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Structure and Stratigraphy of Forearc Regions[1]

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Abstract Active continental margins and the active flanks of island arcs lie in the forearc regions of arc-trench systems generated by plate consumption. Arc-trench systems are initiated by contractional activation of previously rifted continental margins, by reversal of subduction polarity following arc collisions, and as island arcs within oceanic regions. The varied configurations of shelved, sloped, terraced, and ridged forearcs arise partly from differences in initial geologic setting, but mainly from differences in structural evolution during subduction. In regions where large quantities of sediment are delivered, forearc terranes enlarge during subduction through linked tectonic and sedimentary accretion of deformed ocean-floor sediments and igneous oceanic crust, uplifted trench-floor and trench-slope sediments, and the depositional fills of subsiding forearc basins. Where sediment delivery is small, enlargement is subdued or absent, and shortening of the arc-trench gap may be possible. Trench inner slopes typically are underlain by growing subduction complexes composed of imbricate underthrust packets of ocean-basin, trench-floor, and trench-slope sediments in thrust sheets, isoclines, and melanges. The structure of subduction complexes is governed by the thickness and nature of oceanic layers rafted into the subduction zone, variable thicknesses of trench and slope sediments, and the rate and obliquity of plate convergence.

Forearc basins between the magmatic arc and the trench axis include (a) intramassif basins lying within and on basement terranes of the arc massif, (b) residual basins lying on oceanic or transitional crust trapped between the arc massif and the site of initial subduction, (c) accretionary basins lying on accreted elements of the growing subduction complex, (d) constructed basins lying on the arc massif and accreted subduction complex, and (e) a composite of these basins.

Strata deposited in forearc basins are typically immature clastic sediments composed of unstable clasts derived from rapid erosion of volcanic mountains or uplands of plutonic and metamorphic rocks within the arc massif. In equatorial regions reef-carbonate associations are also common. Facies patterns of turbidites, shelf sequences, and fluviodeltaic complexes within forearc basins are governed by the elevation of the basin thresholds, the rate of sediment delivery, and the rate of subsidence of the substratum.

Petroleum prospects in forearc regions typically are limited by the prevalence of small, obscure structures within the subduction complex, the scarcity of good reservoirs in the forearc basin, the limited occurrence of source beds, and low geothermal gradients except within the arc massif where heat flux is commonly excessive.

INTRODUCTION

Plate tectonics has made possible an integrated analysis of the structural development of active continental margins and the flanks of active island arcs. Both areas are associated with plate consumption that generates arc-trench systems. Characteristic features include deep trenches marking the sites of plate consumption, chains or belts of volcanoes standing parallel with the trenches, and inclined seismic zones angling downward from the vicinity of the trenches into the mantle beneath the volcanoes. The seismic zones mark the course of plate descent. The arc volcanism evidently is triggered by descent of the plates of lithosphere into the asthenosphere (Barazangi and Isacks, 1976).

Subduction at the trenches is reflected by deformation of sediments lying on oceanic basement (Silver, 1971a) and may involve the basement as well. The bathymetric floor of the trench itself is underlain by undeformed deep-marine sediments ponded along the axis of the depression in many localities, but in several, little or no ponded sediment is present. The subduction zone of active tectonism thus actually occupies an uplifted belt on the arc side of the trench axis. As subduction continues over a period of time, the deformed rocks often form a mass of growing bulk that can be termed a "subduction complex." The subduction complex typically grows in width

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as successive increments are added by underthrusting at the trench. It also tends to grow upward to become a positive structural feature as a result of progressive tectonic thickening. Tectonic accretion at trenches thus has both lateral and vertical aspects (Karig and Sharman, 1975).

The zone of active igneous activity is termed the "volcanic arc." If emphasis is placed on the historic evolution of an arc-trench system rather than on its instantaneous behavior, the belt of igneous rocks can be called the "magmatic arc." This term includes not only the older volcanic components of the arc, but also the cogenetic plutons emplaced in the crust beneath the surface volcanoes. As the crustal profile of the arc is augmented by both volcanic and plutonic additions, the arc evolves as a positive topographic feature marked by mountainous islands or uplands. With respect to flanking sedimentary accumulations, we here adopt the term "arc massif" to denote the whole arc terrane of volcanic sequences, the underlying plutons, and their metamorphic country rocks.

The consistent parallelism of arcs and trenches implies that arc-trench systems have a high degree of longitudinal continuity coupled with marked lateral asymmetry. A satisfactory general description of their major morphologic and geodynamic elements is thus possible within the two-dimensional framework of a lateral transect. The relative positions of arc and trench in the arc-trench couplet then impart an inherent polarity to the arc-trench system as a whole. We therefore denote as the "forearc" region all those features lying on the trench side of the volcanoes at any given time. The "backarc" region is then behind the volcanoes. If the belt of active volcanoes changes position with time, various crustal elements may shift from the forearc to the backarc, or vice versa. This kind of complexity in the history of an arc-trench system can be expected.

Our purpose here is to discuss the geometry and evolution of structural and stratigraphic relations in the forearc region. We do not discuss in detail any of the backarc features lying behind the volcanic arc. Our focus is also primarily on sequences of sedimentary rocks within the forearc; therefore, we do not discuss in any detail either the igneous activity and metamorphism within the arc massif or the metamorphic and deformational regimes that may prevail at deep crustal levels within the subduction complex. Interested readers are referred to recent summaries of these processes by Ernst (1971, 1972, 1974), Miyashiro (1972, 1973, 1974), Presnall and Bateman (1973), and Ringwood (1974). On an even broader scale, the role of subduction-related phenomena in the overall tectonic evolution of orogenic belts has been well elucidated by Hamilton (1969, 1970, 1973).

In this article our discussion of forearc regions breaks naturally into four main parts. The first is an overview of forearc tectonics; included are summary diagrams for which full justification is delayed until later sections of the text. The second is an analysis of active forearc systems based mainly on data from seismic profiling and other work at sea (mainly by D.R.S.). The third is a synthesis of salient styles of forearc evolution over long periods as inferred from detailed studies of selected forearc regions (mainly by W.R.D.). The last is a brief evaluation of the implications of our analysis for the petroleum geology of forearc regions.

FOREARC TECTONICS

Arc-Trench Systems

The overall geotectonic settings of arc-trench systems are highly varied (Fig. 1). Some volcanic arcs have oceanic lithosphere in the backarc region as well as in front of the trench. The maximum thickness of crust in such intraoceanic arcs is intermediate between standard oceanic and normal continental values. Some intraoceanic arcs display extensional deformation of lithosphere in the backarc region, as interarc basins there open by a type of seafloor spreading that occurs during the time of igneous activity in the arc. In other places, however, the oceanic lithosphere in the backarc is apparently older than any of the arc activity.

Other arcs that lie along the edges of continental blocks have crustal thicknesses comparable to or far in excess of those beneath cratons in continental interiors. Some of these continental-margin arcs display contractional deformation along fold-thrust belts in the backarc region. Others, however, do not. Some display extensional deformation either along or just behind the igneous belt, but reverse faults manifest compression affecting this zone in some arc-trench systems.

In concentrating exclusively here on the forearc region, we are aware that some aspects of forearc evolution may be influenced strongly by geologic circumstances and styles of tectonism in the backarc region. The most important of such factors are probably those that bear on the crustal thickness in the arc massif, and on the rate and kind of sediment delivery to the forearc area. Nevertheless, the dominant controls on forearc evolution are seemingly related to the geodynamic effects of subduction. We are thus able to discuss forearc development as a separate and co-

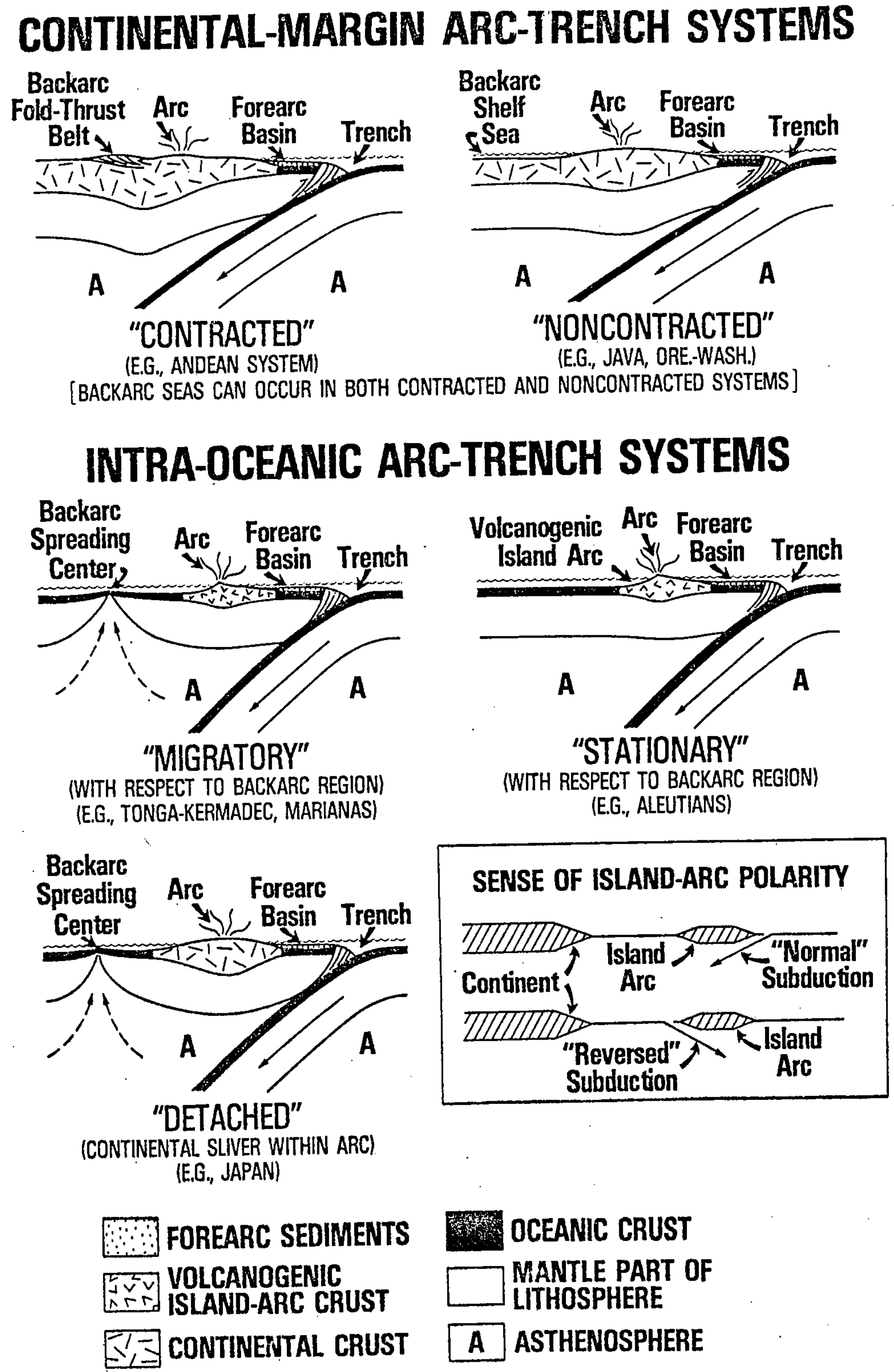

FIG. 1—Settings of arc-trench systems (modified from Dickinson, 1975) defined by backarc features.

herent topic without systematic reference to backarc development.

Accordingly, on Figures 2 and 3 generalized models of geotectonic features within the forearc region are presented. The terminology adopted for use in this paper (Fig. 2) is meant to be valid in any given case only for a particular instant in time. Our own past usages and those of other key authors are also indicated on the diagram. In our nomenclature here, we distinguish between morphologic terms and petrotectonic terms. The former refer to bathymetric and topographic features within the forearc region, whereas the latter refer to rock masses beneath the surface. The nature of the morphologic features is now well determined for many modern arc-trench systems,

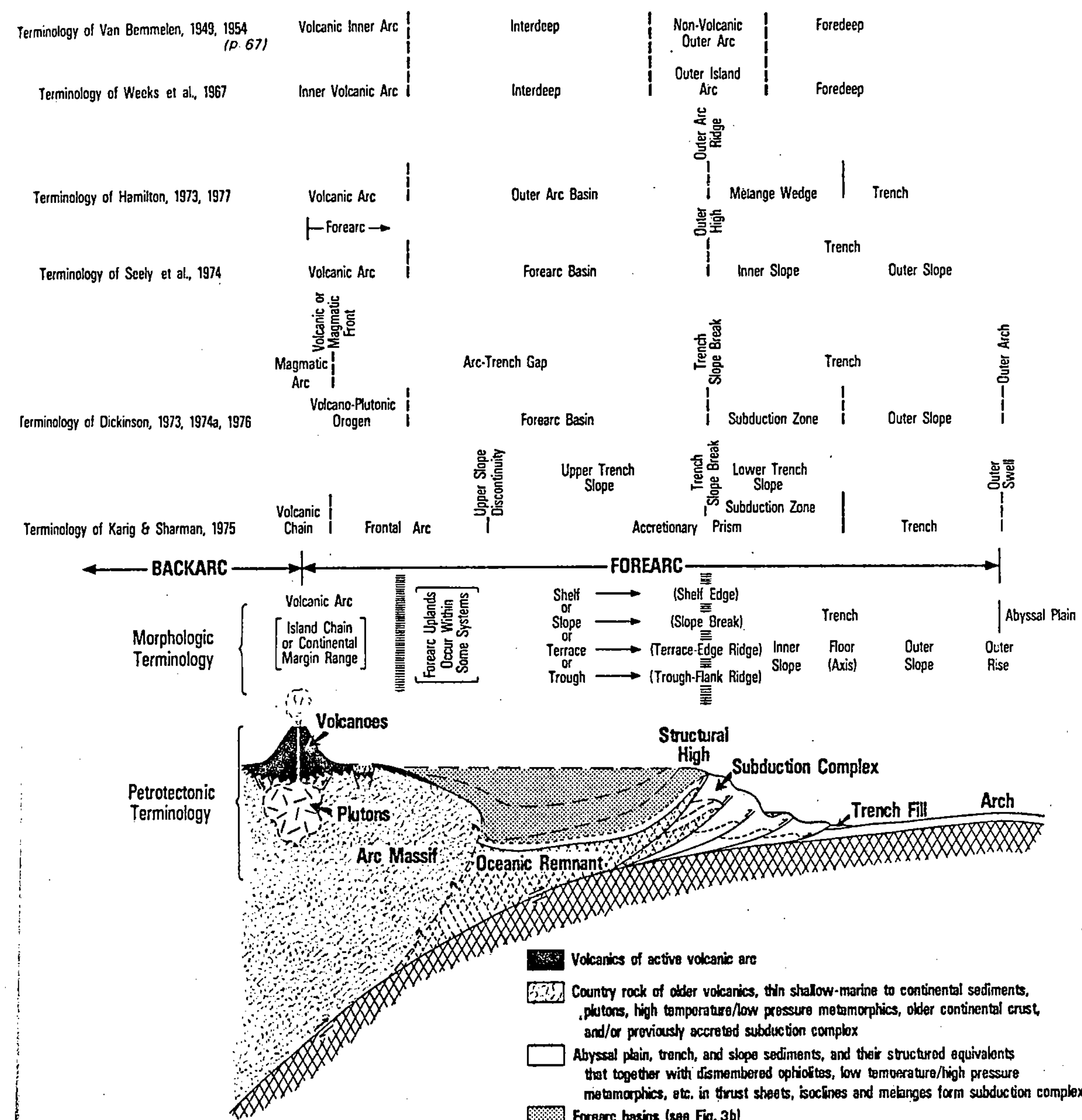

FIG. 2—Generalized instantaneous forearc model (after Ernst, 1970; Grow, 1973; Marlow et al, 1973; and references cited for terminology).

blages is still largely inferential. The petrotectonic assemblages have been studied in detail only for eroded arc-trench systems whose morphologic features can no longer be observed directly. Consequently, the correct correlation of petrotectonic assemblages with morphologic features requires mutual feedback between parallel studies of modern and ancient forearc systems.

