seibold et al., 1973; Wagner and van der Togt, 1973]. Local sediment buildup of the Holocene unit on the Iranian side led to the formation of the northwest trending Central Swell (profile C, Figure 5). Since sea level reached its present position the Holocene transgressive unit has undergone extensive erosion by bottom currents. This erosion is most intense in the Eastern Basin where the Wisconsin regressive surface has been exposed or was never buried (Figure 6).

Crustal Structure

The entire crust beneath the Persian Gulf and the Zagros Mountains is of continental

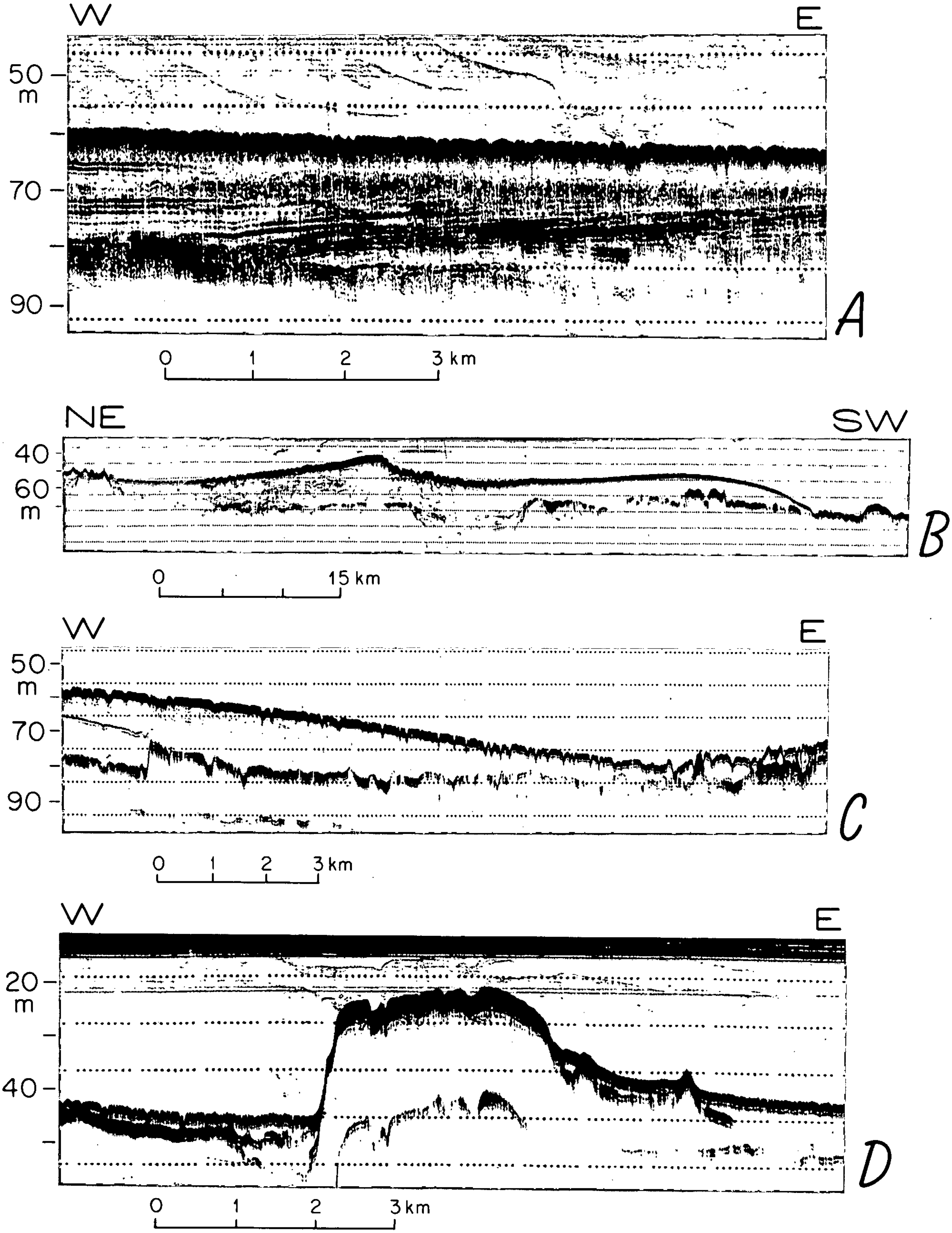

Fig. 5. Examples of 3.5-kHz profiles from the Persian Gulf (see Figure 6 for locations of profiles). Along profile A on the outer shelf a well-stratified unit onlaps the Wisconsin erosional surface and along profile B the erosional surface is covered by an acoustic transparent unit showing evidence of erosion at its shallowest point. Profile C which cuts across the Central Swell demonstrates that this high is depositional in origin. Profile D cuts across one of the topographic highs on the Arabian shelf. Note that in D the Holocene unit is restricted to topographic lows and tends to subdue the topography of the Wisconsin regressive surface.

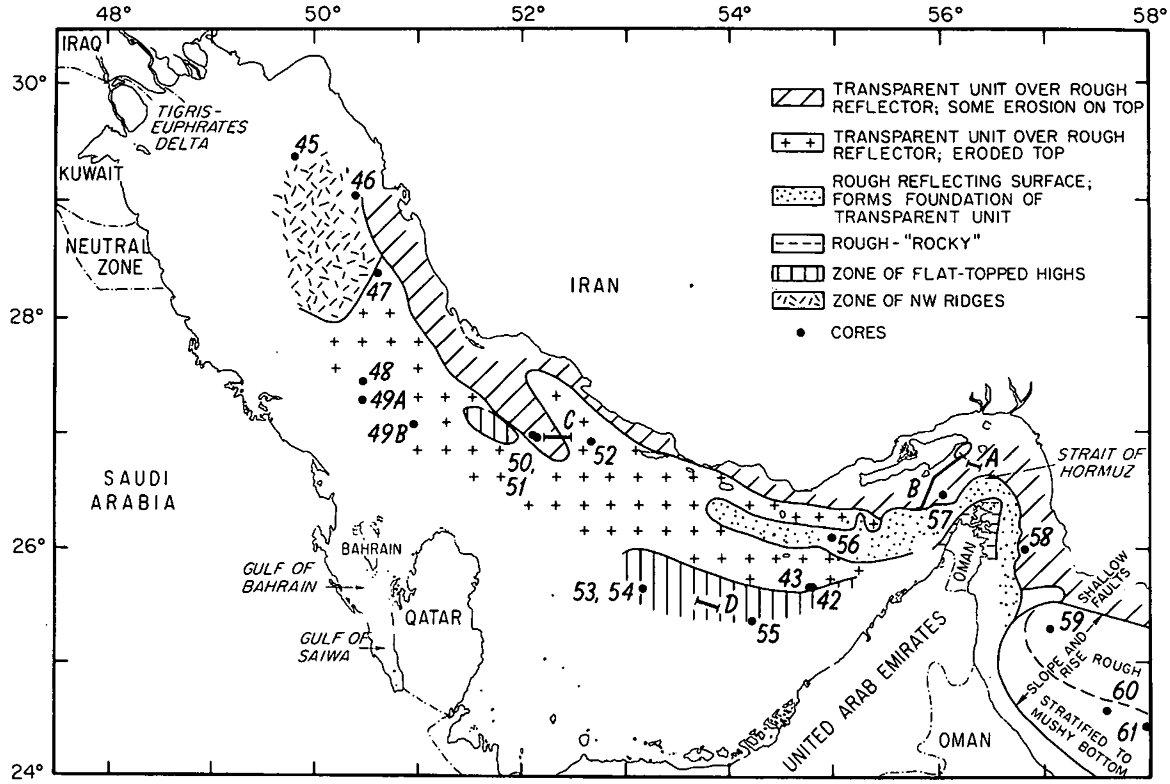

Fig. 6. Wisconsin and Holocene acoustic facies of the Persian Gulf. Compiled from 3.5-kHz profiles recorded during the present investigation.

type, while the crust beneath the Gulf of Oman to the east of the Oman line is of oceanic type (Figure 1). There have been no crustal seismic refraction lines in the Persian Gulf region, so crustal thickness estimates depend on surface wave dispersion characteristics or body wave delay times of teleseismic arrivals, or on even less well constrained inferences from heat flow or gravity anomalies in the region. Controlled source wide-angle seismic profiles have, however, been made in the Gulf of Oman, as will be discussed later.

