PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Fate of Sediments on the Descending Plate at Convergent Margins [and Discussion]

D. E. Karig, R. W. Kay, G. C. Brown and R. L. Armstrong

Phil. Trans. R. Soc. Lond. A 1981 **301**, 233-251
doi: 10.1098/rsta.1981.0108

Phil. Trans. R. Soc. Lond. A **301**, 233–251 (1981)
Printed in Great Britain

Fate of sediments on the descending plate at convergent margins

By D. E. Karig and R. W. Kay
Department of Geological Sciences, Cornell University, Ithaca, New York 14853, *U.S.A.*

The evolution of the continents and of continental crust is strongly dependent on the trajectory of the sedimentary cover on the descending oceanic lithosphere at arc systems. Although direct calculations of accretion are not reliable, indirect evidence strongly suggests that most of the sediment cover is either accreted or underplated (subcreted) to the upper plate. This evidence includes the increased thickening of accretionary prism beneath parts of the inner trench slope that cannot be explained by deformation within the prism and by protolith composition both in subophiolitic metamorphics and in blueschist terrains. That a small fraction of this sediment cover is transported to depths of at least 100 km is demonstrated in several ways. Flux calculations of mass and selected elements through arc systems require addition of a few tens of metres of sediment to the arc magmas. Global correlations of variations between arc magma characteristics and regional geological parameters show: (1) a strong correlation between silica content of average arc magmas and thickness or maturity of crust in the upper plate, attributed to upper plate contamination; (2) regional variations in $^{87}Sr/^{86}Sr$ and Pb isotopic ratios of arc basalts that correlate spatially with isotopic ratios in the non-calcareous components of pelagic sediments. This correlation is argued to reflect the variation of terrigenous material in the basal pelagics that are involved in magma production. Subduction of continental crust to depths of 100 km, either as part of the descending plate or by tectonic erosion of the upper plate is not supported by these correlations. Recycling rates of continental crust by accretion and subcretion and of mantle by subduction of oceanic lithosphere in contemporary arcs are very large compared with the growth rate of continental crust, which appears similar to the magma production rates in arc systems. This present production rate of continental crust is very much smaller than that during early Earth history, and is compatible with Phanerozoic ocean freeboard changes. In addition, average SiO_2 and K_2O contents of contemporary arc magmas are much lower than those estimated for mean continental crust, leading to the conclusion that the magmas being produced at active arcs cannot be used as a model for the development of most of the Earth's continental areas.

Introduction

One of the most significant geological corollaries of the plate tectonic theory is that vast areas of oceanic crust and its sediment cover have disappeared along convergent plate margins. The structural processes by which material responds to that convergence, and the fate of the various components of the descending plate, are of interest from a number of contemporary perspectives. One of the most fundamental of these is the role that the processes at consuming margins play in the evolution of continental crust.

The broad spectrum of opinion concerning the rate of mode of evolution of the continental crust through time (see, for example: Armstrong 1971; Moorbath 1978) in large part is due to disagreement concerning the degree to which sediments carried into ocean basins by erosion and returned to trenches might be carried to depths where they can be reincorporated into the mantle.

Some approaches to the question of continental evolution have been direct. For instance the dating and areal distribution of crustal basement rocks have yielded information on the areal extent of rocks that have survived from various past times to the present. In contrast, other approaches have been indirect, relying on the temporal evolution of crustal radiogenic isotope systems, and continental freeboard (relative sea level) arguments to trace continental growth. Relatively less has been done with data from contemporary convergent margins, attempting to deduce mass trajectories and fluxes at sites of active crustal formation.

At present, studies in various convergent margins have led to different, often contradictory conclusions, all of which might be valid in different arc settings. For instance, data suggest accretion of most of the sediment cover of the descending plate at shallow levels in the western Sunda arc (Karig *et al.* 1980), whereas most workers conclude that continental crust is being actively removed from the upper plate along sections of the Peru–Chile trench (Hussong *et al.* 1976; Kulm *et al.* 1977). Similarly, a wide divergence of opinion concerning the role of subducted sediment and oceanic crust in arc magmatism has existed since Coats (1962) drew attention to the possibility of sediment sources for arc volcanics.

Our purpose here is to review recent data that concern the fate of the sediment cover on the descending plate as far as it affects the large-scale transport of material at convergent margins and thus the evolution of continental crust. Relevant evidence can be drawn from the geometry and distribution of deformation within accretionary prisms, from lithologic variations within exposed accretionary complexes, and from correlation of variations in the geochemistry of arc volcanics with critical geological variations on a global scale.

Transport trajectories of subducted sediment: kinematic evidence

The most direct approach to the problem of material transfer from the descending plate to the upper plate (or the opposite) would be to compare the influx of sediment to the trench with the increase in size of the accretionary prism over some definable time interval. However, the lack of control of input parameters and boundary conditions, as well as the wide range of assumptions or interpretations of available data, not to mention the likelihood that different arcs have different tectonic responses, have produced inconclusive and contradictory results (see, for example: Karig & Sharman 1975; von Huene 1972). More useful information can presently be obtained by indirectly approaching the problem through an integrated interpretation of the geophysical nature of the leading edge of the upper plate and the geological character of exposed accretionary complexes.

Behaviour at the leading edge of the upper plate

There is no doubt that the lower part of the sediment cover on the descending oceanic plate can be carried beneath the leading edge of the upper plate. The fraction of this subducted sediment seems to be governed by a zone of maximum porosity and minimum shear strength that usually occurs near the base of the trench-fill turbidites, where these exist (Karig 1974; Moore 1975). However, the subducted sediment must pass beneath a flatter shallow section of the plate interface, which may be up to several hundred kilometres wide, before reaching mantle depths (figure 1). Some recent analyses (see, for example, Armstrong, this symposium) appear to ignore the possibility of transfer of sediment and even oceanic crustal rocks to the base of the upper plate beneath this shallow part of the interface. It is such a process, for which we have coined the term subcretion, that is of critical importance in determining the degree to which sediment is recycled to the mantle.

Seismic observation of subcreted material appears to be difficult. For instance, several processed multi-channel seismic reflexion profiles have been interpreted as showing detached or upthrust segments of oceanic crust above the surface of the descending oceanic basement beneath the accretionary prism (see, for example, von Huene 1979), but differentiation of thrust blocks from normal faults or from original basement topography on the basis of single seismic profiles is at best highly subjective. Many refraction profiles across inner trench slopes show large volumes of material with compressional velocities of 4.5–5.5 km/s at the base of the accretionary prism, but this material could either be accreted basaltic crust or dewatered and metamorphosed sediment (see, for example, Kieckhefer *et al.* 1980).

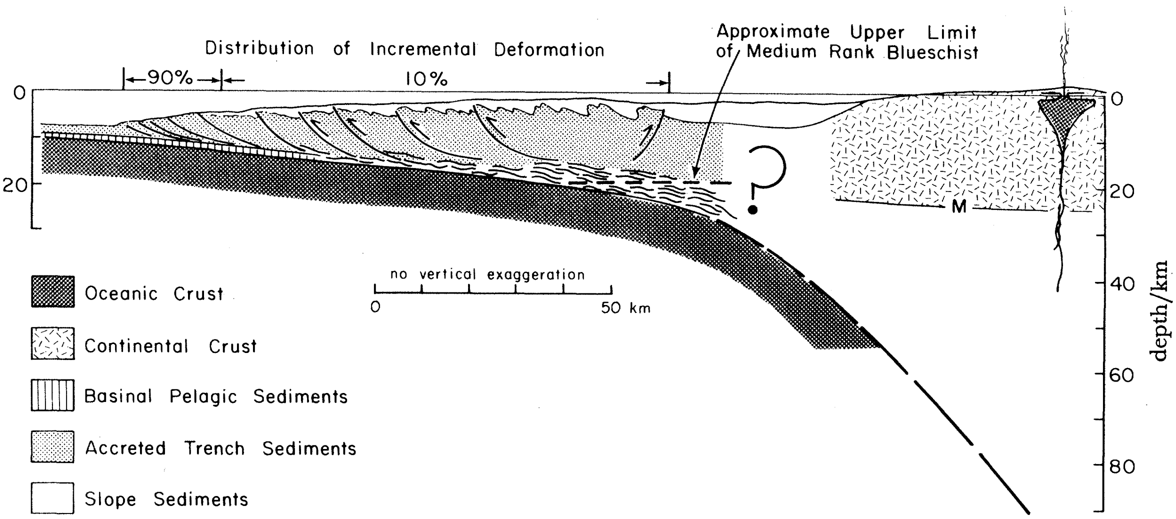

FIGURE 1. True-scale section of an accreting trench slope, illustrating the need for subcretion to explain the relation between the geometry of the accretionary prism and the distribution of deformation across the slope.

Geometry cf. *deformation of the accretionary prism*

Subcretion is more firmly suggested by the arcward thickening of accretionary prisms in excess of that which can be explained by the horizontal shortening of sediments accreted at the base of the trench slope. Deformation in most arcs is concentrated near the base of the inner trench slope (figure 1). Field mapping of structures on the emergent trench slope break of the Sunda arc and reflexion profiles across that inner slope lead to the conclusion that less than 10% of the interplate convergence is absorbed by deformation more than 20 km from the base of the trench slope (Karig *et al.* 1980).