Most of the terminology used here is adopted from and similar to past usage by ourselves or others. However, we do classify forearc basins in an unfamiliar way (Fig. 3). Depending on the nature of the substratum beneath the basin, we recognize four types of forearc basins: (1) intramassif basins, where the strata lie unconformably on rocks of the arc massif; (2) residual basins, where the strata lie depositionally on oceanic or transitional crust trapped between the arc massif and the subduction zone; (3) accretionary basins, where the strata lie directly on the subduction complex; and (4) constructed basins, where the strata lie unconformably across a structural join between the arc massif on the inner side of the basin and deformed accreted strata of the subduction complex on the outer side of the basin. It may well be that a composite of these basin types is the most common occurrence where the forearc basin is broad. For example, the basin that begins as a residual forearc basin on Figure 3 (upper right) develops into a broad composite basin that onlaps the arc massif on one flank and lies on the subduction complex on the other. Although forearc basins are primary depositional sites, major sediment accumulations can also occur as slope, trench-fill, or abyssal-plain deposits.

Forearc Types

The mechanisms that initiate plate consumption at new subduction zones and thus generate arc-trench systems are incompletely known. Four general scenarios are inferred (Fig. 4), two each for intraoceanic and continental-margin arcs (A, C, and B, D, respectively): (1) breakage across a previously intact oceanic plate, (2) activation of a previously passive continental margin, (3) reversal of polarity for an isolated intraoceanic arc, and (4) reversal of polarity following accretion of an intraoceanic arc to a previously passive continental margin by crustal collision.

The thickness and nature of the crust that existed between the arc and the trench at the onset of subduction are difficult to evaluate from active modern arc-trench systems. That region is typically buried beneath the thick sediments of forearc basins. We believe, however, that segments of oceanic crust and lithosphere are commonly either as continuous strips or in pockets present at intervals. Our reasons are twofold (Fig. 5).

1. Analysis of the spacing between arcs and trenches in modern systems indicates a minimum distance of about 100 km between the trench floor and the volcanoes (Dickinson, 1973). This minimum distance is true even for systems where subduction is inferred to have begun as late as the Neogene. The relation evidently stems from the fact that arc volcanoes stand uniformly at least 90 km, and most commonly 100 to 150 km, above the inclined seismic zone (Dickinson, 1975). The seismic zone is thought to mark the cool upper part of the lithosphere descending into the mantle from the subduction zone at the trench (Fig. 5A). The arc could thus never stand immediately adjacent to the trench unless the consumed plate of lithosphere made a sharp bend right at the subduction zone and plunged vertically into the mantle. Behavior of this kind during plate flexure is apparently precluded by the physical properties of lithosphere. Consequently, for all intraoceanic arcs initiated by breakage of an oceanic plate, a continuous strip of oceanic crust should underlie some part of the forearc region where it would lie between the flank of the arc massif and the accreted subduction complex. Similar bands of oceanic crust might be present beneath the forearc regions of other kinds of arcs unless subduction was begun precisely at the interface between oceanic crust and the thicker crustal profile of an island arc or continental margin.

2. Consideration of the geometry of plate consumption makes it clear that some pockets of oceanic crust probably will be trapped between the arc and the trench even where the initial subduction zone follows such a crustal interface as closely as it can. The argument in support of this contention is also geometric. The thickness of plates of lithosphere evidently dictates that flexure generally must occur along hinges that are smooth and have modest curvature in plan view. Nearly all trenches are smoothly arcuate, as the name arc implies. However, the edges of rifted continental margins are jagged in detail (Fig. 5B; Dewey and Burke, 1974). Marginal offsets linking sharp reentrants and projections are common on various scales. The same is probably true, though less well demonstrated, for the rifted flanks of migratory arcs and remnant arcs associated with the opening of interarc basins. Consequently, a newly initiated subduction zone cannot faithfully trace every irregularity in the rifted edge of a crustal block. Nor can lithosphere containing thick blocks of buoyant crust be consumed. The trend of a trench that develops along the jagged edge of

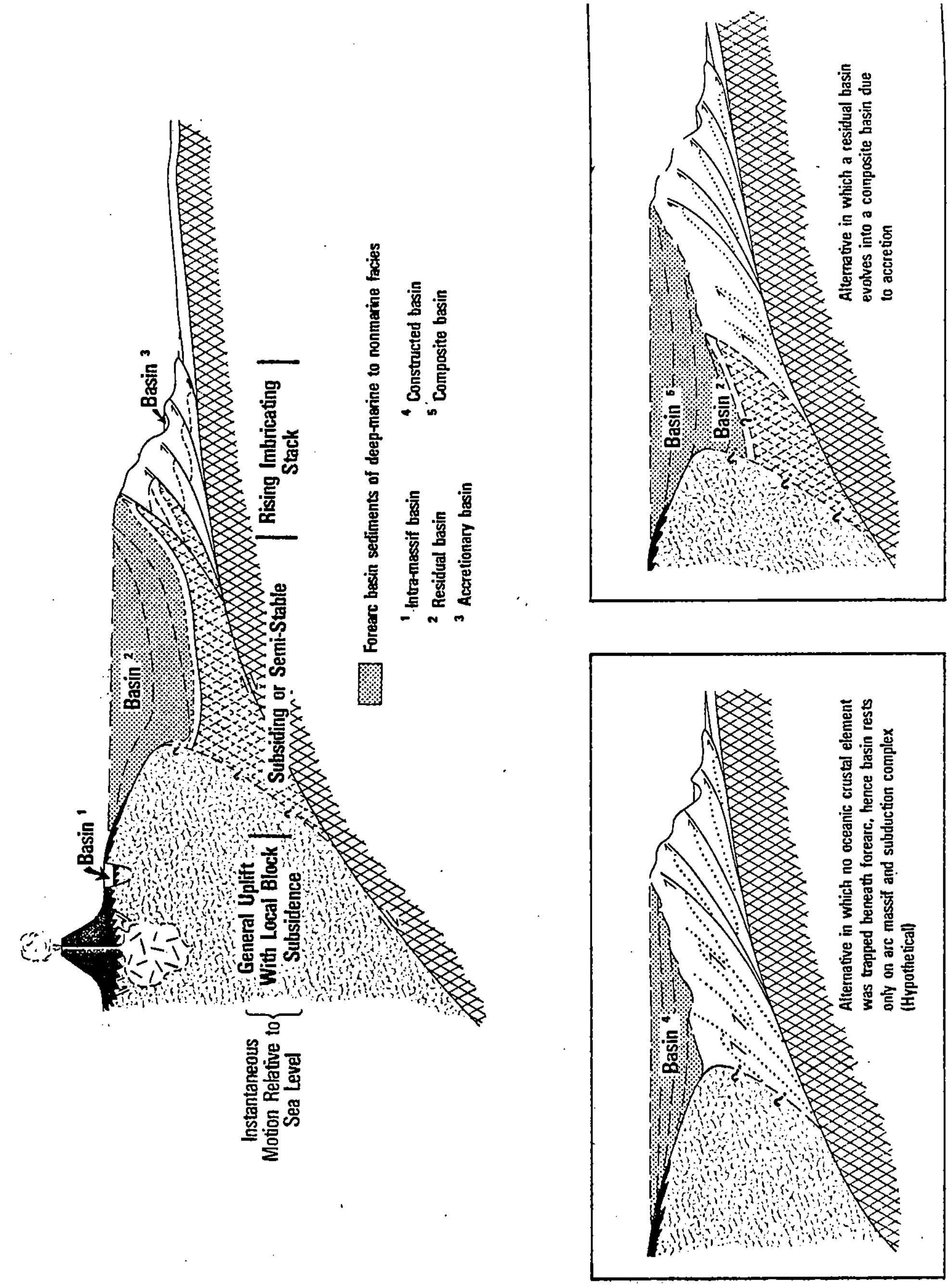

FIG. 3—Forearc basins.

a crustal block must thus conform to the positions of the points of crustal projections extending into the oceanic region (Fig. 5C). Oceanic and transitional crust between the projections will be caught between arc and trench (Fig. 4B [2a]), but this will not occur at the projections (Fig. 4B [2b]).

We strongly suspect that the presence of thin crust of oceanic and transitional character between the arc and the trench is a prerequisite for

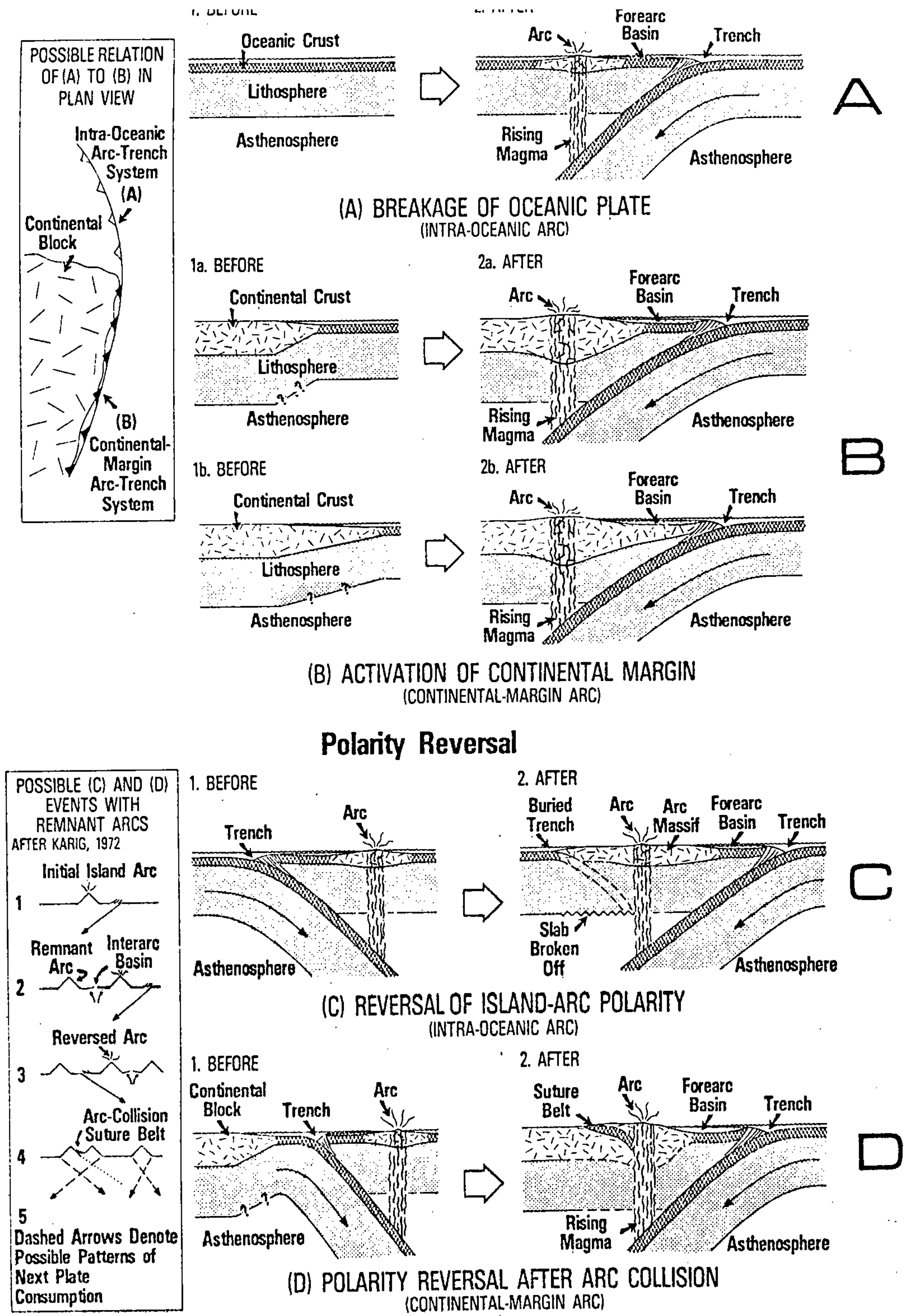

FIG. 4—Models for initiation of arc-trench systems (see text for discussion).