First estimates of the crustal thickness in the Arabian Peninsula (Arabian Shield) were made by studying the dispersion characteristics of surface waves from distant earthquakes that had traveled along a great circle path between two World-Wide Standard Seismograph Network (WWSSN) stations. Niazi [1968] found a mean crustal thickness of 35 km along the 2700-km path between Shiraz, in the Zagros Mountains, and Addis Ababa in North Africa. Knopoff and Fouda [1975], again using Rayleigh wave dispersion along the same path, found the same thickness of 35 km, but added the constraint that it was unlikely to be less than 27 km, nor more than 44 km thick. These estimates unequivocally indicate continental crust, but since the path between Addis Ababa (Ethiopia) and Shiraz (Figure 1) crosses a diverse range of variable crustal blocks, including the thinner crust beneath the Red Sea, they are not necessarily very precise measurements of the crust beneath the Persian Gulf itself.

Two further studies using surface waves have attempted to find crustal structure beneath the Zagros by using dispersion paths running approximately northwest-southwest along the Zagros. Asudeh [1982] reports a mean crustal thickness of 46 km beneath the Zagros using a pair of stations in Shiraz and Tabriz. Bird [1978] also found a thickness of 46 km using Rayleigh wave group velocities from earthquakes in eastern Turkey that had sampled 1600 to 1800 km of travel within the Zagros crust.

A second method of determining crustal thickness using earthquake seismic sources is to study the delays of body wave arrivals at receivers within the Zagros. Eslami [1974] deduced that the crustal thickness beneath Shiraz is 44 ± 5 km by this method, Akasheh [1975] reports a thickness of 48 ± 4 km and Asudeh [1981] finds a crustal thickness of 39 km.

Gravity measurements show there to be a gradient in the Bouguer gravity anomaly decreasing northeastward from the Persian Gulf across the Zagros Mountains. Bird [1978] models this gradient by an increase in crustal thickness of about 17 km from the Persian Gulf to the Zagros thrust. Akasheh [1975] finds a similar result, postulating crustal thickening from about 40 km under the gulf to 55 km on the northeastern edge of the Zagros. This is in agreement with the calculations from seismic arrivals. The heat flow in the center of the Zagros is about 37 mW/m^2 [Bird et al., 1975]. Since the normal heat flow through shield areas is about 50 mW/m^2, then this depressed value is consistent with crustal thinning of up to 34% near the suture in the northeastern Zagros. There are no firm measurements of the thickness of sedimentary rocks in the Zagros, though Lees [1952] estimates it as about 7 km, and Falcon [1969] as 12 km.

At the western end of the Gulf of Oman, Niazi et al. [1980] shot three 180-kg explosive charges into an array of ocean bottom seismometers on the Makran continental margin of Iran, at 24.5°N, 59°W. They found a 12-km pile of sediments overlying oceanic crust with a total depth to the Moho of 19 km. Further east, in a pair of 150-km-long reversed seismic refraction lines using explosives and multiple receivers, White and Louden [1982] showed that the crust beneath the Gulf of Oman is oceanic in nature, with about 6 km of oceanic igneous crust underlying 7 km of normally compacted sediments; from a refraction line on the Makran continental crust they found that the oceanic crust dips northward to an angle of 1.5° to 2° with a steadily thickening overlying sediment wedge. This is in agreement wth simple two-dimensional gravity models across the margin [White, 1979b].

Structural Trends

The Persian Gulf region is dominated by six major structural elements: (1) the Arabian platform underlying the gulf and most of the Arabian Peninsula, (2) the Zagros Mountains, part of the Alpine-Himalayan chain extending along the southwestern and central portions of Iran, (3) the Zagros Crush zone northeast of the mountains, (4) the Musandam Peninsula (5) the Oman line, and (6) accreted sediments and igneous rocks of the Makran north of the Gulf of Oman (Figure 9). As displayed in the seismic reflection profiles (Figures 7 and 8) the Persian Gulf is characterized by strata dipping gently in the direction of the Zagros Mountains interrupted by Cretaceous-Tertiary and mid-Oligocene hiatuses. Continuity of the strata also is disrupted by diapiric Infracambrian salt structures, some of which rise above sea level to form islands (Figure 9). Kassler [1973] states that the Plio-Pleistocene was a period of extensive intrusion of salt domes in the Persian Gulf as well as the Iranian mainland. Mobility of the salt has led to the doming of Mio-Pliocene beds in several of the salt dome islands, and the uplift of Quaternary beaches. Other structural features of the Arabian platform are large arches and low-dipping north-south trending anticlines (Arabian trends; profile 13, Figures 7 and 8) including the Kuwait and Qatar arches. Many of the oil fields in the region are along this trend [Mina et al., 1967]. Some believe that these structures are due to salt diapirism, whereas others postulate that the highs are due to basement uplift. The structures have a history of prolonged upward movement during sedimentation. Kamen-Kaye [1970] states that the growth of the structures began as early as the Jurassic, and that earlier growth is possible if the Hormuz salt proves to be the agent responsible for the growth. If salt mobility can take place with little overburden then growth of the folds could have begun in the Paleozoic.

At a large angle to the platform structural trends are the high-frequency parallel, northwest trending Zagros Mountains, composed of folded Paleozoic to Cenozoic shelf carbonates (Figure 9). This Plio-Pleistocene trend, which is superimposed on the platform fabric, is the result of the shortening and uplift of the seaward edge of the Arabian platform which caused the strata above basement to slide southwestward along the Hormuz salt and younger evaporites. The present southwest edge of this gravitational movement of sediment is marked by the islands off Iran (Figure 9). The crush zone itself is a complex zone of Cretaceous nappes with radiolarites, ophiolites, shallow water carbonates, and Tertiary flysch. Along the northeast flank of the crush zone is the Zagros thrust (Figure 1; see also Tchalenko and Braud [1974]).

At the mouth of the Persian Gulf is the northeast trending Musandam Peninsula. Seismic reflection profile 15 (Figures 9 and 10) shows that this high extends across the Strait of Hormuz. Detailed mapping on land indicates that the high is the result of overthrusting of oceanic crust and oceanic facies from the southeast over a Mesozoic carbonate platform during the Turonian to early Maestrichtian, and compressional deformation in the mid-Tertiary that formed the folds along the west flank of the Musandam Peninsula. After the Miocene, vertical movements were dominant in the region with the Oman Mountains apparently rising and Musandam Peninsula subsiding [Searle et al., 1983]. This subsidence which was most pronounced toward the Strait of Hormuz and the east has been at least 450 m since the Pliocene [Lees, 1928], attaining rates of up to 60 m over the last 10,000 years [Vita-Finzi, 1973, 1979].