The accretionary prism, as outlined by refraction and seismological data, increases in thickness continuously at least as far as the trench slope break (figure 1). That section of the Sunda accretionary prism between 20 and 80 km from the trench, and beneath the slowly deforming section of the lower slope, thickens from 8 to 20 km. This 150% average thickening is significantly more than can be explained by the shortening of the overlying material. A similar arcward thickening of other accretionary prisms implies that subcretion is a common process.

Subophiolite metamorphic protoliths

Corroborative geological evidence for subcretion can be extracted from the protolith assemblages of the metamorphic rocks found beneath ophiolite sheets and within accretionary complexes. In both the Bay of Islands (Williams & Smyth 1973; Malpas 1979) and the Semail (Alleman & Peters 1972; Reinhardt 1969) ophiolites, for which the most extensive data exist, the tectonized metamorphic zone is at a low angle to the base of the ophiolitic crust, generally lying between 3 and 6 km into the mantle. Both reconstruction of the overlying crust and sediment cover (assumed to be no more than several kilometres thick) and mineralogical phase assemblages (Ghent & Stout 1979; Malpas 1979) indicate that the depth to this shallow dipping part of the plate interface was in the vicinity of 10 km below the sediment interface. This depth to the plate interface in young active arcs occurs beneath the lower trench slope, generally not further than 30–40 km from the trench.

Of interest to the problem of subcretion is the dominance, within the metamorphic zone, of pelagic sediments and mafic igneous protoliths, to the near exclusion of trench-fill sediments (see, for example, Woodcock & Robertson 1977). The volcaniclastics that form a large percentage of the metamorphic zones beneath the ophiolites of western Newfoundland (Williams & Smyth 1973; Malpas 1979) are readily interpreted as segments of the volcaniclastic aprons that floor the forward parts of oceanic back-arc basins. The presence of mafic igneous and oceanic sedimentary protoliths beneath the ophiolites is evidence not only that these lithologies were subducted to depths of at least 10 km, but also that they have been transferred from the descending plate to the base of the upper plate.

Blueschist protoliths

Some indication of how much sediment is carried to deeper levels along the plate interface can be gleaned from the distribution of lithologies within emergent and eroded subduction complexes where metamorphic mineral assemblages can be used as palaeodepth indicators. Protolith compositions of high-pressure facies metamorphic rocks in subduction complexes have received only cursory and qualitative attention. Variations in protolith composition could result not only from tectonic gradients but also from temporal and spatial variations of the subduction process under different arc situations. Ideally only rocks of different metamorphic grade in a single area should be compared, but even where such a condition is approached, as in the Franciscan complex, there is little doubt that tectonic transport subsequent to metamorphism has occurred (see, for example, Ernst 1975, p. 10).

Despite these provisos, there seems to be a common and consistent pattern in blueschist-bearing subduction complexes that we interpret to be a reflection of conditions along the interplate zone of shear. Protoliths of the highest-pressure metamorphics are dominantly pelagic sediments and mafic to ultramafic igneous rocks, whereas lower-pressure metamorphics have larger components of greywackes and other coarse clastics having an arc provenance. Examples showing this progression include the Franciscan complex (Coleman & Lee 1963; Ernst *et al.* 1970), the Kodiak–Kenai area of Alaska (Carden *et al.* 1977; Cowan & Boss 1978) and New Caledonia (Brothers & Blake 1973). Other complexes, such as the Torlesse of New Zealand (see, for example, Landis & Bishop 1972) and the Sambagawa of Japan (see, for example, Ernst *et al.* 1970), do not show such a clear correlation, but neither do they display the higher-pressure metamorphic assemblages.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A

As demonstrated by studies in the Franciscan complex (see, for example, Moore & Liou 1979), greywacke and other coarse clastics that are probably trench sediments are commonly metamorphosed to the jadeite–quartz–lawsonite assemblage, often with surprisingly little megascopic deformation. However, the more strongly deformed metamorphosed and recrystallized blueschists, typified by types III and IV rocks of Coleman & Lee (1963), and felt to represent even-higher-pressure environments (Coleman & Lanphere 1971), are very much dominated by metapelagic and mafic to ultramafic metaigneous rocks. From the literature it appears that blueschists of the highest grade and ecologites are dominantly to exclusively composed of oceanic protoliths (see, for example, Coleman & Lamphere 1971).

Ernst (1975) and others have previously appealed to some form of subcretion or 'underplating' to transport material to depth within the accretionary prism, but we would like to draw two inferences pertinent to subcretion from the proposed correlation between protolith composition and depth of metamorphism. First, the jadeite–quartz–lawsonite-bearing metaclastics in the Franciscan complex and in several other subduction complexes are clear evidence that trench and/or slope sediments can be brought at least to depths around 20 km. Accretionary prisms exceed a thickness of 20 km beneath or slightly behind the trench slope break (see, for example, Westbrook 1975; Kieckhefer *et al.* 1980). The metaclastics could be transported to those depths either by thrust-thickening of accreted sediments or by subcretion. The intense deformation of subcreted metamorphics and melange beneath ophiolites suggests to us that the little-deformed Franciscan medium-rank blueschists might have been accreted and structurally depressed rather than subcreted.

A second inference, derived from the dominance of protoliths representing the oceanic sediment cover, crust and upper mantle in the highest-grade blueschists and eclogites, is that this not only demonstrates that oceanic sediments can be subducted to depths approaching 30 km, but that they, and pieces of crust and upper mantle, can be subcreted at those depths (figure 1). Moreover, a decrease of both trench-fill and pelagic sediments with the increase in depth of metamorphism suggests that most of the sediment cover on the descending plate is accreted or subcreted at depths less than 30 km. For depths greater than about 30 km in arc systems, our next observation of material that may be derived from near the subducting slab is from the composition of arc magmas.

Transportation of subducted sediment to depths of 100 km: evidence from arc magmas

Studies of arc volcanics have arrived at a postulated proportion of sediment cf. source magma, whether that be the subducted slab or the overlying mantle wedge. Much less has been learned about the total sediment volume involved or the percentage of sediment fed to the trench that is ultimately fused. Although the former figure is reasonably constrained by regional pelagic stratigraphy and by finite plate motions, we have much poorer control on the latter figure, due to our lack of knowledge about magma production rates.

Magma production rates

More reliable estimates of the rates of arc volcanism range from 3 to less than 15 km²/Ma per kilometre of arc (table 1). A few active arcs, as the segments of the Andean arc with shallow dipping seismic zones, have no associated volcanism. Estimates of production rates in continental arcs are hampered by problems of erosion and basement geometry. Those of oceanic arcs are conceptually more accurate, but the volcanic geometries and porosity variations are only now being adequately outlined. The better-constrained values for the Mariana and Lesser Antilles arcs are probably sufficiently accurate to show a positive correlation between the rates of subduction and eruption, at least among similar arc types.

Rates of intrusion are even more difficult to establish. Individual concentrically zoned plutons, which probably correspond to a single major volcanic centre, are generally thought to be tear-drop- or tadpole-shaped diapirs (see, for example: Hopson *et al.* 1970; Gastil 1975). Mapping of individual plutons (Pitcher 1978; Citron 1980) suggests a diameter near 10 km at shallow depths. Disappearance of the diapiric roots in migmatite zones at depths of the order of 15 km further suggests that the bulk of these intrusives lies at significantly shallower depths.

Table 1. Rates of magmatism and of plate convergence in active arc systems

arc	time span	extrusion rate†/(km^3/Ma)	subduction rate (normal component)/(cm/a)	reference
Mariana	0–5 Ma	10	9	this study; Sample (1980)
Central America	0–0.4 Ma	7	8	Bice (1980); Stoiber & Can (1973)
Japan	'Quaternary'	> 4	8.7	Sugimura *et al.* (1973)
Caribbean	0–0.1 Ma	4	1.7	Sigurdsson *et al.* (1980) with modified porosity values
Cascades	0–10 Ma	4.5	2	McBirney (1976)
central Andes	0–6 Ma	2.9	9.2	Francis & Rundle (1976)
Kurile‡	0–33 a	45	8	Markhinin (1962)

† Includes flows, pyroclastic material and ash.
‡ The time span represented by this observation is probably too short to be representative.

The duration of intrusion at a single diapiric centre has been assumed to be that of the associated volcanic activity. In arcs such as the New Hebrides and Mariana, this appears to be several millions of years. If we model the diapiric intrusives as cones 10 km high and 10 km in diameter, having a spacing of 70 km (Marsh 1979) and an active period of 3 Ma, we arrive at intrusion rates of less than 2 km^3/Ma per kilometre of arc length. This compares well with the 3.1 km^3/Ma calculated by Everden & Kistler (1970) for a 130 Ma period in the Sierra Nevada batholith and with a 2.9 km^3/(km Ma) figure for one model of the coastal batholith of Peru (Francis & Rundle 1976). These estimates lead to a working assumption that rates of intrusion are similar to those of extrusion. The magma production rate that we used (table 1) reflects volcanic activity during the past 10 Ma or less and is higher than a long-term average if one accepts the global volcanic pulses suggested by Kennett *et al.* (1977), McBirney (1978) and others.