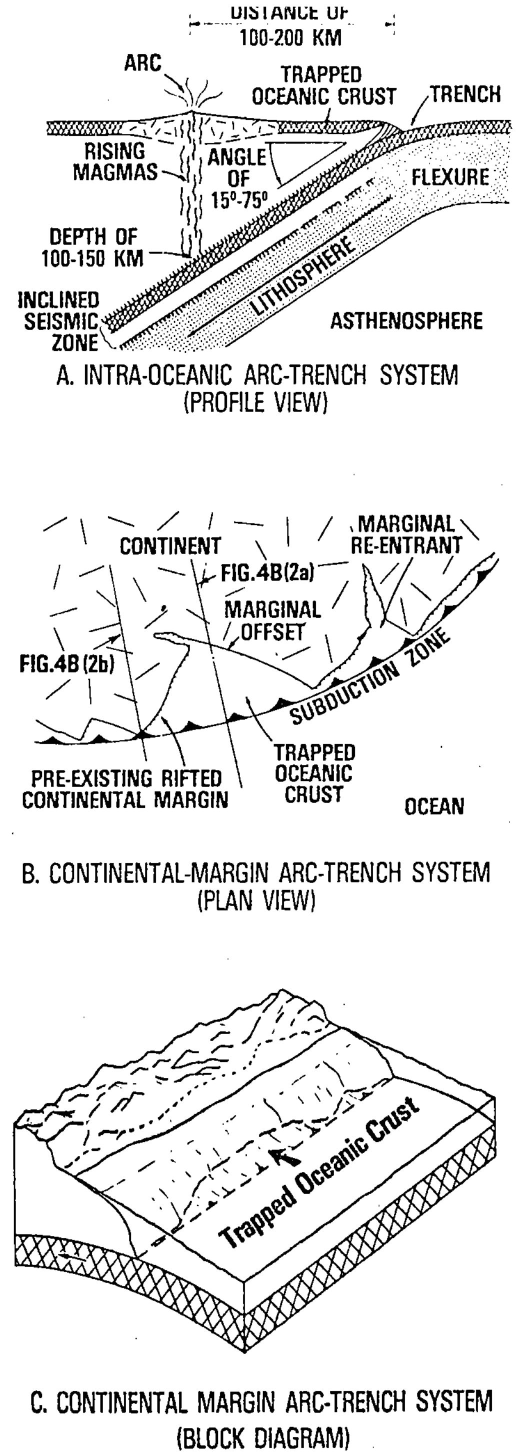

FIG. 5—Origins of trapped oceanic crust in forearc region.

truly deep water or large net subsidence within the forearc region. Deep or thick forearc basins are thus probably initiated as residual forearc basins in our usage here. In effect, the potential for isostatic subsidence under sedimentary loading is essentially the same for such forearc basins as for the sediment prisms deposited off rifted continental margins (cf., Karig et al, 1976). Additional subsidence caused by compressional downfolding of the thin crust also is possible.

Data on the nature and thickness of the crust across the forearc regions of modern arc-trench systems are commonly inadequate to distinguish between residual basins lying on trapped oceanic crust and constructed basins lying on the arc massif and deformed subduction complex (e.g., Grow, 1973). However, recent data for the Sunda and Banda arcs in Indonesia do indicate, as we would infer, that prominent forearc basins lying south of Java and Bali were initiated as residual basins because they mainly overlie oceanic crust (Curray et al, 1977). The crustal profile is somewhat thicker than standard oceanic values and may be transitional to the crustal profile of the arc massif. The younger sedimentary fill in these prominent forearc basins laps partly onto the arc massif on one flank and the subduction complex on the other as evolution is producing a composite basin. All these features are compatible with our views here (see preceding and Figs. 2, 3).

Residual forearc basins can become major repositories for sediment accumulation. The function of the arc massif is to provide a nearby source of sediments. The function of the subduction complex is to serve as a dam to pond sediment in the forearc basin. As subduction proceeds, the morphology of forearc regions may adopt varied configurations that we have classified descriptively as shelved, sloped, terraced, and ridged (Fig. 6). Distinctions between the various types of forearc basins depend on the structural evolution of the subduction complex and the history of accompanying sedimentation. Details of the processes controlling forearc morphology are discussed in the following section on active forearc systems. Long-term changes in the configurations of forearc regions are discussed in the section on forearc evolution.

ACTIVE FOREARC SYSTEMS

Outer Rise

Where present, the outer rise is a broad upwarp that rises gently from the abyssal plain and bends downward more steeply into the trench (Figs. 2, 3). It is commonly elevated a few hundred meters to a kilometer or more above the abyssal plain; lower elevations are associated with arc-trench

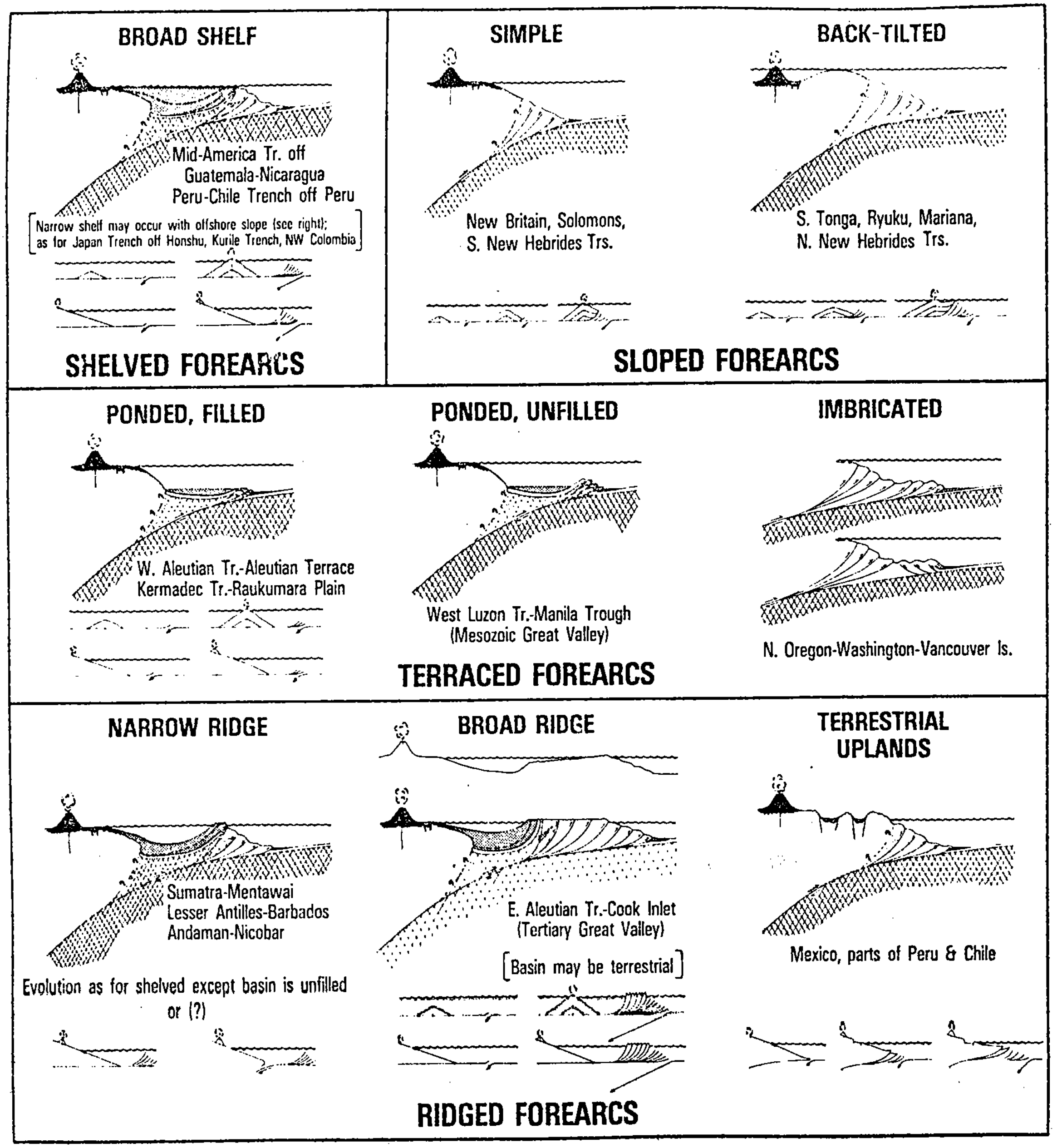

FIG. 6—Configurations of modern forearc regions. Volumes of subduction complexes in modern forearcs are unknown in most areas (note question marks on diagrams), as are subsurface extent and nature of forearc igneous and metamorphic rocks. Answers to these questions are among current objectives of JOIDES investigations. Where subduction complex volumes are small, features of arc massif may be nearer trench than shown.

systems having higher volumes of accreted materials (Karig et al, 1976). The breadth of the rise generally measures hundreds of kilometers, but, because of its asymmetry, its crest is only 100 to 200 km from the trench axis. Configuration of the rise is duplicated by a positive free-air gravity anomaly (Karig, 1973; Watts and Talwani, 1974; Segawa and Tomoda, 1976; Fig. 7).

Causes for an outer rise have been variously interpreted. Downflexing of the oceanic lithosphere into a trench as primarily an elastic phenomenon caused by downward and/or relatively large horizontal stresses acting on the lithosphere in the subduction zone was suggested by Watts and Talwani (1974). The shape of the outer rise has been more recently explained in terms of a bending moment applied to the oceanic lithosphere either by the down-going slab or by relatively small shear stresses at the underthrusting plate boundary (Parsons and Molnar, 1976). Caldwell et al (1976) modeled the outer rise assuming a thin (20 to 29 km) elastic plate and neglecting horizontal applied forces. On the basis of studies of earthquakes and faulting in the lithosphere below the outer rise, and on the difficulty of fitting the deeper parts of Benioff zones to an elastic plate of normal thickness, W. M. Chapple and D. W. Forsyth (personal commun., 1977) concluded that the oceanic lithosphere is an elastic-plastic medium with relatively low yield stresses in its lower half.

In connection with the downflexing, extension is believed to occur in the upper 20 km of lithosphere in the outer rise (or in the downbend if a rise is not present), and the extension may cause normal faulting. Small-scale intrusions into the extensional regime from the lower, compressed, 60-km zone are postulated to cause local gravity anomalies and to explain the anomalously high heat flow (0.2 to 0.4 HFU above normal) observed in the Japan and Izu-Bonin Trenches (Segawa and Tomoda, 1976).

Trench Outer Slope

The outer slope is underlain by abyssal-plain sediments lying on a down-bent and commonly normal-faulted, usually oceanic basement that may carry various types of irregularities (Fig. 7). A prominent variation in this general relation is the Australian margin of the Timor Trough—its outer slope. There the sediments lie on a continental rather than oceanic basement. The continental basement is also being broken by normal faults as it underthrusts the island of Timor.

The thickness and composition of the sediments on the outer slope vary widely as the result of several controls, such as the age of oceanic basement, the latitude, the carbonate compensation depth (van Andel et al, 1976), the proximity to major river mouths, and the obliquity and rate of plate convergence. Young basement ages, middle latitudes, deep carbonate compensation depths, long distances to major river mouths, and high angles and rates of plate convergence all reduce the thickness of the sedimentary cover. Deep carbonate compensation depths and latitudes away from the equatorial organic bloom during sediment accumulation also greatly reduce the amount of calcium carbonate to be expected in sediments on the trench outer slope.

Basement-involved normal faults cutting the sedimentary cover on trench outer slopes have been described from many trenches, including the Japan (Ludwig et al, 1966), Peru-Chile (Hussong et al, 1976), Mid-America (Seely et al, 1974), Kurile (W. H. Geddes, personal commun., 1977), and Puerto-Rico (Maley et al, 1974) trenches, as well as the previously mentioned Timor Trough. Although apparently a common feature of trench outer slopes, basement-involved normal faults are not everywhere present or abundant on them. They are not abundant, for example, on the outer slope of the eastern Aleutian Trench in the area of JOIDES holes 178 to 180 or in the Cascadia basin off northern Oregon, Washington, and Vancouver Island. As both these latter areas are characterized by large-scale tectonic accretion in the forearc region, it may be that the presence of a broad wedge of accreted materials may suppress the tendency for normal faulting by inducing a very gentle slope to develop on the trench flank of the outer rise (cf., Karig et al, 1976).

Basement-involved thrust faults, however, are uncommon features of trench outer slopes. The few found in the lower outer slope of the Peru-Chile Trench (Hussong et al, 1976) are the only ones described in the Pacific. Numerous thrust faults have been interpreted on the outer (southern) slope of the eastern Mediterranean (Hellenic) trench (Rabinowitz and Ryan, 1970). The most prominent thrust system can be viewed as a subsidiary subduction zone that causes the Hellenic Trench to bifurcate into two troughs, the main Pliny Trench on the north and the narrow cleft of the Strabo Trench on the south (Jongsma, 1977). The presence of such thrust faults and the absence of normal faults in other trenches may be due to horizontal compressive stresses acting on the descending lithosphere in the subduction zone (Watts and Talwani, 1974).

Three major types of irregularities on trench outer slopes can play a role in the evolution of arc-trench systems: (1) irregularities in the basalt surface caused by factors controlling emplacement of the basalt originally or by subsequent faulting, (2) bathymetric features such as seamounts, and (3) spreading ridges and other features that segment the seafloor, such as fracture zones or active strike-slip faults.