The Zagros Mountains and crush zone terminate on the east against the north-south trending Oman line [Falcon, 1974; Kadinsky-Kade and Barazangi, 1982]. This is a major boundary between the Zagros structures to the west and the east-west trending Makran Ranges to the east. Not only do the structures change abruptly at the Oman line, but the geology also is markedly different on either side (the Hormuz salt, for example, is truncated at the Oman line). To the west of the Oman line, there is currently continent-continent collision between the Arabian and Eurasian plates, while to the east there is an ocean-continent subduction between the same two plates. Overall movement on the Oman line is uncertain, but fault plane solutions on the Zendan-Minab fault system in Iran which mark the position of the Oman line indicate some eastward thrusting [Kadinsky-Kade and Barazangi, 1982]. This may be caused by eastward displacement of material caused by impaction of the Musandam promontory on the Eurasian plate.

Included within the Makran province are, from north to south, a 100-km-wide andesitic and basaltic submarine volcanic belt; a 50-km-wide subsiding forearc basin; the

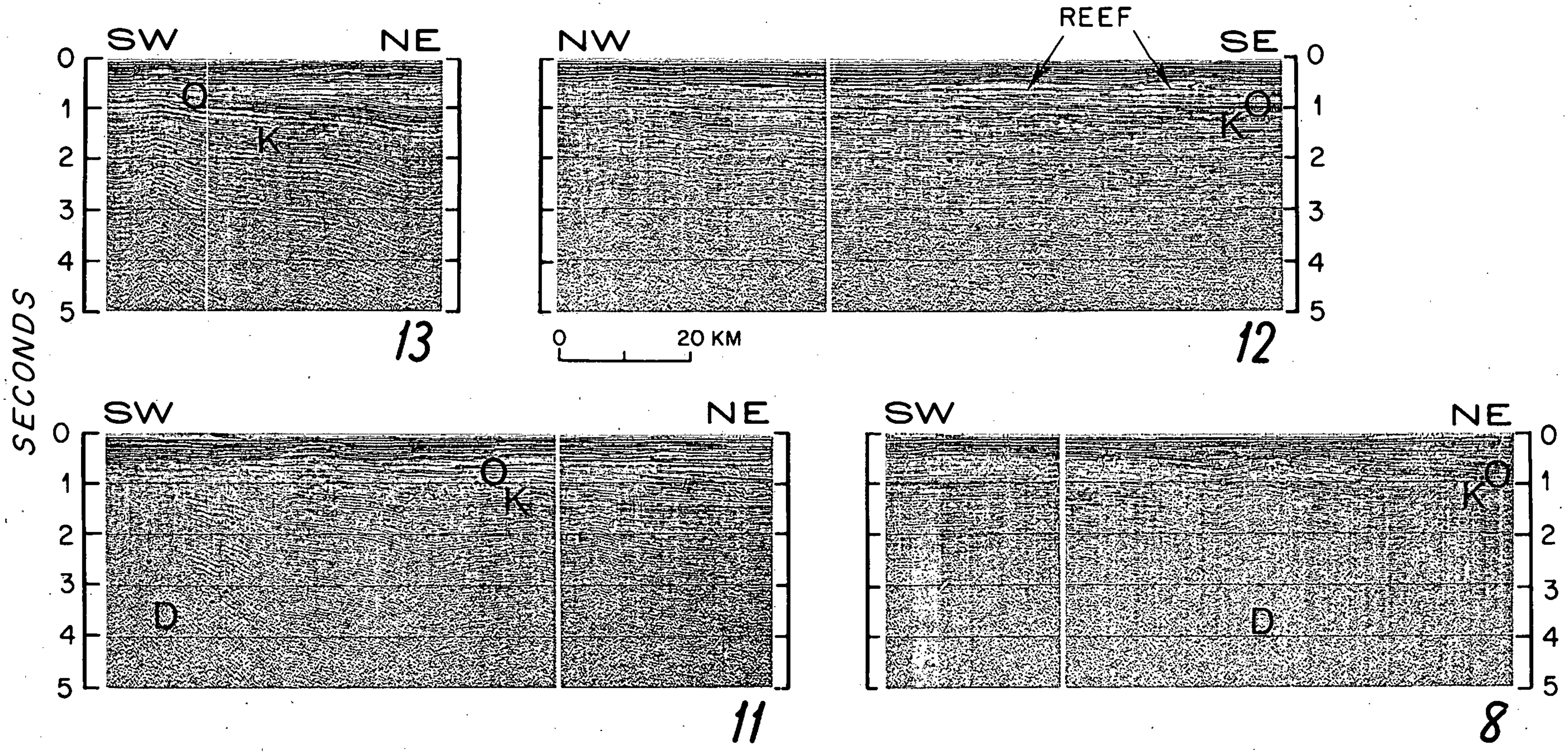

Fig. 7. Multichannel seismic reflection profiles of the Persian Gulf. K = unconformity atop Cretaceous; O = mid-Oligocene unconformity; D = diapir. Note reef atop reflector O along profile 12 and broad warp along profile 13. For positions of profiles see Figure 9.

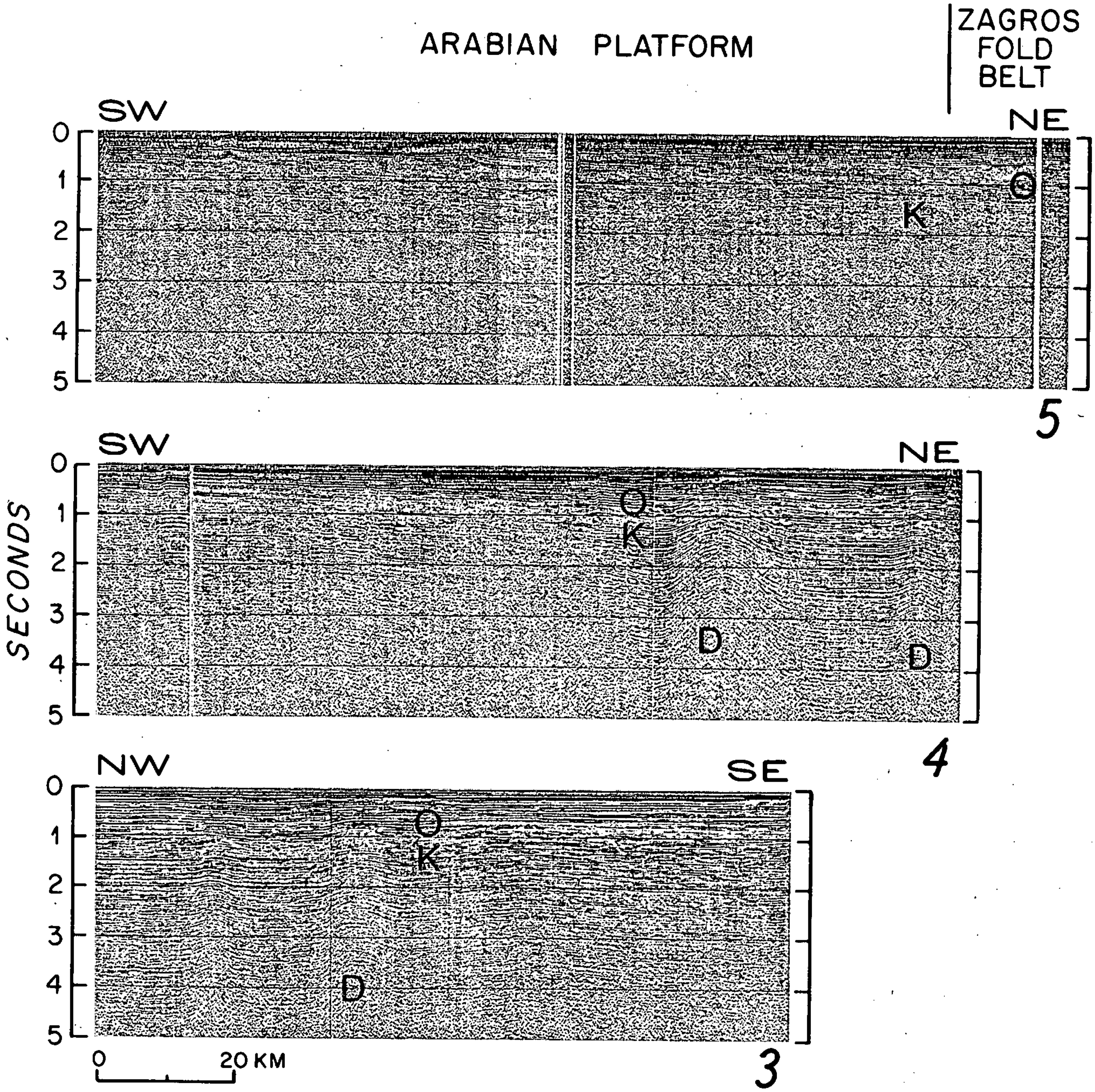

Fig. 8. Multichannel seismic reflection profiles of the Persian Gulf. K = unconformity atop Cretaceous; O = mid-Oligocene unconformity; D = diapir. For positions of profiles see Figure 9.