Geochemical balances through arc systems

Mass-balance calculations can be carried out for the input and output to such well studied arcs as the Marianas (figure 2). More than 500 km^3 of igneous oceanic crust and perhaps 50 km^3 of sediment is being transported to the arc per million years. Both of these figures far exceed the total mass of erupted material or even the total igneous mass added to the arc. The convective flux of mantle overlying the down-going slab (also a possible source for arc magmas) is also large compared to the igneous output of the arc. It appears that even inefficient

(< 10%) extraction of melt from any or all of these sources can account for the total mass. Thus, total mass calculations provide no constraints for the origin of the arc magmas or for the fate of the sediment column.

The contrasting compositions of input and output to the arc implies that some elements, such as potassium, are in much shorter supply than others. Even though the K_2O content of the outputs is higher than that of the oceanic crust, that source as well as the sediment or mantle wedge above the descending plate could supply sufficient K_2O to account for the K_2O in Mariana arc lavas.

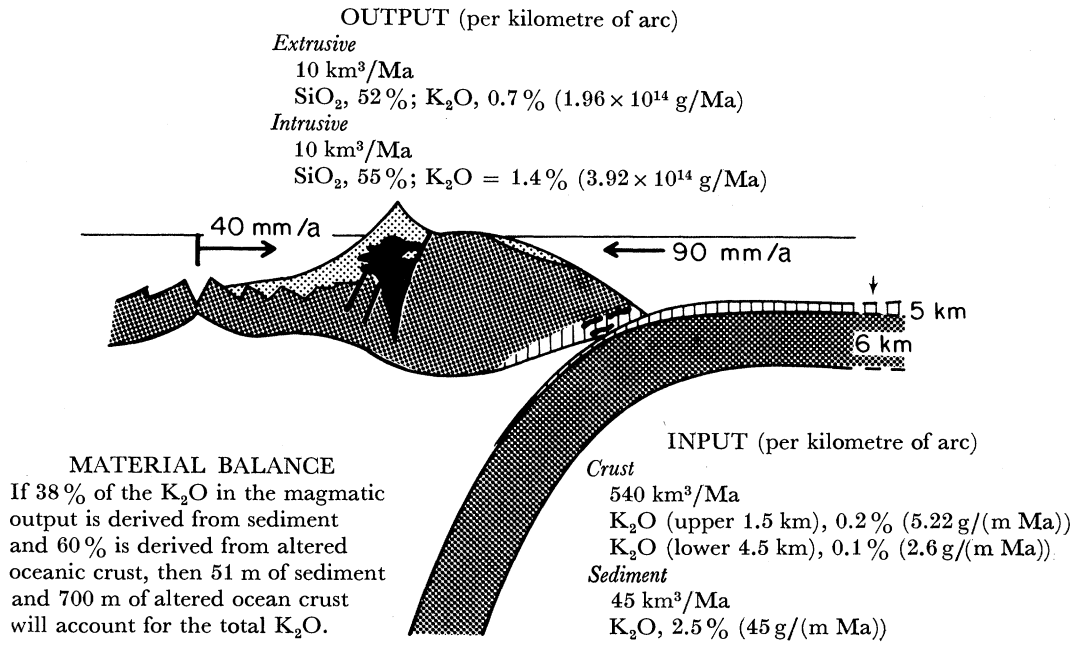

FIGURE 2. Flow diagram of the Mariana arc, illustrating the mass and of K_2O involved during processes of plate convergence and arc magmatism. Use of realistic element ratios in possible sources results in the conclusion that no more than a few tens of metres of the basal pelagic sediments are involved in magma production.

Ratios of elements that are thought to behave similarly during partial melting, and radiogenic to non-radiogenic isotope ratios, offer two ways to discriminate between element sources. To be successful, the element and isotope ratios chosen should be different in the variety of element sources. Kay (1980), following Armstrong (1971), has developed a melting–mixing model for magma genesis in arcs that predicts the proportions of rare-earth elements, Ba, Rb, Pb and K from sediment, seawater alteration of oceanic crust, and igneous sources (residual mantle or oceanic crust). Ratios of elements like K, Ba and Pb to elements like La, Sm and Sr are quite different for the various possible sources, as are the isotope ratios of Sr, Pb and Nd. Ratios of K and Ba to La in Mariana lavas (Dixon & Batiza 1979) indicate that K derived from sedimentary sources contributes about 35% of the total K in those lavas. The remaining 65% is derived from seawater alteration of oceanic crust (45%) and from the K contributed by the unaltered basaltic oceanic crust, and perhaps from overlying mantle peridotite (20%).

The proportional contributions to the total K flux can then be used to determine the required thickness of pelagic sediment and basalts involved in the Mariana magmas (figure 2). About 50 m of sediment and 700 m of altered oceanic crust are sufficient to furnish all the required K.

Chemical and geological variations among arc systems

Most interpretations of chemical and isotopic data from arc systems stress intra-arc variations and their causes. It is also important to develop a broader perspective of interarc variations and correlation of those variations with geological parameters that could conceivably affect variations in arc magmatism.

The recent surge in the availability of both geochemical and geological data from a wide geographical distribution of arcs has emboldened us to attempt a preliminary synthesis and spatial correlation of geological and geochemical variations on a global level. Use of selected geochemical parameters is hazardous because of the local processes that can produce large variations. We have attempted to reduce this risk by using data from arc basalts or the most mafic andesites for comparison, with the assumption that shallow fractionation and contamination effects ought to be minimal and most uniform in these melts. Geological processes that might affect the magmas during their rise through the upper plate are reviewed first to 'remove' the masking of the role of subducted sediment and other processes related to the subducted plate.

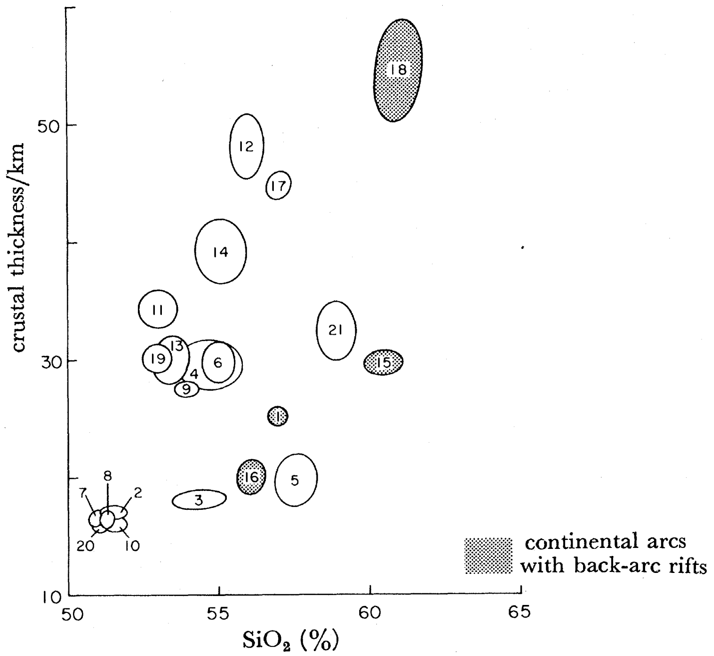

FIGURE 3. Correlation between average SiO_2 contents for contemporary arcs and crustal thickness of those arcs, indicating the contaminating effects of the upper plate on arc magmas. Ellipses associated with each arc qualitatively mark the reliability of data estimated by the authors quoted. Arcs having volcanotectonic rift zones (stippled ellipses) appear to have significantly more contaminating effects from the upper plate. High SiO_2 values in several other arcs (5, 21) can reasonably be attributed to effects of subducted sediment (see text). Identification and sources are as follows: (1) New Zealand (Ewart *et al.* 1977; Cole 1978); (2) Kermadec (Ewart *et al.* 1977; Shor *et al.* 1971); (3) Tonga (Ewart *et al.* 1977); (4) New Britain (Wiebenga 1973; Page & Johnson 1974); (5) Banda (Jacobsen *et al.* 1979; Jezek & Hutchison 1978); (6) Sunda (Nicholls & Whitford 1976); (7) Mariana (Dixon & Batiza 1979); (8) Bonin (Hotta 1970; Larson *et al.* 1975); (9) northeast Japan (Sugisaki 1976); (10) central Aleutian (Marsh 1981); (11) eastern Aleutian (Marsh 1981); (12) northern Cascades (McBirney 1976); (13) central Cascades (McBirney 1976); (14) southern Cascades (McBirney 1976); (15) northern Central America (Pichler & Weyl 1973; Pushkar *et al.* 1972); (16) southern Central America (Pichler & Weyl 1975); (17) northern Andes (Thorpe & Francis 1979); (18) central Andes (Thorpe & Francis 1979); (19) southern Andes (Thorpe & Francis 1979); (20) Scotia (Gledhill & Baker 1973); (21) Caribbean (Westbrook 1975; Brown *et al.* 1977). Crustal thickness data not specifically cited are from Cummings & Shiller (1971).