Although undocumented, irregularities in the basalt surface (Fig. 7) should increase the possibility that parts of the igneous basement will be

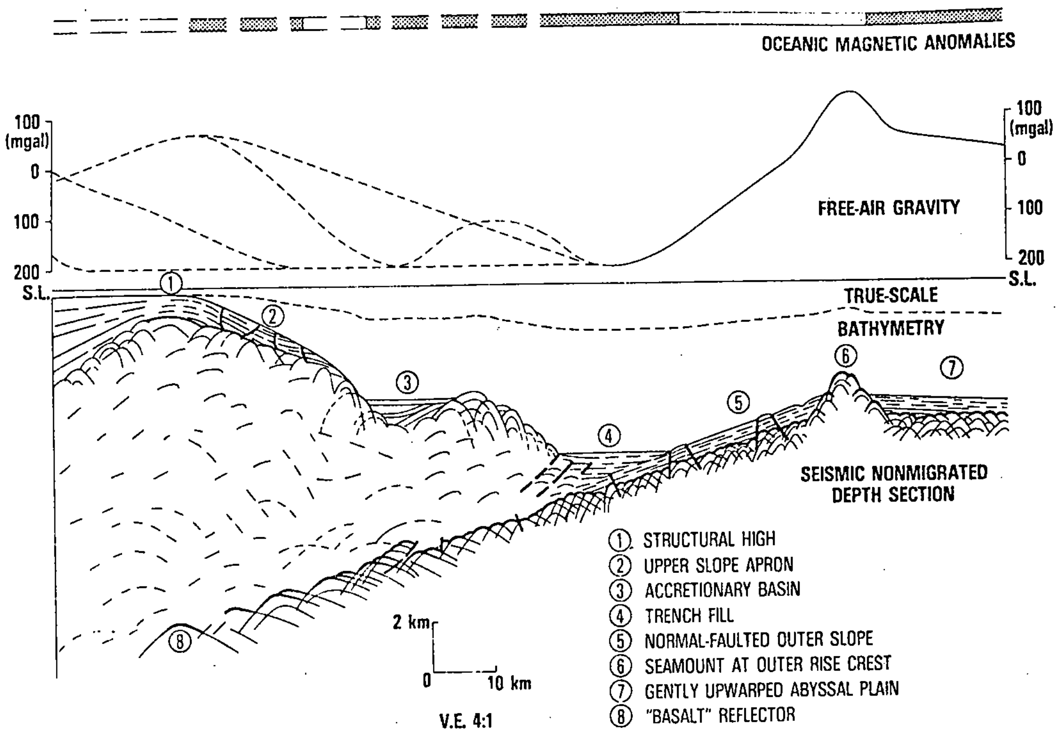

FIG. 7—Composite diagram showing common geophysical expressions of trench; seamount shown at outer rise crest is in transit from abyssal plain to outer slope of trench as oceanic plate moves toward trench.

cut off from the oceanic plate and incorporated into the subduction complex when rocks of the outer slope are thrust under the inner slope (Karig, 1974). The possibility should be greater if the relief of the irregularities is large in relation to the thickness of sediments lying on the outer slope and flooring the trench. An example where the relief is small relative to sediment thickness is in the eastern Aleutian Trench where a buried hill probably composed of igneous crust is present beneath the lower part of the trench inner slope. It has not yet been cut off from the oceanic plate (Seely, 1977).

Collisions between trench inner slopes and prominent bathymetric features such as seamounts and aseismic ridges that are common on the outer slopes of many trenches should have produced anomalous sections along arc-trench systems. Depending on the sizes of the colliding features and whether they represent zones of weakness in the oceanic lithosphere, the anomalies may be confined to the trench slope or may involve most of the system. Though not yet described on modern trench inner slopes—but postulated on an ancient slope of the Oregon-Washington coastal strip (Cady, 1975; Dickinson, 1976)—the collisions of seamounts with an inner slope composed largely of a sedimentary subduction complex should produce local magnetic sources associated with abrupt changes in slope structure that may have bathymetric expression. However, the near coincidence between the collisions of the Nazca and Juan Fernandez Ridges with western South America and offsets of the Benioff zone (Barazangi and Isacks, 1976) suggests that subduction of these types of large features can segment the entire arc-trench system by breaking the underthrust plate into separately subducting pieces. L. D. Kulm (personal commun., 1977) reported that the trench inner slope is also steepened and without basins where the Nazca Ridge collides with it.

Definitive evidence for some of the effects of the intersection of spreading ridges with an arc-trench system has not yet been presented. Principal remaining points of discussion revolve around how far into backarc areas ridge subduction is manifested, to what extent the overriding plate is elevated at the intersection, and what effects increased heat flow and magmatism have at that point. Effects on plate motions at triple junctions of this type have been discussed by McKenzie and Morgan (1969). Generally, relative motions along the trench differ on the two sides of the junction, resulting in a change in characteristics of the arc-trench system at that point (e.g., Chile Rise–Peru-Chile Trench intersection) or its termination (e.g., East Pacific Rise–Mid-America Trench intersection).

Similarly, the presence of relatively small-scale strike-slip faults that have been active during subduction, such as those in the Cascadia basin (the outer slope of what we consider to be the Oregon-Vancouver filled trench) described by Silver (1971b) and in the Gulf of Alaska (R. W. Couch, personal commun., 1977), results in differing relative motions along trenches. The expression of such faults or reactivated fracture zones on overriding plates, like that of spreading ridges, has not been fully documented. Generally, however, the localization and amount of shear experienced by the overriding plate because of slip along these faults should change as underthrusting continues by an amount that depends on their rate of slip relative to the rate of underthrusting and their trend relative to the direction of underthrusting. Higher rates of slip and greater parallelism between their trend and the direction of underthrusting tend to increase and localize the shear exerted on the overriding plate, thus increasing the possibility that the strike-slip faults will be expressed in it.

Trench Axis or Floor

A line connecting the deepest parts of a trench defines its axis (Figs. 2, 3). Where no wedge of turbidites fills the axial area, such a line is readily drawn and generally marks the collision point between underthrusting sediments covering the outer slope and compressing sediments and/or oceanic igneous rocks forming the subduction complex at the toe of the inner slope. Where a trench-fill turbidite wedge is present (Fig. 7), the trench will have a subhorizontal floor with a breadth dependent on the extent of the wedge. This, in turn, depends on the depositional rate of the turbidites, the magnitude of the convergence-rate component perpendicular to the slope, and the inclination of the outer slope. High depositional rates and low perpendicular convergence components favor deposition of large quantities of trench turbidites. A low outer-slope inclination favors development of a broad floor.

Where the low rates of convergence result in part from highly oblique subduction rather than simply slow subduction, the formation of large volumes of trench turbidites is especially favored if the subduction is along a continental margin because of the prolonged transit time of the seafloor in the commonly high-sedimentation-rate fringe of passing land areas. This effect combined with high sedimentation rates related to major nearby river systems appears to be responsible for the large quantities of trench sediments present off the Oregon-Washington-Vancouver coast ("Cascade Trench") and in the northwestern part of the Java (Indonesian) Trench.

About 55% of the combined length of Pacific trenches contains less than 400 m of sediment. A large proportion of this length is adjacent to western and southwestern Pacific volcanic arcs. Thicker trench fill is common in the continental margin trenches of the eastern Pacific (Scholl and Marlow, 1974).

Trench-fill turbidites in the eastern Aleutian and Japan Trenches consist largely of muds and silts with intercalations of medium-grained to very fine-grained, erratic-containing sands. In the eastern Aleutian Trench (Site 180), 470 m of late Pleistocene mud, silty mud, graded silts, and ash beds were drilled. Granule and pebble erratics, a detrital shelf fauna, and diatomaceous strata were identified. Site 180 is located on the seaward side of the trench floor, and the hole may have penetrated overbank deposits from a turbidite channel presently located near the base of the slope (von Huene et al, 1973a). At JOIDES Site 181, a short distance up the inner slope from Site 180, the lower 200 m of section penetrated is considered to be deformed trench sediments (von Huene et al, 1973b). The section is characterized by faintly bedded mudstone. Within the mudstone are thin, well-sorted silts and poorly sorted medium to coarse sands having a clay matrix and containing pebble-sized erratics.

Drilling at JOIDES Site 174 (Kulm et al, 1973) in the Cascadia basin encountered 284 m of upper Pleistocene sediment, which we interpret as trench fill, overlying approximately 600 m of abyssal-plain section. The fill consists of thick-bedded to thin-bedded, moderately sorted, medium-sand to very fine-sand turbidites and lesser amounts of interbedded silty clay and silt. The sands are subangular to angular and have an average composition of 39% quartz, 11% potassium feldspars, 21% plagioclase, 23% lithic material, and 6% heavy minerals. Many were probably deposited on the outer part of a submarine fan having its apex near the foot of the inner slope of the "Cascadia Trench." At the latter locality coarser sediments of the inner fan should be present.

JOIDES Site 298 is located near the foot of the inner slope of the Nankai Trough, but sediments penetrated in the 611-m hole are believed to be deformed trench fill (Ingle et al, 1975). They consist of an upper section of clayey silt, and clayey and silty, medium to fine-grained sand with granule- and cobble-sized erratics of silty claystone, calcareous sandstone, and limestone. The lower section is dominantly clayey silt and silty clay, although graded medium-grained to very fine-grained, poorly sorted sands are also present and are composed of quartz, feldspar, and minor volcanic ash. The two sections are considered to represent proximal and distal turbidites, respectively.

Trench Inner Slope

The trench inner slope, at least in its lower part, appears generally to be underlain by the subduction complex (Figs. 2, 3), a thick section of deformed abyssal-plain, trench, and slope sediments containing variable proportions of inferred oceanic crustal slivers (Seely et al, 1974). This section is covered in places by a relatively thin veneer of lower slope sediments, and the highly irregular surface commonly topping the section forms local structural-bathymetric prominences and accretionary basins (Moore and Karig, 1976), which are discussed in the section on forearc basins. The upper parts of the slopes of several trenches (Mid-America, Seely et al, 1974; Peru-Chile, Hussong et al, 1976; eastern Aleutian, von Huene, 1972; Japan, S. Uyeda, personal commun., 1977; Tonga-Kermadec, Karig, 1970) are covered by a relatively thick sediment apron (Fig. 7) that has spilled across the edge of the shelf or been transported by contour currents.

The toe of the slope is usually the seaward edge of the active belt of thrusts and folds that is deforming the abyssal-plain and trench sediments scraped off the underthrusting plate (Fig. 7), although local slumps may modify the toe to varying degrees. The intensity of surface deformation diminishes progressively upslope from the toe at a rate that appears related to the depth to the top of the underthrust plate and to the rigidity of the overthrust plate (which, in turn, may be related to the proportion of sediment underlying the slope). Lesser depths and rigidities should permit development of a broad zone of active surface deformation, as in the eastern Aleutian and Lesser Antilles Trenches.

In most localities where inner slope structure can be discerned, the thrusts dip landward (Fig. 7), and associated folds have seaward vergence. This produces a prevalence of landward dip at lower structural levels beneath the slope. In some localities progressive landward tilting of the subduction complex produces steeper landward dips beneath the upper slope than beneath the lower slope (Seely et al, 1974). The degree to which this structural "fanning" occurs is related to the spacing and sizes of folds and thrusts, and these dimensions, in turn, are a function of the physical properties of the rocks being deformed. In a few localities the dip relations may be the reverse of those described previously, that is, thrusting and folding produce dominantly seaward dips (Silver, 1972; Barnard, 1973; Kulm et al, 1973; Carson et al, 1974). These relations also may be caused by rock physical properties, especially the presence of low-strength sedimentary layers (Seely, 1977).

Major factors influencing the character of trench inner slopes (Fig. 8), because they influence the physical properties of the rocks being deformed and the stresses acting within them, are (1) the location of the convergent margin at its inception; (2) if located at an older plate margin, the type, or prior history of that margin; and (3) the sequential interaction of basic evolutionary variables. These variables include the age of the seafloor (the major control of the total bathymetric relief of the trench inner slope), the presence of irregularities in the underthrust plate previously discussed (these produce anomalous segments of the trench inner slope), the directions and rates of convergence (these influence structural trends on trench slopes and sediment thicknesses), and the depositional rates, thicknesses, and types of abyssal plain, trench, and slope sediments (these determine the types and amounts of sediments beneath trench inner slopes and are the primary control of slope steepness and the size, spacing, and vergence of inner slope structures).

The upper parts of the trench inner slopes of sloped forearcs and of some of the terrestrial uplands in ridged forearcs are underlain by the more rigid rocks of the arc massif (Fig. 6). The nature of the contact between the arc massif upslope and the subduction complex downslope is not known. It may be relatively abrupt and represent the seaward edge of a packet of subduction complex accreted in an immediately preceding pulse of subduction (Karig, 1974) or the edge of a presubduction continental margin. It may be gradational, resulting, for example, from progressive metamorphism and thickening of the subduction complex. Its significance is discussed in later sections that deal with terraced, shelved, and the other types of ridged forearcs.

Structural High

A prominent structural high is usually present beneath the shelf edge of shelved forearcs and the terrace edge of terraced forearcs, and it forms the narrow and broad ridges of ridged forearcs (Figs. 6, 7). It may cause a singular break in slope, the trench-slope break, in some sloped forearcs, but

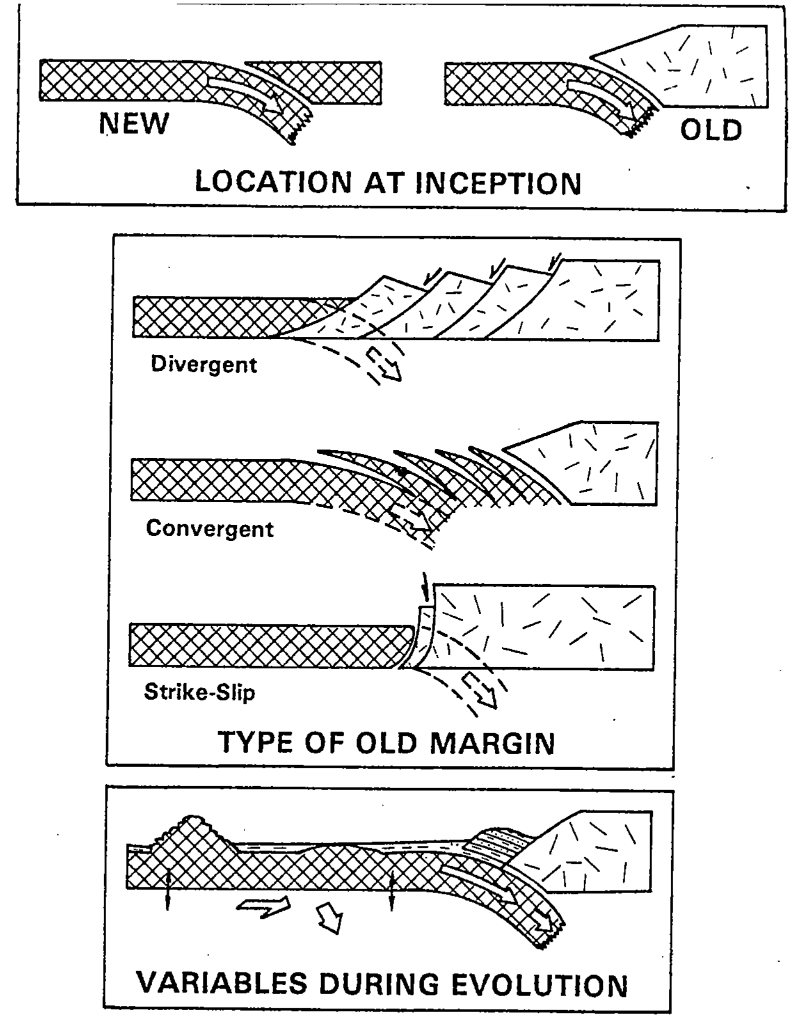

FIG. 8—Major factors influencing character of trench inner slopes.

the absence of such a break in other sloped forearcs suggests it is not present in them. Likewise, the absence of a trench-slope break, terrace, shelf, or trough-flank ridge along parts of the terrestrial uplands forearc of Latin America indicates the absence of a correlative structural high there.