approximately 50-km-wide inner Makran zone consisting of Upper Cretaceous to Miocene turbidites, and an ophiolitic melange of great blocks of ultramafic rocks, serpentinite, pillow lavas and radiolarian cherts; an outer Makran or coastal zone (the 100-km-wide subaerial segment of the accretionary wedge) consisting of folded Pliocene sediments partially buried by less disturbed Quaternary sediments which rests unconformably on the inner Makran; and a continental margin zone (the 150-km-wide submarine segment of the accretionary wedge) consisting of east-west trending folds [Shearman, 1976; White and Klitgord, 1976; Jacob and Quittmeyer, 1979; White and Ross, 1979]. The inner Makran sediments with affinity to some rocks of the Zagros crush zone are juxtaposed across the Oman line against folded sediments of the Zagros Mountains. All of the structural provinces east of the Oman line probably rest on oceanic crust, and are the result of northward subduction of the Arabian plate beneath the Eurasian plate with the deformational front migrating progressively and rapidly southward [White, 1979b, 1981, 1984, Berberian and King, 1981; White and Louden, 1982]. Those west of the Oman line are the result of the northeastward subduction of the Arabian plate, collision of the Arabian and Eurasian plates, and subsequent gravitational gliding in a southwest direction.

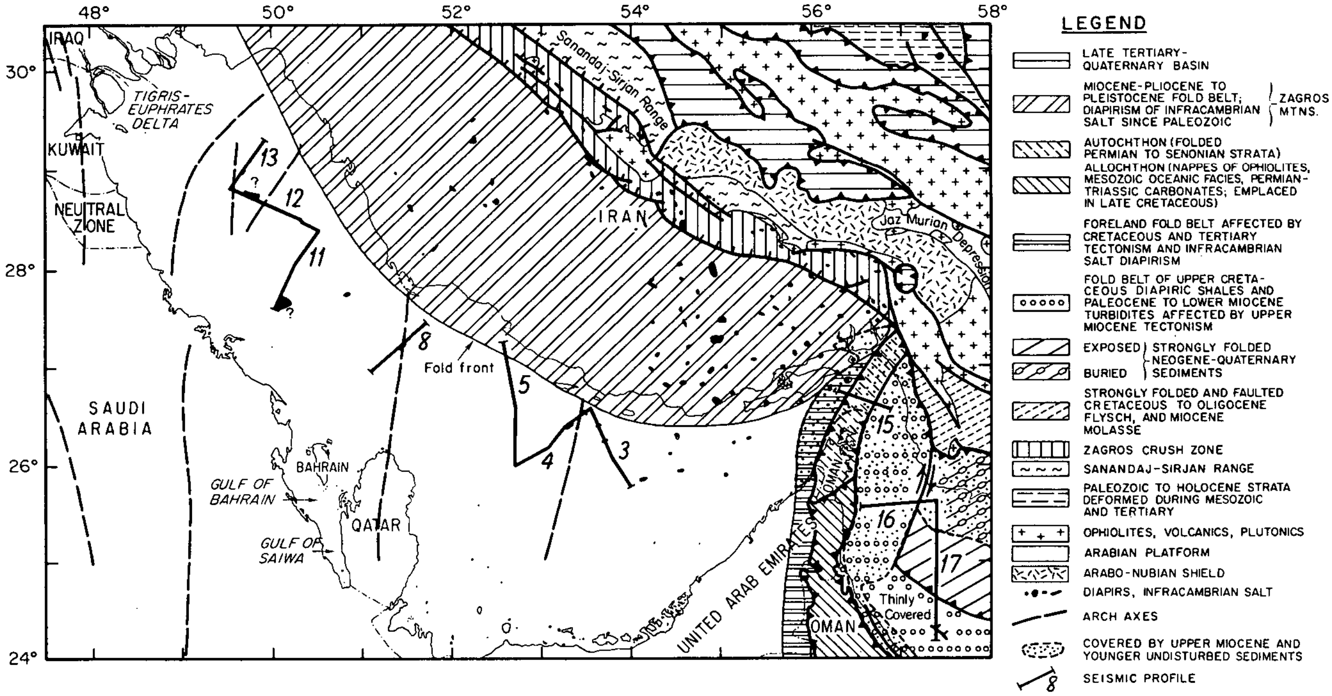

Fig. 9. Tectonic map of the Persian Gulf region. Compiled from data from National Iranian Oil Company [1959], Mina et al. [1967], Kamen-Kaye [1970], Brown [1971], Kassler [1973], Tchalenko and Braud [1974], Petroconsultants S.A. [1977], White and Ross [1979], Ricateau and Riché [1980], Berberian and King [1981], De Jong [1982], and Searle et al. [1983].

At the western end of the Gulf of Oman, east of the Musandam Peninsula, the lower part of the argillaceous section has been deformed into ridges that parallel the coast [Ricateau and Riché, 1980]. These highs may be the result of diapirism of Upper Cretaceous (?) clay which rests either on the cratonic shield covered with carbonates or on remnants of oceanic crust obducted over the Mesozoic carbonate platform on the Musandam Peninsula. Ricateau and Riché [1980] favor the second hypothesis. Between the ridges are Paleocene to lower Miocene turbidites that were emplaced from the north. During the mid-Tertiary both the ridges and the series above were deformed. Resting unconformably upon the deformed units are upper Miocene and younger deltaic complexes consisting of alternating silts and shales or sandy shales and conglomerates which thicken toward the east, reaching 2.5 km on the outer shelf (Figure 9).

Between the Musandam Peninsula and the Oman line, only the Upper Cretaceous to mid-Miocene sediments are deformed while relatively thick upper Miocene and younger sediments resting unconformably above them are undisturbed (Figure 11). East of the Oman line, however, the whole section off the Outer Makran is folded and thrust as it accretes from the underlying subducting plate (Figure 11). Further east, continued convergence of the Arabian and Eurasian plates has brought the continental margins of the Makran and of Oman into contact. Here the folding and deformation characteristic of the Makran continental margin sediments propagates onto the passive margin of the Oman, given a continuity of deformation of the entire sediment sequence (Figure 12). This region is in the initial stages of a continent-continent collision with only a narrow remnant of oceanic crust remaining beneath the Gulf of Oman.

Sediment Velocity Structure

The most impressive feature of the sediment velocity structure in the Persian Gulf is a steep velocity gradient in the uppermost 1 to 2 km below the seafloor (Figure 13). This gradient is higher than usually observed in sedimentary sections and reflects the carbonate facies and rapid increase in degree of its lithification with depth. Seismic refraction velocities immediately beneath the seafloor are high, typically 2.0 to 2.1 km/s.

