PAGEOPH, Vol. 128, Nos. 3/4 (1988)
0033-4553/88/040455-46$1.50 + 0.20/0

Subduction-Channel Model of Prism Accretion, Melange Formation, Sediment Subduction, and Subduction Erosion at Convergent Plate Margins: 1. Background and Description

MARK CLOOS[1] and RONALD L. SHREVE[2]

Abstract—Many geological and geophysical investigations, particularly the Deep Sea Drilling Project, have shown that convergent plate margins are highly diverse features. For example, at some sites of subduction, such as the Lesser Antilles, the bedded sediment atop the incoming oceanic plate is extensively offscraped, whereas at others, such as Mariana, not only is the incoming sediment completely subducted beneath crystalline rock but portions of the overriding plate are undergoing subduction erosion. Earthquakes indicate wide variations in stress distribution within and between sites of plate convergence. Many ancient accretionary complexes include tracts of intensely-deformed subduction melange that contain blocks of mafic greenstones. Some contain bodies of thoroughly recrystallized blueschist that were uplifted from depths of 20 to 30 km. A comprehensive model for convergent plate margins must explain these and numerous other observations. Although the still widely cited imbricate-thrust model for prism accretion qualitatively explains some observations at subduction zones, it does not account for many others, such as deep sediment subduction and subduction erosion.

The subduction-channel model postulates essentially the same basic mechanics for all convergent plate margins that have attained a quasi-steady state (typically reached after about 20 Ma of subduction at speeds of 10 to 20 km Ma^{-1}). It assumes that the subducting sediment deforms approximately as a viscous material once it is dragged into a relatively thin shear zone, or subduction channel, between the downgoing plate and the overriding one. It predicts the overall movement patterns of the sediment deforming within the channel and near its inlet, accounts for most of the observed features at convergent plate margins, and quantifies the processes of sediment subduction, offscraping, and underplating, and the formation of subduction melange. The predicted variations in tectonic behavior depend upon such site-specific variables as the speed of subduction, the supply of sediment, the geometry of the descending plate, and the topography and structure of the overriding block.

Key words: Subduction zone, subduction accretion, subduction erosion, sediment subduction, accretionary prism, melange, offscraping, underplating.

Introduction

The advent of the theory of plate tectonics in the late 1960s provided a new framework that unifies and explains a remarkably diverse array of geological and

---

[1] Department of Geological Sciences and Institute for Geophysics, University of Texas at Austin, Austin, Texas 78713-7909, U.S.A.

[2] Department of Earth and Space Sciences and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, California 90024-1576, U.S.A.

geophysical observations. It showed that oceanic trenches, volcanic arcs, and Benioff seismic zones mark sites of plate convergence, where oceanic lithosphere is subducted at speeds of several tens of kilometers per million years (several centimeters per year). Many geological and geophysical studies, particularly the Deep Sea Drilling Project (DSDP) investigations by the Glomar Challenger, have shown that at some sites of subduction, such as the Lesser Antilles (Figure 1), the sediment carried on the subducting plate is largely accreted onto and under the leading edge of the overriding plate at depths of only a few kilometers, whereas at others, such as Mariana, not only is the incoming sediment completely subducted, apparently to depths as great as 100 km, but also some of the overriding plate is being detached and deeply subducted as well. Thus, subduction erosion as well as accretion can occur. Clearly, convergent plate margins are highly diverse features, whose particular characteristics must reflect differences in such factors as the speed of subduction, the dip of the descending plate, the structure of the overriding block, and the nature and amount of incoming sediment.

Most published reports and quantitative models to date have addressed only the accretion or deformation of prisms. We have developed a model (SHREVE and CLOOS, 1986) that quantitatively describes not only the process of prism accretion

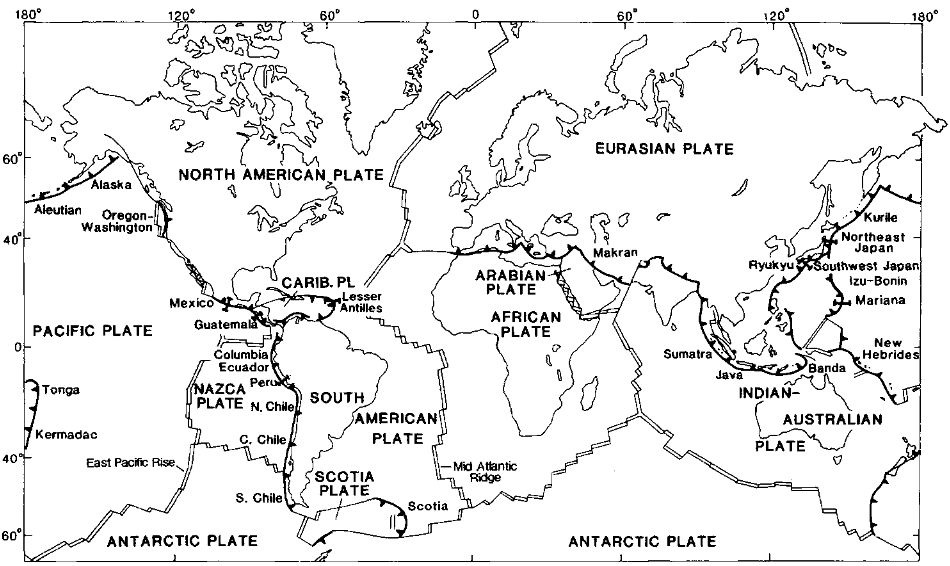

Figure 1
Locations of subduction zones (heavy solid lines, barbs in direction of subduction) where plate convergence is faster than about 20 km Ma^{-1} (2 cm yr^{-1}). Cross bars indicate locations of Deep Sea Drilling Project transects off Mariana (Leg 60), southwest Japan (Legs 31 and 87), northeast Japan (Legs 56, 57, and 87), Mexico (Leg 66), Guatemala (Legs 67 and 84), Peru (Leg 112), and the Lesser Antilles (Legs 78A and 110).

but also of sediment subduction and melange formation and in addition identifies the likely sites of subduction erosion. Basically, we envision subduction-driven deformation as largely concentrated in what we term a *subduction channel*, which is a relatively thin layer of relatively rapidly shearing, poorly consolidated sediment (bulk density 2.2 Mg m^{-3}) dragged by the descending plate beneath the overriding one and, where one is present, its accretionary prism. The sediment in the channel is analogous to the lubricant in a bearing. The principal differences, which make possible the varied behaviors of active margins, are that the overriding block is neither strictly fixed nor rigid but can deform slowly, the buoyancy of the sediment affects the flow, the sediment can underplate onto the hanging wall, and the sediment supply can vary widely relative to the capacity of the system. The model, herein called the *subduction-channel model*, quantitatively explains the whole range of convergent-margin behavior from complete sediment subduction to nearly complete offscraping. It unifies many of the previously proposed models for convergent margins and provides new insights into the interpretation of ancient subduction complexes.

This paper is a nonmathematical companion to our earlier paper (SHREVE and CLOOS, 1986), which derived the subduction-channel model and applied it to representative sites off Mariana, Mexico, the Lesser Antilles, Alaska, and northeast Japan. It is divided into two Parts, which can be viewed as more or less the predecessor and the successor of the earlier paper. In this Part we summarize the basic geological and geophysical observations that all models of subduction zones must explain. Then we review the various published models, paying particular attention to the widely-cited imbricate-thrust model for accretion developed in the early 1970s (SEELY *et al.*, 1974; KARIG, 1974a; KARIG and SHARMAN, 1975). Finally, we describe the subduction-channel model in qualitative terms and discuss certain aspects of it that may seem problematical.

In Part 2 (CLOOS and SHREVE, 1988, this volume) we characterize the five possible types of active convergent margin and discuss in detail how the subduction-channel model, and especially margin type, relates to the zone of compression of the incoming sediment at the base of the trench slope, to the dewatering of subducting and accreted sediments, and to subduction accretion, including the ages of included fossils in relation to time of accretion and the potential for preservation of olistostromes. Then we discuss how the model relates to subduction erosion, to tectonics ranging from vertical movements to subduction-zone earthquakes, to allochthonous bodies such as subducted seamounts, melange diapirs, and serpentinite diapirs or blocks, to the generation and synsubduction uplift of blueschists, and finally to the generation of magmas in volcanic arcs.

Our overall objective in these papers is to show how the subduction-channel model comprehensively unifies and explains many of the diverse geological and geophysical observations that have been made at both modern convergent margins and ancient accretionary complexes.

Fundamental Observations

The reviews of KARIG (1983), VON HUENE (1984a, 1986a), JARRARD (1986), and MOORE, J. C., and SILVER (1987) summarize many of the important characteristics of convergent plate margins. In this section we emphasize the wide range of geological and geophysical observations that theoretical models of such margins must fit and, if possible, explain.

Terminology

A complex terminology has developed for describing features and processes at convergent plate margins. Unfortunately, some authors have given somewhat different meanings to the same terms. Hence, it is necessary to define terms as we use them here.

An *accretionary prism* consists largely of variously deformed sedimentary materials that accumulated at the leading edge of the overriding plate during convergence (Figure 2). The sizes of modern accretionary prisms vary greatly, from essentially nonexistent off Mariana and Guatemala to approximately 200 km wide off Makran, southwest Japan, and the Lesser Antilles. The sediments in prisms are considerably more compacted (bulk densities typically 2.4 to 2.5 Mg m^{-3}) and lithified than are the incoming sediments at the trench axis. Volcanic rocks,

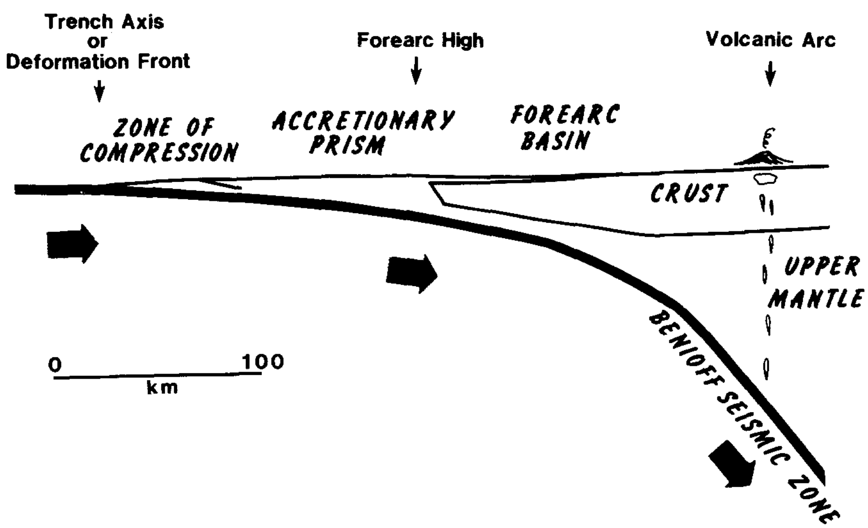

Figure 2
Schematic diagram of major morphotectonic features at an idealized accretionary convergent plate margin. Backstop separating the accretionary prism from the crust and upper mantle may dip away from the prism as shown, or dip toward it, or have a more complex configuration.

thoroughly recrystallized metamorphic ones, and other strong rocks may be present, but with only rare exceptions are volumetrically minor.

The *backstop* is the boundary where accreted sediments abut either the crystalline leading edge of the overriding plate (BRANDON, 1986) or the more highly consolidated remnants of previous episodes of accretion (SHREVE and CLOOS, 1986). Accreted sediments can be emplaced either above it, if it dips trenchward, or below it, if it dips arcward, or both above and below it. Its importance is that it is a lithological discontinuity across which profound mechanical and permeability contrasts can exist.

The *zone of compression*, which is present at all margins, is the region near the base of the trench slope where the incoming sediment pile is shortened and thickened by the bulldozer-like action of the overriding block. It comprises the region between the toe of the overriding block and the trench axis or, where the axis is ill-defined, the *deformation front* at the base of the trench slope. It is not part of the accretionary prism, and, in fact, must be present even in the absence of an accretionary prism, as at Mariana and Guatemala. Within it the incoming sediment pile longitudinally shortens and vertically thickens by thrusting, folding, and penetrative strain. Materials deformed in it may or may not eventually be added to an accretionary prism. The zone of compression commonly is the region where tectonic structures are imaged best in seismic-reflection profiles.

Subduction accretion is the general process of growth of the prism by offscraping at the toe and underplating onto the base (Figure 3). The accreted materials typically are blanketed by the directly-deposited sediments of the trench *slope cover*. *Sediment subduction* is the process by which poorly consolidated sediment is transported beneath the overriding crystalline block and, if one is present, the accretionary prism. Subducted sediment eventually is either underplated, or returned to the inlet and offscraped, or transported to the depths of arc magmagenesis. *Offscraping* is the process of trenchward growth or widening of the prism by accretion at the toe. We use the term in the general manner of SEELY *et al.* (1974) and KARIG and SHARMAN (1975), but we visualize certain details differently. *Underplating* is the process of addition of material to the bottom of the prism, which thickens and uplifts it without additional shortening. We use the term in the general manner of MOORE, J. C., *et al.* (1982a). *Subduction erosion* is the process in which rock is rasped or stoped from the base of the overriding block and, once coupled to the descending plate, dragged to greater depth. It can occur at deeper levels at the same time subduction accretion is occurring at shallower ones, or vice versa.

General Relationships from Seismic Surveys

The topography, structure, and composition of the overriding block varies considerably from one margin to another (see COULBOURN and MOBERLY, 1977;

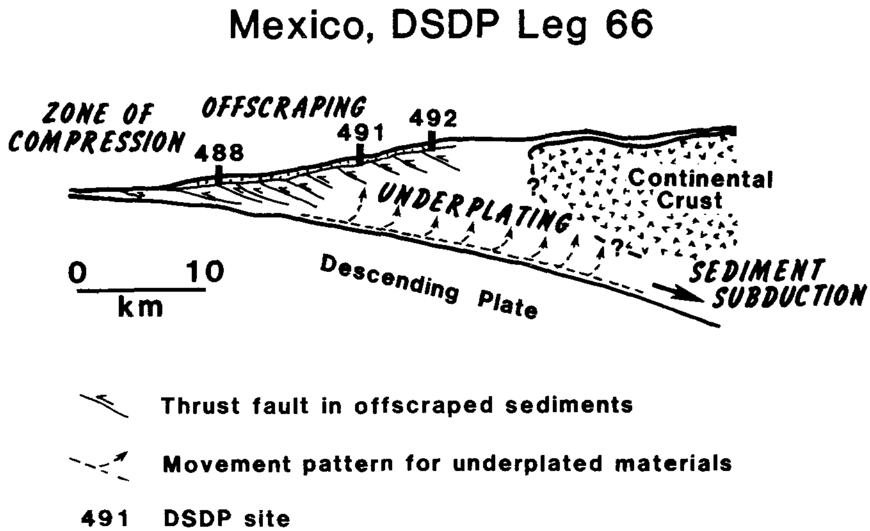

Figure 3
Major tectonic processes at the convergent margin off Mexico inferred from seismic-reflection and deep-sea drilling studies of Leg 66 (after MOORE, J. C., *et al.*, 1982a). The tectonic processes are compression, offscraping, underplating, and sediment subduction. Accreted materials are blanketed by slope cover (stippled pattern), which was cored at DSDP Sites 488, 491, and 492. Depositional ages of the deepest cored sediments are younger toward the trench, indicating trenchward growth of the accretionary prism. Thrust-faulted sediments that are presumed to become offscraped onto the toe of the prism were imaged by seismic-reflection methods. Fossils from the drill cores indicate recent uplift of the slope. This uplift cannot be due to thickening of the prism by longitudinal compression, because the landward-dipping seismic reflectors beneath the slope cover (interpreted by MOORE, J. C., *et al.*, 1982a, to be thrusts) do not progressively steepen up the slope; hence, thickening by underplating is inferred. Mass-balance considerations indicate that a significant fraction of the incoming sediment is subducted to greater depths.

DICKINSON and SEELY, 1979; LUNDBERG, 1983; VON HUENE, 1986b; and cross-sections given by SHREVE and CLOOS, 1986). At some sites, such as Peru-Chile, it consists dominantly of continental crust; at others, such as the Aleutians, it consists of oceanic crust; and at a few, such as Ecuador, the one gives way to the other across or along strike.

Figure 4a shows schematically the major features seen in seismic-reflection profiles across a typical convergent plate margin having a well-developed accretionary prism and a prominent forearc high. Examples are the Lesser Antilles and Sumatra. Although some margins of this type lack a well-defined bathymetric high near the juncture between the accretionary prism and the forearc basin, they generally have at least pronounced trenchward steepenings in slope there. Figure 4b, in contrast, shows a margin lacking an accretionary prism. An example is Mariana. At such margins the basement typically extends with uniform or increasing surface slope from the volcanic arc all the way to the trench bottom.

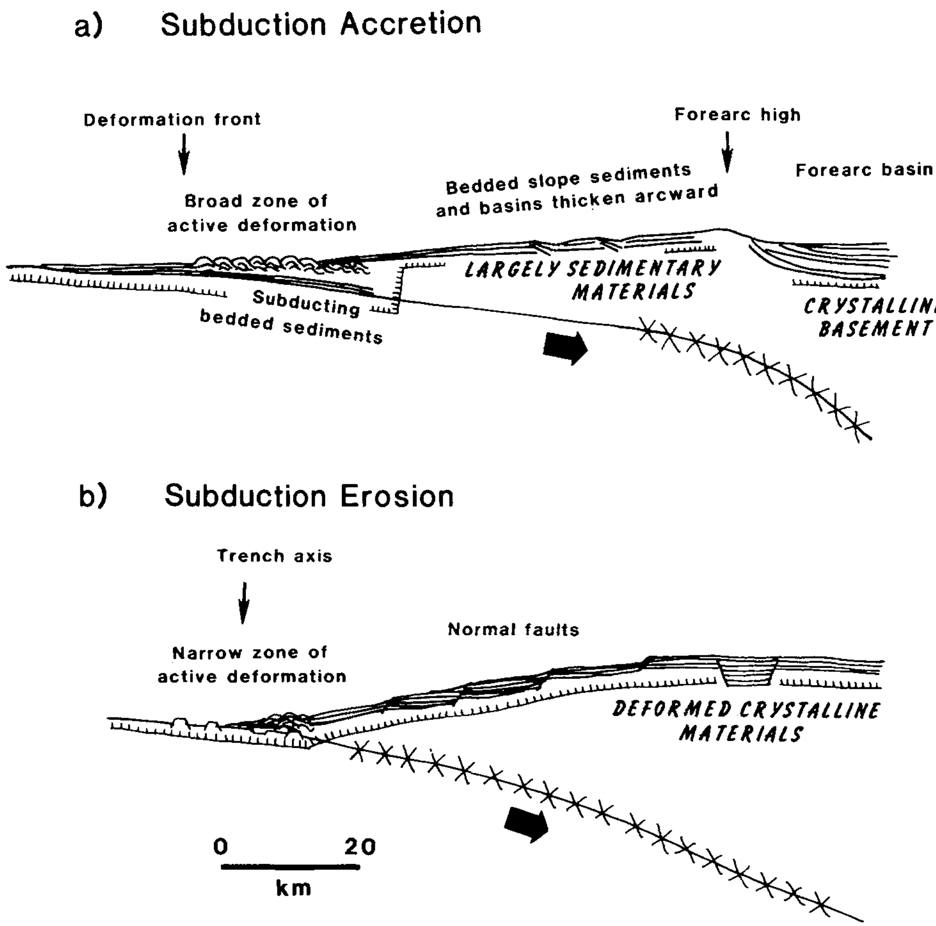

Figure 4
Schematic cross sections illustrating major features imaged by seismic-reflection methods at margins dominated by (a) subduction accretion and (b) subduction erosion. Hachured light line indicates approximate lower boundary of region typically imaged. Pattern of *X*s along line delineating top of downgoing plate indicate region of large thrust-type earthquakes.