Upper plate factors

The most commonly examined upper plate factor suspected to affect volcanism along consuming plate margins is the character of the crust underlying the arc, including both its age and thickness. As age and thickness correlate positively, we have used the more simply determined seismically defined crustal thickness. Comparison of crustal thickness with isotopic ratios of either the average arc volcanic or the basaltic component show contradictory correlations. Within some arcs, both Pb and Sr isotopic ratios become more radiogenic as the underlying crust thickens and becomes more continental. This response is claimed for both Pb and Sr isotopes in the Tonga – Kermadec – New Zealand system (Armstrong & Cooper 1971; Oversby & Ewart 1972; Ewart *et al.* 1977), and for Sr ratios along Java (Hamilton 1979), where the active arc crosses the boundary between an oceanic and an older continental foundation. On the other hand, Kay *et al.* (1978) found little if any variation in Sr and Pb isotopic ratios along the Aleutian arc, from the oceanic central section to the Alaskan Peninsula. Similarly, there is no variation in isotopic ratios along the Cascade arc that could be correlated with the variations in basement (McBirney 1978). Pushkar *et al.* (1972) reached the same conclusion about Central America, where isotopic ratios did not vary as the underlying crust changed southward from continental to nearly oceanic. The very poor interarc correlation between crustal thickness and $^{87}Sr/^{86}Sr$ ratios suggests that other factors are far more important in controlling the global variation of isotopic ratios among arc systems.

On the other hand, there does appear to be a strong positive correlation between the average chemistry of arc volcanics, as measured by the SiO_2 content, and crustal thickness. This has been pointed out along the Sunda arc by Hamilton (1979), in the Aleutians by Forbes *et al.* (1969), and in Central America by McBirney (1969) and by Pichler & Weyl (1973).

The SiO_2 content of average volcanic rock from most of the global arc system, derived from studies specifically oriented toward determination of average arc chemistry, shows a good positive correlation with crustal thickness (figure 3), suggesting crustal contamination. If the data were normalized to arc length the low-SiO_2 oceanic arc values would dominate. A preliminary calculation indicates that the average arc extrusive contains 53–54 % SiO_2.

SiO_2 values from the oceanic Banda and Lesser Antilles arcs are conspicuously anomalous and will be claimed to reflect descending plate effects. Anomalously siliceous extrusives also occur in the volcanic chains of the Middle America, central Andes, and New Zealand arcs, where they are adjacent to extensive ignimbrite fields. The ignimbrites have not been used in the determination of the average arc volcanic product because we feel these rocks to be part of a back-arc extensional suite related to a distinctive and separate thermal process in the underlying mantle (Karig 1974).

The type ignimbrites in question occur as part of a volcanotectonic rift system, located just behind the volcanic arcs or fronts in only a small fraction of the global subduction system. These zones are relatively short-lived and appear to have distinctive evolutionary histories and volcanic compositions. A typical and complete event begins with uplift and subjacent high mantle temperatures, proceeding to the eruption of ignimbrites and then into the crustal rupture phase in which volcanic tectonic rifts may ultimately evolve into marginal basins (Karig & Jensky 1972; Karig 1974; Ewart *et al.* 1977).

A review of all well studied Tertiary back-arc volcanotectonic rift settings reveals variations attributable to the stage of evolution as well as to the nature of the underlying crust. Nevertheless, it seems quite clear from this perspective that the ignimbrites result primarily from anatexis of relatively non-radiogenic lower crust, mobilized by the introduction of subjacent, anomalously hot mantle. This leads us to claim that ignimbrite plateaus and volcanotectonic rifts in the back-arc setting, which occur only sporadically in space and time, form a geological phenomenon separate and relatively independent of volcanic arcs. The more silicic nature of those volcanic arcs associated with back-arc rifting (figure 3) is logically attributed to a higher than normal crustal thermal gradient and thus to a higher degree of crustal contamination.

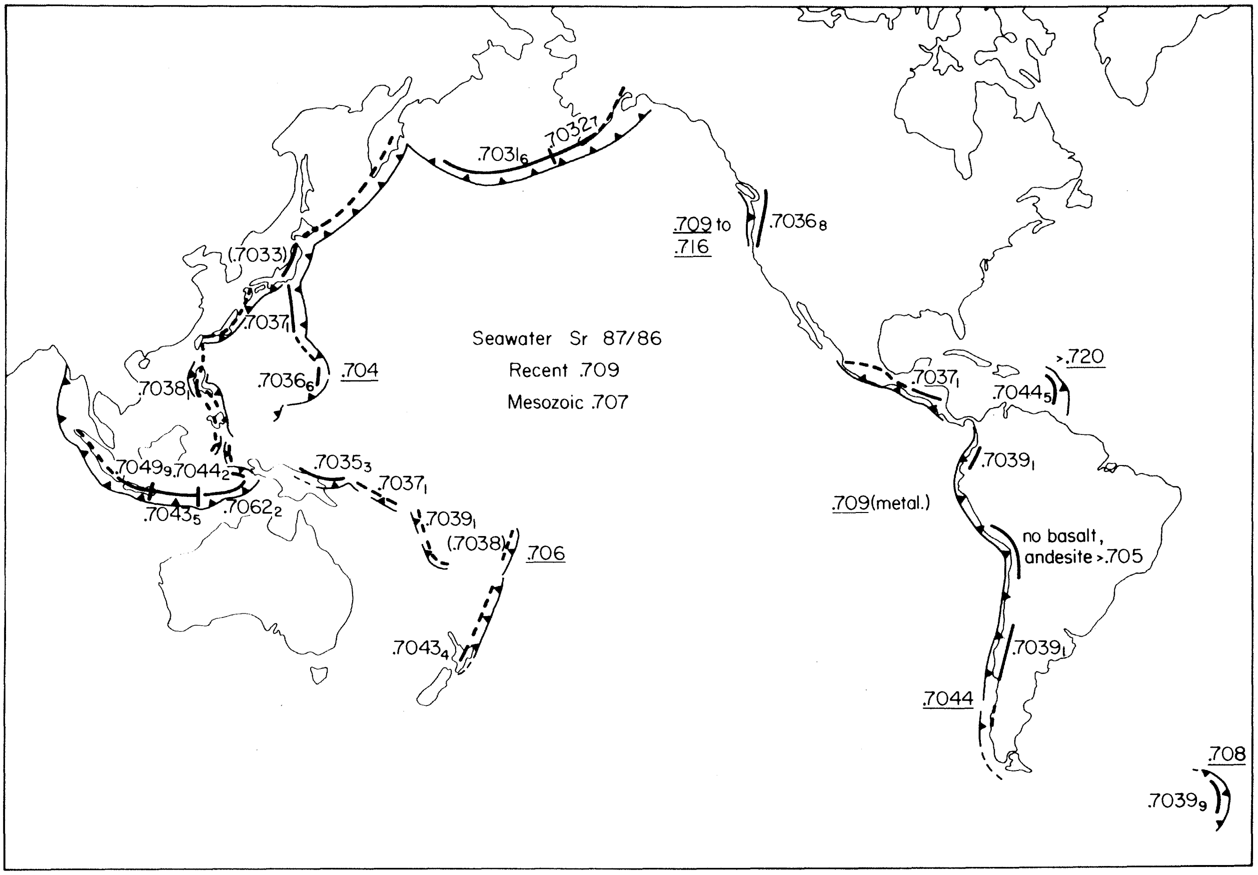

FIGURE 4. Global distribution of $^{87}Sr/^{86}Sr$ ratios in island-arc basalts and in the non-calcareous component of surficial pelagic sediments. The subscript refers to the number of volcanic centres used. This is used as a basis for correlating ratios in the basal pelagic sediments with the adjacent arc magmas (see text for arguments). Data from Ewart & Stipp (1968), Ewart *et al.* (1973), Whitford *et al.* (1977), Page & Johnson (1974), Whitford (1975), Meijer (1976), Dixon & Batiza (1979), Stern (1979), Matsuda *et al.* (1977), Kurasawa (1978), Kay *et al.* (1978), Gledhill & Baker (1973), Hawkesworth *et al.* (1977), Peterman *et al.* (1970), Church & Tilton (1973), Hedge *et al.* (1970), Pushkar (1968), Francis *et al.* (1977), Klerkx *et al.* (1977), Sinha & Hart (1972), Dasch (1969), Dasch *et al.* (1971), Brass (1976), Sun (1980) and several other references cited in the above.

Variations associated with the descending plate

Variations attributable to the subducted plate include the character and quantity of oceanic sediments entering the trench, the rate of subduction and the possible inclusion of material from the upper plate that has been tectonically eroded and subducted. Perhaps the clearest and most coherent variation that we found is the geographic variation in the character of the pelagic sediments, which we fingerprint using Pb and Sr isotopic ratios.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A

Average and modal $^{87}Sr/^{86}Sr$ ratios for arc basalts fall between 0.7038 and 0.7039 (figure 4). Nevertheless, significantly less radiogenic values typify the arc systems of the northern Pacific, regardless of other geological variations. Aleutian basalts with an average $^{87}Sr/^{86}Sr$ ratio of 0.7032 (Kay *et al.* 1978) define the lowest arc value, but significantly submodal arc averages continue westward into or beyond the northern Japan arc and southeastward to the Cascade arc (figure 4). Examples of higher than modal $^{87}Sr/^{86}Sr$ ratios that cannot easily be attributed to local effects include the Lesser Antilles arc, the eastern Sunda arc, and the Mediterranean arcs.