Seismic velocity increases discontinuously with depth rather than in a smooth velocity gradient. Refraction velocities tend to be higher than interval velocities at the same depth, as determined from move-outs across the multichannel array suggesting the presence of horizontal-vertical velocity anisotropy. The reason is probably that the Tertiary and Upper Cretaceous sections are massive limestone and dolomite beds intercalated by sands and shales that have much lower seismic velocities [Mina et al., 1967]. The massive limestone and dolomite beds have high velocities and form good refractors, which are seen well on the sonobuoy profiles, whereas the interval velocities derived from the wide-angle

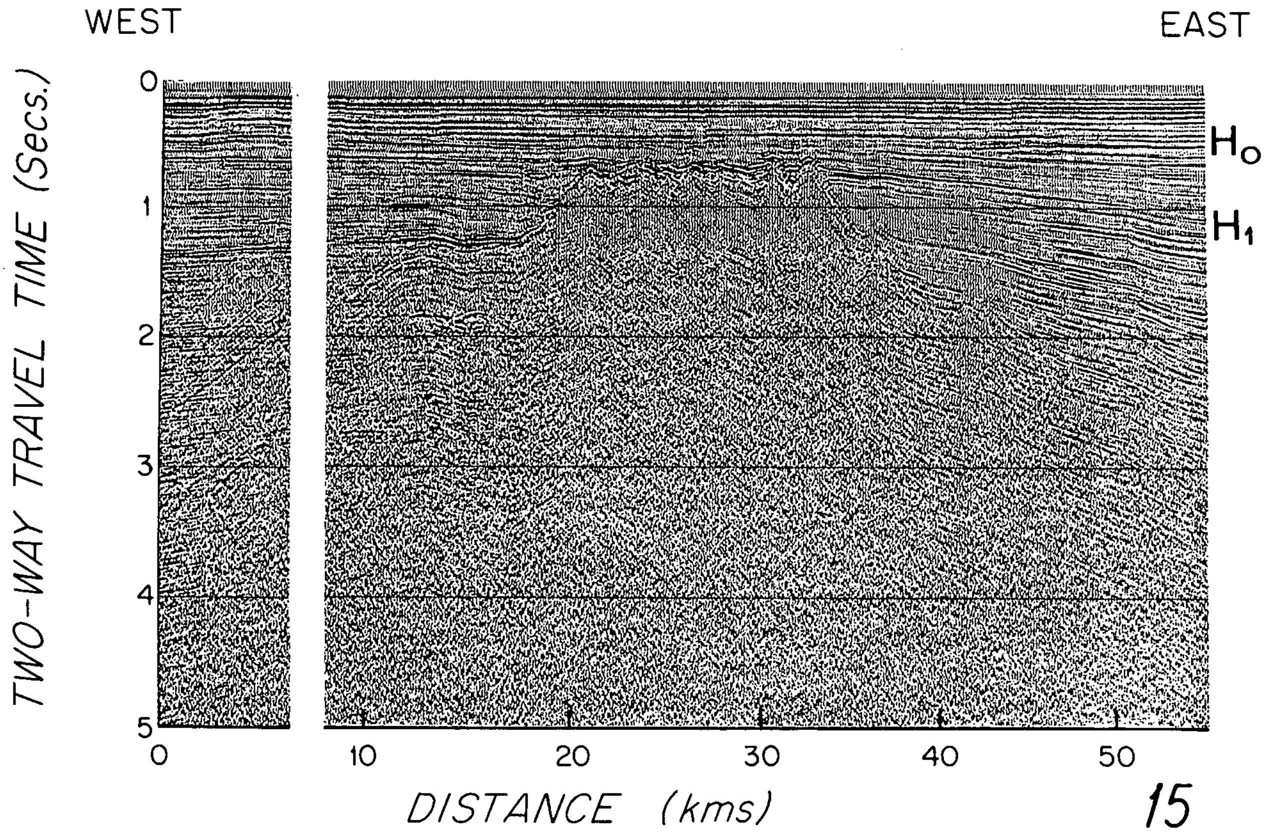

Fig. 10. Multichannel seismic reflection profile of the northern extension of the Musandam Peninsula high [from White and Ross, 1979]. Note deformed lower section along western side of high, and fault on eastern side near 50-km tick. H_0 = Plio-Pleistocene and H_1 = late Miocene hiatuses [Ricateau and Riché, 1980]. For location of profile see Figure 9.

reflections received by the array represent a mean of the entire velocity section, including both the high-velocity massive beds and the lower-velocity intervening sediments. Further support for this interpretation is provided by the short distance over which many of the refractors appear on the sonobuoy profiles. On sonobuoy 6, for example, the first arrival

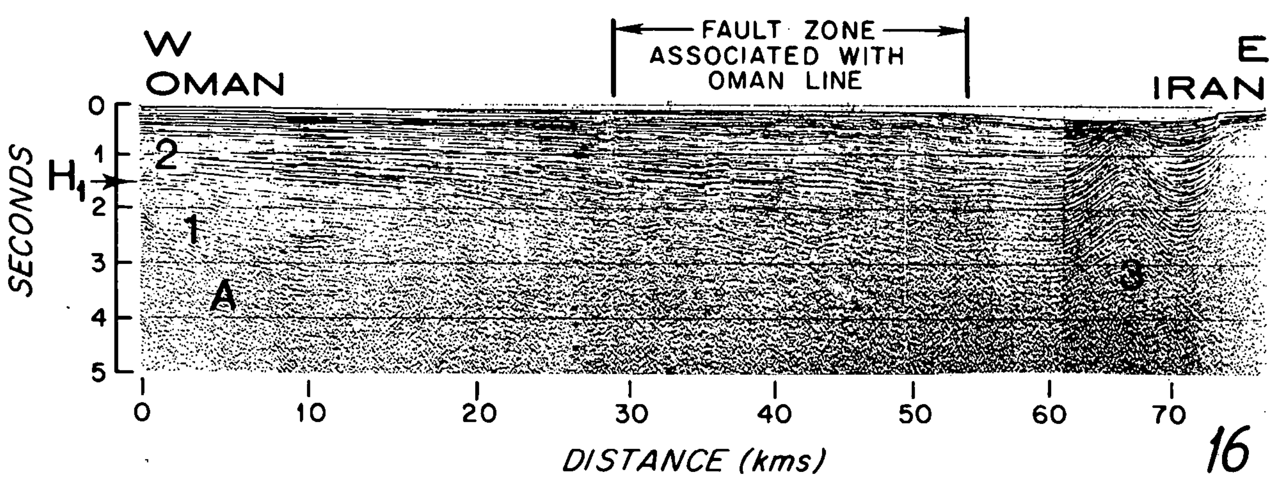

Fig. 11. Seismic reflection section east of the Strait of Hormuz. The reflector marked A is believed to represent the seaward extension of the Musandam block. Unit 1 represents the deformed Upper Cretaceous to mid-Miocene section, and unit 2 represents the undisturbed late Miocene and younger sediment unit. Horizon H_1 is late Miocene in age. The fault zone in the middle of the profile is associated with the Oman line. Note that east of the fault zone (end of profile) all the sediments are deformed (unit 3). For location of profile see Figure 9. Profile from White and Ross [1979].