Important factors to consider about the origin of a specific structural high are whether it represents the momentary upward culmination of a subduction complex that joins the arc massif and is growing upward and seaward with time (Karig, 1974; Seely et al, 1974), or whether it represents something more fundamental that is not migrating, such as the initial rupture zone formed at the onset of subduction or a major thrust plate of some kind within the subduction complex. In the first situation, sediments in the basin landward from the high lie on the subduction complex and, farther landward, onlap the arc massif (Figs. 2, 3) to form a constructed basin. In the second situation, initial sediments in the basin landward from the high may lie on a remnant of oceanic lithosphere or some type of transitional crust to form a residual basin, or they may be entirely contained in an accretionary basin within the subduction complex (Figs. 2, 3). A combination of origins is possible, such as when the structural high grows upward and seaward from the initial rupture within oceanic or transitional lithosphere.

None of these origins have been well documented in modern arcs. Coulbourn and Moberly (1977) pointed out the landward shift of the inner flank of a structural high at the edge of a relatively small terrace in the upper slope off northern Chile, and this phenomenon appears common elsewhere on trench inner slopes (Seely, 1979). The phenomenon, however, does not preclude the possibility of upward and outward growth of a succession of such highs, as postulated by Karig (1974), Seely et al (1974), and in greater detail by Moore and Karig (1976). However, the structural high at the edge of the Guatemalan shelf appears to have formed the edge of a residual forearc basin, because it is probably floored by oceanic lithosphere and has had no seaward migration since the Late Cretaceous (Seely, 1979). Cook Inlet–Shelikof Strait may also be a residual forearc basin and, if so, it probably originated in Mesozoic time (Moore and Connelly, 1977).

Intrusions are exposed on islands of the eastern Aleutian forearc ridge and are possibly present locally in the structural high at the edge of the Guatemalan shelf (Seely et al, 1974). The origin of these intrusions and the role they play in forearc evolution remains to be clarified (e.g., Hudson et al, 1977; Marshak and Karig, 1977).

Forearc Basin

Forearc basins vary in structural style and lithologic content. Arc-massif basins are characterized by block faulting and marine or nonmarine arc- or backarc-derived sediments; accretionary basins are commonly structured by compressional folds and listric thrust faults and are filled with marine sediments derived from uplifted subduction complex or arc terranes; residual and constructed basins are structurally and stratigraphically transitional between arc-massif and accretionary basins, except that residual basins also include a basal sequence of abyssal-plain sediments. These differences are related to the different types of substrata that define them and to the uplift of rock types that vary somewhat systematically in forearc areas. Brittle rocks of the igneous-metamorphic complex and the thick crust comprising the arc massif break into blocks in a stress regime removed from the direct effect of shear exerted by the underthrust plate. Seaward, there is a transition to ductile sediments that lie directly on the underthrust plate and are compressed as a result of shear at the plate boundary. Both the arc massif and the structural high are belts of uplift and, thus, are potential provenance areas for basins within them or proximal to them.

The faults in arc massif basins are generally considered to be normal, perhaps related to surficial extensions above rising plutons or magmatic withdrawal beneath the arc terrane. Reverse faults in the arc massif, however, have been described in the high cordillera of Chile (Reutter, 1974), and compression in the arc massif occurs in southwest Japan (Uyeda, 1977). Some arc-massif reverse faults may be caused by a cessation of magmatism and development of crustal compression during periods when the underthrust plate extends far into former backarc areas before diving into the asthenosphere (Barazangi and Isacks, 1976; Mégard and Philip, 1976).

A wide variety of structures occurs in residual and constructed forearc basins. The brittle fracture or warping of the arc massif as it is bent down to form the inner flank of the basins commonly produces irregularly shaped closures bordered by reverse or normal faults such as on the eastern side of the Sacramento Valley, California, and the west coast of Chile. Surficial normal faults are present also off Guatemala and Honshu. Compressional folds and thrusts occur on the seaward side of the Sacramento Valley, in the Cook Inlet, and on the landward side of the Nicaraguan forearc basin. Longitudinal strike-slip trends can be anticipated in some basins (Fitch, 1972; Karig, 1974) as a response to oblique subduction. Transverse structures also can be expected. Although such structures have not been documented in modern residual or constructed forearc basins, other transverse forearc trends (e.g., transverse trends offsetting the volcanic arc or belts of seismicity) have been reported (Stoiber and Carr, 1973; Spence, 1977). A major transverse element, the Stockton arch, segments the Great Valley of California.

Accretionary basins developing on trench inner slopes (Fig. 7) are structured with folds and thrusts by the compression that is deforming the subduction complex in which they are found. Listric normal faults, however, can also be present if the surficial fill is deposited with an initial seaward dip. Although in places seaward, dips of older basin fill commonly are landward and increase progressively with depth (Karig, 1974; Seely et al, 1974; Coulbourn and Moberly, 1977). The dips record the sequential tilting of the basin substratum apparently resulting from the basin's position within a thrust plate having a seaward edge that is being wedged upward by understuffing (Seely, 1979).

Forearc basin sediments have been described by Dickinson (1970, 1974a, 1974b). Typically, they consist largely of clastic deposits derived from the volcanic arc and its roots, and are deposited in water depths determined by the positions of the basins at origin and the relative rates of sedimentation and subsidence within them. Continental-margin arc-massif basin sediments are commonly nonmarine, whereas intraoceanic arc-massif basin sediments are probably dominantly marine. Residual basins originate in deep water, and constructed basins, hypothetically, originate at various water depths, but basin fill in either is dominantly marine. The sediments deposited in accretionary basins developing on trench inner slopes are relatively deep marine, whereas those in such basins forming on shelves or ridges may include nonmarine sequences.

An idealized residual basin sequence would be composed of deep-water, arc-derived montmorillonitic shales, ash falls, and lesser amounts of fine-sand turbidites lying on abyssal-plain sediments at the base of the sequence and grading upward to more abundant and coarser turbidite, shelf, or deltaic sands derived, in part, from the uplifted roots of the arc. On the inner edge of the basin these would interfinger with lava flows, lahars, tuffs, agglomerates, and fans.

Major exceptions to the idealized sequence exist where significant provenance areas other than the volcanic arc are present, as where long river systems reach the basins, where there are extensive nonvolcanic uplands nearby, or in equatorial areas where reefs can grow on shallow basin margins. Where river systems empty into forearc basins, the basins fill longitudinally and quickly and are composed of large volumes of shallow-marine to nonmarine facies deposited during the subsidence phase(s) of basin evolution (e.g., Cook Inlet, Alaska). Sediments eroded from extensive nonvolcanic uplands can be the dominant constituent of forearc basins (e.g., Peru). In equatorial regions, thick sections of foundered carbonate banks and reefs can accumulate during basin subsidence (e.g., Mentawai Trough).

FOREARC EVOLUTION

Over short periods, perhaps up to 1 million years, progressive tectonic development of the forearc region can be conceived in terms of minor incremental modifications to the instantaneous forearc model (Figs. 2, 3), which is viewed as a steady-state configuration. For time frames of millions or tens of millions of years, however, significant changes in morphology can be expected. Such changes most typically involve a continuous style of evolution during essentially steady tectonic accretion. For example, sloped or terraced forearcs may evolve gradually into shelved or ridged forearcs (Fig. 6). An initial residual forearc basin may later become a composite forearc basin (Fig. 3, top right). However, a prime factor governing forearc evolution is the quantity of sediment delivered to the forearc region. Large-scale lateral accretion can occur only if there are large quantities of trench-fill, abyssal-plain, and/or slope sediments. Where there is an absence of such quantities, lateral accretion cannot occur, and, possibly, shortening of the arc trench gap will ensue.

Other modifications may be induced in the forearc region as vectors of plate convergence vary through time. Key changes also may stem from discontinuous aspects of forearc evolution. Most important, perhaps, are jumps or shifts in the position of the subduction zone induced by tectonic accretion of large crustal blocks, such as intraoceanic arcs or seamount chains. Where such bulky and buoyant crustal elements lodge against a subduction zone, their incorporation into the subduction complex is a form of crustal collision or suturing together of crustal blocks. Subduction is thereby forced out beyond the newly accreted crustal block. Crustal collision may deform one forearc basin, thus terminating its history, and induce a new one to form at a different place (Dickinson, 1976).

The full interpretation of long-term trends in forearc evolution requires detailed knowledge of a staggering range of geologic data. In arc-trench systems, processes of sedimentation, metamorphism, volcanism, plutonism, and diastrophism are inseparably linked as related manifestations of the same geodynamic system. Radiometric data on the ages of igneous rocks in the arc massif and mineralogic evidence on the conditions of metamorphism within the subduction complex may be as important as paleontologic data on the ages of sedimentary beds in the forearc basin for understanding the overall evolution of the forearc region. Nevertheless, we have focused here on the forearc basins and, in particular, on selected well-studied examples in California and Alaska. First, however, we discuss briefly some of the salient evolutionary trends within the arc massif and the subduction complex that form the flanks of major forearc basins.

Arc Massif

In the geologic record, magmatic arcs appear as elongate belts of coordinate volcanic, or metavolcanic, and plutonic rocks of so-called calc-alkalic or orogenic type (Dickinson, 1970). The volcanic rocks are andesitic and dacitic suites, but with important basaltic and rhyolitic associates, and the plutonic rocks are granitic, in the broad sense, with subordinate dioritic and gabbroic associates. Coeval volcanic and plutonic assemblages together form cogenetic igneous suites (Ustiyev, 1965; Hamilton and Myers, 1967). Together with metamorphic wall rocks and underpinnings derived from preexisting crustal elements, these volcano-plutonic suites form the arc massif. The metamorphic rocks of the arc massif constitute the high-temperature terranes of paired metamorphic belts (Miyashiro, 1967). The region of high temperature in the crust forms as an elongate envelope enclosing the intrusions of the arc roots.

The position of the axis of magmatism with respect to the trench governs the extent of the forearc region. Slow retrograde migration of the magmatic axis away from the forearc is common at rates of up to 1 km/m.y. (Dickinson, 1973). Where this type of gradual shift in igneous activity occurs, the arc flank of forearc basins may widen as broad shelves develop across the eroded arc massif.

Aspects of the arc massif that influence the history of forearc regions more indirectly are those that bear on the rate and kind of sediment delivery to forearc basins. Where the crust is thick in the arc massif, there is a potential for isostatic uplift and subsequent erosion. Sediment derived from the plutonic roots of the arc massif may thereby join volcanic debris being deposited in the forearc basin. To some degree, the progressive emplacement of plutons into the arc massif may serve to thicken the crust there. Thus, the amount of uplift is likely to increase with time. Accordingly, the ratio of plutonic to volcanic detritus is likely to increase upward in the stratigraphic section within the forearc basin. However, the proportion of airborne pyroclastic debris in the section there may depend mainly on the direction of prevailing winds during the history of arc volcanism. The petrologic character of the volcanogenic materials produced along an arc tends to evolve from relatively mafic rock types early in the history of the arc to more felsic rock types during later stages of igneous activity.

Subduction Complex

Details of structural relations within the subduction complex are inferred almost entirely from studies of ancient systems whose interiors have been exposed by deep erosion following uplift. Only large-scale imbricate thrusting can be discerned by seismic profiling of still submerged active systems (Seely et al, 1974). The locations of ancient subduction zones are identified generally with the high-pressure terranes of paired metamorphic belts (Ernst, 1973, 1975). Only in subduction zones where surficial materials are transported rapidly to great depth can the requisite conditions of low temperature and high pressure be attained.

Materials displaying three distinct structural styles may be formed as discrete belts or slabs within the subduction complex (e.g., Cowan, 1974): (1) bedded sequences, which may be isoclinally folded, in which stratification is largely preserved and metamorphic fabrics are not widely developed, although metamorphic minerals may largely replace original constituents; (2) metamorphic tectonites, schistose or semischistose, in which a pervasive metamorphic fabric that largely supplants original textures exerts the dominant control on structural relations; and (3) chaotic melanges, in which pervasive mesoscopic shear fractures have disrupted the original rock into isolated tectonic inclusions immersed in a sheared matrix. The relative proportions of these three kinds of terranes vary widely for different subduction complexes and for different parts of the same subduction complex. Presumably, variations in strain rate and pressure-temperature regime within the subduction zone are the controlling factors. Contacts between units characterized by different deformational styles may be gradational locally, or may be clear-cut shear zones across which severe dislocation has shuffled the subduction complex.