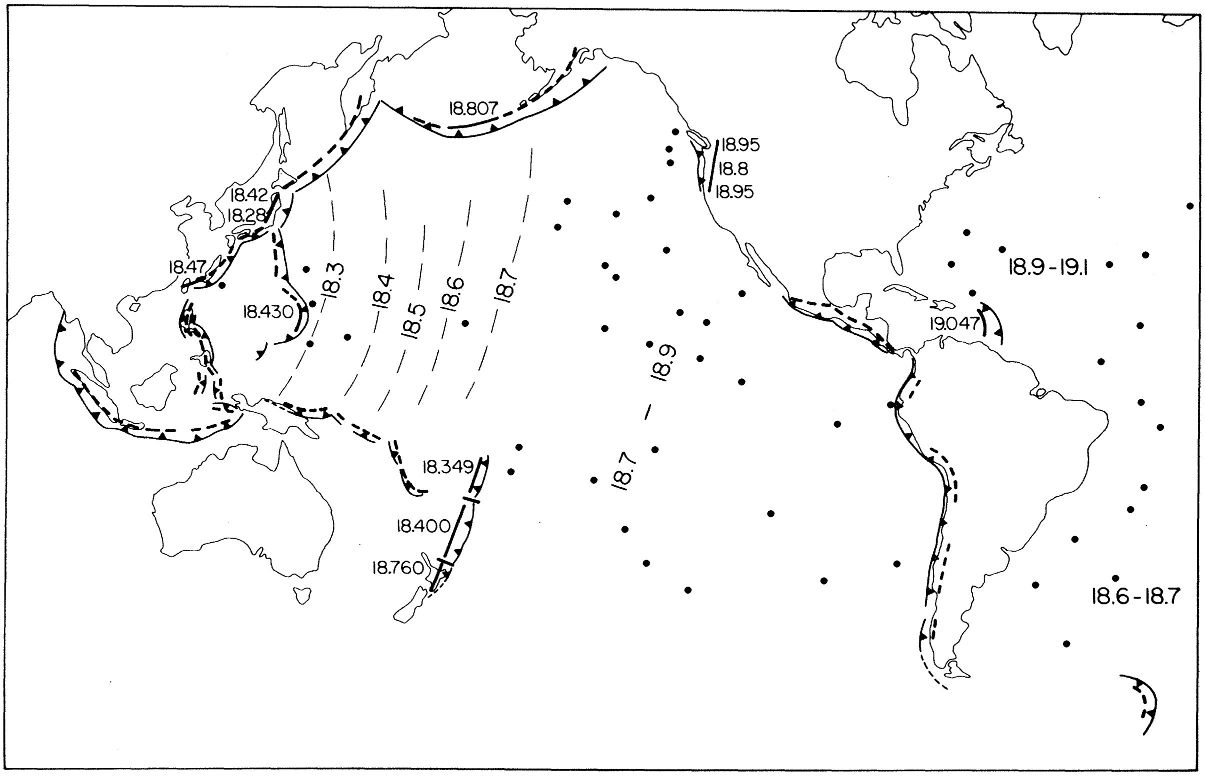

FIGURE 5. Global distribution of $^{206}Pb/^{204}Pb$ ratios in island-arc extrusives and in surficial pelagic sediments. The close spatial correlation is interpreted as supporting a pelagic sediment component in arc magmas. Dots represent sediment data points. Data from Chow & Patterson (1962), Armstrong & Cooper (1971), Meijer (1976), Oversby & Ewart (1972), Sinha & Hart (1972), Sun (1980), Church (1976), and Church & Tilton (1973).

As pointed out by Armstrong (1971), the sparser Pb isotopic data from volcanic arcs show a similar regional variation, exemplified by $^{206}Pb/^{204}Pb$ ratios (figure 5). Compared without regard to magma composition, the average ratios are lowest in the western Pacific, but are lower throughout the Pacific than in the Lesser Antilles arc.

No regional variation in the Pb and Sr isotopes among mid-ocean-ridge basalts have been recognized that appears capable of explaining these regional differences. On the other hand, regional differences in the Pb isotopic ratios of pelagic sediments and in the $^{87}Sr/^{86}Sr$ ratios of the carbonate-free component of those sediments do correlate well with the variation in arc values (figures 4, 5). However, isotopic values from surficial pelagics, which compose the only available data set, must be shown to be valid indicators if only the lowermost (basal) pelagics are likely to be subducted to depths of 100 km or more. More specifically, we must show that there are valid reasons for reliance on regional variations of the Sr isotopic ratios in the non-carbonate fraction rather than, as is usually the case, assuming that the seawater ratio of the carbonate fraction effectively masks other variations. Pelagic sediments are composed dominantly of biogenous and authigenic components, having seawater $^{87}Sr/^{86}Sr$ ratios with a gradual global evolution (see, for example, Brass 1976), and a terrigenous component having $^{87}Sr/^{86}Sr$ ratios ranging from less than 0.704 to more than 0.720 (see, for example, Dasch 1969). The ratios of the terrigenous component correlate very well with the proportions of less radiogenic volcanic clay and glass relative to that of clays derived from more radiogenic continental cratons.

Because the carbonate compensation depth (c.c.d.) is greater than oceanic spreading ridges, a common assumption is that basal pelagic sediments are dominantly calcareous. In fact, a review of those strata drilled during the Deep Sea Drilling Project (D.S.D.P.) at sites seaward of active arcs shows typical carbonate contents of 0–25% although values range to 80%. The shallower c.c.d. earlier than the Late Eocene (Berger & Winterer 1974) is probably a major factor for the small carbonate fraction.

Secondly, the effects of sediment diagenesis on Sr content are ignored by the use of surface data. Studies of D.S.D.P. cores (Hawkesworth & Elderfield 1978; Gieskes *et al.* 1975) reveal significant losses of Sr from the sediment into pore waters by the recrystallization of carbonate and the alteration of volcanic glass. On the other hand, Sr interchange between pore water and the clay fraction is low to negligible (Dasch 1969).

Yet another factor is the variable growth of radiogenic Sr in pelagic sediments before their involvement in magmatism. A Rb/Sr range of 0.25 to more than 1.0 in the non-calcareous fraction (see, for example, Dasch 1969), which correlates with source provenance, as does the $^{87}Sr/^{86}Sr$ ratio, combined with ages of 10^8 a for many subducted sediments, results in differential increases of $^{87}Sr/^{86}Sr$ ratios that could exceed 0.003.

Radiogenic lead contents of Recent sediments (see, for example: Chow & Patterson 1962; Church 1973) display a similar correlation with sediment provenance (figure 5). Although most lead is transported in the oceans in solution (Chow & Patterson 1962), regional source characteristics are preserved despite relatively long residence times (10^4 a). The least radiogenic values in the western Pacific and maximum values in the Atlantic again reflect the contrast between continental and volcanic sources.

The isotopic correlation between pelagic sediments and arc basalts supports the concept that a small amount of sediment, composing the basal strata on the descending oceanic lithosphere can be subducted to the zone where arc magma is being generated.

Another geological process that might affect the character of arc magmatism is the subduction of continental crust. This has been most plausibly suggested as an alternative to sediment subduction in the Banda arc (Whitford & Jezek 1979; Magaritz *et al.* 1978), where mafic lavas with very high $^{87}Sr/^{86}Sr$ ratios occur (figure 4). However, a review of the geological and tectonic variations along the arc suggests that sediment subduction is more plausible.

Along the southern and eastern perimeter of the Banda arc, the lithospheric plate entering the trench carries continental crust, whereas the upper plate is more nearly oceanic. Extrapolation of the Australian continental margin beneath the western section of the arc (see, for example, Hamilton 1979), coupled with local gravity data (Chamalaun *et al.* 1976) indicates that continental crust has been subducted beneath the island of Timor to a depth near 25 km.

However, even at these depths, the descending plate is still directly overlain by continental crust or accretionary material and has not yet begun its steep descent into the mantle.

Geochemical data are available only for Quaternary volcanics in the Banda arc, but, even so, would probably reflect the spatial relations between the volcanic chain and geologic units on the descending lithosphere that existed several million years ago. At present, the western boundary of arc basalts with highly radiogenic Sr is approximately north of the point where the Australian continental margin impinges on the Sunda trench. Two million years ago, assuming a subduction rate of 7 cm/a, the Australian continent should have extended beneath only that sector of the Sunda arc east of Timor. Both the geophysical data and the spatial correlations lead us to believe that, rather than continental crust, the responsible contaminant was the basal strata of the Indian ocean basin, carrying a large terrigenous component because of its proximity to Australia.