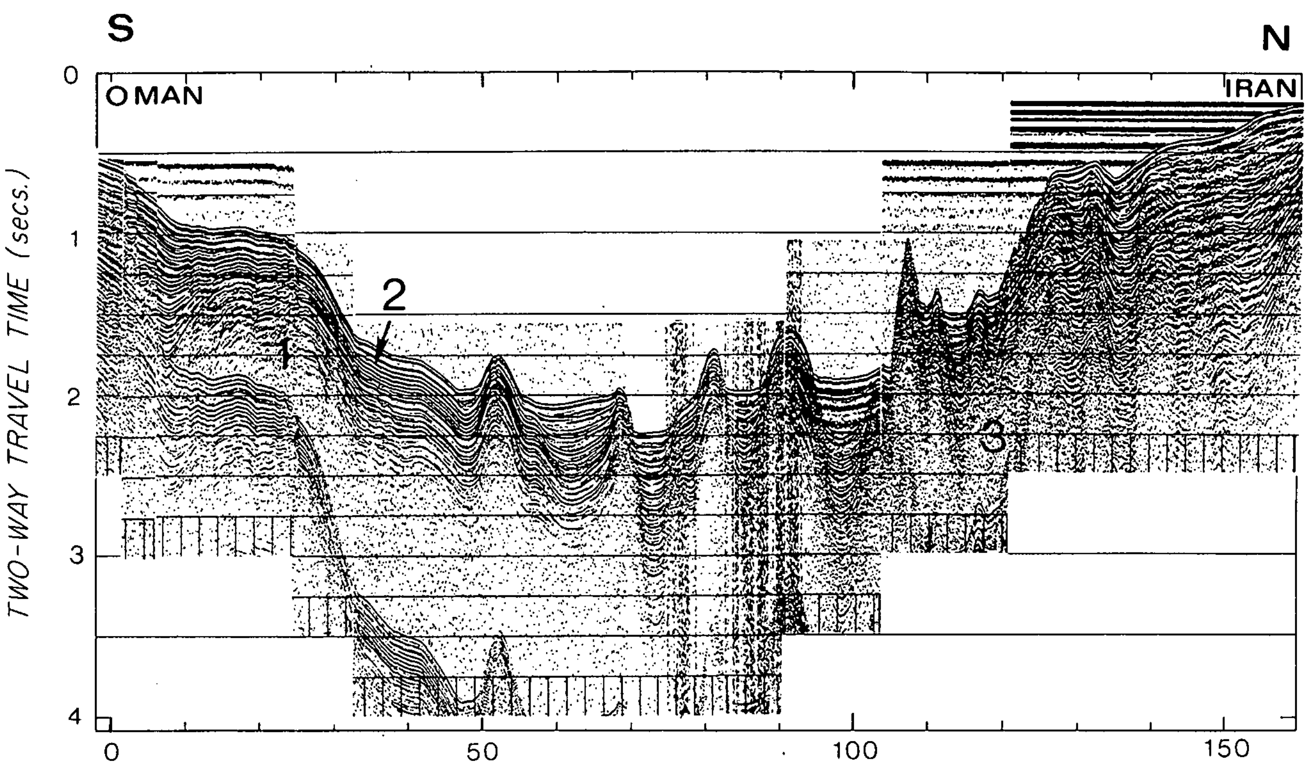

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Fig. 12. Seismic section from the central part of the Gulf of Oman. Along this section the folded Upper Cretaceous to mid-Miocene section (1) is much shallower than along profile 16 (Figure 11). The late Miocene and younger section (2) resting on unit 1 has also undergone folding and faulting. In this area the fold belt of the outer Makran (3) extends across the Gulf of Oman from Makran to Oman. Profile from White and Ross [1979]. For location of profile see Figure 9.

refractions rapidly die away (Figure 14), suggesting either that the refractor is laterally discontinuous, or that there is no positive velocity gradient beneath the refractors since that would give a high-amplitude interference head wave which would continue to be strong over a considerable distance [White, 1979a]. The manner in which successive first arrival refractions step back to later times on sonobuoy 6 is also suggestive of intervening lower-velocity material between the major refractors.

Velocities above the mid-Oligocene unconformity (Figures 7 and 8) are generally about 3 km/s, increasing markedly beneath this unconformity. The deepest velocities are poorly controlled by move-out across the multichannel array which cannot resolve deeper than 1.5 to 2 km due to the relatively short array length (750 m). The highest refraction velocities observed on the sonobuoy profiles range from 4.5 to 5.7 km/s, depending on the signal to noise ratio of the particular profile. Where velocities from beneath the Cretaceous unconformity (Figures 11 and 12) were measured, they were higher than 5.2 km/s.

In the Gulf of Oman a thick (up to 7 km) pile of rapidly accumulated terrigenous sediments overlies the basement. Wide-angle sonobuoy seismic profiles show velocities typical of normally compacting sediments beneath the abyssal plain [White and Klitgord, 1976]. On the Makran continental margin, where sediments are actively being scraped off into an accretionary prism above the subducting Arabian plate, the seismic velocities of the sediments are much higher than on the abyssal plain, indicating 30% to 40% compaction and dewatering due to tectonic consolidation [S.R. Fowler et al., 1986].

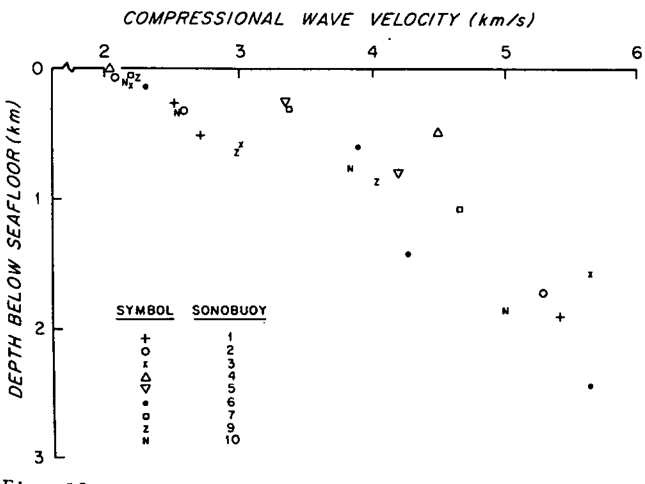

Fig. 13. Plot of true compressional wave velocity (i.e., corrected for dip) versus depth below the seafloor for the sonobuoys recorded in the Persian Gulf. See Figure 1 for locations of buoys.