A slope cover of sediments generally blankets the accretionary prism (or, where no prism is present, the overriding crust). It generally is thin near the trench axis and thickens upslope, locally to as much as a kilometer or so. In seismic profiles it typically appears well-layered, with generally fairly continuous reflecting horizons that lap onto the forearc high or, in some cases, merge with reflectors of the forearc basin that extend trenchward of the high (SEELY *et al.*, 1974; MOORE, G. F., and KARIG, 1976; KARIG, 1977; SEELY, 1979). In places these horizons may terminate abruptly against the arcward sides of areally-restricted local sequences of reflectors as much as 1 to 2 km thick that are interpreted as slope-basin deposits. In some cases these deposits broaden and thicken upslope (MOORE, G. F., and KARIG, 1976; UNDERWOOD and BACHMAN, 1982); and within them the reflectors dip arcward and steepen with depth. Minor deformation has been detected in the basal portions of some slope basins (STEVENS and MOORE, G. F., 1985). These relations indicate that many accretionary complexes grow oceanward by addition of material near the foot of the trench slope and that directly convergence-driven deformation is mostly concentrated near the base of the slope (KARIG *et al.*, 1980).

Seismic-reflection techniques have not yet adequately imaged the internal structure of accretionary prisms. Beyond a few tens of kilometers arcward of the trench axis, they typically show only complex diffractions beneath the slope cover. Two notable exceptions to this generalization are off Mexico (SHIPLEY *et al.*, 1980) and northeast Japan (NASU *et al.*, 1980), where many landward-dipping reflectors appear far upslope from the trench axis. These reflectors extend from near the base of the slope cover to subbottom depths of 2 to 3 km, and have generally been interpreted as tilted beds or thrust faults. Off Mexico (Figure 3) they dip nearly uniformly, from which MOORE, J. C., *et al.* (1982a) concluded that the prism must have thickened by underplating, rather than by longitudinal compression, which would have progressively rotated the dipping horizons.

Normal faults cut the crystalline basement of the overriding block off Mariana (MROZOWSKI and HAYES, 1980), Ecuador (LONSDALE, 1978; SHEPHERD and MOBERLY, 1981), Peru-Chile (TRAVIS, 1953; KATZ, 1971; COULBOURN and MOBERLY, 1977; MOBERLY *et al.*, 1982; VON HUENE *et al.*, 1985), and Guatemala (AUBOUIN *et al.*, 1984). Thus, periods of significant subhorizontal crustal extension approximately perpendicular to the trench have occurred concurrently with convergence in at least some forearc areas.

Uplift and Subsidence at Active Margins

Short-term vertical movements in forearc areas generally are clearly associated with seismic events, as off Alaska during the great earthquake of 1964 (PLAFKER, 1965). Long-term periods of uplift, subsidence, or both in alternation, also occur. The forearc regions off Japan, Mariana, Mexico, and Guatemala, for example, have undergone lengthy intervals of subsidence (VON HUENE, 1981; MCMILLEN and BACHMAN, 1982; UYEDA, 1982). Inasmuch as Japan and Mexico are presently undergoing uplift, it appears that both subsidence and uplift can occur during a single period of convergence. Unconformities in seismic-reflection profiles across many forearc regions suggest that such movements are probably episodic. Their long-term rates are typically around a few hundred meters per million years (Table 1).

Descending Plate

The descending plate is subducted at speeds ranging from about 10 to as much as 110 km Ma^{-1} (1 to 11 cm yr^{-1}) where currently active plate convergence is

Table 1

Vertical movements at convergent plate margins

Location (References)	Movement
Mariana (DSDP Leg 60; HUSSONG and UYEDA, 1982; MROZOWSKI and HAYES, 1980)	Sink 50–100 m Ma^{-1}; weakly-constrained rise, then sink
Mexico (DSDP Leg 66; MCMILLEN and BACHMAN, 1982)	Sink as much as 950 m Ma^{-1}; rise 130–200 m Ma^{-1}
Lesser Antilles (Barbados Island; SPEED and LARUE, 1982)	Rise 200 m Ma^{-1}
Alaska (DSDP Leg 18; VON HUENE and KULM, 1973)	Rise 800–2000 m Ma^{-1} at foot of trench slope
Japan (DSDP Legs 56 and 57; VON HEUNE, 1981)	Sink 250 m Ma^{-1}; rise 125 m Ma^{-1}
Sumatra (Nias Island; MOORE, G. F., *et al.*, 1980)	Rise 200 m Ma^{-1}
Peru (ODP Leg 112; KULM *et al.*, 1981; VON HUENE *et al.*, 1987)	Sink 275 to 500 m Ma^{-1}
Ecuador (LONSDALE, 1978)	Rise 100–250 m Ma^{-1}
Guatemala (DSDP Leg 67; VON HUENE *et al.*, 1980; SEELY, 1979)	Sink, rate uncertain; rise 200–500 m Ma^{-1} in forearc region
Oregon-Washington (DSDP Leg 18; KULM and FOWLER, 1974)	Rise 100–200 m Ma^{-1} at shelf, 1000–2000 m Ma^{-1} at foot of slope

recognized from deep earthquake patterns that define Benioff seismic zones (MINSTER and JORDAN, 1978). It is capped in virtually all cases by oceanic crust whose upper surface can be traced in some multichannel seismic-reflection profiles from 20 to as much as 50 km arcward of the trench. It is depressed by the weight of the overriding plate (KARIG *et al.*, 1976). At margins with accretionary prisms it typically dips arcward a few degrees near the trench axis and around 10° beneath the forearc high. The Benioff zone indicates that in most cases it dips about 45° beneath the volcanic arc (ISACKS and BARAZANGI, 1977).

At some margins, particularly those lacking significant accretionary prisms, the surface of the incoming oceanic crust is broken by normal faults (HILDE, 1983) with scarp heights that are typically tens to hundreds of meters, but rarely more than a thousand meters. The fault blocks can either step predominantly downward toward the trench, as off Japan (LUDWIG *et al.*, 1966), or form horsts and grabens, as off Peru-Chile (SCHWELLER and KULM, 1978a; COULBOURN, 1981; WARSI *et al.*, 1983).

Seamounts up to several thousand meters high widely dot the ocean floors and locally form chains of islands. Similarly, scarps with relief up to about a thousand meters typically are present where oceanic transform faults offset lithosphere less than a few tens of millions of years old. When these features impinge upon the overriding block during subduction, they can substantially modify trench sedimentation by damming the axial flow of turbidity currents.

Characteristics of Incoming Sediment Pile

The basal layers of the incoming sediment generally consist of pelagic sediments that accumulated on the oceanic crust near spreading ridges far from continental margins. They may be quite calcareous, if deposited above the carbonate compensation depth (the CCD), or quite siliceous, if deposited below it, and distinctively metal-rich, if deposited near active hydrothermal springs. Typically, pelagic sediments deposited away from oceanic spreading ridges are less calcareous and less metalliferous (DAVIES and GORSLINE, 1976). Their biogenic content primarily depends upon the organic productivity, the oxygenation, and the depth of the CCD at the site of deposition (LEGGETT, 1985). Likewise, their content of clay and other fine terrigenous detritus depends upon such factors as prevailing winds and the distance to volcanic arcs, which can provide air-fall tuff, or to glaciated regions, which can supply ice-rafted detritus (GRIFFIN *et al.*, 1968). The thickness of the pelagic part of the incoming sediment pile typically is less than a few hundred meters, but varies greatly, because of nondeposition, ponding, and even erosion by bottom currents, which locally can be significant (see, for example, LONSDALE, 1981).

As the incoming sedimentary pile approaches a continent or volcanic arc, the new sediment added to it normally contains an increasingly terrigenous component. At a few margins this component is of minor importance, usually because the trench is too far from any continent, or the local climate precludes significant delivery of sediment to the ocean, or the forearc area is a site of unusually effective ponding. Examples include Mariana, which is in the mid-Pacific (HUSSONG and UYEDA, 1982), and much of Peru-Chile, which adjoins land of extreme aridity (SCHWELLER *et al.*, 1981). At the vast majority of margins, however, the terrigenous component is of major importance. Thus, the basal pelagic sediments normally are succeeded by hemipelagic sediments, in which the terrigenous component is significant, and finally by terrigenous ones, in which it is dominant (LASH, 1985; THORNBURG and KULM, 1987). The terrigenous material is largely deposited from gravity-driven sedimentary flows (UNDERWOOD, 1986a). Where subduction speed is more than about 20 km Ma^{-1} (2 cm yr^{-1}), the trench typically is a well-defined bathymetric low, and flows that reach its floor are axially confined and can travel hundreds to thousands of kilometers (PIPER *et al.*, 1973; UNDERWOOD, 1986b; THORNBURG and KULM, 1987). The resulting deposits resemble elongate submarine fans, in which the character of the stratigraphic section depends strongly upon position with respect to feeder canyons (SCHWELLER and KULM, 1978b; UNDERWOOD and KARIG, 1980; UNDERWOOD and BACHMAN, 1982; BOGGS, 1984). Where subduction speed is slow or sediment supply is large, however, the trench is not so well defined, or may even be absent, the flows can cross it, and more typical submarine-fan complexes can develop.

The hemipelagic section should be thicker and the transition to coarser-grained trench-axis terrigenous strata should be more abrupt where the trench traps the coarse fraction of gravity-driven sedimentary flows but allows plumes of fine-grained suspended sediment to travel farther oceanward (MOORE, J. C., *et al.*, 1982b). Other factors likely to cause changes in lithology, both gradational and abrupt, are migration of depositional lobes, damming of canyons that supply the trench, and damming of the trench itself by seamounts, fracture zones, or other bathymetric highs.

Mechanical Behavior of Incoming Sediment Pile

Three aspects of the mechanical behavior of the incoming sediment pile stand out as particularly important.

- First, as MOORE, J. C., (1975) has pointed out, the siliceous or calcareous pelagic sequences commonly are more lithified than the overlying hemipelagic and terrigenous deposits and therefore tend to be distorted less in the zone of compression.
- Second, the deformational and other properties of incoming terrigenous sediment at margins where the flows are confined to the trench axis should be different than where they are unconfined. Deposits from axially-confined flows will be subducted soon after deposition, whereas those from unconfined ones may be subducted tens of millions of years later, and hence may be more compacted and lithified. In addition, deposits from unconfined sedimentary flows will tend to be more tabular, and therefore their anisotropy differs from where flows are axially confined. These factors can significantly affect stress transmission, dewatering flow, and seismic-reflection characteristics.
- Third, the variations in anisotropy, lithology, lithification, and pore-fluid pressure in the incoming sediments are the dominant factors governing whether initial deformation is localized as a few faults or is distributed more evenly (see also BREEN *et al.*, 1986).

Deformation Patterns from Seismic-Reflection Profiles

Most of the deformation of the incoming sediments and the accretionary prism, as measured by variation in attitudes of the uppermost strata, takes place within 10 to 30 km of the base of the inner trench slope (SEELY *et al.*, 1974; KARIG and SHARMAN, 1975; MOORE, G. F., and KARIG, 1976; KARIG, 1983; VON HUENE, 1986b). At most margins it is due to arcward-dipping thrusting and imbrication, although large-scale folding is an important process at several localities, such as Oregon-Washington, Makran, and southern Chile. At a few places trenchward-dipping thrust faults appear to cut the upper portions of the deforming sediment pile (SEELY, 1977). Where the thickness of the incoming sediments is more than about a kilometer, some of the flat-lying topmost layers merge into those underlying the trench slope. Examples occur at Oregon-Washington (SILVER, 1971; KULM and

FOWLER, 1974; SEELY, 1977), southern Alaska (SEELY, 1977; SCHOLL *et al.*, 1982), the Aleutians (GROW, 1973), the Lesser Antilles (CHASE and BUNCE, 1969; WESTBROOK *et al.*, 1982; MOORE, J. C., *et al.*, 1982c; WESTBROOK and SMITH, 1983), southwest Japan (AOKI *et al.*, 1983; LEGGETT *et al.*, 1985), Makran (FARHOUDI and KARIG, 1977; WHITE and KLITGORD, 1976; WHITE, 1982), Sumatra (KARIG *et al.*, 1980; MOORE, G. F., and CURRAY, 1980), and southernmost Chile (HERRON *et al.*, 1977; CANDE and LESLIE, 1986). Upthrust ridges of basalt (presumably the top of the oceanic crust) detected in sediment-poor regions of the Chile trench (KULM *et al.*, 1973; PRINCE and KULM, 1975) comprise the only concrete evidence of possible detachment of a fragment of the top of the descending basement at an actively convergent margin.

No well-defined trench axis exists where the incoming sediment is especially thick. Instead, the slope base marks the beginning of disruption, or deformation front. The basal few hundred meters of sediment continue with relatively little discernible disruption for tens of kilometers arcward of the front. Examples are the Lesser Antilles and southwest Japan. A gently-dipping fault, or décollement, must separate the upper more-deformed layers from the lower less-deformed ones.

The trench axis generally is a well-defined trough as much as 10 to 20 km wide where the incoming sedimentary succession is thinner than about a kilometer. Typically, all that can be detected in seismic-reflection profiles is tilting or offset of reflectors in the sedimentary section within a few kilometers of the base of the trench slope. Examples are Java (BECK and LEHNER, 1974; HAMILTON, 1979; BREEN *et al.*, 1986), Mexico (SHIPLEY *et al.*, 1980; MOORE, J. C., *et al.*, 1982a), Guatemala (VON HUENE *et al.*, 1980), Costa Rica (SHIPLEY and MOORE, 1986), northeast Japan (NASU *et al.*, 1980; MATSUZAWA *et al.*, 1980; VON HUENE *et al.*, 1982), parts of the Aleutians (GROW, 1973; SCHOLL *et al.*, 1983), most of Peru-Chile (SHEPHERD and MOBERLY, 1981; HUSSONG and WIPPERMAN, 1981; COULBOURN, 1981), and Mariana (MROZOWSKI *et al.*, 1982). Recent multichannel seismic profiles show that the top of the incoming trench fill off Peru (VON HUENE *et al.*, 1985) and the Aleutians (MCCARTHY and SCHOLL, 1985) is complexly imbricated and thickened by thrusting into slices 500 to 1000 m thick over a distance of 20 to 30 km arcward of the deformation front. In these profiles the top of the descending plate, broken by a few horsts and grabens as much as several hundred meters deep, can be traced arcward perhaps as far as 40 to 50 km.

In short, the incoming sediment pile at convergent plate margins thickens dominantly by thrusting in the zone of compression. At all margins the basal few hundred meters of the pile, which consist principally of pelagic sediments, appear to be carried past the base of the trench slope, in some cases many tens of kilometers (WESTBROOK *et al.*, 1982); and, at many, onlap and trenchward thinning of the slope cover indicates trenchward growth of the accretionary prism by offscraping of the upper portion of the thickened pile onto the toe of the overriding plate.

Deformation Patterns from Deep-Sea Drilling Project Cores

Deep-sea drilling on trench slopes has reached maximum subbottom depths of about a kilometer (MOORE, J. C., and LUNDBERG, 1986). Whether it has actually penetrated the base of the slope cover at any site is unclear, inasmuch as core recovery generally is poor and differentiation between deformed slope cover and offscraped trench deposits is difficult. Cores obtained from the lower trench slope off Oregon and Alaska (DSDP Leg 18; VON HUENE and KULM, 1973) and southwest Japan (Nankai Trough, DSDP Leg 31; MOORE, J. C., and KARIG, 1976) consisted of sediment identical to the flat-lying sediment in the trench. Others from off Japan (DSDP Legs 56 and 57; CARSON *et al.*, 1982), Mexico (DSDP Leg 56; LUNDBERG and MOORE, 1981), and Guatemala (DSDP Leg 67; COWAN, 1982) showed increasing disruption with depth. Scaly fabrics due to convergence-driven deformation of mudstone layers were first described from such cores by MOORE, J. C., and GEIGLE (1974), CARSON *et al.* (1974), and CARSON (1977). In their summary of observations on DSDP cores from forearc regions, LUNDBERG and MOORE (1986) found the commonest nonsedimentary structures in sandstone layers to be stratal disruption by disaggregative particulate flowage and variable cataclasis and the commonest ones in mudstone layers to be scaly or spaced foliation. Nearly-vertical extension fractures were common in both.

The only thrust penetrated at an actively convergent margin, as indicated by stratigraphic inversion, is near the base of the trench slope off the Lesser Antilles (Legs 78A and 110; MOORE, J. C., *et al.*, 1982c; MOORE, J. C., *et al.*, 1987). That this feature is a fault is shown by a distinctive scaly fabric in the adjacent sediments, which indicates that the deformation was localized, at least during the initial stages of movement (MOORE, J. C., *et al.*, 1986).