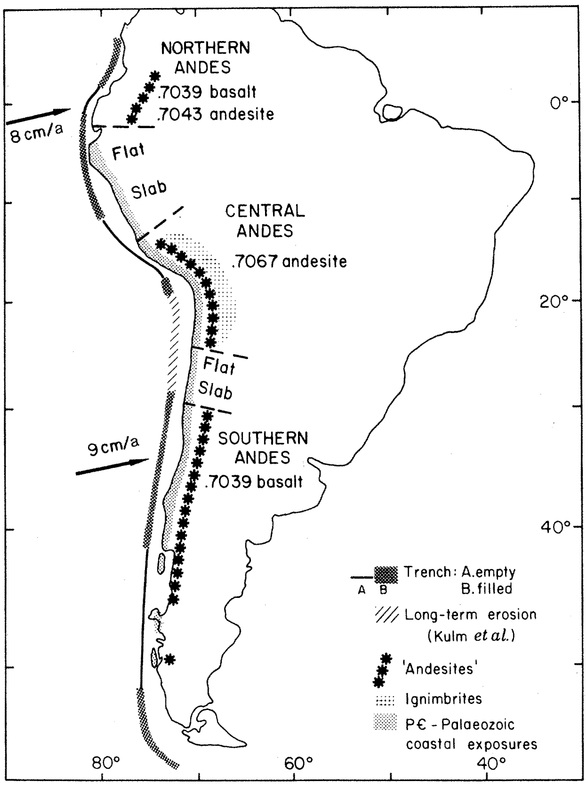

FIGURE 6. Tectonic and volcanic features bearing on the role of tectonic erosion in arc magmatism along the Andean arc. The anomalously high $^{87}Sr/^{86}Sr$ ratios of the central Andean volcanic (andesitic) arc correlate closely with the Altiplano, interpreted to be a back-arc rifting feature. To the contrary, there is little correlation between areas of long-term tectonic erosion and character of arc magmas. Data from Kulm *et al.* (1977), Thorpe & Francis (1979), and Barazangi & Isacks (1976).

Yet another possible contaminant of arc magmas is continental crustal material from the upper plate, tectonically eroded during subduction and carried down to the region of magma generation. A commonly cited example of this process is the central section of the Andean arc (figure 6), where high $^{87}Sr/^{86}Sr$ ratios and modally more silicic arc volcanism are associated with the apparent lack of a sufficiently large accretionary prism (Hussong *et al.* 1976; Kulm *et al.* 1977). Presumably because basalts are relatively rare in the central Andes, no analyses from this compositional range are available, but the dominant andesites have $^{87}Sr/^{86}Sr$ ratios higher than have rocks of similar bulk composition in other arcs. Addition of radiogenic strontium may have occurred along the slab or in the lower crust (see, for example, Briqueu & Lancelot 1979). Strontium originating from the slab would probably be from crustal fragments rather than from sediments, because the detrital component of eastern Pacific sediments is relatively non-radiogenic (figure 4). If tectonic erosion of the magnitude implied by calculations of continental areas removed did occur, a mixture of rocks from the more radiogenic upper crust and less radiogenic lower crust would be expected. Fusion of the lower crust in the upper plate would favour a less radiogenic average contribution.

We can suggest several spatial correlations that raise questions concerning the efficacy of tectonic erosion to transport sialic material to great depth. Quaternary volcanism in the central Andes terminates to the north and south by non-volcanic sectors of the arc that overlie flat slab sections of the seismic zone (figure 6), but the sectors of the volcanic arc in the north and southern Andes appear to be isotopically normal (figure 4).

Criteria postulated to favour active tectonic erosion characterize only that part of the trench opposite the southern part of the central Andean volcanic province and a small part of the flat slab area to the south (Kulm *et al.* 1977). It was recognized, however, that the tectonic response at the base of the trench slope could well be transient and rapidly changing. Most of the arc, from northern Peru to south-central Chile, displays geological criteria cited for the long-term retreat of the continental margin (see, for example, Katz 1971). These longer-term processes ought to have a strong imprint on the chemical response if any degree of temporary and spatial homogenization occurs during the mechanical transfer of subducted material along the plate interface. The question seems more that of why anomalously silicic and radiogenic arc volcanics are restricted to the central Andes. Their very close spatial correlation with the Altiplano, with its ignimbrite eruptions and anomalous, probably very hot mantle (see Chinn *et al.* 1980, and references therein), leads us to prefer to attribute the anomalous arc volcanism of the central Andes to back-arc thermal activity, similar to that in New Zealand and Central America.

Recent D.S.D.P. results, indicating a lack of requisite space in accretionary prisms for the sediment influx to the associated trench, have spurred models of deep subduction of these sediments (see, for example, Armstrong, this symposium). If significantly more sediment is being subducted to depths greater than 100 km in this type of arc than in that where most sediment is accreted at shallow levels, some differences in the respective volcanic arcs ought to be observable. Good reflexion profiles support shallow accretion in the Sunda (Karig *et al.* 1980) and Cascades (Seely *et al.* 1974) arcs. Partial to total subduction in postulated in the Japan arc (D.S.D.P. Leg 57 Staff 1978), the Mariana arc (D.S.D.P. Leg 60 Staff 1978) and parts of the Aleutian (von Huene 1979) and southern Middle America (von Huene *et al.* 1980) arcs among others. Although the sample is small, we fail to see an increase in bulk chemistry or isotopic ratios that might correlate with partial or total subduction. In particular we note that the SiO_2 and isotopic ratios in the northeastern Japan arc are at least as low as those of the Cascades arc (figures 3, 4) and are lower than those in Sumatra. This lack of correlation lends support to the earlier arguments in favour of the subcretion of most sediments at relatively shallow depths.

Conclusions

The fate of the overwhelming proportion of trench sediments and a large proportion of pelagic sediments appears to currently be one of shallow level accretion and subcretion, with a minor fraction being returned to the crust in arc magmas. Tracing the flow of material through an active margin clearly illustrates that large amounts of continental detritus are recycled to the continents and that even larger amounts of oceanic crust are recycled to the mantle via subduction. Transfer of mantle-derived material by arc magmatism to the continents is, in comparison, small. Because of the small amount of deeply recycled sediment in the arc magmas, arc magma production rates approximately equal gross crustal growth rates. Net crustal growth rate very nearly equals gross crustal growth rate, since the amount of sediment returned to the mantle is small, and tectonic erosion does not appear to return significant amount of crustal material to the mantle. The extent of crustal destruction and resorption to the mantle in collision zones or along basement thrusts is unassessed, but, from a survey of contemporary plate boundaries, these processes appear to be relatively minor.

The conclusion that magma production rate approximately equals crustal growth rate might be assessed by comparing the global magma production rate of arcs with the volume of continental crust. If we assume a linear relation between subduction rate and a time-averaged production rate about half that of the present (based on the value for the Mariana arc), the production rate of all arc systems, calculated individually, is about 2×10^{21} g/Ma. Accepting a continental mass of 22.4×10^{24} g (Hurley & Rand 1969), we observe that the present rate of magma production and continental growth is less than one-half of the growth rate averaged over 4.5 Ga. If, as is likely, continental growth was more rapid during the Archaean (see, for example, Windley 1977), the present growth rate is only a small fraction of that early rate.

The continental freeboard argument for large scale recycling of sediments to the mantle (Armstrong, this symposium) might be reviewed against such a low growth rate, which can probably be extrapolated through most of the Phanerozoic. Our calculated growth rate implies only a small change in crustal volume during the Phanerozoic that is well within the limits of uncertainty indicated by Wise (1974), especially considering the possible long term trends in the Earth's thermal régime (see, for example, Turcotte & Burke 1979).

Not only is the present rate of production of continental crust much less than that during early Earth history, but also the character of the magmas also appears to be significantly different. The average SiO_2 value of contemporary arc magmatism (intrusion plus extrusion) appears to be near 55%. If the more silicic arc-magma suites (figure 3) are attributed to crustal contamination or melting (see, for example. Eichelberger 1978), then the newly added material is even more mafic. In contrast, recent estimates of mean crustal composition based on seismic velocity data (see, for example, Smithson 1978) and of exposed deep crustal terranes (see, for example, Holland & Lambert 1972) suggest SiO_2 values near 63%. Although refinement of both data sets is necessary, it appears clear that the bulk of continental crust cannot have been constructed from the material now being derived from the mantle in island arcs.

Fluxes for specific elements through the arc system also imply that the present arc magmatism has little compositional affinity with the bulk crustal composition. For instance, a K flux calculation (Kay 1980) shows that the K content, representing non-recycled K in arc magmas, divided by the total mass of mantle-derived material added to the arcs is about 0.5%, much lower than the *ca.* 2% in most crustal models.

Thus, we see no compelling reason for the use of present arc systems as a quantitative model for the development of most of the Earth's continental crust. Although continental crust seems to be growing at present, the rate of growth is too slow, and the composition of material is too mafic.