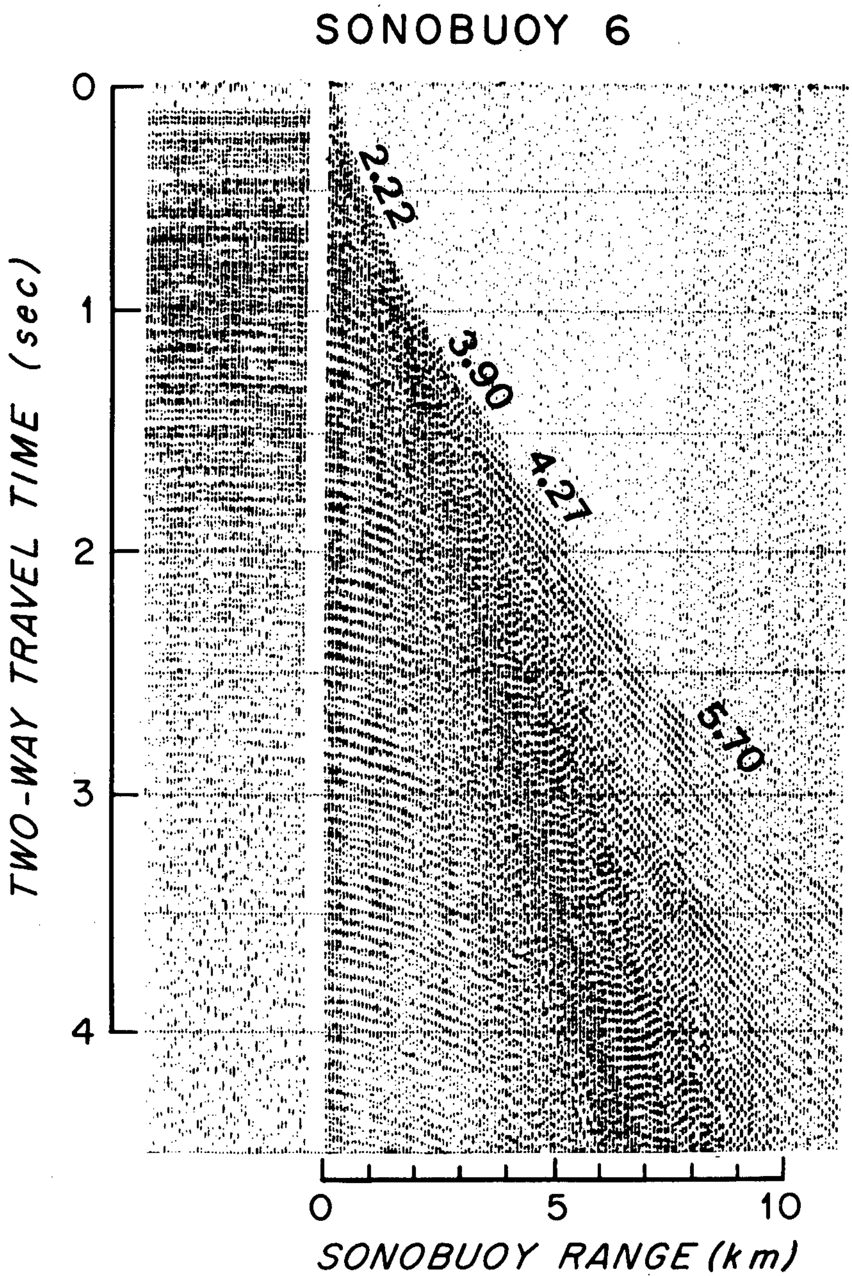

Fig. 14. Variable density record of sonobuoy 6 (see Figure 1 for location). Left-hand side of the profile is single-channel normal incidence reflection profile prior to sonobuoy launch. True refraction velocities (corrected for dip) are shown in kilometers per second. Filter and gain settings were adjusted throughout the profile.

Neotectonics

The entire Persian Gulf and Gulf of Oman area is, at present, a convergent region at the northern margin of the Arabian plate. The Arabian plate is colliding with the stable continental Eurasian plate to the north. There are no direct measurements of the magnitude of motion between Arabia and Eurasia, so these vectors have to be calculated from the poles of rotation of other plates in the vicinity [McKenzie, 1972], or from global present-day instantaneous plate motions [Chase, 1978; Minster and Jordan, 1978]. As listed in Table 1 the convergence between Arabia and Eurasia is in a direction of about 020°, at a rate of 30 to 50 km/Ma (Figure 15).

Seismicity throughout the Zagros is high. It is bounded to the north by the Main Recent fault (northwest part of the Zagros thrust) and to the south extends into the Persian Gulf (Figure 16). The Oman line marks an abrupt reduction of seismicity to the east. In the Makran region there is a gradual deepening of seismicity away from the coast where shallow activity begins and continues inland for about 170 km. From here the activity deepens more sharply, reaching a depth of about 80 km south

TABLE 1. INSTANTANEOUS POLES OF RELATIVE MOTION BETWEEN ARABIA, EURASIA AND IRAN.

Plate Boundary	Reference	Latitude	Longitude	Rate Deg/Ma
Arabia-Eurasia	McKenzie [1972]	32.2°N	5.2°W	0.544
	Chase [1978]	34.9°N	7.2°E	0.493 ± 0.104
	Minster and Jordan [1978]	29.82°N	2.64°N	0.357 ± 0.054
Eurasia-Iran	Jackson and McKenzie [1984]	27.5°N	65.8°E	0.560
Iran-Arabia*	Jackson and McKenzie [1984]	34.5°N	39.8°E	0.955

Angular rotation rates about the pole position give the motion of the first named plate with respect to the second. Rotations are positive for clockwise motions when viewed from the center of the earth.

*Vector addition of Arabia-Eurasia pole from Chase [1978] and Eurasia-Iran pole from Jackson and McKenzie [1984].

of the volcanic zone [Jacob and Quittmeyer, 1979]. Only a few events have been reported below a depth of 80 km. Such a deepening zone of seismicity is characteristic of a subducting oceanic lithosphere. Within the Zagros, the earthquakes are almost all thrusts, with fault planes dipping at 40° to 50° [Jackson and McKenzie, 1984]. The slip vectors indicate consistent motion at azimuths of 030° to 040°, with the exception of a very few earthquakes associated with structural features crosscutting the Zagros, such as the Kzerun line, and of earthquakes on the Oman line where the Musandam promotory is impacting on the Eurasian plate and pushing material sideways to the east. The seismicity is caused by basement shortening beneath the Zagros. None of the shocks are subcrustal, and micro-earthquake surveys show that small earthquakes also occur in the sedimentary column as it responds to movement in the underlying basement. A noticeable feature of the seismicity is that the larger shocks occur toward the southern part of the Zagros, between the High Zagros and the Simple Folded belt. This is probably because the crustal stresses are greatest not where the topography is highest, but where it exhibits the largest gradients [Jackson and McKenzie, 1984]. As convergence continues and the crust is thickened back to and beyond its prerifting state, folding is likely to continue to migrate southward across the Persian Gulf.

To the north of the Zagros, Jackson and McKenzie [1984] show that central Iran is relatively aseismic, and so may be considered as a stable block between Eurasia and Arabia. Jackson and McKenzie suggest that the convergent motion taken up across the Zagros is rather less than the total Arabian-Eurasian motion computed from global plate motions (Figure 15).

To the east of the Oman line, continued input and offscraping of the thick pile of sediment on the Arabian plate is causing the front of the accretionary prism to migrate southward at a rate of about 10 km/Ma [White and Louden, 1982; White, 1984]. In the western Gulf of Oman the continental part of the Arabian plate is just coming into contact with the opposing continental margin, and in the future will presumably continue to shorten and imbricate the huge volume of sediments accumulated in the Makran accretionary prism.

Geologic History

The present morphology of the Persian Gulf region is the result of a platform phase in late Paleozoic, rifting in the Permian and Triassic, seafloor spreading in the Jurassic and Early Cretaceous, and convergence in the Late Cretaceous culminating with the closure of the southern Tethys by the Oligocene or early Miocene [Koop and Stoneley, 1982]. Continuing convergence of the Arabian platform (Arabian plate) and the Eurasian plate has been accommodated by progressive thickening and shortening of continental crust [Berberian and King, 1981] leading to the uplift of the northeast edge of the platform along the Zagros Crush zone, pronounced subsidence south of the crush zone, and emplacement of the Zagros Mountains by gravitational sliding during the Miocene to Plio-Pleistocene [Koop and Stoneley, 1982]. Deformation of the eastern edge of the platform (Musandam Peninsula) by compression from the east and southeast took place in the Late Cretaceous and mid-Tertiary, and in the Makran throughout the Late Cretaceous to the present. Other factors that have produced the present morphology of the region include salt tectonics, broad upwarping of the platform cover by uplift of basement or salt tectonics, and Neogene-Quaternary uplift and subsidence of the coastal region. This tectonic pattern has been modified little by Pleistocene sedimentation, fluvial erosion during the Wisconsin regression, and marine deposition during the Holocene.

Stratigraphic evidence indicates that Eurasia became separated from the Arabian platform sometime in the Permian to form what Berberian and King [1981] call the "High Zagros Alpine Ocean." Searle et al. [1983] state that fragmentation of a Paleozoic

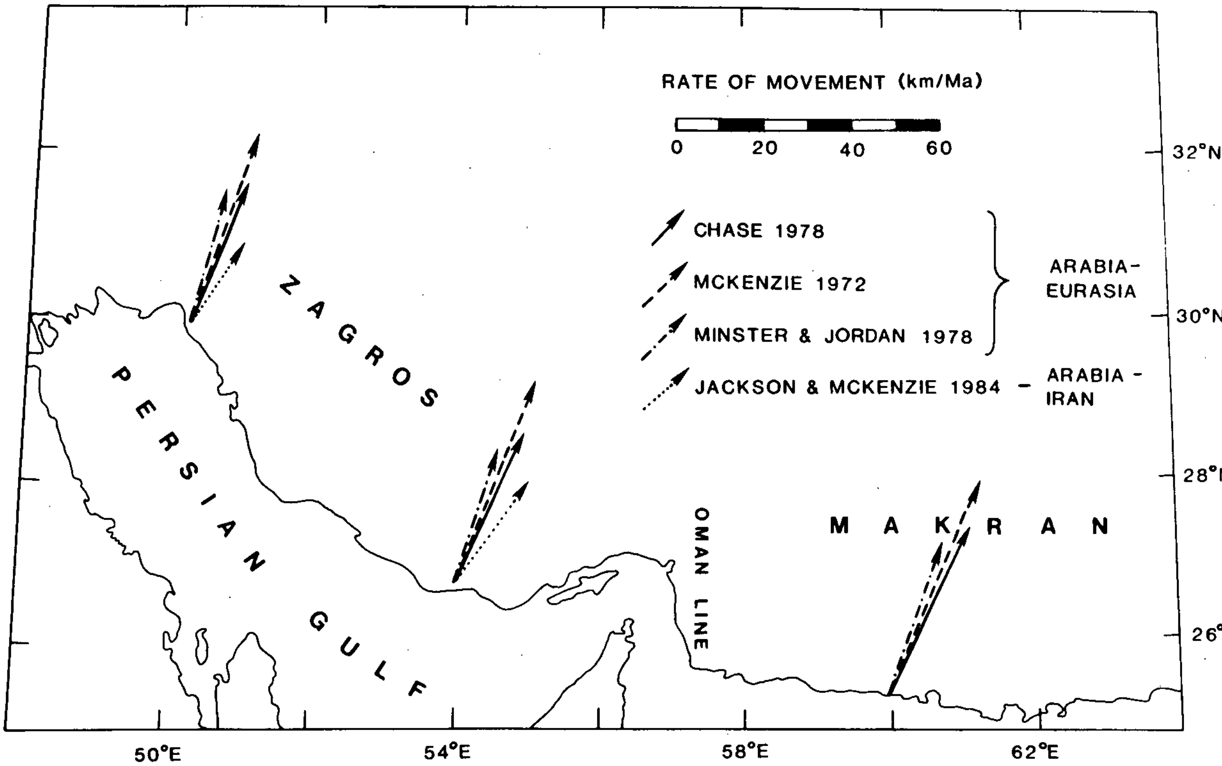

Fig. 15. Present-day motions of the Arabian plate at its northern edge in the Zagros and Makran mountains. Instantaneous poles of rotation are listed in Table 1.

carbonate platform of the Musandam Peninsula occurred in the Middle and Late Triassic. Thus by the end of the Triassic the broad Arabian platform, the Zagros shelf of Dewey et al. [1973], extended from the southern end of the Arabian Peninsula to the Levant transform fault to the north and the Zagros Crush zone to the east. Deposition on the platform was controlled by changes in sea level (regressions at the Permian-Triassic, Triassic-Jurassic, and Jurassic-Cretaceous boundaries, Neocomian-Barremian, mid-Aptian, Cenomanian-Turonian, Turonian-Coniacian, Cretaceous-Tertiary boundary, mid-Oligocene, and Pleistocene; see Setudehnia [1978] and Murris 1980]). Other factors controlling deposition on the platform include a continuous 600- to 400-m drop in sea level since the Cretaceous [Watts, 1981], salt tectonics, growth of broad north trending anticlines, and climatic changes. During the transgressions, carbonate deposition developed to its maximum, and during the regressions clastic sediments from the southwest prograded across the platform. At times of arid conditions, evaporite deposition dominated large areas of the platform (Middle Triassic, latest Jurassic, Paleocene-Eocene, and early Miocene). During more humid times, conditions were at an optimum for carbonate deposition on the platform (see paleogeographic maps in the work by Murris [1980]). Near the end of the Jurassic, depositional patterns in the platform were further affected by tectonic differentiation of the platform. Where conditions were favorable small bioherms developed along the seaward edge of the platform and within the platform itself.

Weak deformational metamorphism, intrusion in the Sanandaj zone (Figure 9), and general deformation and incipient metamorphism through central Iran indicate that seafloor spreading ceased in the High Zagros Alpine Ocean in the Late Jurassic and the Arabian platform began to converge on the Eurasian plate [Dewey et al., 1973]. Nappe emplacement along the Zagros crush zone during the subduction phase was accompanied by wildflysch and olistostrome deposition near the nappes and flysch sedimentation farther basinward. As described by Murris [1980] the nappes tended to override their own preceding debris.

Telescoping and destruction of the Arabian platform margin appear to have temporarily ceased at the end of the Maestrichtian, and

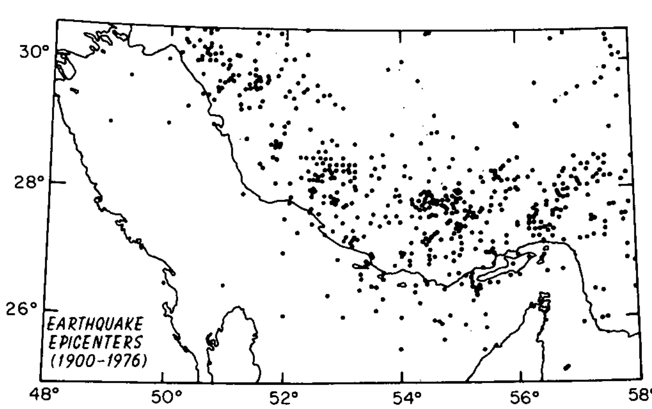

Fig. 16. Epicenter map of the Persian Gulf-Gulf of Oman region (1900-1976). Compiled from data from Berberian [1976].

from the late Maestrichtian to the Eocene, shallow carbonates were deposited unconformably over the allochthon. The Pabdeh foredeep persisted in the northern United Arab Emirates and Musandam Oman. During the mid-Oligocene regression (or late Eocene according to Berberian and King [1981]) the greater part of the Persian Gulf and southeast Iran (Zagros basin) became emergent. By the late Oligocene the sea again covered most of the Arabian platform to the Zagros Crush zone. The second phase of folding and thrusting in the Musandam Peninsula occurred in the late Oligocene to Miocene, and collision of the Arabian platform and the Eurasian plate took place in the Oligocene or early Miocene. Emplacement of the Zagros Mountains in the Miocene-Plio-Pleistocene was the result of compression along the Zagros Crush zone during the closure of the High Zagros Alpine Ocean (southern Tethys). This compression was accompanied by shortening, thickening, and elevation of the continental crust and its sediment cover. As the result of the uplift the sediments slid southeastward along the Hormuz salt and early Miocene evaporites to form the Zagros fold belt. Except for an amplification of the tilting of the platform the rest of the Arabian platform to the southwest shows little trace of tectonism [Kassler, 1973]. Tectonism along the present boundary of the Arabian plate ranges from subduction of oceanic crust in the southeast (Gulf of Oman) to continental collision in the northwest (Zagros Mountains).

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D. A. Ross and E. Uchupi, Woods Hole Oceanographic Institution, Woods Hole, MA 02543.

R. S. White, Bullard Laboratories, Madingley Road, University of Cambridge, Cambridge, England, CB3 OEZ.

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