Deformation Patterns from Field Studies of Accretionary Complexes

Bodies of chaotically-mixed, mud-matrix melange exist in many active and ancient accretionary complexes (RAYMOND and TERRANOVA, 1984; SUZUKI, 1986). Examples include the huge Central Belt of the Franciscan Complex of California (MAXWELL, 1974), the Uyak melange on Kodiak Island off Alaska (CONNELLY, 1978), the Oyo melange on Nias Island off Sumatra (MOORE, G. F., and KARIG, 1980), portions of the Shimanto Belt of Japan (TAIRA *et al.*, 1982), portions of the Basal Complex on Barbados Island in the Lesser Antilles (SPEED and LARUE, 1982), and numerous less-studied localities in Indonesia (HAMILTON, 1979). Many of these melanges have a distinctive scaly cleavage similar to that found in the fault zones cored by the DSDP (MOORE, J. C., *et al.*, 1986). Because scaly fabrics can develop in surficial landslides (e.g., LARUE and HUDDLESTON, 1987), however, generation of these melanges entirely by tectonic processes seems problematical to many workers. Many melanges include exotic bodies of greenstone (locally pillowed, brecciated, or layered in flows), serpentinite, chert, or limestone, typically amounting to 10% or less by volume. These bodies are exotic in the sense that they were not simply deposited *in situ* with the voluminous terrigenous sediments surrounding them.

Melanges, even if near the surface, are difficult to image by seismic-reflection techniques, because of both lack of bedding continuity and scattering from rounded blocks. It seems unlikely that chaotic, mud-matrix melange comprises large portions of the shallow levels of accretionary complexes where seismic imaging is good, as off Makran, southwest Japan, Oregon-Washington, and the Lesser Antilles. At the deeper levels that are not imaged and even in the shallower levels of most active margins, however, its volume is unknown.

Fault-bounded packets of well-bedded sediments are found in all accretionary complexes; and in many places they are much more extensive than bodies or belts of chaotically-mixed melange. Some of them are remnants of the slope cover or of slope basins (MAXWELL, 1974; MOORE, G. F., and KARIG, 1976; MOORE, G. F., *et al.*, 1980). Others are fragments of offscraped trench fills. Although offscraped trench sediments can readily be distinguished from modern slope-basin fills in seismic-reflection profiles across the younger portions of active margins, the distinction is much more difficult in the older portions and in ancient complexes (LASH, 1985). Differentiation may be possible only where interbedded pelagic carbonate strata indicate deposition above the carbonate compensation depth (CCD), which typically is less than 5 km and therefore is at the middle of the trench slope at most active convergent margins.

Bedded sequences interpreted to be offscraped trench-fill are usually folded, commonly with steeply dipping limbs. Examples include portions of the Shimanto Belt of southwest Japan (TAIRA *et al.*, 1982), the Ghost Rocks Formation of Kodiak Island and the Stilikiak Formation of Stilikiak Island, both off Alaska (MOORE, J. C., and ALLWARDT, 1980; BYRNE, 1982, 1984), the Chugach Formation (NILSEN and ZUFFA, 1982) and the Orca Group (HELWIG and EMMET, 1981) of southeast Alaska, and parts of the Coastal Belt of the Franciscan Complex of northern California (BACHMAN, 1978, 1982; AALTO, 1982). Portions of all these folded terranes contain zones of intense deformation that can appropriately be classified as melange (UNDERWOOD, 1984; FISHER and BYRNE, 1987; MACKENZIE *et al.*, 1987; HIBBARD and KARIG, 1987). Although, as already noted, broad, open folds are often detected in sediments being offscraped near the trench axis, the far tighter folds commonly seen in outcrop in ancient accretionary complexes cannot be detected, if present, in seismic sections of modern ones, because of their much steeper dips.

In short, many ancient accretionary complexes contain mappable units largely comprised of variously-deformed coherent trench fills or slope cover, or both, and of intensely deformed block-bearing melange.

Sediment Subduction and Subduction Erosion

Although most early attention focused on offscraping, numerous observations have made it increasingly evident that sediment subduction and subduction erosion are also significant processes at convergent margins.

The volumes of rock derived from pelagic siliceous oozes and metalliferous sediments are much smaller in accretionary complexes than expected were the incoming plate continually scraped clean of sediment during convergence (SCHOLL and MARLOW, 1974; SCHOLL *et al.*, 1977). Apparently, these materials, which form the basal 200 m or so of the incoming sediment pile, are rarely accreted at levels that eventually become exposed by uplift and erosion; hence, they must be subducted to depths of 30 km or more. MOORE, J. C. (1975) suggested that they are "selectively subducted" to greater depths than the overlying trench-fill turbidites because the cherty materials comprising them couple to the descending plate more strongly than normal. Subduction of metalliferous sediments all the way to the depths of magmagenesis (about 100 km) would provide a simple explanation for the source of the trace metals in some volcanic arcs (SILLITOE, 1972).

Arc-type volcanic or plutonic rocks are present in the forearc regions of Mariana (HUSSONG and UYEDA, 1982; BLOOMER, 1983), Japan (VON HUENE *et al.*, 1982), Alaska (Kodiak Island; HILL *et al.*, 1981), and much of Peru-Chile (COUCH *et al.*, 1981; ZIEGLER *et al.*, 1981). Some of these occurrences may be explained by the subduction of spreading ridges (MARSHAK and KARIG, 1977), but at Mariana and Peru-Chile extensive subsidence has occurred in the forearc area (HUSSONG and UYEDA, 1982; VON HUENE *et al.*, 1987). Subduction erosion of the base of the overriding plate seems to be the only explanation.

Moreover, continental crust 1000 to 2000 million years old exists within a few tens of kilometers of the trench axis off Central and South America (RUTLAND, 1971); and similar relations are apparent for past convergence along parts of western California (YEATS, 1968; PAGE, 1970). The accretionary complexes that have accumulated along much of the Peru-Chile margin are remarkably small, in light of the long history of continuous convergence there (HUSSONG and WIPPERMAN, 1981; COULBOURN, 1981; VON HUENE *et al.*, 1987). Indeed, if present rates of convergence and thicknesses of trench fill have existed since the beginning of the Tertiary, considerable volumes of trench sediment are unaccounted for at many margins around the Pacific (HILDE, 1983). The same conclusion follows from comparison of sediment volume in the Pacific Basin with the volume of sediment of equivalent age on the floor of the Atlantic (GILLULY, 1969; GILLULY *et al.*, 1970). Although small accretionary complexes and truncated margins might be explained by transform faulting that carried the missing bodies elsewhere (KARIG, 1974b; KARIG *et al.*, 1978), this explanation does not seem adequate, particularly in the case of western South America, where an enormous mass of sediment is missing.

JONES *et al.* (1978), SCHWELLER and KULM (1978a), UYEDA and KANAMORI (1979), COULBOURN (1981), HUSSONG and WIPPERMAN (1981), HILDE (1983), and others have argued from these relations that sediment subduction to considerable depths and locally even subduction erosion of the overriding plate are processes comparable in importance to accretion by offscraping (Figure 2; SCHOLL *et al.*, 1980; VON HUENE, 1984a, 1986a). The mechanism of these processes is particularly obscure, however, because they produce no rock record.

Diapirism of Sediment, Mud-Matrix Melange, and Serpentinite

SUPPE (1973) and MAXWELL (1974) suggested that diapirism of unlithified, undercompacted sediment in the accretionary prism was an important deformational process responsible for the formation of some melange in the Franciscan complex of northern California. BECKER and CLOOS (1985) described a body of blueschist-bearing mud-matrix melange that appears to have risen diapirically into a slope-basin deposit in the Franciscan in central California. SPEED and LAURE (1982), WILLIAMS *et al.* (1984), and COWAN (1985) argued that shale diapirism is an important mechanism of melange formation in active accretionary terranes. Mud volcanoes have been found on the trench slope of the southern Lesser Antilles (BIJU-DUVAL *et al.*, 1982; STRIDE *et al.*, 1982) and near Java (BREEN *et al.*, 1986). COLEMAN (1980) has described serpentinite diapirs that pierced the Great Valley forearc basin in California; and FRYER *et al.* (1985) have described some that are spreading as submarine flows on the trench slope off Mariana. The processes of mud volcanism and of diapirism of sediment, mud-matrix melange, and serpentinite have been documented in both modern and ancient accretionary complexes, but their size range and volumetric importance are uncertain.

Blueschists and Subduction-Zone Thermal Structure

The presence of blueschist-facies metamorphic rocks is one of the more diagnostic criteria for recognizing sites of former plate convergence (ERNST, 1970; DEWEY and BIRD, 1970). Extensive tracts of blueschist-facies metamorphic rocks are found on the arcward side of late Mesozoic and younger subduction complexes in new Caledonia, Japan, Alaska, California, and the Alps (ERNST, 1975; see numerous papers in memoir edited by EVANS and BROWN, 1986). Classical blueschists contain minerals such as lawsonite, aragonite, sodic amphibole, and sodic pyroxene which indicate that metamorphic recrystallization occurred at pressures above 300 to 500 MPa (3 to 5 kbar, corresponding to depths of approximately 12 to 20 km) but at temperatures of 250 to 450°C, which are unusually low for such depths of burial (ERNST, 1971). The recent discovery of blueschist-facies metasedimentary rocks bearing coesite and pyrope in the western Alps shows that subduction of sediment layers to depths of 90 km or so followed by uplift and re-exposure at the surface does occur (CHOPIN, 1984).

Conditions for formation of blueschists develop in subduction zones because plate convergence at speeds of tens of kilometers per million years (centimeters per year) carries cold lithosphere downward faster than the interior heat of the earth is conducted upward through it. As a result, after a few tens of millions of years of subduction, the leading part of the overriding plate cools and in the forearc region the geothermal gradients become greatly reduced compared to typical crustal-level gradients in the continents and ocean basins (less than 10 compared to 25 to 35°C km^{-1}).

Conductive thermal models of BIRD (1978), HSUI and TOKSÖZ (1979), HONDA and UYEDA (1983), WANG and SHI (1984), and VAN DEN BEUKEL and WORTEL (1986) show temperatures less than 150°C at depths of 20 to 30 km or more (Figure 5) where plate convergence has been fast for more than 10 to 20 million years and frictional or shear heating has been minor because shear stresses are less than a few tens of megapascals (a few hundred bars). These calculated temperatures are below those commonly estimated for "classical" blueschist-facies rocks that recrystallized in this depth range (300 to 400°C; see CLOOS, 1985). Nevertheless, they should approximate the actual temperatures, because shear stresses are actually only a few megapascals (a few tens of bars) when pore-fluid pressures are near lithostatic values (see WANG and SHI, 1984). Hence, it appears that the thermal conditions conducive to production of classical blueschist-facies mineral assemblages in materials subducted to depths of 20 to 30 km exist only while the hanging wall is still relatively warm during the early stages of subduction, or when convergence is extremely slow (less than 10 km Ma^{-1}, or 1 cm yr^{-1}).

Water derived from subducted sediments that percolates through the overlying block will advect heat and perturb the thermal structure from that calculated from a conductive model. Upward water flow makes the near-surface temperature gradients higher than calculated and the deeper ones lower (perhaps less than 2 or 3°C km^{-1}; RECK, 1987). Nonetheless, the total volume of water available from subducted sediments and hydrated oceanic crust is rather small, and the deep-seated thermal structure (at depths of tens of kilometers) should not be radically different from that calculated from the purely conductive models (PEACOCK, 1987).

Once subduction ceases, the reestablishment of near-normal geothermal gradients will occur over a few tens of millions of years if the cold subducted plate beneath the blueschists remains in place (OXBURGH and TURCOTTE, 1974) or much more rapidly if it sinks and hot asthenospheric material wells into its place (DICKINSON and SNYDER, 1979). In either case, heating of deep-seated blueschists will occur because the heating from below exceeds the cooling from above due to erosion of the cover rocks. It causes hydrous low-temperature minerals to dehydrate and the blueschists to be replaced by greenschists and amphibolite-facies mineral assemblages (e.g., YARDLEY, 1982; DRAPER, 1986; SCHLIESTEDT and MATTHEWS, 1987) as the thermal regime returns to normal (DRAPER and BONE, 1980; ENGLAND and THOMPSON, 1984). Wherever blueschists are preserved with little

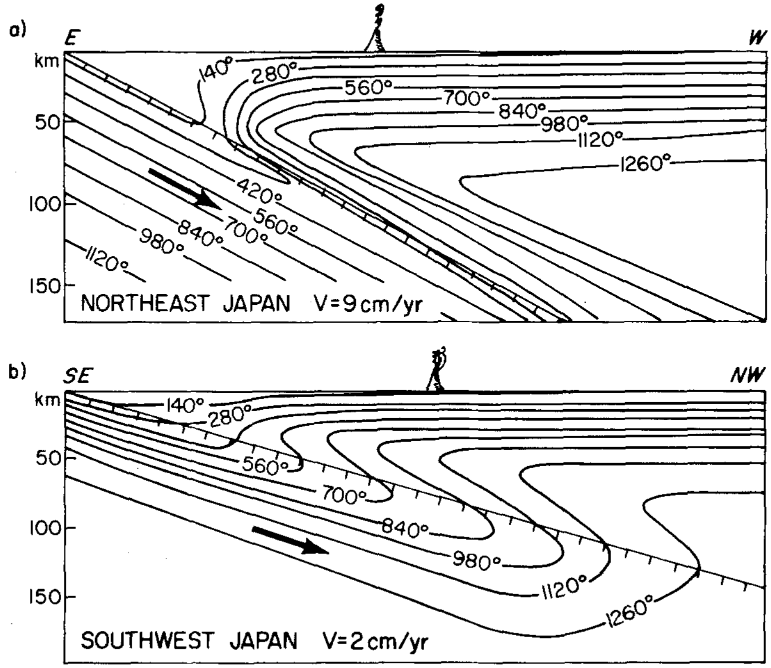

Figure 5

Patterns of isotherms calculated for northeast Japan and southwest Japan by HONDA and UYEDA (1983). Hachured line indicates top of descending plate. In these calculations, heat flow was assumed to be dominantly conductive, and frictional or shear heating along the plate interface was assumed to be negligible. This approximation is reasonable because pore-fluid pressures are probably very nearly lithostatic within the layer of subducting sediments dragged beneath the overriding plate by the descending one. Heat advection by upward-moving water originating from dewatering at the hanging wall will modify the pattern of isotherms, however, by making near-surface geothermal gradients steeper and deeper-level ones gentler than those shown (see RECK, 1987), but will not change the overall pattern. At northeast Japan the subduction speed is fast (90 km Ma^{-1}, or 9 cm yr^{-1}), the descending Pacific plate is relatively old and cold, and the plate dip is steep. This causes such extreme downbowing of the isotherms that most thermally-driven recrystallization occurs only below depths of 40 km or more. At southwest Japan the subduction speed is slow (20 km Ma^{-1}, or 2 cm yr^{-1}), the descending Philippine plate is relatively young and warm, and the plate dip is gentle. Although the isotherms are downbowed, thermally-driven recrystallization to create schistose blueschists can occur at depths as small as 15 to 20 km.

greenschist-facies alteration, as in the Franciscan of California (CLOOS, 1986), and particularly where metamorphic aragonite is present (CARLSON and ROSENFELD, 1981), either sysubduction uplift to depths shallower than about 10 km or extremely rapid postsubduction unroofing of much of the cover rock is indicated.

Although the generation of blueschists in subduction zones seems so well understood that their presence is now a primary criterion used to recognize ancient sites of subduction, the relationship between the uplift and exposure of blueschists and the growth of accretionary prisms is particularly obscure. Part of the problem is that no blueschist bedrock or fragment has yet been found by drilling or dredging on the trench slope of any actively convergent margin. Thus, it appears that the emergence of blueschist bedrock over a substantial area at the surface typically occurs after subduction ceases. In any case, the widespread preservation of blueschists in ancient accretionary complexes seems to indicate directly that subduction zones can be "two-way streets," along which sediment commonly is subducted down to, and then uplifted back from, depths of 20 to 30 km (SUPPE, 1972).

The formation and preservation of blueschists place strong constraints on thermal structure, which clearly is a primary factor controlling the geodynamic evolution of a subduction zone, because of its strong effect on the mechanical properties of rock. Thermal modeling indicates that quasisteady low-temperature thermal conditions are attained and thermal effects are minor down to depths of 30 km or more after a few tens of millions of years of fast plate convergence. Prior to that time, however, large horizontal and vertical variations in temperature strongly control the mechanical behavior and petrologic character of the subducted rocks (see CLOOS, 1984, 1985).

Subduction-zone Seismicity

The pattern of ground breakage during the great Alaska earthquake of 1964 provided conclusive evidence for movement on a gently-dipping thrust during a major Benioff-zone earthquake (PLAFKER, 1965). The development of plate-tectonic theory explained thrust-type motions as due to plate convergence (ISACKS *et al.*, 1968), which many workers now consider to be primarily driven by slab pull (see SPENCE, 1987). Since then, numerous studies have shown that subduction-zone earthquakes vary greatly in number, magnitude, distribution, and rupture patterns from one margin to another, and even along a single margin (LAY *et al.*, 1982; TAJIMA and KANAMORI, 1985; HABERMANN *et al.*, 1986; JARRARD, 1986). "Asperities," whose failure causes a rupture that propagates into weak areas, and "barriers," which stop propagation of ruptures that nucleated in weak areas, along the interplate shear zone are thought to play a critical role in determining the detailed nature of subduction-zone seismicity (RUNDLE *et al.*, 1984).

Of crucial importance to understanding the geodynamics of the forearc region is recognition that the ruptures in great thrust-type earthquakes nucleate at depths of 30 to 50 km and propagate updip. Seismic activity typically is small above depths of 20 km. In fact, the margins with the largest accretionary prisms, such as Makran, Oregon-Washington, and the Lesser Antilles, seem to have the fewest and smallest earthquakes. Hence, it now appears that earthquake-seismological studies do not closely constrain the tectonic stress state and large-scale movement patterns within much of the thickness of large accretionary prisms (see review by CHEN *et al.*, 1982; FROHLICH *et al.*, 1982). Taken together with the evidence for extension in forearc regions with little or no accretion, such as Mariana, Guatemala, and parts of

Peru-Chile, this observation indicates that the state of tectonic stress in the forearc region is not so simple as the subhorizontally-directed compression that has often been assumed for convergent plate margins (see also VON HUENE, 1984b, 1986).

Arc Volcanism

The origin of volcanic arcs around the Pacific has been a focal point for most tectonic models for the earth. COATS's (1962) paper on the Aleutian arc is remarkably prescient in light of present-day plate-tectonic theory and our current understanding of the origin of arcs. He proposed that eugeosynclinal sediments and basalts of the ocean floor are carried down to depths of at least 100 km along a major thrust marked by the Benioff seismic zone, where they are added to molten mantle material, which then rises to form the volcanoes. ISACKS *et al.* (1968) showed how Coats's picture for the origin of volcanic arcs is readily incorporated as part of plate-tectonic theory. The importance of corner flow induced in the asthenospheric wedge in heating the cold descending plate and causing melting near its top is now recognized as a fundamental process at convergent plate margins (MCKENZIE, 1969; HSUI *et al.*, 1983). Research has focussed on determining what fractions of arc magmas, which typically are basaltic to andesitic in composition, are derived from subducted oceanic crust, from subducted sediments, and from the overlying mantle wedge. One point that is clear is that, where magmas pass through continental crust, extensive modification of their composition by crustal assimilation is typical (LEEMAN, 1983). This makes recognition of the primary source of these arc magmas difficult, if not impossible, to determine.