References (Karig & Kay)

Alleman, F. & Peters, T. 1972 *Eclog. geol. Helv.* **65**, 657–697.
Armstrong, R. 1971 *Earth planet. Sci. Lett.* **12**, 137–142.
Armstrong, R. L. & Cooper, J. A. 1971 *Bull. volcan.* **35**, 27–63.
Barazangi, M. & Isacks, B. 1976 *Geology* **4**, 686–692.
Berger, W. H. & Winterer, E. L. 1974 *Spec. Publ. int. Ass. Sediment*, no. 1, pp. 11–48. Oxford: Blackwell.
Bice, D. 1980 *Trans. Am. geophys. Un.* **61**, 70. (Abstract.)
Brass, G. W. 1976 *Geochim cosmochim. Acta* **40**, 721–730.
Briqueu, L. & Lancelot, J. 1979 *Earth planet. Sci. Lett.* **43**, 385–396.
Brothers, R. N. & Blake, Jr, M. C. 1973 *Tectonophysics* **17**, 337–358.
Brown, G. M., Holland, J. G., Sigurdsson, H., Tomblin, J. F. & Arculus, R. J. 1977 *Geochim. cosmochim. Acta* **41**, 785–801.
Carden, J. R., Connelly, W., Forbes, R. B. & Turner, D. L. 1977 *Geology* **5**, 529–533.
Chamalaun, F. H., Lockwood, K. & White, A. 1976 *Tectonophysics* **30**, 241–259.
Chinn, D. S., Isacks, B. L. & Barazangi, M. 1980 *Geophys. Jl R. astr. Soc.* **60**, 209–244.
Chow, T. J. & Patterson, C. C. 1962 *Geochim. cosmochim Acta* **26**, 263–308.
Church, S. E. 1976 *Earth planet. Sci. Lett.* **29**, 175–188.
Church, S. E. & Tilton, G. R. 1973 *Bull. geol. Soc. Am.* **84**, 431–454.
Citron, G. 1980 Ph.D. thesis, Cornell University.
Coats, R. R. 1962 *Am. geophys. Un. Monogr.* no. 6, pp. 92–109.
Cole, J. W. 1978 *J. Volcan. geotherm. Res.* **3**, 121–153.
Coleman, R. G. & Lee, D. E. 1963 *J. Petr.* **4**, 260–301.
Coleman, R. G. & Lamphere, N. A. 1971 *Bull. geol. Soc. Am.* **82**, 2397–2412.
Cowan, D. S. & Boss, R. F. 1978 *Bull. geol. Soc. Am.* **89**, 155–158.
Cummings, D. & Schiller, G. I. 1971 *Earth Sci. Rev.* **7**, 97–125.
DSDP Leg 57 Staff 1978 *Geotimes* **23**, 16–21.
DSDP Leg 60 Staff 1978 *Geotimes* **23**, 19–22.
Dasch, E. J. 1969 *Geochim. cosmochim. Acta* **33**, 1521–1552.
Dasch, E. J., Dymond, J. R. & Heath, G. R. 1971 *Earth planet. Sci. Lett.* **13**, 175–180.
Dixon, R. & Batiza, R. 1979 *Contr. Miner. Petr.* **70**, 167–181.
Eichelberger, J. 1978 *Nature, Lond.* **275**, 21–27.
Ernst, W. G. (ed.) 1975 *Benchmark Pap. Geol.* **19**.
Ernst, W. G., Seki, Y., Onuki, H. & Gilbert, M. C. 1970 *Mem. geol. Soc. Am.* **124**.
Everden, J. & Kistler, R. 1970 *Prof. Pap. U.S. geol. Surv.*, no. 623.
Ewart, A. & Stipp, J. T. 1968 *Geochim. cosmochim. Acta* **32**, 699–736.
Ewart, A., Bryan, W. B. & Gill, J. B. 1973 *J. Petr.* **14**, 429–465.
Ewart, A., Brothers, R. & Mateen, A. 1977 *J. Volcan. geotherm. Res.* **2**, 205–250.
Forbes, R. B., Ray, D. K., Katsura, T., Matsumoto, H., Haramura, H. & Furst, M. J. 1969 *Bull. Ore. St. Dep. Geol. miner. Ind.* **65**, 111–120.
Francis, R. & Rundle, C. 1976 *Bull. geol. Soc. Am.* **87**, 474–480.
Francis, P. W., Moorbath, S. & Thorpe, R. S. 1977 *Earth planet. Sci. Lett.* **37**, 197–202.
Gastil, R. G. 1975 *Geology* **3**, 361–363.
Ghent, E. D. & Stout, M. Z. 1979 *Eos, Wash.* **60**, 961.
Gieskes, J. M., Kastner, M. & Warner, T. B. W. 1975 *Geochim. cosmochim. Acta* **39**, 1385–1394.
Gledhill, A. & Baker, P. E. 1973 *Earth planet. Sci. Lett.* **19**, 369–372.
Hamilton, W. 1979 *Prof. Pap. U.S. geol. Surv. no.* 1078.
Hawkesworth, C. J., O'Nions, R. K., Pankhurst, R. J., Hamilton, P. J. & Evensen, N. M. 1977 *Earth planet. Sci. Lett.* **36**, 253–262.
Hawkesworth, C. J. & Elderfield, H. 1978 *Earth planet. Sci. Lett.* **40**, 423–432.
Hedge, C. E., Hildreth, R. A. & Henderson, W. T. 1970 *Earth planet. Sci. Lett.* **8**, 434–438.
Holland, J. & Lambert, R. 1972 *Geochim. cosmochim. Acta* **36**, 673–683.
Hopson, C. A., Carter, F. W. & Crowder, D. F. 1970 *Geol. Soc. Am. Abstr. Prog.* **2**, 104.
Hotta, H. 1970 *J. Phys. Earth* **18**, 125–142.
Hurley, P. & Rand, J. 1969 *Science, N.Y.* **164**, 1229–1242.
Hussong, D. M., Edwards, P. B., Johnson, S. H., Campbell, J. F. & Sutton, G. H. 1976 *Am. geophys. Un. geophys. Monogr.* **19**, 71–86.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Jacobsen, R. S., Shore Jr, G. G., Kieckhefer, R. M. & Purdy, G. M. 1979 *Am. Ass. Petr. Geol. Mem.* **29**, 209–222.
Jezek, P. A. & Hutchison, C. S. 1978 *Bull. volcan.* **41**, 586–608.
Karig, D. E. 1974 *A. Rev. Earth planet. Sci.* **2**, 51–75.
Karig, D. E. 1980 *J. Geol.* **88**, 27–39.
Karig, D. E. & Jensky, W. 1972 *Earth planet. Sci. Lett.* **17**, 169–174.
Karig, D. E. & Sharman, G. F. 1975 *Bull. geol. Soc. Am.* **86**, 377–389.
Karig, D. E., Moore, G. F., Curray, J. R. & Lawrence, M. B. 1980 *Am. Geophys. Un. geophys. Monogr.* **23**, 179–208.
Katz, H. R. 1971 *Bull. Am. Ass. Petrol Geol.* **55**, 1753–1758.
Kay, R. W. 1981 *J. Geol.* **88**, 497–522.
Kay, R., Sun, S.-S. & Lee-Hu, C.-N. 1978 *Geochim. cosmochim. Acta* **42**, 263–273.
Kennett, J., McBirney, A. & Thunell, R. 1977 *J. Volcan. geotherm. Res.* **2**, 145–163.
Kieckhefer, R. M., Shor Jr, G. G. & Curray, J. R. 1980 *J. geophys. Res.* **85**, 863–889.
Klerkx, J., Deutsch, S., Pichler, H. & Zeil, W. 1977 *J. Volcan. geotherm. Res.* **2**, 49–71.
Kulm, L. D., Schweller, W. J. & Masias, A. 1977 *Am. geophys. Un. M. Ewing Ser.* **1**, 285–301.
Kurasawa, H. 1978 *U.S. geol. Surv. open-File Rep.* no. 78-701, pp. 235–237.
Landis, C. A. & Bishop, D. G. 1972 *Bull. geol. Soc. Am.* **83**, 2267–2284.
Larson, E. E., Reynolds, R. L., Merrill, R., Levi, S., Ozima, M., Aoki, Y., Kinoshita, H., Zasshu, S., Kawai, N., Nakajima, T. & Hirooka, K. 1975 *Bull. volcan.* **38**, 361–377.
McBirney, A. R. 1969 *Bull. St. Ore. Dep. Geol. miner. Ind.* **65**, 185–189.
McBirney, A. 1976 *Can. Miner.* **14**, 245–254.
McBirney, A. R. 1978 *A. Rev. Earth planet. Sci.* **6**, 437–456.
Magaritz, M., Whitford, D. & James, D. 1978 *Earth planet. Sci. Lett.* **40**, 220–230.
Malpas, J. 1979 *Can. J. Earth Sci.* **10**, 2086–2101.
Markhinin, E. 1968 *Am. geophys. Un. Monogr.* **12**, 413–422.
Marsh, B. 1979 *J. Geol.* **87**, 687–713.
Marsh, B. D. 1981 In *Orogenic andesites* (ed. R. A. Thorpe). New York: John Wiley. (In the press.)
Matsuda, J.-I., Zashn, S. & Ozima, M. 1977 *Tectonophysics* **37**, 141–151.
Meijer, A. 1976 *Bull. geol. Soc. Am.* **87**, 1358–1369.
Moorbath, S. 1978 *Phil. Trans. R. Soc. Lond.* A **288**, 401–413.
Moore, J. C. 1975 *Geology* **3**, 530–532.
Moore, D. E. & Liou, J. G. 1979 *Bull. geol. Soc. Am.* **90**, 1089–1091.
Nicholls, I. A. & Whitford, D. J. 1976 In *Volcanism in Australasia* (ed. R. W. Johnson), pp. 77–90. Amsterdam: Elsevier.
Oversby, V. M. & Ewart, A. 1972 *Contr. Miner. Petr.* **37**, 181–210.
Page, R. W. & Johnson, R. W. 1974 *Lithos* **7**, 91–100.
Peterman, Z. E., Carmichael, I. S. E. & Smith, A. L. 1970 *Bull. geol. Soc. Am.* **81**, 311–318.
Pichler, H. & Weyl, R. 1973 *Geol. Rdsch.* **62**, 357–396.
Pichler, H. & Weyl, R. 1975 *Geol. Rdsch.* **64**, 457–475.
Pitcher, W. 1978 The anatomy of a batholith. *J. geol. Soc. Lond.* **135**, 157–182.
Pushkar, P. 1968 *J. geophys. Res.* **73**, 2701–2714.
Pushkar, P., McBirney, A. R. & Kudo, A. M. 1972 *Bull. volcan.* **35**, 265–294.
Reinhardt, B. M. 1969 *Schweizer. miner. petrogr. Mitt.* **49**, 1–30.
Sample, J. 1980 Senior thesis, Cornell University.
Seely, D. R., Vail, P. R. & Walton, G. G. 1974 In *The geology of continental margins* (ed. C. A. Burk & C. L. Drake), pp. 249–260. New York: Springer-Verlag.
Shor, G. G., Kirk, H. K. & Menard, H. W. 1971 *J. geophys. Res.* **76**, 2562–2586.
Sigurdsson, H., Sparks, R., Carey, S. & Huang, T. 1981 *Bull. geol. Soc. Am.* (In the press.)
Sinha, A. K. & Hart, S. R. 1972 *Carnegie Instn Wash. Yb.* **71**, 309–312.
Smithson, S. 1978 *Geophys. Res. Lett.* **5**, 749–752.
Snyder, G. 1959 *Bull. U.S. geol. Surv.* no. 1028-H, pp. 169–210.
Stern, R. J. 1979 *Contr. Miner. Petr.* **68**, 207–219.
Stoiber, R. & Carr, M. 1973 *Bull. volcan.* **37**, 1–22.
Sugimura, A. & Uyeda, S. 1973 *Island arcs, Japan and its environs. Developments in geotectonics*, vol. 3. Amsterdam: Elsevier.
Sugisaki, R. 1976 *Lithos* **9**, 17–30.
Sun, S.-S. 1980 *Phil. Trans. R. Soc. Lond.* A **297**, 409–445.
Thorpe, R. S. & Francis, P. W. 1979 *Tectonophysics* **57**, 53–70.
Turcotte, D. & Burke, K. 1978 *Earth planet. Sci. Lett.* **41**, 341–346.
von Huene, R. 1972 *Bull. geol. Soc. Am.* **83**, 3613–3626.
von Huene, R. 1979 *Am. Ass. Petrol. Geol. Mem.* **29**, 261–272.
von Huene, R., Aubouin, J. *et al.* 1980 *Bull. geol. Soc. Am.* **91**, 421–432.
Westbrook, G. K. 1975 *Geophys. Jl R. astr. Soc.* **43**, 201–242.
Whitford, D. J. 1975 *Geochim. cosmochim. Acta* **39**, 1287–1302.
Whitford, D. J., Compston, W., Nicholls, I. S. & Abbott, M. J. 1977 *Geology* **5**, 571–575.
Whitford, D. J. & Jezek, P. A. 1979 *Contr. Miner. Petr.* **68**, 141–150.