The most problematic of the three components are the melanges (Hsu, 1968). The mixing of disparate rock types and the scaly fabric of the foliate matrix have been ascribed both to tectonic brecciation and to massive landsliding. In principle, either process could occur at subduction zones. In fact, the compromise notion that the characteristic melanges are strongly sheared olistostromes has been widely entertained. Unfortunately, the ordinary criteria used in other geologic settings to distinguish between mesoscopic structures of depositional and tectonic origin lose much of their force when applied to subduction zones. For example, soft-sediment deformation can provide no proof of strictly surficial origin by mass movements because unconsolidated sediments are continually rafted into a site of purely tectonic deformation. However, sharply bounded blocks of consolidated sedimentary rock do not preclude olistostromal origins because compacted sediments may be continually uplifted to form tectonically unstable submarine slopes. Moreover, unfamiliar processes may be responsible for some or all melanges. Where saturated sediments are drawn rapidly into a subduction zone and begin to dewater under tectonic overburdens, unusually high strain rates may be imposed on markedly overpressured sediments that still retain high porosity. The products of deformation under these circumstances in which materials lack cohesion, but are confined, might well display a struc-

ly associate with tectonic shearing on the one hand and olistostromal sliding on the other. Whatever the true origin of melange, it appears now to be a reliable, though not ubiquitous, guide to the location of ancient subduction zones.

Whether they are present as intact bedded sequences, as foliated tectonites, or as blocks in melanges, the rocks of subduction complexes may represent a wide range of oceanic environments. Gabbros and diabases of the oceanic crust and ultramafic rocks of the oceanic mantle may be parts of intact ophiolite sequences or detached thrust slices. Any volcanic rocks present are typically basaltic greenstones representing either deep seafloor or seamount edifices. Shallow-marine carbonate rocks may be associated with volcanic piles that served as pedestals for carbonate platforms. Turbidites may include the deposits of widespread abyssal plains carried into the subduction zone, local trench fill spread longitudinally along the belt of subduction, and subsea fans built transversely across the site of subduction. Turbidites and other deposits ponded in perched accretionary basins on the trench inner slope may also be incorporated into the subduction complex as deformation continues. Finer grained sedimentary rocks may include oceanic pelagites and hemipelagites of the open seafloor as well as analogous trench-slope deposits. The heterogeneous stratal components of subduction complexes ordinarily are arranged as successively underthrust packets in a complexly imbricated array (e.g., Maxwell, 1974). Any combination of rocks—turbidites and pelagites, melanges and isoclines—among the range potentially present may locally be juxtaposed across faults. Perhaps no other geologic setting inherently creates such structural complexity as does subduction.

Contacts between the intensely deformed strata of the subduction complex and the orderly strata of major forearc basins may be either depositional or tectonic. Both are noted in the following examples discussed. Depositional contacts are formed where strata along the flank of a forearc basin lap across uplifted subduction complex forming the structural high (see Figs. 2, 3). Such contacts are commonly unconformable. Tectonic contacts are formed where structural telescoping is in progress along or beneath the trench flank of the forearc basin as deposition continues. Such contacts are thrust surfaces or shear zones in outcrop.

Forearc Basin

For most forearc basins regardless of type (Figs. 2, 3), the general trend in paleobathymetry is toward shallower water depths for successively higher stratigraphic zones. Of course, the sedimentary dynamics of depositional systems, such as subsea fans or deltas, may cause local reversals to the trend. In gross, however, the filling of initial sediment traps and the uplift of accreted subduction complex that may partly underlie the basins bring the sediment-water interface progressively closer to sea level from a position of maximum depth at the onset of forearc sedimentation. There are exceptions where tectonic evolution of the forearc has involved discontinuous changes in behavior. Typically, then, but not in all places, turbidites and associated deep-marine strata preserved at depth within forearc basins eventually give way upward to deltaic or shelf deposits if sedimentation continues long enough.

Major subsea fans and deltas built into forearc basins originate on the arc flank from which most sediment is derived. Where volcanic processes are most active in the arc, clastic depositional systems of marine environments within the forearc basin can pass gradationally into laterally equivalent volcaniclastic systems of terrestrial environments within the arc massif (Kuenzi et al, 1975). The transverse dispersal of sediment into the forearc basin may be restricted to specific river mouths or submarine canyons located along the arc flank of the basin. From these points of delivery, longitudinal dispersal may be dominant along the floor of a turbidite trough or elongate fluvial plain occupying the axial part of the basin. Dispersal patterns in the trough can be quite complex owing to the bathymetric relief created by actively growing structures.

Carbonate buildups may occur at suitable latitudes either fringing the arc massif or atop the structural high along the trench flank of the basin. The lack of clastic dilution favors sites along the structural high, which is in a position sedimentologically analogous to that of the shelf break off rifted continental margins.

Accumulation, on a large scale, of organic-rich sapropelic sediments within a forearc basin is dependent on either (1) persistent silled conditions leading to euxinic basinal deposits, or (2) abundant slope facies deposited within the oceanic oxygen-minimum zone. Uplift of the structural high to form ridged forearcs (Fig. 6) can lead, when coupled with suitable transverse structures, to restricted basins where organic-rich muds can accumulate on anoxic bottoms. Either transverse or longitudinal progradation of deltaic complexes across or along the length of a deep forearc trough can allow a large aggregate volume of organic-rich slope facies to accumulate. In effect, these shaly strata are prodelta slope deposits that form the center of a sedimentary sandwich, where they are caught between sandier strata of underlying turbidite facies and of overlying deltaic facies. Coals may be present in the deltaic and fluvial-plain deposits that form the upper beds within some forearc basins.

In the following sections, we outline briefly the evolution of forearc regions of northern California and southern Alaska where the important forearc basins of the Sacramento Valley and Cook Inlet are located. These two forearc regions are selected from among the many in the world because of their domestic locations, the relatively high quality of available stratigraphic data, and the fact that both include important hydrocarbon provinces. We highlight them here as illustrative examples of forearcs with a long depositional history. They are not representative, however, of the extensive forearc regions to which there has been relatively little sediment delivered (e.g., central-western South America and western Pacific island arcs, such as the Marianas). Description of these latter regions must await the acquisition of detailed geophysical and subsurface data from offshore areas, because sufficient accretion has not occurred to cause surface exposures recording their evolution.

Northern California

The present terrestrial Great Valley of California is the vestige of a large marine forearc basin that developed during the late Mesozoic and early Cenozoic. In the late Cenozoic, subduction along the continental margin was replaced by transform shear along the San Andreas and related fault systems (Atwater, 1970). Prior to the establishment of the present transform regime, plate consumption dominated the tectonics of the continental margin (Hamilton, 1969), which was thus the site of an arc-trench system (Dickinson, 1970).

In the Cretaceous, when most of the Great Valley sequence was deposited, the tectonic elements of California resembled those of the present Sunda arc and Java Trench in both kind and dimensions (Fig. 9). The plutons of the present Sierra Nevada represent the roots of the old arc. The Great Valley sequence on the west accumulated in a forearc basin that was originally of residual type but evolved into composite type. Details of the sedimentary petrology of sandstones at different horizons within the Great Valley sequence correlate well with petrologic features belonging to igneous rocks of comparable ages in the Sierra Nevada where the intrusive sequence is known from independent radiometric studies (Dickinson and Rich, 1972; Ingersoll, 1978a). The thrust contact between the Great Valley sequence and the Franciscan Complex in the northern Coast Ranges farther west separates forearc basin sediments from deformed rocks of the old subduction zone (Ernst, 1970). Paleontologic and radiometric dates for Franciscan rocks, which represent the old subduction complex, range from Late Jurassic to Paleogene (Suppe and Armstrong, 1972; Blake and Jones, 1974; Evitt and Pierce, 1975). The same span of time saw the deposition of the Great Valley sequence and overlying Paleogene marine strata. Emplacement of plutons in the Sierra Nevada magmatic arc, deposition of the Great Valley sequence in the forearc basin, and both deposition and deformation of the Franciscan subduction complex thus were all broadly coeval, as required by the arc-trench interpretation.

Figure 10 is a generalized geologic map of that segment of the forearc region including the Sacramento Valley. In the subsurface, the Sacramento basin is delimited sharply on the south by the Stockton fault along the flank of the Stockton arch, a major transverse structure in the basement. On the north, the basin fill thins gradually against the southern flank of the Klamath Mountains terrane, which occupies another major transverse salient of basement rocks. The main exposures of the Great Valley sequence form a long homoclinal belt dipping eastward along the western side of the Sacramento Valley. Outcrops are present also farther west as klippen lying structurally on the Franciscan Complex, and subcrops are present beneath much of the Sacramento Valley. Only local exposures are preserved lying unconformably on granitic and metamorphic rocks of the Sierra Nevada foothills along the eastern side of the Sacramento Valley. Figure 10 also shows the eastern limits of mid-Cretaceous and Paleogene rocks within the Franciscan Complex. Lines marking these limits reflect in a general way the progressive westward growth of the subduction complex, as successive accretionary belts, during the time span of forearc deposition.

Figure 11, a series of idealized tectonic profiles drawn transverse to the Sacramento Valley, depicts the evolution of the Sacramento forearc basin as inferred here (Ingersoll, 1978b). The Late Jurassic time frame for the oldest diagram reflects the oldest known radiometric dates for the recrystallization of Franciscan blueschists, the paleontologic age of the oldest known fossiliferous strata within the Franciscan Complex, and the age of the ophiolite sequence that concordantly underlies the thicker parts of the Great Valley sequence (Lanphere, 1971). The last diagram, for the Neogene, is included mainly to show that the evolution depicted does lead successfully to the presently observed configuration of the chief tectonic elements. The diagram for the Paleogene, however, represents the last stage of forearc evolution prior to uplift of most of the forearc region.

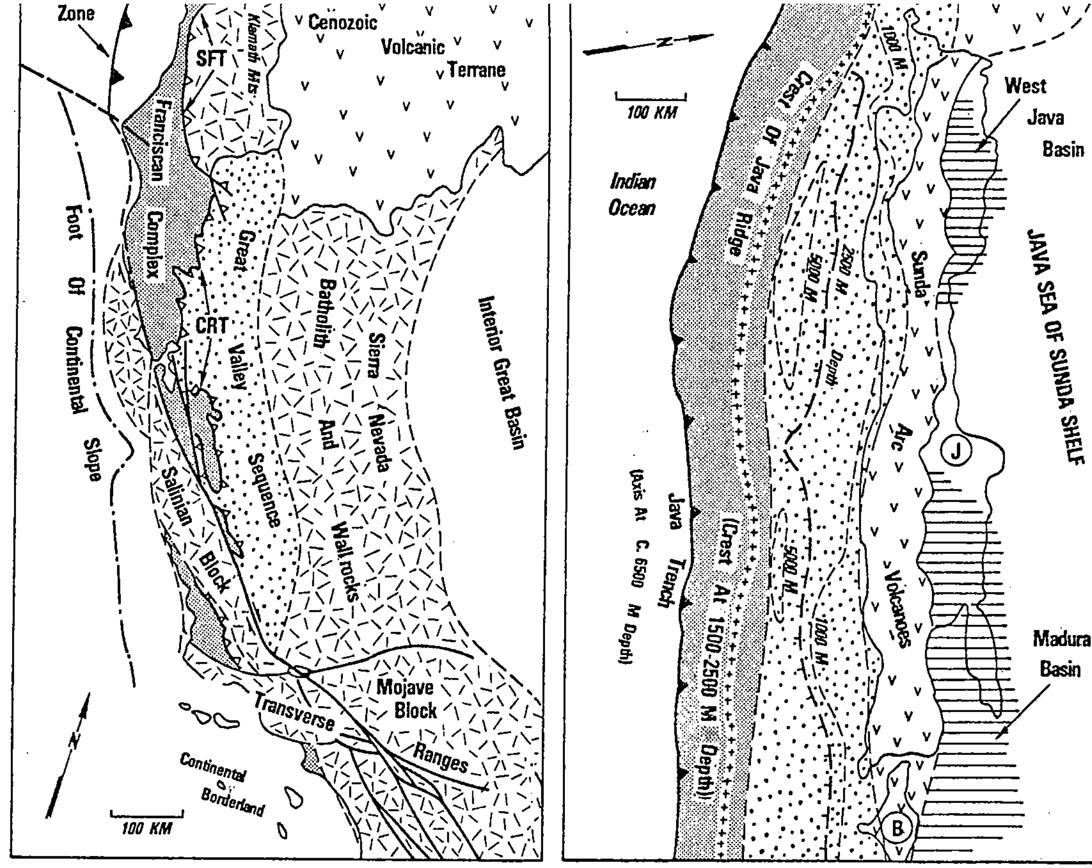

FIG. 9—Comparison (at same scale) of forearc regions during late Mesozoic in California (left, after King, 1969) and during late Cenozoic off Java (right, after Hamilton, 1974). Wide-spaced dots denote fill of forearc basins (shown on left as now preserved following Cenozoic erosion); close-spaced dots denote coeval subduction complexes (shown on right as inferred from seismic profiling records). Other patterns indicate plutonic-metamorphic roots and volcanic cover of arc massif. On left, *SFT* is South Fork Mountain thrust between Klamath Mountains terrane (arc roots) and Franciscan Complex (subduction complex), and *CRT* is Coast Range thrust between Great Valley sequence (forearc basin) and Franciscan Complex; heavy lines are San Andreas and related Neogene faults that offset parts of the arc massif and adjacent tectonic elements. On right, hachured line is 2,500-m bathymetric contour on arc flank of present forearc trough. Dashed lines (1,000 m and 5,000 m) are isopachs of sediment fill in forearc basin; crest of Java Ridge is trough-flank ridge atop uplifted subduction complex, and arc volcanoes form backbone of Java, *J*, and Bali, *B*. Continental crust underlies backarc area in both systems, but extensional deformation has formed terrestrial graben-horst system in Great Basin on left and marine basins along coast between Java and Sunda shelf on right.

In the Late Jurassic terraced forearc (Fig. 6), slope facies (Ingersoll, 1977) prograded westward into a residual forearc basin trapped behind the structural dam of an incipient subduction complex (Fig. 11A). During the Late Jurassic, subduction first began on the site of the northern Coast Ranges following the accretion of an island arc to the continental margin along a suture belt that marks the site of a prior subduction zone in the present Sierra Nevada foothills (Schweickert and Cowan, 1975). The residual nature of the incipient forearc basin is indicated clearly by the fact that the Upper Jurassic strata of the Great Valley sequence everywhere overlie an ophiolite sequence interpreted as oceanic crust (Bailey et al, 1970).

The terraced forearc is shown in a more

evolved state for the Early Cretaceous (Fig. 11B). A structural and bathymetric high at the subduction complex is inferred indirectly. Paleocurrents in Upper Jurassic and Lower Cretaceous turbidites of the main homoclinal outcrop belt (Ojakangas, 1968) and in klippen near Clear Lake (Swe and Dickinson, 1970) in the northern Coast Ranges (see Fig. 10) are consistently southerly, or longitudinal with respect to the overall trend of the arc-trench system (see Fig. 9). As these turbidites are largely basin-plain deposits with minor outer fan components (Ingersoll et al, 1977), they evidently accumulated along the floor of a deep marine trough. The arc massif of the Sierra Nevada block clearly formed the eastern flank of the trough, although no marginal facies older than Late Cretaceous have been penetrated in the subsurface by drilling. An uplifted subduction complex would seem to be the only feature that could have formed the western flank of such a trough. Along the southern margin of the Klamath Mountains at the northern end of the trough, Lower Cretaceous strata including slope and shelf facies lap unconformably across only slightly older plutonic rocks of the arc massif. From this relation, we infer that the part of the forearc basin in which thick strata piled up in deep water was confined to the region farther south where trapped oceanic crust (Fig. 5) underlies the forearc region.

The supposedly discrete structural event shown in Figure 11C is the most problematic aspect of our diagrams. We use it here as an artifice to account for roughly mid-Cretaceous—or at least intra-Cretaceous—telescoping across the foreign region. We ascribe two features to this structural telescoping. Most important is the Coast Range thrust (Bailey et al, 1970) along which the Franciscan Complex was apparently underthrust at least 25 to 50 km beneath the ophiolite at the base of the Great Valley sequence (see Fig. 10). This displacement, representing crustal contraction along the trench flank of the forearc basin, occupied an unknown span of Cretaceous time, and may even have begun earlier and continued later. We assign it here mainly to some mid-Cretaceous interval from evidence suggesting that lesser but important structural telescoping along the arc flank of the forearc basin took place in mid-Cretaceous time.

Roughly mid-Cretaceous faulting between the forearc basin and the arc massif is an attractive hypothesis to account for discrepant thicknesses of Cretaceous strata on the two flanks of the Sacramento Valley (Brown and Rich, 1967). Lateral equivalents of Upper Cretaceous strata exposed in the homoclinal belt along the west side of the valley are present in the subsurface beneath the eastern part of the valley, but ... Cretaceous beds are not (e.g., Fig. 11F). The inference that a buried down-to-basin fault zone is present in the subsurface is seemingly required to reconcile the known geometric relations between outcrop and subcrop sections. We suspect reverse faulting rather than normal faulting from the geometry of a set of splay faults active prior to the Late Cretaceous (Jones et al, 1969) near the northern end of the valley (see Fig. 10).

The splay faults diverge from the complex juncture between the Coast Range thrust, which carries Franciscan Complex beneath the Great Valley ophiolite, and the South Fork Mountain thrust, which carries Franciscan Complex beneath rocks of the arc massif in the Klamath Mountains (see Fig. 10). These faults now have steep dips suggestive of strike slip (Jones and Irwin, 1971). However, the manner in which the faults trend obliquely across the strike of steeply dipping beds in the Great Valley sequence (Jones and Bailey, 1973) indicates that they may originally have had gentle dips, suggestive of thrusting, when those beds were subhorizontal (Ingersoll et al, 1977). Continuation of these faults toward the southeast might link them with the down-to-basin faults inferred for the subsurface of the Sacramento basin proper. Such a linked system of thrust and reverse faults along the arc flank of the forearc basin could then accommodate some degree of mid-Cretaceous crustal contraction along the tectonic join between the residual forearc basin and the arc massif.

A widespread mid-Cretaceous hiatus and presumed unconformity in the southern Coast Ranges (Bailey et al, 1964) indirectly strengthens the case for regional mid-Cretaceous tectonism, but does not bear directly on relations shown farther north by Figures 10 and 11. Our rather tenuous argument that the mid-Cretaceous was a time of especially marked structural telescoping across the forearc region may well not withstand the weight of either additional data or other chains of logic that might be pursued. As an alternative, our mid-Cretaceous diagram (Fig. 11C) could be deleted, and the structural telescoping it involves could be apportioned in some measure to the Early Cretaceous (Fig. 11B) and the Late Cretaceous (Fig. 11D). The age relations of the rocks involved across the key structures would then dictate that faulting between arc massif and forearc basin was mainly Early Cretaceous, but that faulting between subduction complex and forearc basin was mainly Late Cretaceous.

During the Late Cretaceous (Fig. 11D), the residual forearc basin expanded into a composite forearc basin as strata were spread unconform-

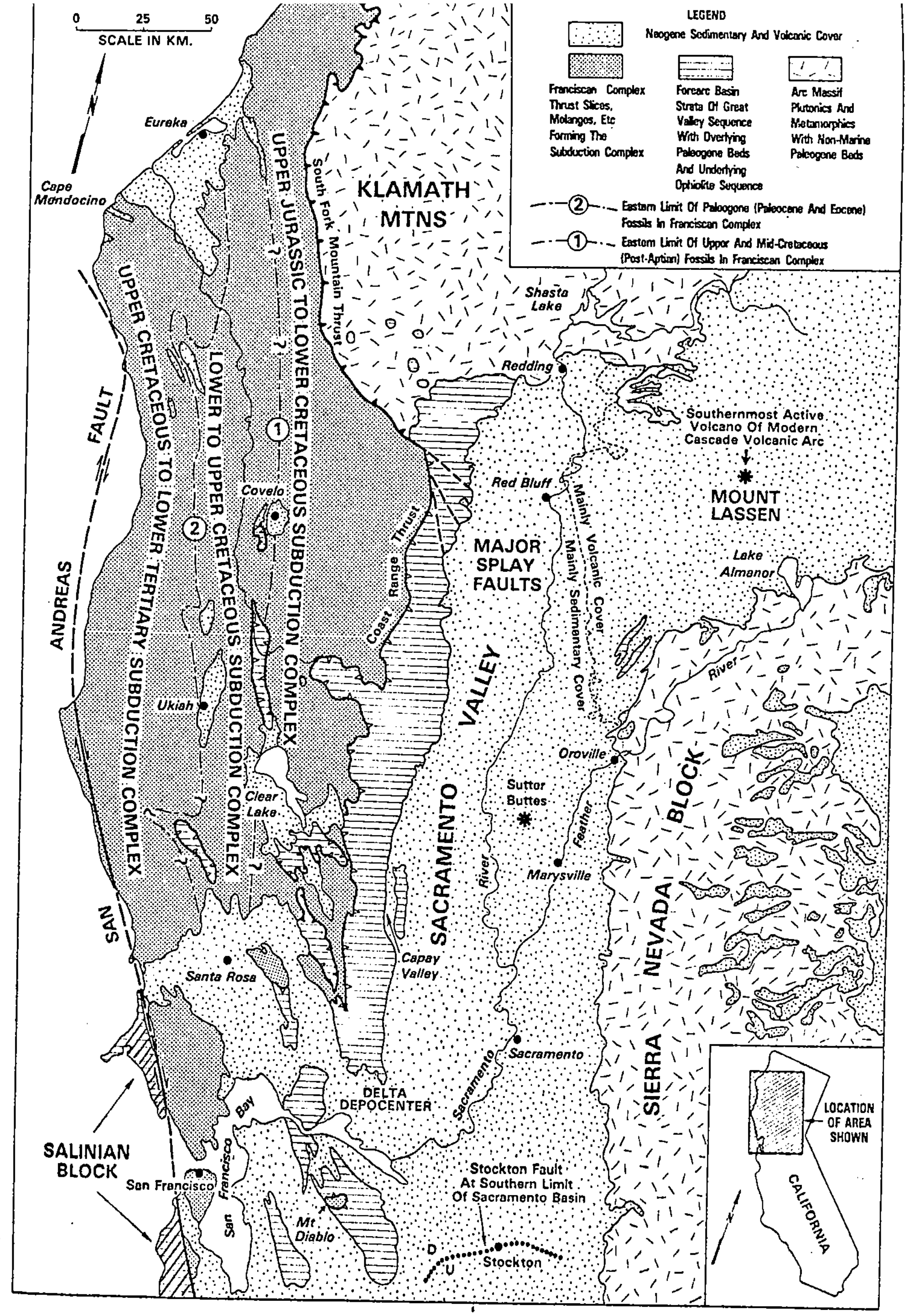

FIG. 10—Generalized geologic map of part of northern California showing structural relations of strata deposited in late Mesozoic and early Cenozoic forearc basin of Sacramento Valley and adjacent uplands of northern Coast Ranges on west. Upper Jurassic to Upper Cretaceous Great Valley sequence (see legend and Fig. 9) lies on ophiolite sequence (see text) above Coast Range thrust, but elsewhere lies unconformably on rocks of arc massif.

ably by depositional onlap across the eroded flank of the arc massif. However, on the other side of the forearc basin, all known contacts with the subduction complex are shear zones. Presumably, at least local deformation along the structural high at the threshold of the forearc basin continually disrupted any depositional contacts that may have formed at times by depositional onlap. Upper Cretaceous strata are dominantly turbidites deposited in subsea-fan systems that were built into the forearc basin by both transverse and longitudinal paleocurrents (Ingersoll, 1976). Subordinate basin-plain and slope facies are also present; the latter are especially widespread near the head of the forearc trough fringing the Klamath Mountains. Toward the end of the Cretaceous, major deltaic systems with aggregate stratigraphic thicknesses on the order of 1 km built shallow marine platforms partly across the previously deep forearc trough from sources in the Sierra Nevada (Drummond et al, 1976). Strandline and associated sandy facies of the delta platforms, and sandy turbidites deposited near the toes of the prodelta slopes, form the major gas reservoirs of the Sacramento basin. Uppermost Cretaceous shelf deposits that locally capped the structural high in the northern Coast Ranges provide evidence for coordinate shoaling of the other flank of the forearc basin by tectonic uplift of the subduction complex. During the Late Cretaceous, therefore, the forearc region evolved from a terraced forearc, with a deep forearc trough, to a shelved forearc (Fig. 6).

A thin veneer of Paleogene shelf deposits was spread across most of the forearc shelf basin (Fig. 11E). Only in the delta depocenter (Zieglar and Spotts, 1976) at the southern end of the Sacramento basin (Fig. 10) did a residual pocket of deep water persist. Several large submarine canyons that scored the Paleogene shelf on the north and east locally funneled turbidites toward the delta depocenter. Widespread emergence of the structural high along the west flank of the forearc basin is suggested by erosional unconformities between Cretaceous and Paleogene strata within the klippen of the northern Coast Ranges (Fig. 10). At least local emergence is confirmed by the presence of Franciscan detritus in the Paleogene beds exposed there, and also in the Paleogene components of the subduction complex farther west. Markedly angular unconformities below Miocene strata of the northern Coast Ranges point to wholesale uplift of the structural high to form a ridged forearc (Fig. 6) by mid-Cenozoic time. Only nonmarine Neogene strata are present in the Sacramento basin. The evolution of the ridged forearc cannot be reconstructed in detail because of later deformation associated with the San Andreas transform system, which sliced off some indeterminate part of the forearc region beginning in the Pliocene (Fig. 11F).

At present (Fig. 11F), the whole western part of the forearc basin has been uplifted, together with the subduction complex, and largely eroded away. The remnant of the basin is preserved as a huge megasyncline whose axis lies beneath the present Great Valley. Original depositional relations of the eastern part of the forearc basin along the flank of the arc massif are preserved almost intact, but the western flank of the megasyncline is wholly structural.

Southern Alaska

Critical information bearing on the tectonic evolution of the forearc region in southern Alaska is much less complete than for northern California. We include a generalized tectonic map and profile for southern Alaska (Fig. 12) mainly because they illustrate a mature ridged forearc in an advanced stage not represented in California. Massive tectonic accretion in Alaska has formed perhaps the widest forearc region in the world. Accretionary belts of upper Mesozoic, Paleogene, and Neogene subduction complex are each 75 to 100 km wide (Fig. 12).

The subduction complex is so wide that two discrete Cenozoic forearc basin areas are present, one in Cook Inlet and the other beneath the offshore shelf. The two are separated by the uplifted Kodiak-Kenai tectonic ridge. The basins beneath the shelf are accretionary forearc basins that developed during the Cenozoic and may contain mainly shelf and slope deposits. Beneath Cook Inlet, however, Cenozoic strata are underlain by the thick Mesozoic section of the so-called Matanuska geosyncline (Kirschner and Lyon, 1973). We infer that the Cook Inlet basin is a residual forearc basin, although the ophiolite base shown in Figure 12 is entirely speculative. Although part of the Mesozoic sequence was deposited in deep water, high sedimentation rates during Cenozoic subsidence maintained generally shallow-water to nonmarine levels.

FOREARC PETROLEUM POTENTIAL

Petroleum has been encountered in the residual and/or composite basins (Fig. 6) of broad-ridged forearcs (Cook Inlet and Sacramento Valley, Fig. 13); a narrow-ridged forearc (Mentawai Trough); a terrestrial-uplands, ridged forearc (Talara, Peru); and a shelved forearc (Honshu shelf). Of these, the total recoverable reserves are greatest in the Talara and Cook Inlet areas (each having 1 billion bbl or more of estimated recoverable oil

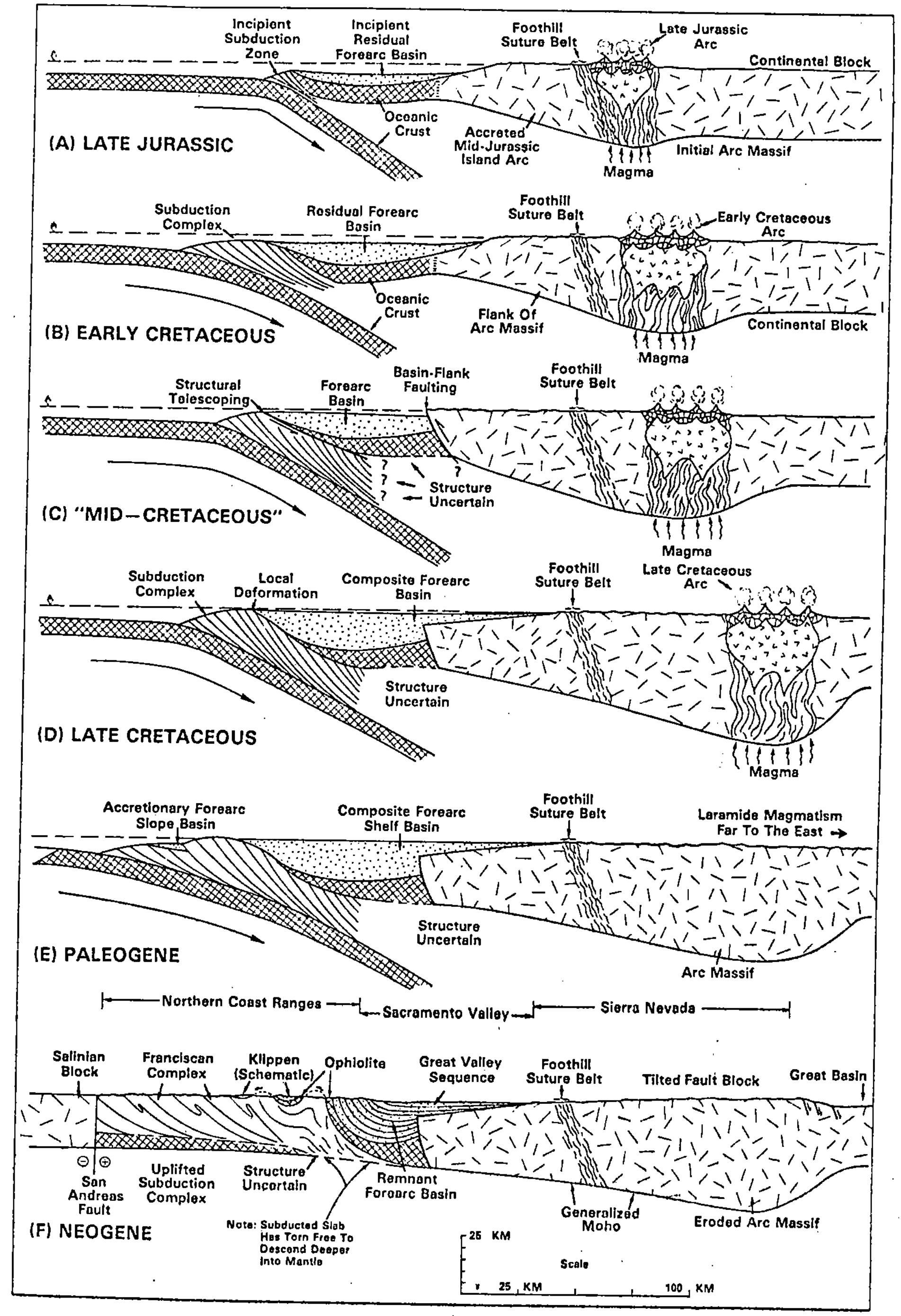

FIG. 11—Schematic diagrams (no vertical exaggeration) illustrate tectonic evolution of Great Valley forearc basin and related features in northern California. Each diagram depicts conditions at end of time span indicated. Crustal profile generalized from Eaton (1966) and Cady (1975). Stratigraphic thicknesses after Swe and Dickinson (1970), Dickinson and Rich (1972), and Drummond et al (1976). Shifting magmatism after Evernden and Kistler (1970). Structural telescoping is depicted here arbitrarily as discrete mid-Cretaceous event, but doubtless took place over some finite span of Cretaceous time.

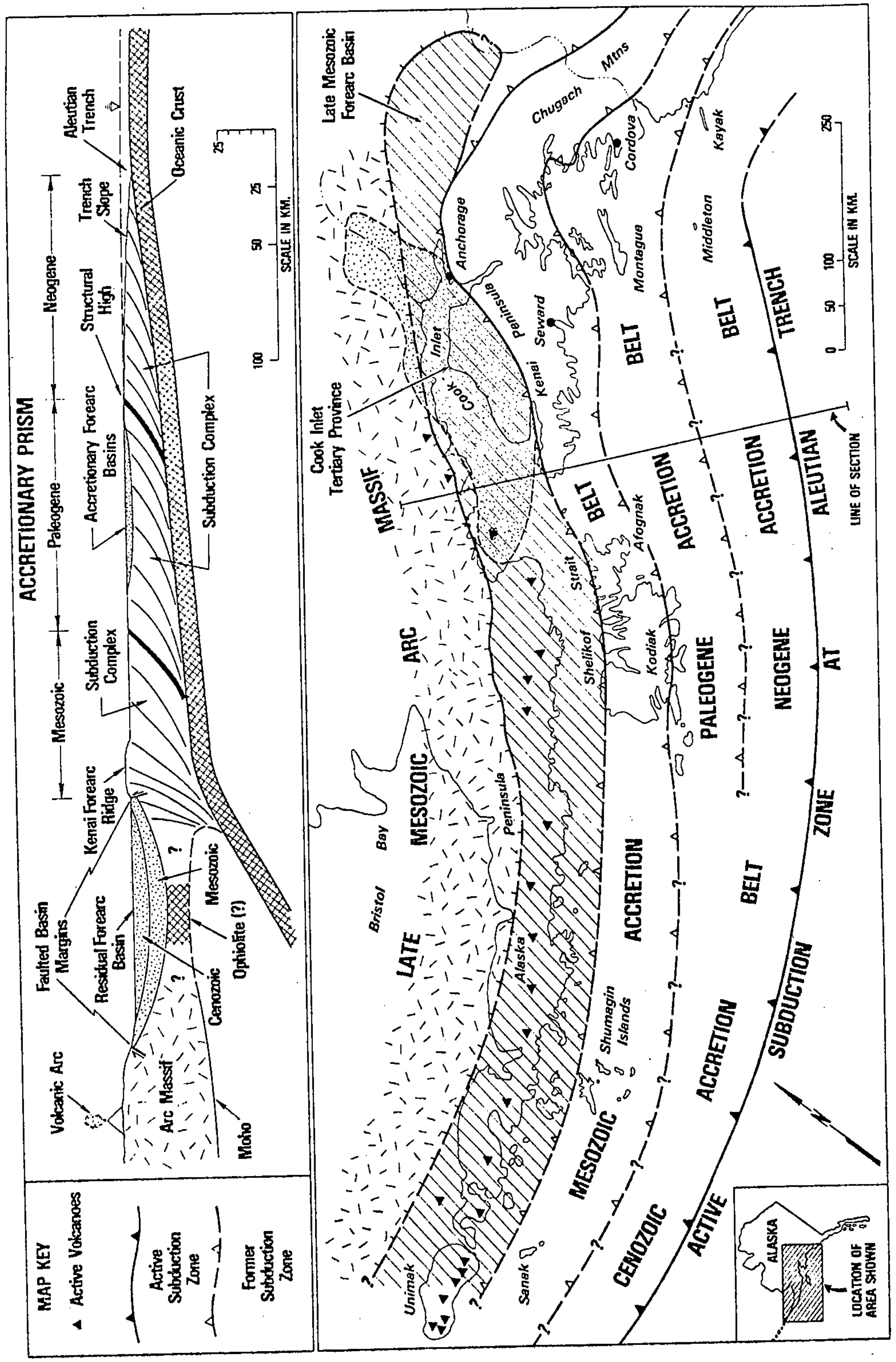

FIG. 12—Generalized cross section (top) and tectonic sketch map (bottom) of forearc region in southwestern Alaska. Map after Beikman (1974a, 1974b, 1975) and Moore and Connelly (1977). Section after von Huene and Shor (1969), Plafker (1972), Kirschner and Lyon (1973), and Karig et al (1976).

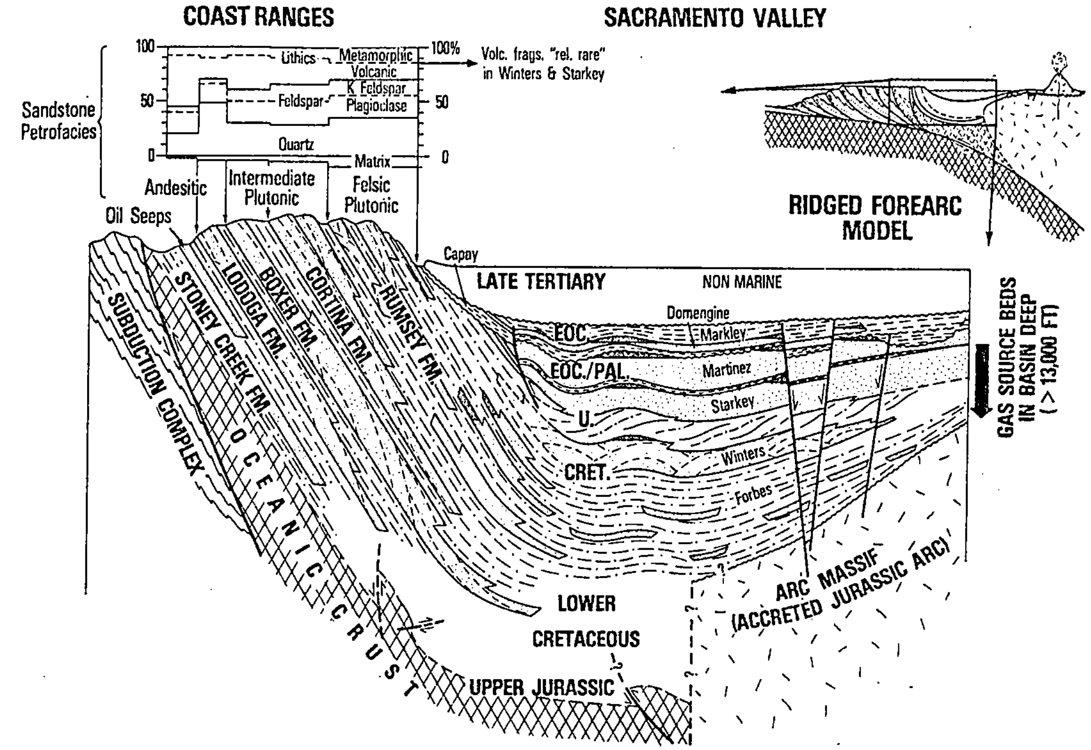

FIG. 13—Sacramento Valley petroleum occurrence summary diagram. Structural and stratigraphic positions of gas occurrence are from T. P. Harding (personal commun., 1977). Outcrop sandstone petrofacies are from Ingersoll et al (1977). Volcanic content of Winters and Starkey and progradation between them are from Drummond et al (1976). Interpretation of Upper Cretaceous source beds is from Zieglar and Spotts (1976). Stratigraphic and structural relations are not to scale. Thrusting of oceanic crust on subduction complex (left side of diagram) occurred after deposition in Cretaceous forearc basin.

and gas equivalent) and are noncommercial in the Mentawai and Honshu basins. The reservoir facies of the occurrences are nonmarine, nearshore, fluvial-deltaic, turbidite, and reef. Traps are fault blocks, both reverse and normal; thrust-faulted folds; and stratigraphic. Gas is present in all, whereas oil is not.

The subduction complex does not appear to be a good exploration target. It is commonly composed of overcompacted trench and abyssal-plain sediments that have been so highly structured that traps are small and difficult to define seismically. Faulting and early expulsion of interstitial fluids during compaction in the lower parts of the trench inner slope reduce the possibility of entrapment. Reduction of porosity by diagenesis under tectonic loads and disruption of strata by widespread shear deformation further reduce the likelihood that attractive reservoirs could be preserved during the accretionary evolution of the subduction complex. Moreover, potent source beds are unlikely to be important components of subduction complexes in which turbidites of abyssal plains or trench fills are dominant, but may be present where shaly slope deposits form a significant fraction of the total bulk. The low heat flux that is characteristic in the vicinity of subduction zones, which are heat sinks in the global scheme of plate motions, means that any source beds present are apt to be thermally immature.

The principal obstacles to a high petroleum potential in forearc basins are (1) the common occurrence of immature, clay-plugged sands owing to rapid deep-water deposition and diagenesis of volcanic components, (2) discontinuity of reservoirs owing to tectonic instability and shifting foci of deep-water fans during deposition, and (3) the youth of many basins, which reduces the possibility of mature source beds. On the positive side, however, the wide variability of the types of reservoirs and traps in which forearc hydrocarbons have been found indicates the broad range of geology possible and suggests that new combinations of conditions favorable for petroleum occurrence in forearc basins are yet to be found. In general, petroleum potential should be enhanced for forearc basins in which some of the following conditions are met.

1. The clastic sediment delivered to the basin is derived in large part from uplifted batholiths in the arc massif or from large rivers draining nearby continental blocks, rather than from the volcanic chain of the arc. The resulting quartzose feldspathic sands may then form clastic reservoirs that can maintain the requisite porosity despite diagenetic influences.

2. The sediment fill of the basin includes marginal carbonate buildups within which desirable carbonate reservoirs can develop.

3. The clastic facies within the basin include shoreline, deltaic, or subsea-fan associations in which stratigraphic traps can be anticipated.

4. The tectonic structuring within the basin involves large-scale folding and local faulting to develop adequate structural traps.

5. The basin is of residual or composite type having a broad, thick, and old prism of sediment in whose lower zones the threshold conditions for maturation can be reached.

6. The strata within the basin include semi-euxinic slope or silled-basin facies which might harbor attractive source beds.

The favorable conditions listed here amount almost to platitudes for petroleum exploration. Mention of them in the context of forearc basins makes the point, however, that successful exploration in such basins will be based on the usual sound principles that obtain results elsewhere. It will prove difficult, in our opinion, to find the proper combination of circumstances in many forearc basins. However, we do not doubt that they will be found in certain places. To a considerable degree, realization of the full hydrocarbon potential of the Circum-Pacific rim depends on development of the ability to explore effectively in forearc basins.

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