In the early 1970s, the available isotopic data were commonly interpreted as indicating little incorporation of subducted oceanic sediment into arc volcanic rocks (less than a few percent; CHURCH, 1973). Moreover, subduction of altered oceanic crust (e.g., SPOONER and FYFE, 1973) or contamination by crustal rocks could explain the presence of anomalous ratios of elements or isotopes in arc magmas. In addition, KARIG and SHARMAN (1975) concluded that the volume of material within accretionary prisms at subduction zones can only be explained by complete offscraping of all incoming sediment and probably even slabs of oceanic crust from the top of the descending plate. The generally-held conclusion was that extensive, if not complete, offscraping of incoming sediment occurred at all trenches.

ARMSTRONG (1971), however, was notable for arguing during this period that significant sediment subduction to the depths of magmagenesis could explain the geochemical characteristics of several arcs. Now, much trace-element and isotopic data on Pb, Sr, and Nd in volcanic deposits at many arcs seems to indicate a minor to substantial terrigenous or pelagic sedimentary component (see, for example, KAY, 1980, ARMSTRONG, 1981; ARCULUS and POWELL, 1986). Examples of oceanic arcs with such geochemical evidence include Scotia (BARREIRO, 1983), Lesser Antilles (DAVIDSON, 1986; WHITE and DUPRE, 1986), the Aleutians (MYERS *et al.*, 1986), and Mariana (HOLE *et al.*, 1984). Because some of the compositional variations at these and other arcs can be explained by assimilation of sediments contained within the volcanic pile or its underlying crust (DAVIDSON, 1987), by melting of subducted oceanic crust that has been altered by high-temperature hydrothermal processes at spreading ridges (MOTTL, 1983), by lower-temperature interaction of basalt with seawater as it is conveyor-belted away from the ridge (STAUDIGEL *et al.*, 1981), or by some combination of these processes, deep sediment subduction to the depths of arc magmagenesis has been disputed.

The recent discovery that the short-lived radioisotope ^{10}Be is detectable in arc volcanic rocks from northeast Japan, Guatemala, the Aleutians, and Peru (BROWN *et al.*, 1982; TERA *et al.*, 1986), however, seems to indicate directly the involvement of subducted sediments in magmagenesis at these arcs. This conclusion follows because ^{10}Be is present in measurable quantities only in igneous rocks or sediments less than about 10 million years old and its near-surface incorporation into the magmas from groundwater or sufficiently young sediments appears unlikely.

In addition, a small sedimentary component heterogeneously distributed about the upper mantle seems necessary in order to explain some of the geochemical variations found in ocean island basalts (ZINDLER *et al.*, 1984; WEAVER *et al.*, 1986) and ocean ridge basalts (ALLEGRE *et al.*, 1984) and to account for the entire pattern of crust-mantle chemical evolution (ARMSTRONG, 1981; COHEN and O'NIONS, 1982; PATCHETT *et al.*, 1984; O'NIONS, 1987). The general conclusion seems to be that some sediment apparently is subducted past the site of arc magmagenesis and assimilated into the asthenosphere.

Subduction of significant quantities of sediment to the depths of magmagenesis and even deeper at many, if not most, arcs now seems certain. The volumetric importance of such deep sediment subduction is still uncertain, however.

Models for Convergent Plate Margins

Imbricate-Thrust Model

In the early 1970s the imbricate-thrust model for accretion at trenches was developed through synthesis of seismic-reflection profiles, deep-sea drilling data, and field studies of emergent and ancient accretionary complexes (SEELY *et al.*, 1974; KARIG, 1974a; KARIG and SHARMAN, 1975). This model is still widely cited as a general model for the tectonics of subduction zones. It explained (or, in certain cases, seemed to explain) numerous observations, such as the following.

- Large thrust-type earthquakes are characteristic of many convergent margins and occur at depths as great as 60 km (ISACKS *et al.*, 1968). Hence, thrust-type motions are a dominant aspect of the deformation at convergent plate margins.
- At many convergent margins, gravity, magnetic, and reflection- and refraction-seismic studies indicated the presence of broad, thick bodies of sediment, which

generally appeared to be greater in volume the longer the period of convergence (DICKINSON, 1973). These bodies could be explained as accretionary prisms formed by progressive offscraping of incoming trench fill. Seismic-reflection profiles were readily interpreted in terms of this process off the Lesser Antilles (CHASE and BUNCE, 1969), the Aleutians (GROW, 1973), and Java (BECK and LEHNER, 1974).

- Deep-sea drilling off Oregon and Alaska (VON HUENE and KULM, 1973) and southwest Japan (MOORE, J. C., and KARIG, 1986) found sediment on the lower trench slope identical to flat-lying sediment in the trench, thereby providing direct evidence of uplift that could be explained by imbricate thrusting during offscraping.
- At many margins a well-defined bathymetric barrier, the *forearc high*, separates the forearc basin from the trench slope. In several areas, such as Guatemala and Sumatra, it is a region of recent uplift (SEELY *et al.*, 1974; KARIG and SHARMAN, 1975) that can be explained by thickening of the prism due to progressive emplacement of wedge-shaped thrust sheets.
- In the early 1970s (in contrast to the present) most isotopic data indicated little if any incoporation of subducted oceanic sediment into arc volanic rocks (see, for example, CHURCH, 1973). Near-total offscraping of the incoming sediment explained this observation.

Since 1975 the Deep Sea Drilling Project has obtained a wealth of new data concerning the nature of the convergent margins off northeast Japan (DSDP Legs 56, 57, and 87), Mariana (Leg 60), Mexico (Leg 66), Guatemala (Leg 67), the Lesser Antilles (Legs 78A and 110), and Peru (Leg 112). These investigations, coupled with recent seismological and geochemical studies, have thrown new light on the general applicability of the imbricate-thrust model. The major new observations include the following.

- The Tertiary parts of the accretionary prisms off Guatemala and northeast Japan are much smaller than previously thought (VON HUENE, 1981). In fact, the Guatemala margin, which had been cited by SEELY *et al.* (1974) as a prime example of subduction accretion through imbricate thrusting of incoming trench-fill, appears to have nearly no prism (VON HUENE *et al.*, 1980), and the northeast Japan prism, though very broad, is mostly Cretaceous in age (KARIG and KAGAMI, 1983).
- Although the imbricate-thrust model explains uplift of the forearc, it cannot explain the subsidence or the combination of uplift and subsidence at Mexico and northeast Japan or the probable subduction erosion at Mariana, Guatemala, and Peru-Chile.
- Trace-element and isotopic studies of Pb, Sr, and Nd in volcanic deposits at arcs seems to indicate the presence of recycled continental material, which may be subducted sediment, in the primary magma of many of them (ARMSTRONG, 1981). The presence of the short-lived radioisotope ^{10}Be in arc volcanic rocks

from Guatemala, the Aleutians, and Peru (BROWN *et al.*, 1982; TERA *et al.*, 1986) is a direct indicator of the subduction of sediments to the depths of arc magmagenesis. The recent discovery of metasedimentary blueschists in the Alps, which were subducted to depths of 90 km or so (CHOPIN, 1984), indicates that layered sediments can be carried by the descending plate nearly to the depths of arc magmagenesis.

- Seismological studies do not directly constrain the stress state and large-scale movement patterns within broad, thick accretionary prisms at depths less than 20 km or so, because few, if any, major earthquakes appear to be located within them. This, combined with the evidence for extension at some sites, indicates that the state of tectonic stress in the forearc region is not so simple as the subhorizontally-directed compression generally concluded from seismological models for convergent margins and assumed in the imbricate-thrust model (see VON HUENE, 1984b, 1986).

Other Models

Recently, SILVER *et al.* (1985), LEGGETT *et al.* (1985), PLATT *et al.* (1985), and SAMPLE and FISHER (1986) have modified the imbricate-thrust model by suggesting that accretionary prisms thicken primarily by the emplacement of elongate thrust-bounded packets of sediment (duplexes) beneath them instead of by offscraping of wedge-shaped bodies near their toes. They made this modification in order to explain the localization of deformation near the base of the trench slope, the passive uplift of margins, and the thickening of prisms without shortening.

Following an approach outlined by CHAPPLE (1978), DAVIS *et al.* (1983) and DAHLEN *et al.* (1984) used an empirical Coulomb-failure criterion modified to account for the weakening effects of pore-fluid pressure to model plastic deformation in the upper 10 to 15 km of accretionary prisms at convergent margins and in active fold-and-thrust belts such as Taiwan. By assuming a reasonable value for the coefficient of internal friction, they predicted the basal dip and the pore-fluid pressures from the known surface slopes. This analysis indicates that pore-fluid pressures are near lithostatic values in many accretionary prisms.

Similarly, STOCKMAL (1983) developed a model for deformation of the entire accretionary prism that assumes perfect-plastic rheology. He used it to calculate the pattern and rates of uplift in the prism off Sumatra.

COWAN and SILLING (1978) simulated convergence and accretion in "clay-cake" experiments and concluded that faulting and imbrication occur at shallow depths whereas distributed flowage occurs at greater depths. They found upward flow that would explain the localization of blueschists near the arcward sides of accretionary complexes.

Over the last few years a number of workers have approximated deforming accretionary prisms as Newtonian-viscous bodies. EMERMAN and TURCOTTE

(1983), for example, by analyzing the bathymetric profiles of the Kurile, Ryukyu, and Aleutian prisms deduced that they have equivalent viscosities between 10^{17} and 10^{18} Pa s (10^{18} and 10^{19} poise). ENGLAND and HOLLAND (1979) investigated the effects of the downward shearing and opposing buoyancy in a highly simplified subduction shear zone of uniform thickness and concluded that for a thickness of 1 km low-density sediment would be deeply subducted if its viscosity were greater than 10^{16} Pa s (10^{17} poise). They also concluded that buoyancy was important during the postsubduction rise of deeply-buried eclogite-bearing metasedimentary rocks in the Alps. CLOOS (1982) proposed a *corner-flow model*, from which the subduction-channel model initially evolved, in which upflow occurs in the sediment dragged into a long, narrow corner of predetermined geometry between the overriding and descending plates. He concluded from a quantitative investigation that such upwelling can explain the thorough mixing characteristic of mud-matrix melanges, including the incorporation and uplift of blueschist blocks such as those found in the mud-matrix melanges of the Central Belt of the Franciscan Complex of California.

Although we differ with these authors regarding some specific details, we believe that the imbricate-thrust model and all of the others are valid for particular aspects of certain margins.

Subduction-Channel Model

The subduction-channel model (SHREVE and CLOOS, 1986) provides a quantitative theoretical framework that unifies most of the principal concepts of these subduction models, accords with most of the fundamental observations on specific sites, and elucidates the increasingly apparent diversity of actively convergent margins. It applies where subduction occurs under the sea, unconsolidated sediment is present, and convergence rate exceeds about 20, and perhaps 10, km Ma^{-1} (2, and perhaps 1, cm yr^{-1}). Slower convergence provides more time for heating (HONDA and UYEDA, 1983; WANG and SHI, 1984) and dewatering, which can cause changes in density and viscosity due to recrystallization and lithification that are not included in the present version of the model.

The model addresses the processes of subduction accretion (both offscraping and underplating) and sediment subduction and identifies margins at which subduction erosion is likely. It treats the basic mechanics of these processes as being essentially the same at all convergent margins having Benioff seismic zones. It predicts wide variations in tectonic behavior, depending upon such site-specific variables as the subduction speed, the rate of sediment supply to the trench, the dip of the descending plate, and the topography and structure of the overriding one. It quantitatively predicts the conditions for and rates of offscraping and deep sediment subduction, the magnitudes and distribution of shear stress in the subduction channel, the thickness of the channel, the rate of thickening of the accretionary prism by underplating, the requirement for and the depths of upflow of melange, and numerous other details. It supplies a basis for classification of convergent margins and provides quantitative constraints on, and detailed insight into, such aspects as the patterns and rates of uplift from the trench slope through the forearc region, the mechanisms of blueschist uplift and metamorphic zonation during continued subduction, the distribution and magnitude of earthquakes, and the volumes of deeply subducted sediment that can contribute to arc magmas.

The Subduction-Channel Model

Basic Assumptions and New Terminology

A central assumption of the subduction-channel model is that the subducting sediment deforms approximately as a (not necessarily Newtonian) viscous fluid as it is dragged by the descending plate beneath the overriding crystalline plate and, if one is present, the accretionary prism. The layer of deforming sediment comprises a kind of shear zone, called the *subduction channel* (or *flow channel*). The name emphasizes that the kinematic patterns within the channel differ significantly from those in typical shear zones (Figure 6), especially where upflow occurs. Sediment subduction occurs because the downward-directed shearing overcomes the buoyancy and the (normally) adverse pressure gradient that act upon the low-density sediment. The exact pattern of flow and the thickness of the subduction channel

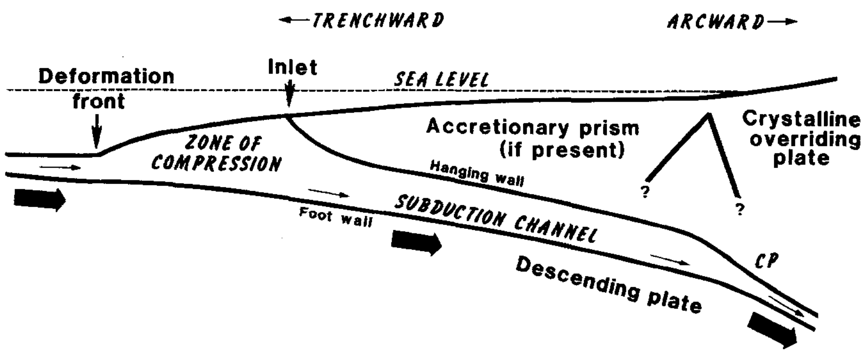

Figure 6
Major features of the subduction-channel model, shown schematically for a margin with an accretionary prism. The subduction channel is the sediment-filled shear zone in which deformation is concentrated. For clarity, its thickness is greatly exaggerated. The deformation front and the inlet bound the zone of compression, which is the region of bulldozer-like action in front of the overriding prism. A control point (*CP*) is located where the channel capacity decreases as a result of a sharp increase in the density of the overlying rocks, in the slope of the topographic surface, or in the dip of the descending plate, or in a combination of the three. A zone of upflow of subduction melange can develop when the amount of sediment arriving at the control point exceeds its capacity.

depend upon the sediment supply and density, the subduction speed, and the pressure gradient along the channel.

The top and bottom of the channel are termed the *hanging wall* and *foot wall*, in analogy with fault terminology. Where underplating is occurring, the hanging wall, as we envision it, is a gradational zone, perhaps a few hundred meters or less thick, across which the bulk deformation rate decreases by at least an order of magnitude, because of compaction and lithification caused by dewatering. The overriding crystalline plate and, where one is present, accretionary prism are thus treated as effectively rigid compared to the undercompacted, weakly lithified sediments in the zone of compression and in the channel, although, of course, they do slowly deform. The models of DAVIS *et al.* (1983), STOCKMAL (1983), EMERMAN and TURCOTTE (1983), and THARP (1985) can address this slow deformation of the overlying block. The subduction-channel model can provide some of the boundary conditions they require.

The opening, or *inlet*, of the subduction channel (Figure 3) is located at the leading edge of the overriding crystalline plate or accretionary prism, typically 10 to 20 km, but as much as 50 km, from the deformation front at the base of the trench slope. Where the incoming sediment is completely subducted, the inlet may be sharply demarcated. Where sediment or melange is offscraped, however, it doubtless is gradational between the unconsolidated sediment of the zone of compression and the lithified rock of the accretionary prism. In the latter case, it is a zone across which the rate of deformation decreases and the degree of dewatering, compaction, and lithification increases. Where offscraping is particularly extensive (for example, where incoming sediments are thick and subduction speed is slow, as at Lesser Antilles, southwest Japan, and Makran), the inlet is unusually far upslope and is probably a diffuse zone several tens of kilometers wide.

The subduction-channel model as currently implemented does not give the details of sediment movement in the zone of compression nor immediately arcward of the inlet, because of the complications there due to strong longitudinal shortening, variable dewatering, bedding anisotropy, large nonuniformities in lithification, and high but variable pore-fluid pressures (SHI and WANG, 1985). Seismic-reflection profiles and deep-sea drilling show that in this region slow deformation is accommodated largely by faulting and folding. Nevertheless, because flow rates at the inlet must match those a short distance down the channel, where the model does apply, it can give the rate of offscraping, which is simply the excess of the incoming sediment supply over the near-inlet channel capacity, hereafter termed for brevity the *inlet capacity*. DAVIS and VON HUENE (1987) have modeled aspects of the deformation during compression in this region.

The incoming sediment typically has a density of about 2.0 Mg m^{-3} and a porosity of 40 to 50%. We account for the partial dewatering and compaction in the zone of compression by taking the bulk density of the sediment entering the inlet to be 2.2 Mg m^{-3}, a value that corresponds to a porosity of 20 to 25%. Within the channel, the porosity and water content are very nearly uniform, and pore-fluid pressures are at or very close to lithostatic.

Underplating of material onto the roof of the channel uplifts and thickens the overriding block with or without concurrent shortening. Although this process has been inferred for many margins (MOORE, J. C., *et al.*, 1982a; KARIG, 1983; SILVER *et al.*, 1985; LEGGETT *et al.*, 1985; PLATT *et al.*, 1985; PLATT, 1986; CLOWES *et al.*, 1987), its cause and mechanism are unclear. We envision that it occurs because dewatering of downgoing sediment or upwelling melange at the roof of the channel causes its yield strength to increase drastically, initially because of compaction due to the water loss, but also because the resultant slight decrease in pore-fluid pressure causes lithification by such processes as pressure solution (see MOORE, J. C., and BYRNE, 1987) and related stress-activated low-temperature diagenetic or metamorphic processes. This consolidation causes further intraplate shearing to concentrate within the underlying sedimentary materials that have higher porosity and nearly lithostatic pore-fluid pressures. Essentially, underplating is due to the compaction and lithification of subducted sedimentary material caused by water loss into the hanging wall. This process occurs from the top of the channel downward, which is opposite the situation in a sedimentary basin.

In detail, our picture is that underplating may occur either grain by grain where the material is a penetratively-deformed melange (see MOORE, J. C., and BYRNE, 1987), or by deactivation of slip surfaces spaced up to tens of meters apart, generating structures similar to duplexes where parts of the accreted sediments retain relict bedding and other sedimentary features (see SAMPLE and MOORE, 1987), or by a combination of the two processes (Figure 7).

We assume that the fluid pressures in the subduction channel are lithostatic, that is, equal to the overburden pressure, and that water content, and hence density, is essentially uniform except very near the hanging wall, where loss of pore water to the hanging wall by percolation upward through the overriding block causes underplating of the adjacent sediment or melange. Our choice of permeabilities for the rocks of the overriding block accord with published values and give rates of thickening (SHREVE and CLOOS, 1986) in agreement with reported long-term rates of uplift, which typically are around a few hundred meters per million years (Table 1). We assume further that dewatering and underplating are inhibited, and subduction erosion is likely, where the shear stress, or *drag*, acting on the sediment or melange near the hanging wall exceeds a critical value, taken to be 1.5 MPa (15 bars). Although no direct observational justification for this particular choice can be given, it gives plausible results, such as the lack of underplating near the inlet at Mariana (SHREVE and CLOOS, 1986).

The forces acting on the hanging-wall rock just above the channel are gravity, which acts downward, the pore-pressure gradient due to the flux of water, which acts upward, and the drag, which acts parallel to the hanging wall and tends to induce tension at a 45° angle to it. Our tentative picture is that subduction erosion

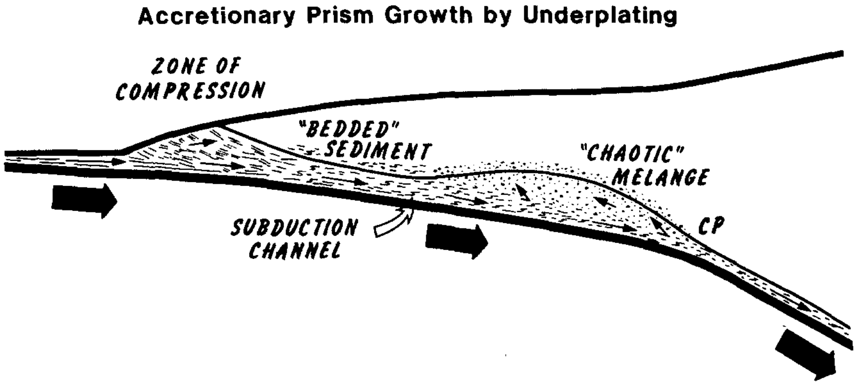

Figure 7
Idealized patterns of underplating onto the hanging wall at a margin where upflowing subduction melange (turned back by the control point *CP*) is entirely underplated onto the base of the prism before reaching the inlet. Longitudinal compression of the incoming sediments in the zone of compression causes deformation ranging from faulting and folding to particulate flowage. Sediments subducted past the inlet of the subduction channel undergo large, mainly bedding-parallel simple shear. The degree of penetrative deformation largely depends upon the depth at which accretion occurs and upon the variations in lithology, bedding anisotropy, and pore-fluid pressure that determine the distribution of shearing within the deforming sediment. When slip surfaces are spaced centimeters or less, shearing is thoroughly penetrative and tectonic melange is generated. When they are spaced meters to tens of meters, on the other hand, deactivation of slip zones by dewatering and compaction causes discrete packets of moderately-deformed bedded sediment to be underplated. Farther arcward, upwelled material is underplated. It is a chaotically-mixed melange, termed subduction melange, which is formed by the intense shearing where it reversed direction near the control point. It lacks most bedding features and probably is underplated more piecemeal than is sediment that retains significant bedding anisotropy.

occurs when the most tensile principal component of the resultant effective stress exceeds the strength of the hanging wall rock, which we take to be about 1.5 MPa (15 bars), or an order of magnitude less than typical tensile strengths of sound laboratory specimens (McLintock and Walsh, 1962). The pressurized sediment from the channel penetrates and extends the resultant cracks, prying off blocks of the wall, until stopped by the increasing disparity between the lithostatic pressure and the pre-crack pore pressure, which increases both the stress difference needed to propagate the crack farther and the rate of dewatering, and hence stiffening, of the sediment in the crack.

We think that, where the hanging wall consists of highly-permeable crystalline rock, as at Mariana, short episodes of rapid erosion may alternate with much longer periods of slow underplating of relatively impermeable sediment, which drastically reduces water input to and pore pressure in the overlying rock. The length of the cycle is governed by the speed at which water can move into and out of the crystalline rock as its crack-dominated porosity and permeability adjust in response to the alternating swings in water pressure at its base.

Control Points and Upflow

Under appropriate conditions downgoing material can be turned back, forming a thick zone of reverse flow, or *upflow*, along the roof of the channel. Such material is intensely deformed, because of the large, complex, penetrative strain it undergoes during the reversal of direction. It is termed *subduction melange* (so qualified to distinguish if from other fragmented or mixed materials that may also, by standard usage of the term, be properly called melange). It can entrain small blueschist or other blocks and raise them to shallow levels (CLOOS, 1982). In addition, it exerts trenchward-directed drag on the hanging wall that might move larger sheets of previously-accreted rock, including early-formed coherent blueschist terranes, toward the surface.

Normally, no more than a certain finite amount of sediment can move downward past any particular point in the subduction channel. This maximum, or *local channel capacity*, varies along the channel and gives rise to variations in flow pattern and channel thickness. A point of minimum local capacity, or *local control point*, governs both the amount of sediment that can pass arcward of it and the pattern of flow that can form trenchward of it. When sediment supply to a control point exceeds its capacity, the accumulation of sediment thickens the channel, causing the adverse pressure gradient and the buoyancy of the sediment to become dominant over the downward shearing. This in turn causes a zone of upflow to form and subduction melange to move trenchward below the roof of the channel. The control point with the least capacity of any in the channel, or *global control point*, governs the maximum amount of sediment that can be deeply subducted, presumably to the depths of arc magmagenesis. It is usually located at the point of greatest adverse pressure gradient along the subduction channel, which typically is where the slope of the topographic surface or the dip of the descending plate increases sharply or where the rear of an accretionary prism abuts crystalline rocks. The capacity of the global control point is the *global channel capacity* for the system. If the incoming sediment supply is less than the global capacity, at least some, and possibly all, of it is deeply subducted; the rest, if any, is underplated; and no upflow occurs.

Several control points with successively lower capacities can be present, in which case particularly complex flow patterns can arise. In the commonest arrangement, however, only two are present: one at the inlet that determines the amount of sediment entering the channel; and one at depth that determines the global capacity and hence the maximum amount of deep sediment subduction possible.

Our calculations indicate that the subduction channel is generally less than about a kilometer thick where upflow is absent, but can be as much as several kilometers thick where it is present. These thicknesses accord with those observed in the regions of most-active deformation (the zone of compression) along lower trench slopes. In the region of upflow a layer of sediment next to the floor of the channel moves downward past the control point at the global-capacity rate and underlies a layer of downgoing sediment and possibly melange that reverses direction before it reaches the control point to become a thick layer of upwelling melange next to the roof. At most sites the upflow peters out some distance upstream by loss to underplating; at some it reaches the inlet, where part of it is offscraped and the rest turns back downward; and, in a few cases, it simply cycles up and down in a whirlpool-like gyre within the subduction channel.

In rare cases the pressure gradient acts to drive subducting sediment down instead of up in a certain, usually short, zone of the channel, called a *zone of unlimited channel capacity*. This can occur where the topographic surface has a sufficiently steep slope to arcward or where the overriding block has an unusual density distribution. Melange that upwells to such a site cannot rise farther; hence, it accumulates and, once its volume grows large enough, may diapirically intrude the overlying block. A possible example is Barbados Island in the Lesser Antilles (SPEED and LARUE, 1982; SHREVE and CLOOS, 1986).

Conditions Governing Subduction Accretion and Erosion

Subduction erosion can occur extensively where the drag on the roof of the channel is strong enough to cause piecemeal stoping of the overriding block or locally where asperities on the downgoing plate are high enough to enable rasping of it. Although the amount of hanging-wall material that can be deeply subducted is probably controlled by the same factors that control the process for sediment, the rate of erosion at different points along the top of the channel probably depends strongly on local drag and lithology. It seems likely, for example, that the "erodibility" of a granitic hanging wall would significantly differ from that of a peridotitic one or an old accretionary prism. Granite may tend to be stoped as large blocks, whereas serpentinized peridotite or sedimentary rock may be removed more piecemeal. The rate of alteration of crystalline hanging-wall rock to clay and the nature of pre-existing discontinuities must also be important factors. The addition of material to the subduction channel will thicken it and should tend to decrease shear stresses on the hanging wall and hence to diminish subduction erosion. Moreover, erosion can occur in one part of the channel at the same time accretion by underplating is occurring elsewhere. In fact, subduction erosion beneath forearc basins may commonly occur simultaneously with extensive offscraping and underplating at shallower levels.

Clearly, the relationship of sediment supply to channel capacity is a key factor governing whether subduction accretion or deep sediment subduction or even subduction erosion characterize a margin. The tendency for subduction accretion will increase as the sediment supply increases, whereas the proportion of deeply-subducted incoming sediment and the possibility of subduction erosion will increase as the supply decreases.

The primary factors governing channel capacity, and hence the ability of a margin to subduct sediment, are the speed of subduction and the pressure gradient along the channel. The faster the subduction speed and the smaller the pressure gradient, the larger the channel capacity. Thus, the capacity is greater for gentler descending-plate dips, gentler surface slopes, and lower-density overriding blocks (accretionary prism or continental crust as opposed to oceanic crust or mantle). No single factor necessarily dominates at any specific margin, because sediment supply, subduction speed, descending-plate dip, surface slope, and hanging-wall structure are all involved. We do not regard it as coincidental that steep plate dip, plentiful sediment supply, and an oceanic hanging wall characterize many of the Indonesian subduction zones that KARIG (1974a) and others have described as sites of subduction accretion, whereas gentle plate dip, sparse sediment supply, and a continental hanging wall characterize much of the South American margin that RUTLAND (1971) and many others have described as sites of substantial subduction erosion.

Approximation of Sediment in Subduction Channel as a Viscous Fluid

The incoming sediment pile is stratified, nonuniformly lithified, and mechanically anisotropic. In the zone of compression between the trench axis (or deformation front) and the inlet it is thickened by bulldozer-like longitudinal compression primarily by thrust imbrication and by folding. Tectonism enhances the dewatering and compaction of sediment due to burial and causes its bulk density to increase to about 2.2 Mg m^{-3}. Numerous faults form, with spacing ranging from the scale of grains (particulate flowage) to hundreds of meters (thrust sheets), and then become deactivated as porosity or fluid pressure or both diminish (KARIG, 1986; MOORE, J. C., and BYRNE, 1987). Continued convergence causes new faults to develop elsewhere in the sediment pile (see experiments by CARSON and BERGLUND, 1986), or new slip surfaces to develop nearby, forming thicker sheared zones (MOORE, J. C., and BYRNE, 1987), or deformation to become penetrative. The density of active and inactive faults and shear zones progressively increases toward the inlet. Concomitantly, the strong influence of bedding anisotropy in localizing deformation progressively decreases.

The low permeability of the hanging wall block (discussed below) severely restricts draining of sediment dragged past the inlet, so that fluid pressures become nearly lithostatic (see VON HUENE and LEE, 1983; and SHI and WANG, 1985; also MORROW *et al.*, 1984). The mechanical effect is that slip surfaces or shear zones deactivated within the zone of compression because of decreasing pore-fluid pressure can be reactivated. Reactivation of old faults and development of new ones causes the intraplate shearing to become distributed through much of the thickness of the channel. We envision that the active faults or shear zones are oriented largely subparallel to the channel walls and are spaced anywhere from millimeters in mud-rich zones to perhaps tens of meters in lithified sand-rich ones.

In striking contrast to the strong longitudinal compression trenchward of the inlet, deformation in the channel is primarily simple shear, which quickly rotates any tilted slip surfaces, such as faults or clay seams, into parallelism with the walls. Thus, even though deformation in the channel is not likely to be truly continuous, we envision it, even in the most discontinuous case, as consisting of slip on a multiplicity of wall-parallel faults or shear zones whose spacing is small in comparison to the channel thickness, and thus as being effectively continuous.

More importantly, we assume that in this effectively continuous deformation the bulk deviatoric stress is an increasing function of the bulk strain rate. Thus, we assume that the sediment in the channel deforms viscously, much as in slow-moving, nonturbulent earthflows (for a somewhat similar concept in the case of continental thrust belts see KEHLE, 1970, and ELLIOT, 1976). This assumption is vital to the subduction-channel model. Unfortunately, although extensive data exist on the engineering failure of materials resembling the subducted sediments, none exists on stresses and strain rates in large enough specimens after large simple-shear deformation under appropriate effective stresses, so a direct test of this assumption is not yet possible.

In implementing the subduction-channel model, we approximated the deforming sediment in the channel as a Newtonian-viscous fluid, in which the deviatoric stress increases linearly with strain rate. While computationally convenient, this particular approximation is not essential to the model. A power-law or other nonlinear law of viscosity could be used, or even a linear or nonlinear law of viscosity that applies only at stresses above a certain yield stress, as in a generalized Bingham material, could also be used (see IVERSON, 1985, for an example in weathered Franciscan melange). The essential requirement is the increase of deviatoric stress with strain rate.

The Newtonian-viscous approximation (and it is just that, a useful approximation) is thus the most reasonable choice in light of present knowledge. When better data become available, we can refine the calculations accordingly. In any case, although the quantitative results may change, the essential physics of the problem will not.

We selected the value 7×10^{17} Pa s (7×10^{18} poise) for the viscosity used in the calculations (SHREVE and CLOOS, 1986). Significantly higher values lead to too little offscraping, too much sediment subduction, implausibly thick channels, unreasonably large shear stresses on the channel walls, and too slow sinking of dense blocks, such as blueschist, in upwelling melange, whereas significantly lower values lead to the opposite disparities. The value adopted is comparable to other estimates for similar materials or conditions (see KEHLE, 1970; ENGLAND and HOLLAND, 1979; CLOOS, 1982; and EMERMAN and TURCOTTE, 1983). Somewhat different values would probably be appropriate for sites where the subduction speed is unusually fast or slow or the incoming sedimentary succession is unusually sand- or carbonate-rich.

Support for the viscous approximation comes from the ease with which the corner-flow model of CLOOS (1982) explains many of the perplexing relations that characterize the belt of blueschist-block-bearing, chaotically-mixed, mud-matrix melange in the Franciscan Complex of California. As already mentioned, similar melanges that differ only in their lack of blueschist blocks are found in many other accretionary complexes. Two interrelated effects cause the chaotic mixing of clasts in the upflowing fluid-rich, undercompacted melange. First, the melange undergoes immense shear strains that widely disperse originally adjacent particles, particularly where material turns back toward the inlet; and, second, blocks of different size and density sink or rise at different rates through the nonuniform flow field. Adventitious events, such as additions of seamounts from the descending plate and blueschists from the overriding one, can contribute further to the chaotic character of these melanges.

The subduction-channel model treats the bulk viscosity of the sediment in the channel as not varying with distance from the inlet. This approximation is consistent with the assumption that sediment within the channel down to depths of 20 to 30 km or more is deforming under conditions of constant volume and near-lithostatic pore-fluid pressures, and at such low temperatures (less than 150°C) that thermally-activated processes, such as dislocation creep, are minor. It is consistent with the conclusion that underplating occurs by loss of pore water and pore pressure at the hanging wall. In short, we believe that, where convergence has been continuous for a few tens of million years at speeds faster than about 20 km Ma^{-1} (2 cm yr^{-1}), the thermal effects on sediment within the channel are relatively minor down to depths of 30 km or more (see Figure 5). The nature and magnitude of thermally-activated changes in sediment viscosity are still poorly known. If such changes occur arcward of the global control point, the model will give only qualitative insights at deeper levels, although it will still give quantitative results at all shallower ones.

If heating and recrystallization cause bulk viscosity to increase, perhaps because of a change in the dominant deformation mechanism or an increase in grain size, the result will simply be enhanced coupling with the descending plate and all material will continue to go downward, with only an increase in the shear stresses on the channel walls. If, as seems probable, however, heating and recrystallization cause bulk viscosity to decrease, channel capacity will decrease. If, in turn, this causes a deeper control point to become operative, ductilely deformed, high-pressure metamorphic rock will upwell, creating a gyre. Rocks caught in the gyre will be intensely deformed, and their compositional layering and included clasts will be severely stretched and attenuated, much as in flowing ice or salt. Boudinage, folding and refolding, and transposition of layering will occur, and the deformation history will be exceedingly complex. This may be a common phenomenon deep within subduction zones (at depths of 30 to 40 km or more) and may explain the synmetamorphic ductile deformation and part of the uplift of some coherent blueschist terranes.

Permeability of Overriding Block

Pore-fluid pressures within the subduction channel are at or very close to lithostatic values. Water migration out of the channel, and hence underplating, requires passage through the hanging-wall block, in which, we assume, Darcy-type flow is the rate-controlling process. The key parameter is therefore the permeability of the overriding block. In weak rocks, such as sediments of an accretionary prism, water moves primarily through interconnected pores, rather than cracks. Permeability is typically on the order of microdarcies (BRYANT *et al.*, 1975; BRACE, 1980). In crystalline rocks and in well-cemented and lithified sedimentary ones, water moves primarily through cracks that tend to close and lose continuity as confining pressures become elevated above pore pressures (WALSH and BRACE, 1984). Although stray chips or offset asperities may help hold cracks open in crystalline rocks or well-lithified sedimentary ones when fluid pressure falls significantly below the least-compressive principal stress (WALSH, 1981), they also become less effective with increasing confining pressure. Furthermore, temperatures are higher and chemically active fluids are more important at depth. Plugging of pores and cracks by cements can reduce permeability (KRANTZ and BLACIC, 1984). The permeabilities of crystalline rocks are on the order of millidarcies near the surface and decrease rapidly with increased confining pressure (BRACE, 1980, 1984; WALSH and BRACE, 1984).

In short, we believe that, although crack permeability makes crystalline rocks more permeable than sedimentary ones at shallow levels of the overriding block, it rapidly diminishes at higher confining pressures. The pore permeability in accreted sediments, on the other hand, is mainly a function of density, and decreases much more modestly with depth.

Hydraulic fractures can develop in any rock type whenever the pore-fluid pressure exceeds the least compressive stress plus the tensile strength of the rock. Such fractures can enhance permeability as long as pore-fluid pressures remain great enough, particularly at the shallowest levels, where pore-fluid pressures significantly less than lithostatic may be sufficient to keep them open. The potential for hydraulic fracturing in the overriding block depends upon the state of stress, which varies with the type of margin (see *Hydraulic fracturing and dewatering of subducted sediment* in CLOOS and SHREVE, 1988, this volume).

Conclusion

A general theory for the tectonics of convergent plate margins must apply equally to margins like Mariana, which has a long-term history of complete sediment subduction and subduction erosion, and margins like the Lesser Antilles, which has a history of extensive sediment offscraping and growth of an enormous accretionary prism. Stratigraphic, structural, and petrologic observations made on outcrops or in Deep Sea Drilling Project cores at scales ranging from microscopic to hundreds of meters must be taken into account and integrated with geophysical observations made by seismic-reflection and other methods at scales ranging from kilometers to tens of kilometers. Patterns and rates of uplift and subsidence in the forearc must be related to the width of the zone of compression and active thickening. The large variation in subduction-zone seismicity from margin to margin and its relation to sediment thickness in the trench axis requires explanation, as does the generation of large bodies of chaotically-mixed melange and its relation to the type of overriding plate. The localization of large-scale sediment diapirism and the uplift of blueschist terranes must be accounted for in the model, and even related to the composition of arc magmas. The thermal history of a convergent margin and the pore-fluid pressures within subducting sediments must be considered. The widely cited imbricate-thrust model for accretion, even as subsequently modified to account for underplating and as quantified by Coulomb-wedge theory, directly addresses very few of these phenomena.

The subduction-channel model of SHREVE and CLOOS (1986) postulates that the basic processes are essentially the same at all convergent plate margins. It assumes that sediment dragged between the descending plate and overriding block deforms effectively as a viscous fluid and that its bulk rheology to depths of 30 km or more is essentially the same at all sites, provided convergence has been steady at a few tens of kilometers per million years (a few centimeters per year) for at least 10 to 20 million years. It not only predicts or explains a diverse array of geological and geophysical features of margins varying in character from Mariana to Lesser Antilles, as well as of ancient accretionary complexes, but it also quantifies the fundamental tectonic processes of offscraping, underplating, and deep sediment subduction. It shows that the broad spectrum of tectonic response to plate convergence depends upon the interplay of site-specific variables, such as the speed of subduction, the supply of sediment, the geometry of the descending plate, and the topography and structure of the overriding block. In Part 2 (CLOOS and SHREVE, 1988, this volume) we discuss the implications of the model for a wide range of geological, geophysical, and geochemical aspects of convergent plate margins.

Acknowledgments

T. Dumitru, M. Helper, S. R. Lipshie, J. C. Moore, W. R. Muehlberger, B. Reck, A. Salvador, and T. H. Shipley gave reviews and discussions that are particularly appreciated. Publication of this paper was supported in part by the National Science Foundation, grants EAR85–18149 (Cloos) and EAR85–17272 (Shreve). University of Texas Institute for Geophysics contribution 654; UCLA Institute of Geophysics and Planetary Physics publication 2818.

References

AALTO, K. R. (1982), *The Franciscan Complex of Northernmost California: Sedimentation and Tectonics*, Geological Society of London Special Publication *10*, 419–432.

ALLEGRE, C. J., HAMELIN, B., and DUPRE, B. (1984), *Statistical Analysis of Isotopic Ratios in MORB: The Mantle Blob Cluster Model and the Convective Regime of the Mantle*, Earth and Planetary Science Letters *71*, 71–84.

AOKI, Y., TAMANO, T., and KATO, S. (1983), *Detailed Structure of the Nankai Trough from Migrated Seismic Section*, American Association of Petroleum Geologists Memoir *34*, 309–324.

ARCULUS, R. J. and POWELL, R. (1986), *Source Component Mixing in the Regions of Arc Magma Generation*, Journal of Geophysical Research *91*, 5913–5926.

ARMSTRONG, R. L. (1971), *Isotopic and Chemical Constraints on Models of Magma Genesis in Volcanic Arcs*, Earth and Planetary Science Letters *12*, 137–142.

ARMSTRONG, R. L. (1981), *Radiogenic Isotopes: The Case for Crustal Recycling on a Near-Steady-State No-Continental-Growth Earth*, Philosophical Transactions of the Royal Society of London *A 301*, 443–472.

AUBOUIN, J., BOURGOIS, J., and AZEMA, J. (1984), *A New Type of Active Margin: The Convergent-Extensional Margin, as Exemplified by the Middle America Trench off Guatemala*, Earth and Planetary Science Letters *67*, 211–218.

BACHMAN, S. B. (1978), *A Cretaceous and Early Tertiary Subduction Complex, Mendocino Coast, Northern California*, In *Mesozoic Palegeography of the Western United States* (eds. Howell, D. G., and McDougall, K. A.) (Pacific Coast Paleogeography Symposium *2*) pp. 419–430.

BACHMAN, S. B. (1982), *The Coastal Belt of the Franciscan: Youngest Phase of Northern California Subduction*, Geological Society of London Special Publication *10*, 401–418.

BARREIRO, B. (1983), *Lead Isotopic Compositions of South Sandwich Island Volcanic Rocks and Their Bearing on Magmagenesis in Intra-Oceanic Island Arcs*, Geochimica et Cosmochimica Acta *47*, 817–822.

BECK, R. M. and LEHNER, P. (1974), *Oceans, New Frontier in Exploration*, American Association of Petroleum Geologists Bulletin *58*, 376–395.

BECKER, D. G. and CLOOS, M. (1985), *Melange Diapirs into the Cambria Slab: A Franciscan Trench Slope Basin near Cambria, California*, Journal of Geology *93*, 101–110.

BIJU-DUVAL, B., LE QUELLEC, P., MASCLE, A., RENARD, V., and VALLERY, P. (1982), *Multibeam Bathymetric Survey and High Resolution Seismic Investigations on the Barbados Ridge Complex (Eastern Caribbean): A Key to the Knowledge and Interpretation of an Accretionary Wedge*, Tectonophysics *86*, 275–304.

BIRD, P. (1978), *Stress and Temperature in Subduction Shear Zones: Tonga and Mariana*, Geophysical Journal of the Royal Astronomical Society *55*, 411–434.

BLOOMER, S. H. (1983), *Distribution and Origin of Igneous Rocks from the Landward Slopes of the Mariana Trench: Implications for Its Structure and Evolution*, Journal of Geophysical Research *88*, 7411–7428.

BOGGS, S. (1984), *Quaternary Sedimentation in the Japan Arc-Trench System*, Geological Society of America Bulletin *95*, 669–685.

BRACE, W. F. (1980), *Permeability of Crystalline and Argillaceous Rocks*, International Journal of Rock Mechanics and Mining Science and Geomechanics Abstracts *17*, 241–251.

BRACE, W. F. (1984), *Permeability of Crystalline Rocks: New In-Situ Measurements*, Journal of Geophysical Research *89*, 4327–4330.

BRANDON, M. T. (1986), *Comment and Reply on "Comments on the Growth of Accretionary Wedges,"* Geology *14*, 184–186.

BREEN, N. A., SILVER, E. A., and HUSSONG, D. M. (1986), *Structural Styles of an Accretionary Wedge South of the Island of Sumba, Indonesia, Revealed by SeaMARC II Side Scan Sonar*, Geological Society of America Bulletin *97*, 1250–1261.

BROWN, L., KLEIN, J., MIDDLETON, R., SACKS, I. S., and TERA, F. (1982), *^{10}Be in Island-Arc Volcanoes and Implications for Subduction*, Nature *229*, 718–720.

BRYANT, W. R., HOTTMAN, W., and TRABANT, P. (1975), *Permeability of Unconsolidated and Consolidated Marine Sediments, Gulf of Mexico*, Marine Geotechnology *1*, 1–14.

BYRNE, T. (1982), *Structural Evolution of Coherent Terranes in the Ghost Rocks Formation, Kodiak Island, Alaska*, Geological Society of London Special Publication *10*, 229–242.

BYRNE, T. (1984), *Early Deformation in Melange Terranes of the Ghost Rocks Formation, Kodiak Islands, Alaska*, Geological Society of America Special Paper *198*, 21–52.

CANDE, S. C. and LESLIE, R. B. (1986), *Late Cenozoic Tectonics of the Southern Chile Trench*, Journal of Geophysical Research *91*, 471–496.

CARLSON, W. D. and ROSENFELD, J. L. (1981), *Optical Determination of Topotactic Aragonite-Calcite Growth Kinetics: Metamorphic Implications*, Journal of Geology *89*, 615–638.

CARSON, B. (1977), *Tectonically Induced Deformation of Deep-Sea Sediments off Washington and Northern Oregon: Mechanical Consolidation*, Marine Geology *24*, 289–307.

CARSON, B. and BERGLUND, P. L. (1986), *Sediment Deformation and Dewatering under Horizontal Compression: Experimental Results*, Geological Society of America Memoir *166*, 135–150.

CARSON, B., VON HUENE, R., and ARTHUR, M. (1982), *Small-Scale Deformation Structures and Physical Properties Related to Convergence in Japan Trench Slope Sediments*, Tectonics *1*, 277–302.

CARSON, B., YUAN, J., MYERS, P. B., and BARNARD, W. D. (1974), *Initial Deep-Sea Sediment Deformation at the Base of the Washington Continental Slope: A Response to Subduction*, Geology *2*, 561–564.

CHAPPLE, W. M. (1978), *Mechanics of Thin-Skinned Fold-and-Thrust Belts*, Geological Society of America Bulletin *89*, 1189–1198.

CHASE, R. L. and BUNCE, E. T. (1969), *Underthrusting of the Eastern Margin of the Antilles by the Floor of the Western North Atlantic Ocean and the Origin of the Barbados Ridge*, Journal of Geophysical Research *74*, 1413–1420.

CHEN, A. T., FROHLICH, C., and LATHAN, G. V. (1982), *Seismicity of the Forearc Marginal Wedge (Accretionary Prism)*, Journal of Geophysical Research *87*, 3679–3690.

CHOPIN, C. (1984), *Coesite and Pure Pyrope in High Grade Blueschists of the Western Alps: A First Record and Some Consequences*, Contributions to Mineralogy and Petrology *86*, 107–118.

CHURCH, S. E. (1973), *Limits of Sediment Involvement in the Genesis of Orogenic Volcanic Rocks*, Contributions to Mineralogy and Petrology *39*, 17–32.

CLOOS, M. (1982), *Flow Melanges: Numerical Modeling and Geologic Constraints on Their Origin in the Franciscan Subduction Complex*, Geological Society of America Bulletin *93*, 330–345.

CLOOS, M. (1984), *Flow Melanges and the Structural Evolution of Accretionary Wedges*, Geological Society of America Special Paper *198*, 71–80.

CLOOS, M. (1985), *Thermal Evolution of Convergent Plate Margins: Thermal Modeling and Reevaluation of Isotopic Ar Ages for Blueschists in the Franciscan Complex of California*, Tectonics *4*, 421–433.

CLOOS, M. (1986), *Blueschists in the Franciscan Complex of California: Petrotectonic Constraints on Uplift Mechanisms*, Geological Society of America Memoir *164*, 77–93.

CLOOS, M. and SHREVE, R. L. (1988), *Subduction-Channel Model of Prism Accretion, Melange Formation, Sediment Subduction, and Subduction Erosion at Convergent Plate Margins: 2. Implications and Discussion*, Pure and Applied Geophysics *128, 3/4*, 501–545.

CLOWES, R. M., BRANDON, M. T., GREEN, A. G., YORATH, C. J., SUTHERLAND-BROWN, A., KANASEWICH, E. R., and SPENCER, C. (1987), *Lithoprobe—Southern Vancouver Island: Cenozoic Subduction Complex Imaged by Deep Seismic Reflections*, Canadian Journal of Earth Sciences *24*, 31–51.

COATS, R. R. (1962), *Magma Type and Crustal Structure in the Aleutian Arc*, American Geophysical Union Monograph *6*, 92–109.

COHEN, R. S. and O'NIONS, R. K. (1982), *Identification of Recycled Continental Material in the Mantle from Sr, Nd, and Pb Isotope Investigations*, Earth and Planetary Science Letters *61*, 73–84.

COLEMAN, R. G. (1980), *Tectonic Inclusions in Serpentinites, Société de Physique et d'Histoire Naturelle de Genève*, Archives des Sciences *33*, 89–102.

CONNELLY, W. (1978), *Uyak Complex, Kodiak Islands, Alaska: A Cretaceous Subduction Complex*, Geological Society of America Bulletin *89*, 755–769.

COUCH, R., WHITSETT, R., HUEHN, B., and BRICENO-GUARUPE, L. (1981), *Structures of the Continental Margin of Peru and Chile*, Geological Society of America Memoir *154*, 569–586.

COULBOURN, W. T. (1981), *Tectonics of the Nazca Plate and the Continental Margin of Western South America, 18° S to 23° S*, Geological Society of America Memoir *154*, 587–618.

COULBOURN, W. T. and MOBERLY, R. (1977), *Structural Evolution of Forearc Basins off Southern Peru and Northern Chile*, Canadian Journal of Earth Sciences *14*, 102–116.

COWAN, D. S. (1982), *Origin of "Vein Sturcture" in Slope Sediments on the Inner Slope of the Middle America Trench off Guatemala*, Initial Reports of the Deep Sea Drilling Project *67*, 645–650.

COWAN, D. S. (1985), *Structural Styles in Mesozoic and Cenozoic Melanges in the Western Cordillera of North America*, Geological Society of America Bulletin *96*, 451–462.

COWAN, D. S. and SILLING, R. M. (1978), *A Dynamic Scaled Model of Trenches and Its Implications for the Tectonic Evolution of Subduction Complexes*, Journal of Geophysical Research *83*, 5389–5396.

DAHLEN, F., SUPPE, J., and DAVIS, D. (1984), *Mechanics of Fold-and-Thrust Belts and Accretionary Wedges: Cohesive Coulomb Theory*, Journal of Geophysical Research *89*, 10,125–10,133.

DAVIDSON, J. P. (1986), *Isotopic and Trace Element Constraints on the Petrogenesis of Subduction-Related Lavas from Martinique, Lesser Antilles*, Journal of Geophysical Research *91*, 5943–5962.

DAVIDSON, J. P. (1987), *Crustal Contamination Versus Subduction Zone Enrichment: Examples from the Lesser Antilles and Implications for Mantle Source Compositions of Island Arc Volcanic Rocks*, Geochimica et Cosmochimica Acta *51*, 2185–2198.

DAVIES, T. A. and GORSLINE, D. S. (1976), *Oceanic Sediments and Sedimentary Processes*, Chemical Oceanography *5*, 1–80.

DAVIS, D., SUPPE, J., and DAHLEN, F. A. (1983), *Mechanics of Fold-and-Thrust Belts and Accretionary Wedges*, Journal of Geophysical Research *88*, 1153–1172.

DAVIS, D. M. and VON HUENE, R. (1987), *Inferences on Sediment Strength and Fault Friction from Structures at the Aleutian Trench*, Geology *15*, 517–522.

DEWEY, J. F. and BIRD, J. M. (1970), *Mountain Belts and the New Global Tectonics*, Journal of Geophysical Research *75*, 2625–2647.

DICKINSON, W. R. (1973), *Widths of Modern Arc-Trench Gaps Proportional to Past Duration of Igneous Activity in Associated Magmatic Arcs*, Journal of Geophysical Research *78*, 3395–3417.

DICKINSON, W. R. and SEELY, D. R. (1979), *Structure and Stratigraphy of Forearc Regions*, American Association of Petroleum Geologists Bulletin *62*, 2–31.

DICKINSON, W. R. and SNYDER, W. S. (1979), *Geometry of Subducted Slabs Related to San Andreas Transform*, Journal of Geology *87*, 609–627.

DRAPER, G. (1986), *Blueschists and Associated Rocks in Eastern Jamaica and Their Significance for Cretaceous Plate-Margin Development in the Northern Caribbean*, Geological Society of America Bulletin *97*, 48–60.

DRAPER, G. and BONE, R. (1980), *Denudation Rates, Thermal Evolution, and Preservation of Blueschist Terrains*, Journal of Geology *89*, 601–613.

ELLIOT, D. (1976), *The Motion of Thrust Sheets*, Journal of Geophysical Research *81*, 949–963.

EMERMAN, S. H. and TURCOTTE, D. L. (1983), *A Fluid Model for the Shape of Accretionary Wedges*, Earth and Planetary Science Letters *63*, 379–384.

ENGLAND, P. C. and HOLLAND, T. J. B. (1979), *Archimedes and the Tauren Eclogites: The Role of Buoyancy in the Preservation of Exotic Eclogite Blocks*, Earth and Planetary Science Letters *44*, 287–294.

ENGLAND, P. C. and THOMPSON, A. B. (1984), *Pressure-Temperature-Time Paths of Regional Metamorphism I. Heat Transfer during the Evolution of Regions of Thickened Continental Crust*, Journal of Petrology *25*, 894–928.

ERNST, W. G. (1970), *Tectonic Contact between the Franciscan Melange and Great Valley Sequence, Crustal Expression of a Late Mesozoic Benioff Zone*, Journal of Geophysical Research *75*, 886–902.

ERNST, W. G. (1971), *Do Mineral Parageneses Reflect Unusually High Pressure Conditions of Franciscan Metamorphism?*, American Journal of Science *271*, 879–914.

ERNST, W. G. (1975), *Systematics of Large-Scale Tectonics and Age Progressions in Alpine and Circum-Pacific Blueschist Belts*, Tectonophysics *26*, 229–246.

EVANS, B. and BROWN, E. H. (eds.) (1986), *Blueschists and Related Rocks*, Geological Society of America Memoir *164*, 423.

FARHOUDI, G. and KARIG, D. E. (1977), *Makran of Iran and Pakistan as an Active Arc System*, Geology *5*, 664–668.

FISHER, D. and BYRNE, T. (1987), *Structural Evolution of Underthrusted Sediments, Kodiak Islands, Alaska*, Tectonics *6*, 775–793.

FROHLICH, C., BILLINGTON, S., ENGDAHL, E. R., and MALAHOFF, A. (1982), *Detection and Location of Earthquakes in the Central Aleutian Subduction Zone Using Island and Ocean Bottom Seismograph Stations*, Journal of Geophysical Research *87*, 6853–6864.

FRYER, P., AMBOS, E. L., and HUSSONG, D. M. (1985), *Origin and Emplacement of Mariana Forearc Seamounts*, Geology *13*, 774–777.

GRIFFIN, J. J., WINDOM, H., and GOLDBERG, E. D. (1968), *The Distribution of Clay Minearals in the World Ocean*, Deep-Sea Research *15*, 433–459.

GROW, J. A. (1973), *Crustal and Upper Mantle Structure of the Central Aleutian Arc*, Geological Society of America Bulletin *84*, 2169–2192.

GILLULY, J. (1969), *Oceanic Sediment Volumes and Continental Drift*, Science *166*, 992–993.

GILLULY, J., REED, J. C., and CADY, W. M. (1970), *Sedimentary Volumes and Their Significance*, Geological Society of America Bulletin *81*, 353–376.

HABERMANN, R. E., MCCANN, W. R., and PERIN, B. (1986), *Spatial Seismicity Variations Along Convergent Plate Boundaries*, Geophysical Journal of the Royal Astronomical Society *85*, 43–68.

HAMILTON, W. (1979), *Tectonics of the Indonesian Region*, U.S. Geological Survey Professional Paper *1079*, 345.

HELWIG, J. and EMMET, P. (1981), *Structure of the Early Tertiary Orca Group in Prince William Sound and Some Implications for the Plate Tectonic History of Southern Alaska*, Journal of the Alaska Geological Survey *1*, 12–35.

HERRON, E. M., BRUHN, R., WINSLOW, M., and CHUAQUI, L. (1977), *Post Miocene Tectonics of the Margin of Southern Chile*, American Geophysical Union Maurice Ewing Series *1*, 273–284.

HIBBARD, J., and KARIG, D. E. (1987), *Sheath-Like Folds and Progressive Fold Deformation in Tertiary Sedimentary Rocks of the Shimanto Accretionary Complex, Japan*, Journal of Structural Geology *9*, 845–857.

HILDE, T. W. C. (1983), *Sediment Subduction Versus Accretion around the Pacific*, Tectonophysics *99*, 381–397.

HILL, M., MORRIS, J. and WHELAN, J. (1981), *Hybrid Granodiorites Intruding the Accretionary Prism, Kodiak, Shumagin, and Sanak Islands, Southwest Alaska*, Journal of Geophysical Research *86*, 10,591–10,606.

HOLE, M. J., SAUNDERS, A. D., MARRINER, G. F., and TARNEY, J. (1984), *Subduction of Pelagic Sediments: Implications for the Origin of Ce-Anomalous Basalts from the Mariana Islands*, Journal of the Geological Society of London *141*, 453–472.

HONDA, S. and UYEDA, S. (1983), *Thermal Process in Subduction Zone—A Review and Preliminary Approach on the Origin of Arc Volcanism*, In *Arc Volcanism: Physics and Tectonics* (eds. Shimozuru, D., and Yokoyama, I.) (Terra Scientific Publishing Company (TERRAPUB), Tokyo, 1983) pp. 117–140.

HSUI, A. T., MARSH, B. D., and TOKSÖZ, M. N. (1983), *On Melting of the Subducted Oceanic Crust: Effects of Subduction Induced Mantle Flow*, Tectonophysics *99*, 207–220.

HSUI, A. T. and TOKSÖZ, M. N. (1979), *The Evolution of Thermal Structures beneath a Subduction Zone*, Tectonophysics *60*, 43–60.

HUSSONG, D. M. and WIPPERMAN, L. K. (1981), *Vertical Movement and Tectonic Erosion of the Continental Wall of the Peru-Chile Trench near 11°30′ S Latitude*, Geological Society of America Memoir *154*, 509–524.

HUSSONG, D. M. and UYEDA, S. (1982), *Tectonic Processes and the History of the Mariana Arc: A Synthesis of the Results of Deep Sea Drilling Project Leg 60*, Initial Reports of the Deep Sea Drilling Project *60*, 909–929.

ISACKS, B. and BARAZANGI, M. (1977), *Geometry of Benioff Zones: Lateral Segmentation and Downards Bending of the Subducted Lithosphere*, American Geophysical Union Maurice Ewing Series *1*, 99–114.

ISACKS, B., OLIVER, J. and SYKES, L. R. (1968), *Seismology and the New Global Tectonics*, Journal of Geophysical Research *73*, 5855–5899.

IVERSON, R. M. (1985), *A Constitutive Equation of Mass-Movement Behavior*, Journal of Geology *93*, 143–160.

JARRARD, R. D. (1986), *Relations among Subduction Parameters*, Reviews of Geophysics *24*, 217–284.

JONES, G. M., HILDE, T. W. C., SHARMAN, G. F., and AGNEW, D. C. (1978), *Fault Patterns in Outer Trench Walls and Their Tectonic Significance*, Journal of the Physics of the Earth *26*, 85–101.

KARIG, D. E. (1974a), *Evolution of Arc Systems in the Western Pacific*, Annual Reviews of Earth and Planetary Sciences *21*, 51–75.

KARIG, D. E. (1974b), *Tectonic Erosion at Trenches*, Earth and Planetary Science Letters *21*, 209–212.

KARIG, D. E. (1977), *Growth Patterns on the Upper Trench Slope*, American Geophysical Union Maurice Ewing Series *1*, 175–186.

KARIG, D. E. (1983), *Deformation in the Forearc: Implications for Mountain Belts*, In *Mountain Building Processes* (ed. Hsu, K. J.) (Academic Press, London, 1983) pp. 59–72.

KARIG, D. E. (1986), *Physical Properties and Mechanical State of Accreted Sediments in the Nankai Trough, Southwest Japan Arc*, Geological Society of America Memoir *166*, 117–133.

KARIG, D. E. and KAGAMI, H. (1983), *Varied Responses to Subduction in Nankai Trough and Japan Trench Forearcs*, Nature *304*, 148–151.

KARIG, D. E. and KAY, R. W. (1981), *Fate of Sediments on the Descending Plate at Convergent Margins*, Philosophical Transactions of the Royal Society of London *A 301*, 233–251.

KARIG, D. E. and SHARMAN, G. F. (1975), *Subduction and Accretion in Trenches*, Geological Society of America Bulletin *86*, 377–389.

KARIG, D. E., CALDWELL, J. G., and PARMENTIER, E. M. (1976), *Effects of Accretion on the Geometry of the Descending Lithosphere*, Journal of Geophysical Research *81*, 6281–6291.

KARIG, D. E., CARDWELL, R. K., MOORE, G. F., and MOORE, D. G. (1978), *Late Cenozoic Subduction and Continental Truncation along the Northern Middle America Trench*, Geological Society of America Bulletin *89*, 265–276.

KARIG, D. E., MOORE, G. F., CURRAY, J. R., and LAWRENCE, M. B. (1980), *Morphology and Shallow Structure of the Lower Trench Slope off Nias Island, Sunda Arc*, American Geophysical Union Monograph *23*, 179–208.

KATZ, R. (1971), *Continental Margin in Chile—Is the Tectonic Style Compressional or Extensional?*, American Association of Petroleum Geologists Bulletin *55*, 1753–1758.

KAY, R. W. (1980), *Volcanic Arc Magmas: Implications of a Melting-Mixing Model for Element Recycling in the Crust-Upper Mantle*, Journal of Geology *88*, 497–522.

KEHLE, R. O. (1970), *Analysis of Gravity Sliding and Orogenic Translation*, Geological Society of America Bulletin *81*, 1641–1664.

KRANTZ, R. L. and BLACIC, J. D. (1984), *Permeability Changes during Time-Dependent Deformation of Silicate Rock*, Geophysical Research Letters *11*, 975–978.

KULM, L. D. and FOWLER, G. A., *Oregon Continental Margin Structure and Stratigraphy: A Test of the Imbricate-Thrust Model*, In *Geology of the Continental Margins* (eds. Burk, C. A., and Drake, C. L.) (Springer-Verlag, New York, 1974) pp. 261–283.

KULM, L. D., PRINCE, R. A., FRENCH, W., JOHNSON, S., and MASIAS, A. (1981), *Crustal Structure and Tectonics of the Central Peru Continental Margin and Trench*, Geological Society of America Memoir *154*, 445–468.

KULM, L. D., SCHEIDEGGER, K. F., PRINCE, R. A., DYMOND, J., MOORE, T. C., and HUSSONG, D. M. (1973), *Theoleiitic Basalt Ridge in the Peru Trench*, Geology *1*, 11–14.

LARUE, D. K. and HUDDLESTON, P. J. (1987), *Foliated Breccias in the Active Portuguese Bend Landslide Complex, California: Bearing on Melange Genesis*, Journal of Geology *95*, 407–422.

LASH, G. G. (1985), *Recognition of Trench Fill in Orogenic Flysch Sequences*, Geology *13*, 867–870.

LAY, T., KANAMORI, H., and RUFF, L. (1982), *The Asperity Model and the Nature of Large Subduction Zone Earthquakes*, Earthquake Prediction Research *1*, 3–71.

LEEMAN, W. P. (1983), *The Influence of Crustal Structure on Compositions of Subduction-Related Magmas*, Journal of Volcanology and Geothermal Research *18*, 561–588.

LEGGETT, J. K. (1985), *Deep-Sea Pelagic Sediments and Palaeo-Oceanography: A Review of Recent Progress*, Geological Society of London Special Publication *18*, 95–121.

LEGGETT, J., AOKI, Y., and TOBA, T. (1985), *Transition from Frontal Accretion to Underplating in a Part of the Nankai Trough Accretionary Complex off Shikoku (SW Japan) and Extensional Features of the Lower Slope*, Marine and Petroleum Geology *2*, 131–141.

LONSDALE, P. (1978), *Ecuadorian Subduction System*, American Association of Petroleum Geologists Bulletin *62*, 2454–2477.

LONSDALE, P. (1981), *Drifts and Ponds of Reworked Pelagic Sediment in Part of the Southwest Pacific*, Marine Geology *43*, 153–193.

LUDWIG, W. H., EWING, J. I., EWING, M., MURAUCHI, S., DEN, N., ASANO, S., HOTTA, H., HAYAKAWA, M., ASANUMA, T., ICHIKAWA, K., and NOGUCHI, I. (1966), *Sediments and Structure of the Japan Trench*, Journal of Geophysical Research *71*, 2121–2137.

LUNDBERG, N. (1983), *Development of Forearcs of Intraoceanic Subduction Zones*, Tectonics *2*, 51–61.

LUNDBERG, N. and MOORE, J. C. (1981), *Structural Features of the Middle America Trench Slope off Southern Mexico, Deep Sea Drilling Project Leg 66*, Initial Reports of the Deep Sea Drilling Project *66*, 793–814.

LUNDBERG, N. and MOORE, J. C. (1986), *Macroscopic Structural Features in Deep Sea Drilling Project Cores from Forearc Regions*, Geological Society of America Memoir *166*, 13–44.

MACKENZIE, J. S., NEEDHAM, D. T., and AGAR, S. M. (1987), *Progressive Deformation in an Accretionary Complex: An Example from the Shimanto Belt of Eastern Kyushu, Southwest Japan*, Geology *15*, 353–356.

MARSHAK, R. S. and KARIG, D. E. (1977), *Triple Junctions as a Cause for Anomalously Near-Trench Igneous Activity between the Trench and Volcanic Arc*, Geology *5*, 233–236.

MAXWELL, J. C. (1974), *Anatomy of an Orogen*, Geological Society of America Bulletin *85*, 1195–1204.

MATSUZAWA, A., TAMANO, T., AOKI, Y., and IKAWA, T. (1980), *Structure of the Japan Trench Subduction Zone, from Multi-Channel Seismic-Reflection Records*, Marine Geology *35*, 171–182.

MCCARTHY, J. and SCHOLL, D. W. (1985), *Mechanisms of Subduction Accretion along the Central Aleutian Trench*, Geological Society of America Bulletin *96*, 691–701.

MCKENZIE, D. P. (1969), *Speculations on the Consequences and Causes of Plate Motions*, Geophysical Journal of the Royal Astronomical Society *18*, 1–32.

MCLINTOCK, F. A. and WALSH, J. B. (1962), *Friction of Griffith Cracks in Rocks under Pressure*, In *Proceedings of the Fourth U.S. National Congress of Applied Mechanics, 2* (ed. Rosenberg, R. M.) (American Society of Mechanical Engineers, New York, 1962) pp. 1015–1021.

MCMILLEN, K. J. and BACHMAN, S. B. (1982), *Bathymetric and Tectonic Evolution of the Southern Mexico Active Margin, Deep Sea Drilling Project Leg 66*, Initial Reports of the Deep Sea Drilling Project *66*, 815–822.

MINSTER, J. B. and JORDAN, T. H. (1978), *Present Day Plate Motions*, Journal of Geophysical Research *83*, 5331–5354.

MOBERLY, R., SHEPHERD, G. L., and COULBOURN, W. T. (1982), *Forearc and Other Basins, Continental Margin of Northern and Southern Peru and Adjacent Ecuador and Chile*, Geological Society of London Special Publication *10*, 171–190.

MOORE, G. F. and KARIG, D. E. (1976), *Development of Sedimentary Basins on the Lower Trench Slope*, Geology *4*, 693–697.

MOORE, G. F. and KARIG, D. E. (1980), *Structural Geology of Nias Island, Indonesia: Implications of Subduction Zone Tectonics*, American Journal of Science *280*, 193–223.

MOORE, G. F. and CURRAY, J. R. (1980), *Structure of the Sunda Trench Lower Slope off Sumatra from Multichannel Seismic Reflection Data*, Marine Geophysical Researches *4*, 319–340.

MOORE, G. F., BILLMAN, H. G., HENHANUSSA, P. E., and KARIG, D. E. (1980), *Sedimentology and Paleobathymetry of Neogene Trench Slope Deposits, Nias Island, Indonesia*, Journal of Geology *88*, 161–180.

MOORE, J. C. (1975), *Selective Subduction*, Geology *3*, 530–532.

MOORE, J. C. and ALLWARDT, A. (1980), *Progressive Deformation of a Tertiary Trench Slope, Kodiak Islands, Alaska*, Journal of Geophysical Research *85*, 4741–4756.

MOORE, J. C., BIJU-DUVAL, B., BERGEN, J. A., BLACKINTON, G., CLAYPOOL, G. E., COWAN, D. S., DUENNEBIER, F., GUERRA, R. T., HEMLEBEN, C. H. J., HUSSONG, D., MARLOW, M. S., NATLAND, J. H., PUDSEY, C. J., RENZ, G. W., TARDY, M., WILLIS, M. E., WILSON, D., and WRIGHT, A. A. (1982c), *Offscraping and Underthrusting of Sediment at the Deformation Front of the Barbados Ridge: Deep Sea Drilling Project Leg 78A*, Geological Society of America Bulletin *93*, 1065–1077.

MOORE, J. C. and BYRNE, T. (1987), *Thickening of Fault Zones: A Mechanism of Melange Formation in Accreting Sediments*, Geology *15*, 1040–1043.

MOORE, J. C. and GEIGLE, J. E. (1974), *Slaty Cleavage: Incipient Occurrences in the Deep Sea*, Science *183*, 504–508.

MOORE, J. C. and KARIG, D. E. (1976), *Sedimentology, Structural Geology, and Tectonics of the Shikoku Subduction Zone*, Geological Society of America Bulletin *87*, 1259–1268.

MOORE, J. C. and LUNDBERG, N. (1986), *Tectonic Overview of Deep Sea Drilling Project Transects of Forearcs*, Geological Society of America Memoir *166*, 1–12.

MOORE, J. C. and SILVER, E. A. (1987), *Continental Margin Tectonics: Submarine Accretionary Prisms*, Reviews of Geophysics *25*, 1305–1312.

MOORE, J. C., MASCLE, A., TAYLOR, E., ANDREIEFF, P., ALVAREZ, F., BARNES, R., BECK, C., BEHRMANN, J., BLANC, G., BROWN, K., CLARK, M., DOLAN, J., FISHER, A., GIESKES, J., HOUNSLOW, M., MCLELLAN, P., MORAN, K., OGAWA, Y., SAKAI, T., SCHOONMAKER, J., VROLIJK, P., WILKENS, R., and WILLIAMS, C. (1987), *Expulsion of Fluids from Depth along a Subduction-Zone Décollement Horizon*, Nature *326*, 785–788.

MOORE, J. C., ROESKE, S., LUNDBERG, N., SCHOONMAKER, J., COWAN, D. S., GONZALES, E., and LUCAS, S. E. (1986), *Scaly Fabrics from Deep Sea Drilling Project Cores from Forearcs*, Geological Society of America Memoir *166*, 55–73.

MOORE, J. C., WATKINS, J. S., MCMILLEN, K. J., BACHMAN, S. B., LEGGETT, J. K., LUNDBERG, N., SHIPLEY, T. H., STEPHAN, J., BEGHTEL, F. W., BUTT, A., DIDYK, B. M., NITSUMA, N., SHEPHARD, L. E., and STRADNER, H. (1982b), *Facies Belts of the Middle America Trench and Forearc Region, Southern Mexico: Results from Leg 66, DSDP*, Geological Society of London Special Publication *10*, 77–94.

MOORE, J. C., WATKINS, J. S., SHIPLEY, T. H., MCMILLEN, K. J., BACHMAN, S. B., and LUNDBERG, N. (1982a), *Geology and Tectonic Evolution of a Juvenile Accretionary Terrane along a Truncated Convergent Margin: Synthesis of Results from Leg 66 of the Deep Sea Drilling Project, Southern Mexico*, Geological Society of America Bulletin *93*, 847–861.

MORROW, C. A., SHI, L. Q., and BYERLEE, J. D. (1984), *Permeability of Fault Gouge under Confining Pressure and Shear Stress*, Journal of Geophysical Research *89*, 3193–3200.

MOTTL, M. J. (1983), *Metabasalts, Axial Hot Springs, and the Structure of Hydrothermal Systems at Mid-Ocean Ridges*, Geological Society of America Bulletin *94*, 161–180.

MROZOWSKI, C. L. and HAYES, D. E. (1980), *A Seismic Reflection Study of Faulting in the Mariana Forearc*, American Geophysical Union Monograph *23*, 223–234.

MROZOWSKI, C. L., HAYES, D. E., and TAYLOR, B. (1982), *Multichannel Seismic Reflection Surveys of Leg 60 Sites, Deep Sea Drilling Project*, Initial Reports of the Deep Sea Drilling Project *60*, 57–69.

MYERS, J. D., FROST, C. D., and ANGEVINE, C. L. (1986), *A Test of Quartz Eclogite Source for Parental Aleutian Magmas: A Mass Balance Approach*, Journal of Geology *94*, 811–828.

NASU, N., VON HUENE, R., ISHIWADA, Y., LANGSETH, M., BURNS, T., and HONZA, E. (1980), *Interpretation of Multichannel Seismic Reflection Data, Legs 56 and 57, Japan Trench Transect, Deep Sea Drilling Project*, Initial Reports of the Deep Sea Drilling Project *57*, 489–503.

NILSEN, T. H. and ZUFFA, G. G. (1982), *The Chugach Terrane, a Cretaceous Trench-Fill Deposit, Southern Alaska*, Geological Society of London Special Publication *10*, 213–228.

O'NIONS, R. K. (1987), *Relationships between Chemical and Convective Layering in the Earth*, Journal of the Geological Society of London *144*, 259–274.

OXBURGH, E. R. and TURCOTTE, D. L. (1974), *Thermal Gradients and Regional Metamorphism in Overthrust Terrains with Special Reference to the Eastern Alps*, Schweizerische Mineralogische und Petrographische Mitteilungen *54*, 641–662.

PAGE, B. M. (1970), *Sur-Nacimiento Fault Zone of California—Continental Margin Tectonics*, Geological Society of America Bulletin *81*, 667–689.

PATCHETT, P. J., WHITE, W. M., FELDMANN, H., KIELINCZUK, S., and HOFMANN, A. W. (1984), *Hafnium/Rare Earth Element Fractionation in the Sedimentary System and Crustal Recycling into the Earth's Mantle*, Earth and Planetary Science Letters *69*, 365–378.

PEACOCK, S. M. (1987), *Thermal Effects of Metamorphic Fluids in Subduction Zones*, Geology *15*, 1057–1060.

PIPER, D. J. W., VON HUENE, R., and DUNCAN, J. R. (1973), *Late Quaternary Sedimentation in the Active Eastern Aleutian Trench*, Geology *1*, 19–22.

PLAFKER, G. (1965), *Tectonic Deformation Associated with the 1964 Alaska Earthquake*, Science *148*, 1675–1687.

PLATT, J. P. (1986), *Dynamics of Orogenic Wedges and the Uplift of High-Pressure Metamorphic Rocks*, Geological Society of America Bulletin *97*, 1037–1053.

PLATT, J. P., LEGGETT, J. K., YOUNG, J., RAZA, H., and ALAM, S. (1985), *Large-Scale Sediment Underplating in the Makran Accretionary Prism, Southwest Pakistan*, Geology *13*, 507–511.

PRINCE, R. A. and KULM, L. D. (1975), *Crustal Rupture and Initiation of Imbricate Thrusting in the Peru-Chile Trench*, Geological Society of America Bulletin *86*, 1639–1653.

RAYMOND, L. A. and TERRANOVA, T. (1984), *The Melange Problem—A Review*, Geological Society of America Special Paper *198*, 1–5.

RECK, B. H. (1987), *Implications of Measured Thermal Gradients for Water Movement through the Northeast Japan Accretionary Prism*, Journal of Geophysical Research *92*, 3683–3690.

RITGER, S. D. (1985), *Origin of Vein Structures in the Slope Deposits of Modern Accretionary Prisms*, Geology *13*, 437–439.

RUNDLE, J. B., KANAMORI, H., and MCNALLY, K. C. (1984), *An Inhomogeneous Fault Model for Gaps, Asperities, Barriers, and Seismicity Migration*, Journal of Geophysical Research *89*, 10,219–10,231.

RUTLAND, R. W. R. (1971), *Andean Orogeny and Ocean Floor Spreading*, Nature *233*, 252–255.

SAMPLE, J. C. and FISHER, D. M. (1986), *Duplex Accretion and Underplating in an Ancient Accretionary Complex, Kodiak Islands, Alaska*, Geology *14*, 160–163.

SAMPLE, J. C. and MOORE, J. C. (1987), *Structural Style and Kinematics of an Underplated Slate Belt, Kodiak and Adjacent Islands, Alaska*, Geological Society of America Bulletin *99*, 7–20.

SCHLIESTEDT, M. and MATTHEWS, A. (1987), *Transformation of Blueschist to Greenschist Facies Rocks as a Consequence of Fluid Infiltration, Sifnos (Cyclades), Greece*, Contributions to Mineralogy and Petrology *97*, 237–250.

SCHOLL, D. W. and MARLOW, M. S. (1974), *Sedimentary Sequences in Modern Pacific Trenches and the Deformed Circum-Pacific Eugeosyncline*, Society of Economic Paleontologists and Mineralogists Special Publication *19*, 193–211.

SCHOLL, S. W., MARLOW, M. S., and COOPER, A. K. (1977), *Sediment Subduction and Offscraping at Pacific Margins*, American Geophysical Union Maurice Ewing Series *1*, 199–210.

SCHOLL, D. W., VALLIER, T. L., and STEVENSON, A. J. (1982), *Sedimentation and Deformation in the Amlia Fracture Zone Sector of the Aleutian Trench*, Marine Geology *48*, 105–134.

SCHOLL, D. W., VALLIER, T. L., and STEVENSON, A. J. (1983), *Arc, Forearc, and Trench Sedimentation and Tectonics: Amlia Corridor of the Aleutian Ridge*, American Association of Petroleum Geologists Memoir *34*, 414–439.

SCHOLL, D. W., VON HUENE, R., VALLIER, T. L., and HOWELL, T. G. (1980), *Sedimentary Masses and Concepts about Tectonic Processes at Underthrust Ocean Margins*, Geology *8*, 564–568.

SCHWELLER, W. J. and KULM, L. D. (1978a), *Extensional Rupture of Oceanic Crust in the Chile Trench*, Marine Geology *28*, 271–291.

SCHWELLER, W. J. and KULM, L. D. (1978b), *Depositional Patterns and Channelized Sedimentation in Active Eastern Pacific Trenches*, In *Sedimentation in Submarine Canyons, Fans, and Trenches* (eds. Stanley, D. J., and Kelling, G.) (Dowden, Hutchison, and Ross, 1978) pp. 311–329.

SCHWELLER, W. J., KULM, L. D., and PRINCE, R. A. (1981), *Tectonics, Structure, and Sedimentary Framework of the Peru-Chile Trench*, Geological Society of America Memoir *154*, 323–349.

SEELY, D. R. (1977), *The Significance of Landward Vergence and Oblique Structural Trends on Trench Inner Slopes*, American Geophysical Union Maurice Ewing Series *1*, 187–198.

SEELY, D. R. (1979), *The Evolution of Structural Highs Bordering Major Forearc Basins*, American Association of Petroleum Geologists Memoir *29*, 245–260.

SEELY, D. R., VAIL, P. R., and WALTON, G. G., *Trench Slope Model*, In *Geology of the Continental Margins* (eds. Burk, C. A., and Drake, C. L.) (Springer-Verlag, New York, 1974) pp. 249–260.

SHEPHERD, G. L. and MOBERLY, R. (1981), *Coastal Structure of the Continental Margin, Northwest Peru and Southwest Ecuador*, Geological Society of America Memoir *154*, 351–392.

SHI, Y. and WANG, C. Y. (1985), *High Pore Pressure Generation in Sediments in Front of the Barbados Ridge Complex*, Geophysical Research Letters *12*, 773–776.

SHIPLEY, T. H. and MOORE, G. F. (1986), *Sediment Accretion, Subduction, and Dewatering at the Base of the Trench Slope off Costa Rica: A Seismic Reflection View of the Décollement*, Journal of Geophysical Research *91*, 2019–2028.

SHIPLEY, T. H., MCMILLEN, K. J., WATKINS, J. S., MOORE, J. C. SANDOVAL-OCHOA, J. H., and WORZEL, J. L. (1980), *Continental Margin and Lower Slope Structures of the Middle America Trench near Acapulco (Mexico)*, Marine Geology *35*, 65–82.

SHREVE, R. L. and CLOOS, M. (1986), *Dynamics of Sediment Subduction, Melange Formation, and Prism Accretion*, Journal of Geophysical Research *91*, 10,229–10,245.

SILLITOE, R. H. (1972), *Relation of Metal Provinces in Western America to Subduction of Oceanic Lithosphere*, Geological Society of America Bulletin *83*, 813–818.

SILVER, E. A. (1971), *Transitional Tectonics and Late Cenozoic Structure of the Continental Margin off Northernmost California*, Geological Society of America Bulletin *82*, 1–22.

SILVER, E. A., ELLIS, M. J., BREEN, N. A., and SHIPLEY, T. H. (1985), *Comments on the Growth of Accretionary Wedges*, Geology *13*, 6–9.

SPEED, R. C. and LARUE, D. K. (1982), *Barbados: Architecture and Implications for Accretion*, Journal of Geophysical Research *87*, 3633–3643.

SPENCE, W. (1987), *Slab Pull and the Seismotectonics of Subducting Lithosphere*, Reviews of Geophysics *25*, 55–69.

SPOONER, E. T. C. and FYFE, W. S. (1973), *Sub-Sea-Floor Metamorphism, Heat and Mass Transfer*, Contributions to Mineralogy and Petrology *42*, 287–304.

STAUDIGEL, H., HART, S. R., and RICHARDSON, S. H. (1981), *Alteration of Oceanic Crust: Processes and Timing*, Earth and Planetary Science Letters *52*, 311–327.

STEVENS, S. H. and MOORE, G. F. (1985), *Deformational and Sedimentary Processes in Trench Slope Basins of the Western Sunda Arc, Indonesia*, Marine Geology *69*, 93–112.

STOCKMAL, G. S. (1983), *Modeling of Large-Scale Accretionary Wedge Deformation*, Journal of Geophysical Research *88*, 8271–8287.

STRIDE, A. H., BELDERSON, R. H., and KENYON, N. H. (1982), *Structural Grain, Mud Volcanoes and Other Features on the Barbados Ridge Complex Revealed by GLORIA Long Range Side-Scan Sonar*, Marine Geology *49*, 187–196.

SUPPE, J. (1972), *Interrelationships of High-Pressure Metamorphism, Deformation and Sedimentation in Franciscan Tectonics, U.S.A.*, International Geological Congress, 24th Session, Section 3, 552–559.

SUPPE, J. (1973), *Geology of the Leech Lake—Ball Mountain Region, California*, University of California Publications in the Geological Sciences *107*, 82 p.

SUZUKI, T. (1986), *Melange Problem of Convergent Plate Margins in the Circum-Pacific Regions*, Memoirs of the Faculty of Science, Kochi University, Series E, Geology, *7*, 23–48.

TAIRA, A., OKADA, H., WHITAKER, J. H. M., and SMITH, A. J. (1982), *The Shimanto Belt of Japan: Cretaceous—Lower Miocene Active Margin Sedimentation*, Geological Society of London Special Publication *10*, 5–26.

TAJIMA, F. and KANAMORI, H. (1985), *Aftershock Area Expansion and Mechanical Heterogeneity of Fault Zone within Subduction Zones*, Geophysical Research Letters *12*, 345–348.

TERA, F., BROWN, L., MORRIS, J., SACKS, I. S., KLEIN, J., and MIDDLETON, R. (1986), *Sediment Incorporation in Island-Arc Magmas: Inferences from ^{10}Be*, Geochimica et Cosmochimica Acta *50*, 535–550.

THARP, T. M. (1985), *Numerical Models of Subduction and Forearc Deformation*, Geophysical Journal of the Royal Astronomical Society *80*, 419–437.

THORNBURG, T. M. and KULM, L. D. (1987), *Sedimentation in the Chile Trench: Depositional Morphologies, Lithofacies, and Stratigraphy*, Geological Society of America Bulletin *98*, 33–52.

TRAVIS, P. B. (1953), *LaBrea-Parinas Oil Field, Northwestern Peru*, American Association of Petroleum Geologists Bulletin *37*, 2093–2118.

UNDERWOOD, M. B. (1984), *A Sedimentologic Perspective on Stratal Disruption within Sandstone-Rich Melange Terranes*, Journal of Geology *92*, 369–386.

UNDERWOOD, M. B. (1986a), *Transverse Infilling of Central Aleutian Trench by Unconfined Turbidity Currents*, Geo-Marine Letter *6*, 7–13.

UNDERWOOD, M. B. (1986b), *Sediment Provenance within Subduction Complexes—An Example from the Aleutian Forearc*, Sedimentary Geology *51*, 57–73.

UNDERWOOD, M. B. and BACHMAN, S. B. (1982), *Sedimentary Facies Associations within Subduction Complexes*, Geological Society of London Special Publication *10*, 537–550.

UNDERWOOD, M. B. and KARIG, D. E. (1980), *Role of Submarine Canyons in Trench and Trench-Slope Sedimentation*, Geology *8*, 432–436.

UYEDA, S. (1982), *Subduction Zones: An Introduction to Comparative Subductology*, Tectonophysics *81*, 133–159.

UYEDA, S. and KANAMORI, H. (1979), *Back-Arc Opening and the Mode of Subduction*, Journal of Geophysical Research *84*, 1049–1061.

VAN DEN BEUKEL, J. and WORTEL, R. (1986), *Thermal Modelling of Arc-Trench Regions*, Geologie en Mijnbouw *65*, 133–143.

VON HUENE, R. (1981), *Review of Early Results from Drilling of the IPOD-1 Active Margin Transects across Japan, Mariana, and Middle America Convergent Margins*, Society of Economic Paleontologists Special Publication *32*, 57–66.

VON HUENE, R. (1984a), *Tectonic Processes along the Front of Modern Convergent Margins—Research of the Past Decade*, Annual Reviews of Earth and Planetary Sciences *12*, 359–381.

VON HUENE, R. (1984b), *Structural Diversity along Modern Convergent Margins and the Role of Overpressured Pore Fluids in Subduction Zones*, Bulletin de la Société Géologique de France, Series *7*, *26*, 207–219.

VON HUENE, R. (1986a), *To Accrete or Not Accrete, That Is the Question*, Geologische Rundschau *75*, 1–15.

VON HUENE, R. (1986b), *Seismic Images of Modern Convergent Margin Tectonic Structure*, American Association of Petroleum Geologists, Studies in Geology *26*, 60.

VON HUENE, R. and KULM, L. D. (1973), *Tectonic Summary of Leg 18*, Initial Reports of the Deep Sea Drilling Project *18*, 961–976.

VON HUENE, R. and LEE, H. (1983), *The Possible Significance of Pore Fluid Pressure in Subduction Zones*, American Association of Petroleum Geologists Memoir *34*, 781–791.

VON HUENE, R., AUBOUIN, J., AZEMA, J., BLACKINTON, G., CARTER, J. A., COULBOURN, W. T., COWAN, D. S., CURIALE, J. A., DENGO, C. A., FASS, R. W., HARRISON, W., HESSE, R., HUSSONG, D. M., LADD, J. W., MUZYLOV, N., SHIKI, T., THOMPSON, P. R., and WESTBERG, J. (1980), *Leg 67: the Deep Sea Drilling Project Mid-America Trench Transect off Guatemala*, Geological Society of America Bulletin *91*, 421–432.

VON HUENE, R., KULM, L. D., and MILLER, J. (1985), *Structure of the Frontal Part of the Andean Convergent Margin*, Journal of Geophysical Research *90*, 5429–5442.

VON HUENE, R., LANGSETH, M., NASU, N., and OKADA, H. (1982), *A Summary of Cenozoic Tectonic History along IPOD Japan Trench Transect*, Geological Society of America Bulletin *93*, 829–846.

VON HUENE, R., SUESS, E., and EMEIS, K. (1987), *Convergent Tectonics and Coastal Upwelling: A History of the Peru Continental Margin*, Episodes *10*, 87–93.

WALSH, J. B. (1981), *Effect of Pore Pressure and Confining Pressure on Fracture Permeability*, International Journal of Rock Mechanics and Mining Science and Geomechanics Abstracts *18*, 429–435.

WALSH, J. B. and BRACE, W. F. (1984), *The Effect of Pressure on Porosity and the Transport Properties of Rock*, Journal of Geophysical Research *89*, 9425–9431.

WANG, C. Y. and SHI, Y. L. (1984), *On the Thermal Structure of Subduction Complexes: A Preliminary Study*, Journal of Geophysical Research *89*, 7709–7718.

WARSI, W. E. K., HILDE, T. W. C., and SEARLE, R. C. (1983), *Convergence Structures of the Peru Trench between 10°S and 14°S*, Tectonophysics *99*, 313–329.

WEAVER, B. L., WOOD, D. A., TARNEY, J., and JORON, J. L. (1986), *Role of Subducted Sediment in the Genesis of Ocean-Island Basalts: Geochemical Evidence from South Atlantic Ocean Islands*, Geology *14*, 275–278.

WESTBROOK, G. K. and SMITH, M. J. (1983), *Long Décollements and Mud Volcanoes: Evidence from the*

Barbados Ridge Complex for the Role of High Pore-Fluid Pressure in the Development of an Accretionary Complex, Geology *11*, 279–283.

WESTBROOK, G. K., SMITH, M. J., PEACOCK, J. H., and POULTER, M. J. (1982), *Extensive Underthrusting of Undeformed Sediment beneath the Accretionary Complex of the Lesser Antilles Subduction Zone*, Nature *300*, 625–628.

WHITE, R. S. (1982), *Deformation of the Makran Accretionary Sediment Prism in the Gulf of Oman (Northwest Indian Ocean)*, Geological Society of London Special Publication *10*, 357–372.

WHITE, R. S. and KLITGORD, K. (1976), *Sediment Deformation and Plate Tectonics in the Gulf of Oman*, Earth and Planetary Science Letters *32*, 199–209.

WHITE, W. M., and DUPRE, B. (1986), *Sediment Subduction and Magma Genesis in the Lesser Antilles: Isotopic and Trace Element Constraints*, Journal of Geophysical Research *91*, 5927–5941.

WILLIAMS, P. R., PIGRAM, C. J., DOW, D. B., and AMIRUDDIN (1984), *Melange Production and the Importance of Shale Diapirism in Accretionary Terrains*, Nature *309*, 145–146.

YARDLEY, B. W. D. (1982), *The Early Metamorphic History of the Haast Schists and Related Rocks of New Zealand*, Contributions to Mineralogy and Petrology *81*, 317–327.

YEATS, R. S. (1968), *Southern California Structure, Sea Floor Spreading, and the History of the Pacific Basin*, Geological Society of America Bulletin *79*, 1693–1702.

ZIEGLER, A. M., BARRETT, S. F., and SCOTESE, C. R. (1981), *Palaeoclimate, Sedimentation and Continental Accretion*, Philosophical Transactions of the Royal Society of London *A 301*, 253–264.

ZINDLER, A., STAUDIGEL, H., and BATIZA, R. (1984), *Isotopic and Trace Element Geochemistry of Young Pacific Seamounts: Implications for the Scale of Upper Mantle Heterogeneity*, Earth and Planetary Science Letters *70*, 175–195.

(Received August 15, 1986, revised February 23, 1988, accepted March 18, 1988)