Wiebenga, W. A. 1973 In *The western Pacific: island arcs, marginal seas, geochemistry* (ed. P. Coleman), pp. 163–177. New York: Crane, Russak and Co.
Williams, H. & Smyth, W. R. 1973 *Am. J. Sci.* **273**, 594–621.
Windley, B. 1977 *Nature, Lond.* **270**, 426–428.
Wise, D. U. 1974 In *The geology of continental margins* (ed. C. A. Burke & C. L. Drake) pp. 45–58. New York: Springer-Verlag.
Woodcock, N. H. & Robertson, A. H. G. 1977 *Geology* **5**, 373–376.

Discussion

G. C. Brown (*Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes MK7 6AA, U.K.*). I appreciated Dr Karig's enlightening summary of the evidence for pelagic sediment subduction, and should like to make one comment and to ask two related questions. First, I am not sure that a correlation between magmatic silica concentrations and crustal thickness should be used as evidence of continental erosion, subduction and remelting. There are other plausible explanations for such a correlation if we consider the greater chances of *either* remelting (to produce more siliceous liquids) in a thick crust *and/or* fractionation of silica-poor mineral phases during ascent through thick crust.

Secondly, am I correct in supporting that the $^{87}Sr/^{86}Sr$ ratio of subducted pelagic sediments will be like that of ocean water, *ca.* 0.709 at the present day, as a *minimum* value? Such subducted sediments must surely melt and then, I wonder, why are not the initial strontium isotope ratios of 'plate margin' magmas much higher than observed? Is there any known correlation between magmatic initial ratios and the sediment thickness on the adjacent plate? I would argue not (see Brown, this symposium),

If, in answer to the strontium question (cf. Armstrong, this symposium), it is argued that pelagic sediments only contribute a small fraction of the required melt, then this *either* (*a*) puts a severe restraint on the amount that is subducted or (*b*) means that much must become refractory and be lost to the deep mantle. In the former case (*a*) it must follow that the crust is still growing by the addition of the mantle-derived component of plate margin magmas. Such a conclusion might also be consistent with (*b*), depending on relative volumes, and so I should welcome any comment regarding modern crustal growth based on his observations.

R. L. Armstrong (*Department of Geological Sciences, University of British Columbia, Vancouver V6T 2B4, Canada*). In response to Dr Brown's question I would agree that the silica concentration – crustal thickness correlation might have an explanation entirely independent of sediment subduction.

The Sr isotope composition of subducted sediment is potentially quite variable, from 0.704 to greater than 0.730, depending on provenance (Dasch 1969). Our own analyses of Mesozoic pelite, argillite, and greywacke from the western edge of North America show ratios typically of 0.705–0.707; a small percentage of this material buffered by the large Sr reservoir in the mantle would be a suitable prescription for the observed initial ratios of arc magmas. My feeling is that much more clastic and volcaniclastic material is subducted than marine carbonate (which is recycled to platform areas during much of the geological record). The material of accretion wedges is a strong argument for this view, if one assumes that much of the sediment being subducted is the same sort of material as is accumulated in these wedges.

I would expect more correlation of magma isotope ratios with sediment provenance than with ocean-floor sediment thickness. We tried to show this for Pb, and I think it also can be argued for Sr isotopes in the Lesser Antilles.

MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES
THE ROYAL SOCIETY A
PHILOSOPHICAL TRANSACTIONS OF

Even if the fluxes of sediment and magma are equal the direct sediment contribution to magmas is small because of the massive chemical buffer provided by the mantle. The sediment may be 100% 'involved' in magma genesis but the magmas reaching the surface will show nearly 100% equilibration with mantle chemistry. Only the elements highly depleted in the mantle can be used as tracers of the subduction process.

The argument can be made by analogy: if a bucket of pure water is thrown into the ocean and a few minutes later a bucket of water taken out, the ocean water now in the bucket would show no measurable trace of the water thrown in. The water thrown in vanishes nearly instantly. If, instead, that bucket first contained an exotic substance such as tritium oxide then its contamination plume could be readily detected in buckets of water collected months, and even years after the event. The consumption of sediment by the mantle is an analogous process. I doubt that recognizable refractory inclusions of sediment survive; the atoms of the subducted sediment become dispersed in newly formed minerals of a less-depleted mantle. It is one step of an endless cycle.

The apparent percentage of sediment in magmas provides no measure of continental growth or stability.

Reference

Dasch, E. J. 1969 *Geochim. cosmochim. Acta* **33**, 1521–1552.

D. E. Karig and R. W. Kay. Several of the points raised by Dr Brown, as, for instance, isotopic nature of pelagic sediments, were covered but probably not emphasized in our oral presentation; we hope that they have been adequately answered in the preceding printed version. We agree with Dr Brown that contamination by lower continental crust is the major cause for the positive correlation of SiO_2 with crustal thickness. In response to his question concerning effects of sediment thickness and to expand on the nature of subducted sediments, we rely upon observation of sediment behaviour at trenches. We strongly dispute Dr Armstrong's assumption that subducted sediments are similar to those observed in accreted terranes. Because of selective subduction, the shallower, and generally more volcanogenic, part of the sediment cover entering trenches is incorporated into accretionary prisms, whereas the deeper, pelagic strata are preferentially subducted and involved in arc magmatism. Selective subduction also explains why no correlation is observed between the character of arc volcanism and the thickness of the sediment cover.

The strong regional correlation between the isotopic character of arc magmas and that of incoming pelagic sediment, which both Armstrong and we recognize, mitigates against the mixing of a limited amount of sediment with a vast reservoir of mantle, as does the physical nature of the subduction zone. The cool, rigid descending lithosphere precludes free mixing with the mantle beneath, and there is serious doubt that significant mantle flow exists in the wedge between the descending slab and the upper plate in the region of melting. It seems more realistic to envisage mixing of a very limited volume of oceanic crust or mantle during fusion of the sediments, which, we feel, ought to be nearly complete.

The magmatic production rate at arc systems far exceeds the limited sediment subduction rate consistent with selective subduction and our observations concerning subcretion. Therefore we conclude that at present the continental crust is growing. We emphasize, however, that the present rate of growth is insufficient to account for the total volume of continental crust and that the composition of new crust is less silicic than that of the average crust.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES