Deep marine arc apron deposits and syndepositional magmatism in the Alisitos group at Punta Cono, Baja California, Mexico

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ABSTRACT

Rocks exposed at Punta Cono include very fine-grained to coarse-grained tuffs, lapilli tuffs, and tuff breccias deposited in a deep marine environment. Syndepositional basaltic intrusive activity was common. In one locality a hyaloclastite-peperite complex formed. Slumped sections with fluidal basalt 'clasts', derived from intrusions that entered the sediment pile from below, are present elsewhere. Abundant soft-sediment folds in fine-grained laminated subaqueous fall-out tuff suggest steep gradients; these are cut by shallow channels filled with coarse-grained tuff, lapilli tuff, and rare tuff breccia. The combination of marine fossils, extreme textural immaturity, abundant slump features, and syndepositional magmatism indicates deposition upon the submarine flanks of an active volcano.

Recognition of magma–wet sediment interaction is hampered in volcaniclastic rocks because of the similarity between host and intrusive fragments. Products of magma–water–sediment interactions at Punta Maria include: (1) jigsaw-puzzle hyaloclastite, formed by non-explosive hydroclastic fragmentation of magma upon contact with water and water-bearing sediment; (2) peperites, produced by mixing of magma with sediment; and (3) an unusual tuff breccia unit, the result of non-explosive mixing of 'wisps' of lava with sediment during remobilization of an unconsolidated section. Low-explosivity magma–water–sediment interactions are favoured by relatively high hydrostatic pressures in sub-wave base settings.

INTRODUCTION

The Alisitos Group includes Aptian/Albian age volcaniclastic and volcanic rocks deposited in an island arc setting (Beggs, 1984). They form the western wall of the Peninsular Range Batholith in the state of Baja Norte, Mexico (Fig. 1). Deformation of the Alisitos Group is minimal away from the batholith, and is commonly restricted to open folding and variable amounts of high-angle faulting (Gastil, Phillips & Allison, 1975). Metamorphic grade within the Alisitos Group is generally lower away from the batholith (e.g. Busby-Spera & White, 1986; 1987), though widespread greenschist facies metamorphism of the basaltic to andesitic oceanic island arc (e.g. Silver, 1986) rocks is characteristic (Gastil *et al.*, 1975; Beggs, 1984).

Localized alteration is associated with smaller intrusions. The presence of syndepositional intrusions, multiple sets of dikes, and fossiliferous tuffs suggests that the abundant epidote and interstitial calcite at Punta Cono (Fig. 2) may have resulted from decarbonation of calcareous units within the section. Multiple generations of chlorite fill vesicles, and fine-grained matrix is commonly chloritized. Most plagioclase is albitized, and fine-grained albite probably makes up low-birefringence parts of the matrix, as reported by Beggs (1984) for other meta-andesites in the Alisitos Group. Glass shards are not preserved due to recrystallization, although they presumably made up the bulk of the fine-grained tuffs and much of the matrix of coarser grained units. Fiamme (flattened pumice), euhedral plagioclase phenocrysts, and highly angular volcanic lithic fragments nevertheless clearly indicate a pyroclastic, rather than epiclastic, origin for the section. Despite recrystallization of microtextures, primary sedimentary structures and textures are well-preserved and, for convenience, the prefix 'meta-' will be dropped from 'meta-tuff', 'meta-andesite,' etc.
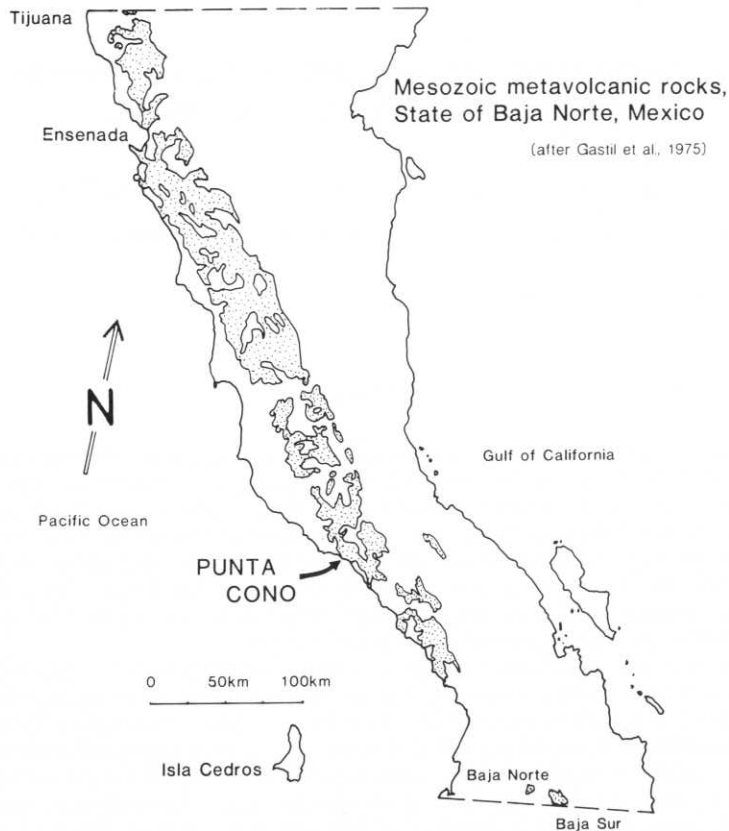
Fig. 1. Index map showing the location of Punta Cono and distribution of pre-batholithic meta-volcanic rocks (stippled) in Baja California Norte, Mexico. Most of the metavolcanics lie just west of the Peninsular Ranges Batholith and belong to the Alisitos Group.

Volcaniclastic rocks exposed in the sea cliffs at Punta Cono, previously mapped as pre-batholithic volcanic rocks (Fig. 2; Gastil *et al.*, 1975), are here included in the Alisitos Group, because of the presence of Early Cretaceous ammonite fossils ('probably 2 genera . . . not as old as Jurassic nor younger than Albian'; LouElla Saul, private communication). These sea cliffs provide an excellent opportunity to study deep marine depositional products on the steep flanks of an island arc volcano, and to examine the products of interaction of magma with wet sediment in this setting. Examination of coarse-grained, near-source submarine volcaniclastic sequences is particularly important because evaluation of sedimentary structures and stratification sequences is extremely difficult in analogous modern settings (e.g. Sigurdsson *et al.*, 1980), largely due to the difficulty of coring coarse-grained sediment. Moreover, processes controlling

deposition and remobilization of sediment in slope settings play important roles in determining depositional style in more widespread basinal settings.

SEDIMENTOLOGY

Very fine-grained tuffs to coarse-grained tuffs, lapilli tuffs, and tuff breccias (Fisher & Schmincke, 1984) make up the bulk of the section at Punta Cono (Fig. 3). Marine fossils and branching, horizontal burrows are present. Absence of wave-generated sedimentary structures, such as symmetrical ripples or bi-directional palaeocurrent indicators, indicates deposition below storm-wave base. Calcareous beds (tuff with shell fragments) occur locally. Shallow-level intrusive rocks, in complex contact relationships with the pyroclastic rocks, are also present.
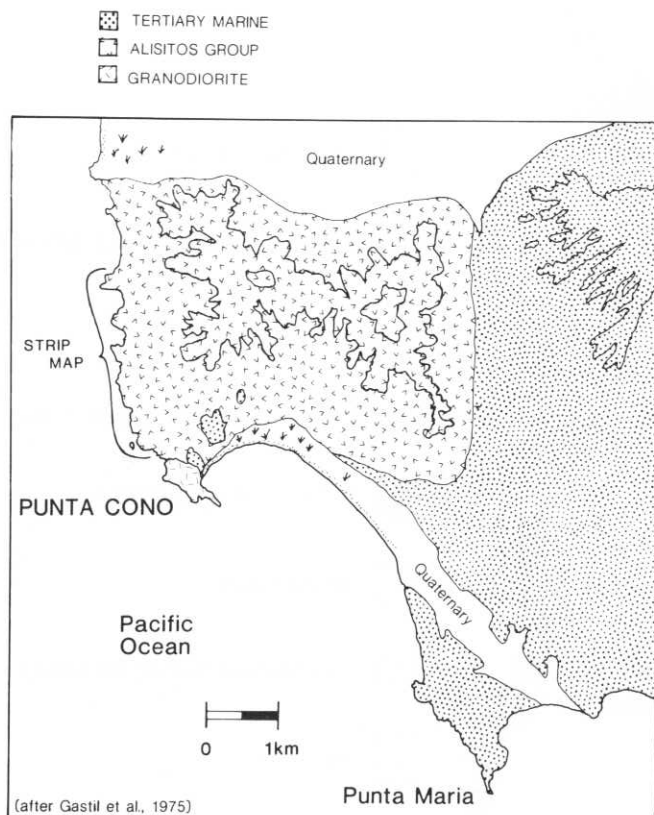
GEOLOGICAL MAP OF
PUNTA CONO AREA

Fig. 2. Map of Punta Maria-Punta Cono area, showing distribution of Alisitos Group rocks of Aptian/Albian (early Late Cretaceous) age.

Fine-grained tuff

Fine- to very fine-grained tuffs make up about a quarter of the section at Punta Cono (Fig. 4). Recrystallization obscures fine grain-size trends, but most of the well-sorted, thinly laminated, flat-bedded tuffs (Fig. 5) are characterized by convolute bedding (Fig. 6), cut and fill lamination, and rare fluid-escape features, indicating rapid deposition and current activity. Finely laminated, unrippled tuffs with non-erosive bases are present locally. Truncated slump folds were produced at the sea-floor (Fig. 4). The shallow, centimetre-scale scour and fill of the majority of fine-grained tuffs suggests erosion by the head or body of turbidity currents that bypassed the area, followed by deposition from their dilute tails. Palaeo-

currents indicated by rare ripple cross-lamination suggests that the currents travelled directly downslope (westward, perpendicular to most slump-fold axes), and were not deposited from overbank flows moving outward from channels. Some of the beds may have been deposited from dilute, slow-moving ash-bearing lutite flows (McCave, 1972) that developed when ash was delivered to the marine environment by run-off during storms, or when ash in shallow marine environments was re-suspended by storm waves (Moore, 1969). Some of the finely laminated, unrippled tuffs with non-erosional bases probably represent subaqueous fallout from ash emissions of neighbouring arc volcanos. Coarse-grained tuff and lapilli tuff beds occupy channels eroded into the flat-based, fine-
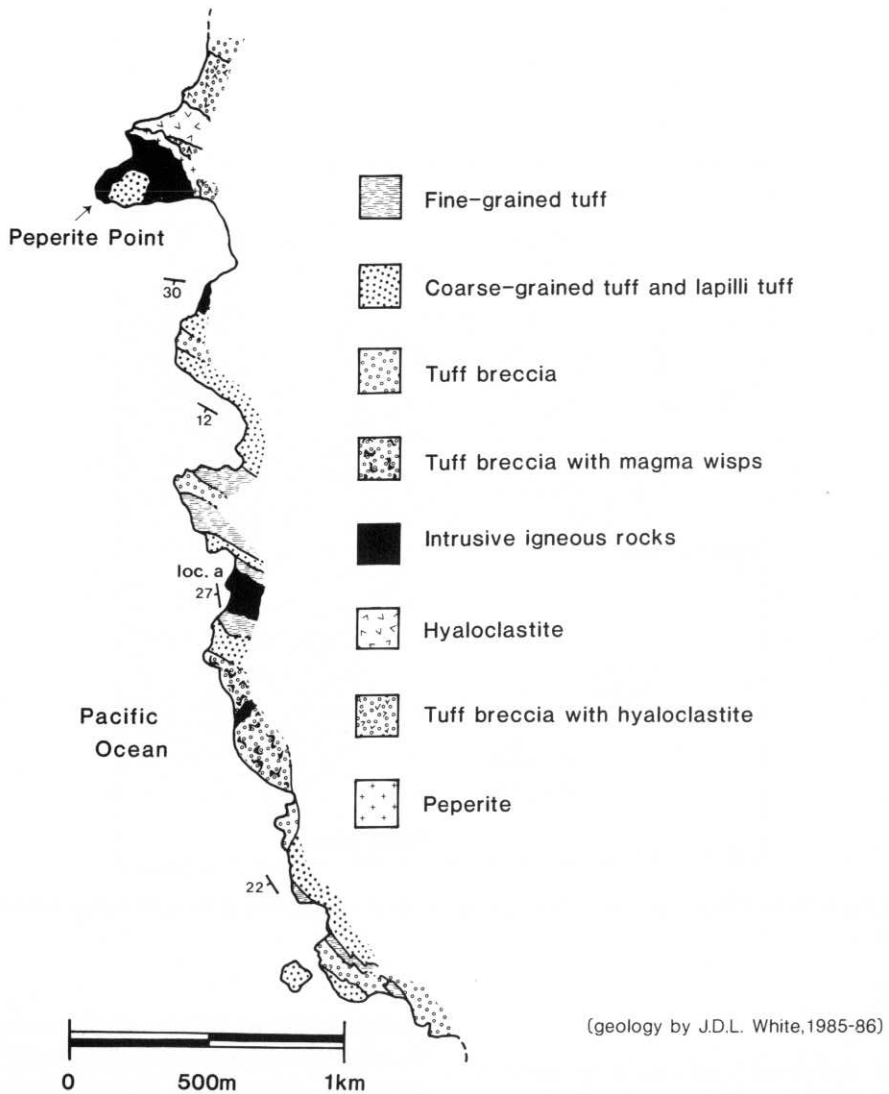
COASTAL STRIP MAP OF PUNTA CONO AREA

Fig. 3. Coastal strip map of the Punta Cono area.

grained tuffs, which remain as lenticular erosional remnants.

Coarse-grained tuffs and lapilli tuffs

Sequences of thin- to medium-bedded coarse-grained tuff and lapilli tuff occupy broad, shallow channels (width/depth ratios > 20:1; Fig. 7) and, rarely, steep-sided channels containing many nested, shallower channels (e.g. at 230 m, Fig. 4). The rocks consist of unrounded, aphanitic or plagioclase–phyric volcanic lithics. Outsize clasts, such as cobble-sized clasts in granule-grade lapilli tuff, are not uncommon. Bedding commonly pinches out, either erosionally or against channel margins (Fig. 7). Channels overlap and are incised into one another with little to no preserved finer-grained material, comprising multilateral channel complexes (Galloway, 1981). High-concentration
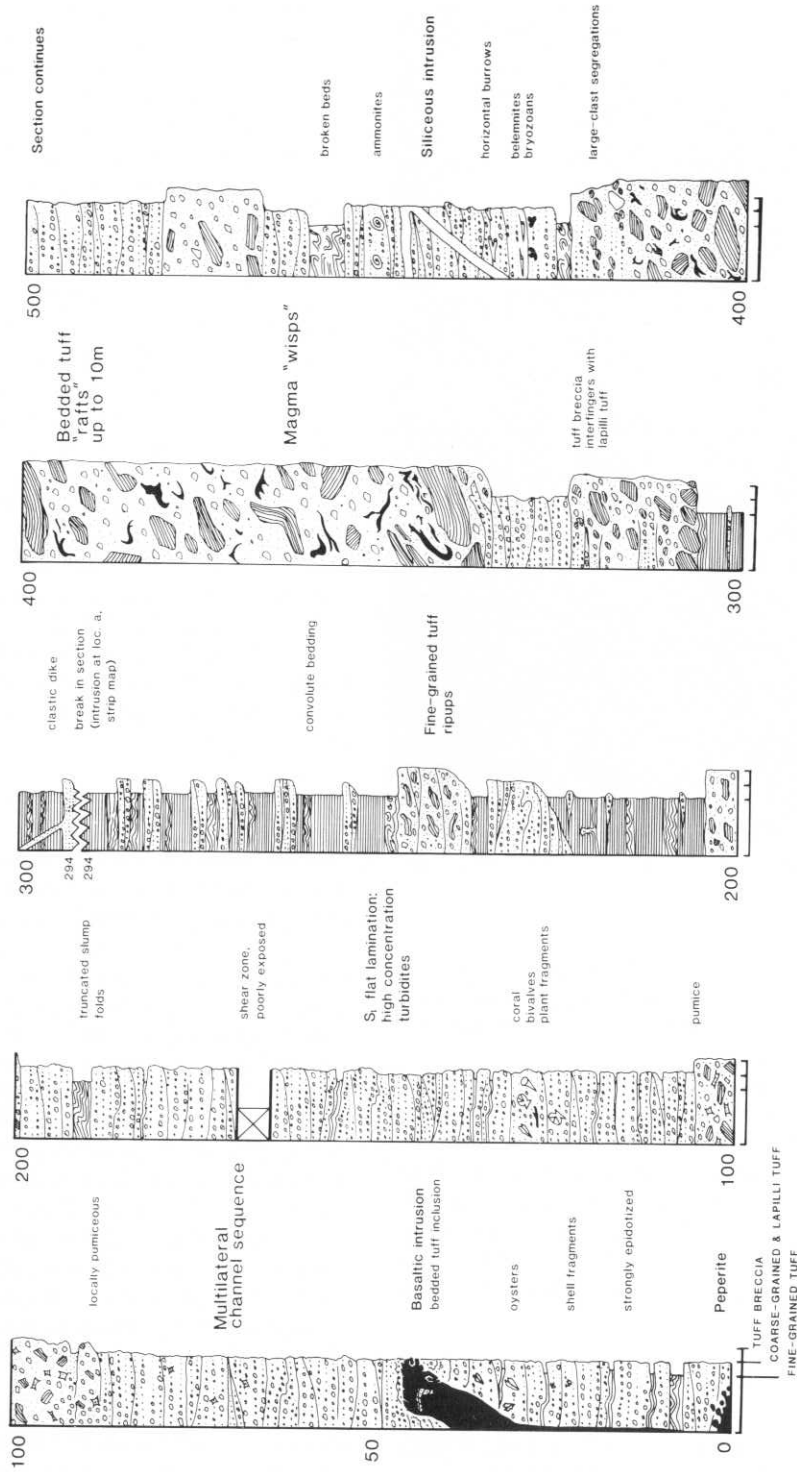
PUNTA CONO MEASURED SECTION

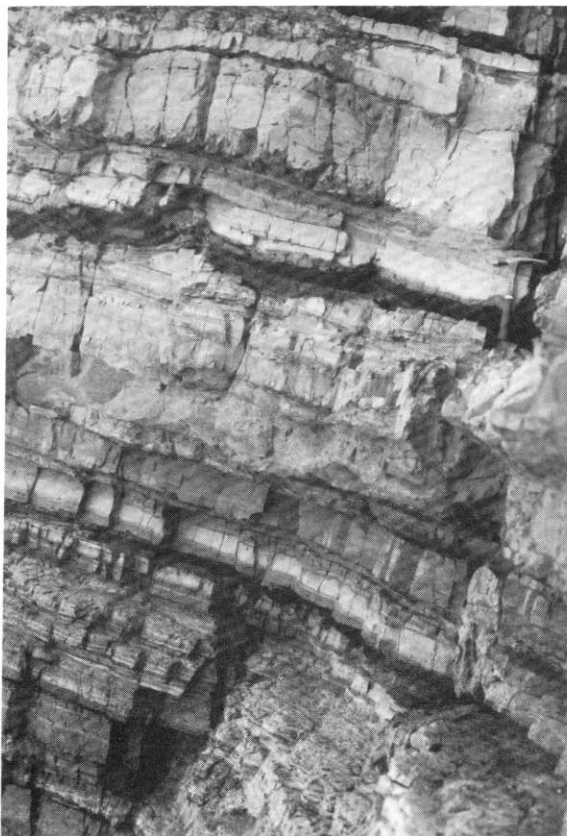**Fig. 4.** Measured section (in metres), Punta Cono.

Fig. 5. Flat-bedded, finely laminated fine-grained tuffs (hammer is 0·32 m).

turbidity currents deposited the majority of coarse-grained tuffs and lapilli tuffs as S_1 layers (Lowe, 1982), characterized by discontinuous flat lamination with gradational boundaries.

The nested channels and lack of associated fine-grained rocks within the channel complexes suggests deposition from migrating, ephemeral, bedload-dominated channels arranged in braided patterns. These are broadly similar to the deposits of braided systems dominated by high-concentration streamfloods and sheetfloods (e.g. Galloway, 1981; Schumm, 1960, 1977), or braided channels on deep-sea fans (e.g. Walker, 1975).

Tuff breccias

Tuff breccias at Punta Cono contain dark- to light-coloured aphanitic volcanic lithic fragments, plagioclase–phyric, dark-coloured volcanic lithic fragments,

and, in some cases, pumice. Grading is absent, and sorting is extremely poor in these matrix-supported tuff breccias. Grain size varies continuously from ash (recrystallized) to decimetre blocks. They are invariably rich in intraformational fragments of fine-grained tuff, which make up as much as 20% of each deposit (Fig. 4), and are commonly somewhat larger than associated lithic blocks.

Tuff breccia is commonly associated with lapilli tuff in channel complexes, and at one locality, amalgamated, matrix-depleted intraclast breccia and tuff breccia grade laterally into bedded lapilli tuffs (Fig. 8) over about 10 m. Tuff intraclasts are abundant only in the tuff breccias, and are rare in the lapilli tuffs. This may indicate that only the very coarsest-grained sediment gravity flows were capable of eroding or destabilizing underlying strata as they moved down the slope. Alternatively, the bedded tuff may have been incorporated into the flows when slumping occurred. The bedded tuff was transported intact within the matrix-rich debris flows that deposited the tuff breccias. Bedded tuff clasts are not associated with the lapilli tuff units because high-concentration turbidity currents either destroyed large bedded tuff intraclasts or segregated them from lapilli-size material.

'Wisp-bearing' tuff breccia

An extremely (90 m) thick tuff breccia bed (Figs 3 and 4) is inferred to have been produced when magma intruded, caused remobilization of, and became mixed with unconsolidated sediments in the section. This thick, polymict tuff breccia contains light and dark aphyric and plagioclase phyric volcanic rock fragments supported in a dark, ashy matrix of similar composition. It is notable for containing numerous 'wisps' of basalt in the shape of small (commonly decimetre to metre size), plastically deformed sheets, or rags (Fig. 9). Wisp margins are cryptocrystalline, presumedly due to chilling effects, whereas wisp interiors are aphyric to sparsely plagioclase-phyric basalt. Lithic fragments are commonly enveloped by the wisps (as beyond hammer's head in Fig. 9a). The tuff breccia also includes bedded, relatively undeformed 'rafts' of fine-grained tuff, up to 20 m long, that moved within the flow, as well as more deformed, smaller intraclasts.

The upper part of the redeposited tuff breccia includes discontinuous layers of large (commonly 0·3–1 m) clasts, either forming framework layers or supported by a poorly-sorted, fines-depleted, matrix

Fig. 6. Horizon of broken beds of fine-grained tuff. Note inconsistent vergence directions in folded bedding remnants (hammer is 0·32 m).
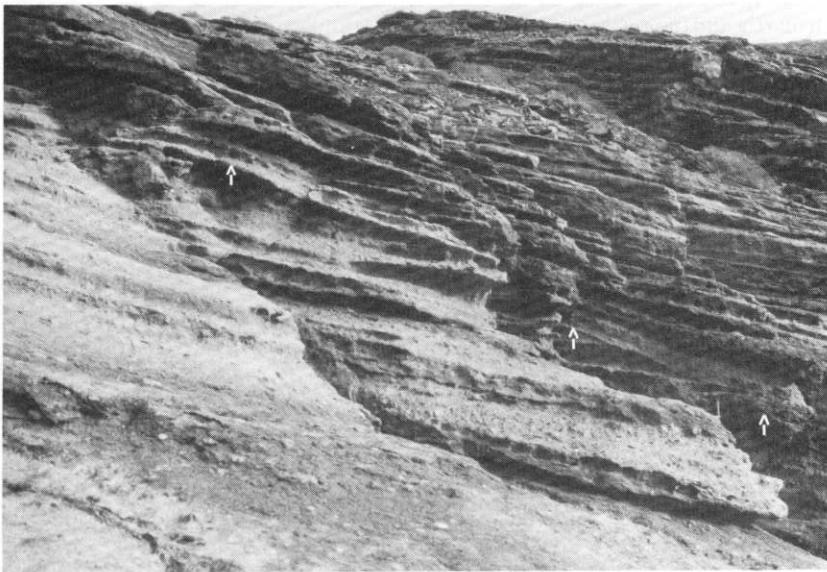

Fig. 7. Coarse-grained tuff and lapilli tuff occupying shallow, multilateral scours. Arrows denote base of larger channel: lower parts of the margin show one or more draped layers of lapilli tuff, infill beds pinch out directly against upper channel margins. Multilateral depositional geometries suggest bedload deposition in a system of ephemeral, braided channels (hammer is 0·32 m).
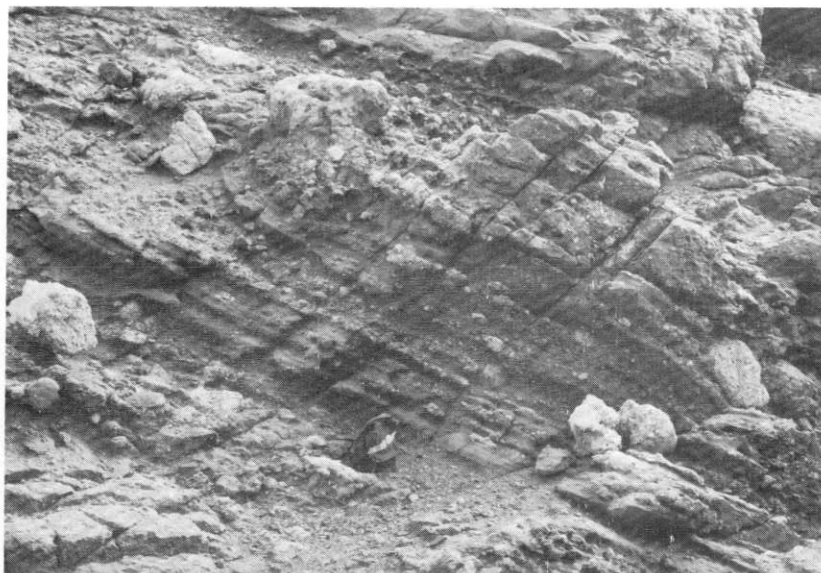
Fig. 8. Lapilli tuff interfingers with tuff breccia. 5 m north (to the right) of this exposure, lapilli tuff beds pinch out and tuff breccia beds are amalgamated in a >10 m thick channel fill (backpack, approximately 0·35 m high, for scale).

of tuff breccia (Fig. 4). The layers are from one to several clasts thick, with rare clasts up to twice the layers' thickness, and have gradational upper and lower contacts. Basalt wisps are very small (centimetres) and rare at this level of the flow, but the similarity of both matrix and (non-wisp) clasts, as well as the absence of a distinct bedding surface separating these lenses from underlying tuff breccia with abundant wisps, supports deposition from a single flow.

It is inferred that, in the initial phase, magma intruded wet sediment, moving in multiple lobes and

fingers into a host further weakened by expansion as steam or supercritical fluids were generated along the magma's margin. This process allowed a considerable amount of magma to become interfingered with and mixed into the sediment pile (Fig. 10b). Furthermore, movement and expansion of steam or supercritical fluids within the sedimentary host may have produced thixotropic conditions, a process described by Koke-laar (1982). This, accompanied by physical jostling by the intruding magma, destabilized the sedimentary pile, initiating a large slump (Fig. 10c). Only the fine-

Fig. 9. Examples of lava wisps (dark) in remobilized tuff breccia (development of wisps discussed in text). Light coloured, angular clasts are felsic volcanic rock fragments. Note extremely irregular shapes of wisps, suggesting emplacement as hot, ductile fragments (hammer is 0·32 m).
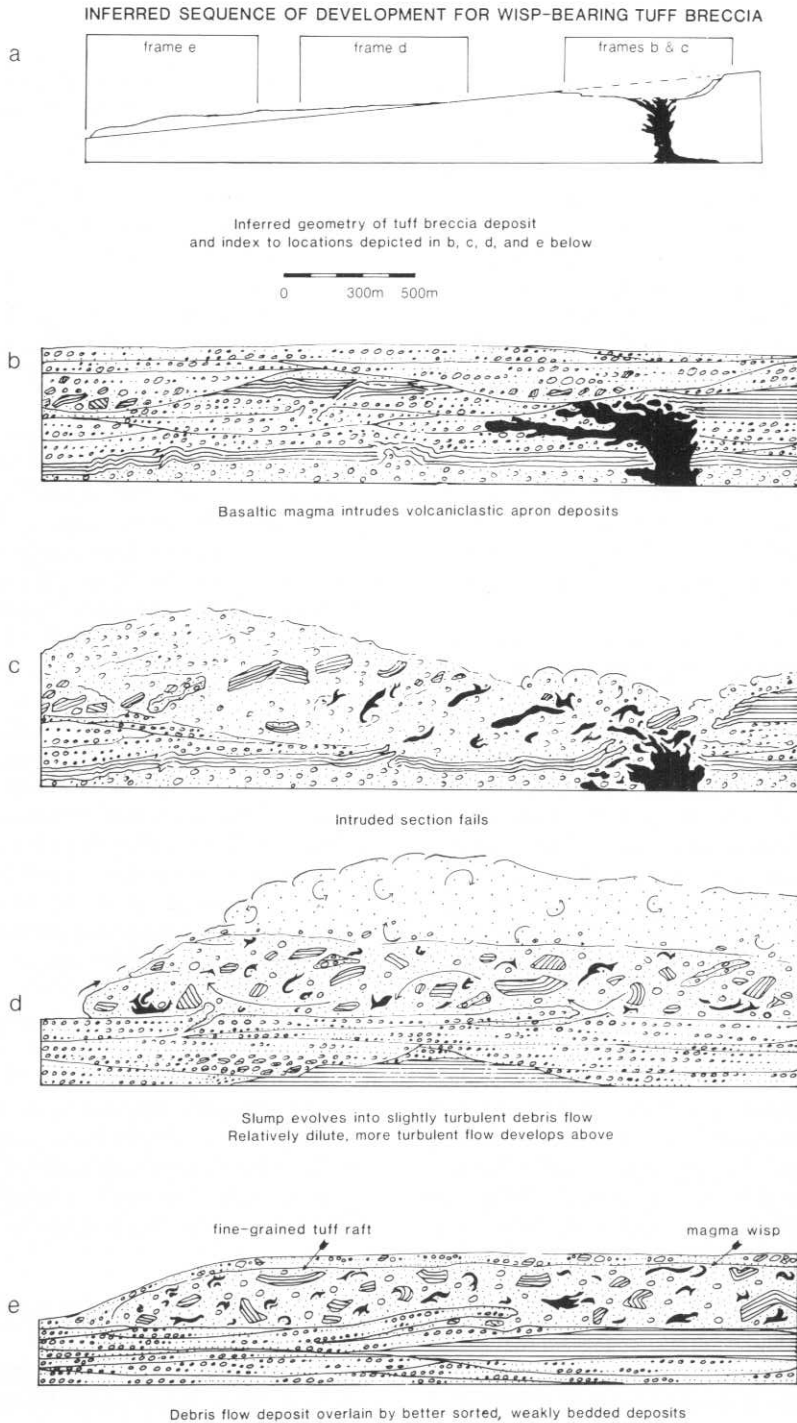
INFERRED SEQUENCE OF DEVELOPMENT FOR WISP-BEARING TUFF BRECCIA

a

frame e   frame d   frames b & c

Inferred geometry of tuff breccia deposit
and index to locations depicted in b, c, d, and e below

0   300m   500m

b

Basaltic magma intrudes volcaniclastic apron deposits

c

Intruded section fails

d

Slump evolves into slightly turbulent debris flow
Relatively dilute, more turbulent flow develops above

e

fine-grained tuff raft   magma wisp

Debris flow deposit overlain by better sorted, weakly bedded deposits

Fig. 10. (a) Sketch showing inferred depositional geometry of wisp-bearing breccia deposit, with locations of subsequent figure frames indicated. The inferred sequence of events responsible for development of the wisp-bearing tuff breccia is illustrated in frames (b–e), not to scale. (b) Magma intrudes base of mixed section, forms fingers and lobes extending through several metres of section. (c) Continued intrusion destabilizes section, and slumping occurs. (d) During downslope movement, the flow dilates, and weak turbulence mixes intraformational fragments and magma wisps throughout most of flow. (e) Flow comes to rest. Wisps and intraformational fragments throughout breccia except for uppermost few metres, where alternating coarse/ fine zones indicate winnowing and sorting by more highly turbulent zone developed above main flow (elutriation and gravitational segregation).

grained tuff beds remained cohesive, and the slump summarily disaggregated, producing a weakly turbulent high-concentration sediment-gravity flow (Fig. 10d). The large, bedded rafts (Fig. 10e) indicate that the flow travelled a relatively short distance (Fig. 10a). Fingers and lobes of the basalt intrusion remained fluidal as they moved downslope with the slump, becoming torn into small, plastically deformed 'wisps' that were mixed through much of the sediment gravity flow (Fig. 10d, e). Although magma wisps have large surface area/volume ratios that would favour rapid cooling, they clearly remained fluidal until the time of deposition, further suggesting limited transport. The degree of size segregation developed in the top of the flow suggests that the flow came to rest in pulses, the deposit of each pulse including one fines-depleted horizon (Fig. 10e). Dilution of the flow may have occurred by ingestion of water at the flow's head (*cf.* Hampton, 1972).

The extremely fluidal shapes of the magma wisps indicate that hydroclastic fragmentation, whether by steam explosions, spalling, or thermal shock, was ineffective within the flow. Steam explosions are damped by increased confining pressures until the critical pressure is reached, beyond which explosive expansion of water to steam does not occur. Hydrostatic pressures for sub-wave base (> 200 m) depositional sites are in excess of 20 bars, sufficient to reduce water/steam volumetric expansion by more than 2/3 compared with subaerial values (see Kokelaar, 1982), greatly reducing the explosivity of magma–water–sediment interactions. Because one of the major processes responsible for hydroclastic fragmentation is inhibited, hydroclastic fragmentation is in general less common in deep marine settings than in shallow marine settings (below Pressure Compensation Level; Fisher, 1984), although studies from submersibles (*e.g.* Lonsdale & Batiza, 1980) suggest that steam fragmentation occurs to water depths of at least 2000 m. The association of unfragmented basalts with hydroclastic basalts (described below) in the Punta Cono section and on modern seamounts supports Lonsdale & Batiza's (1980) interpretation that another factor is more important. These authors suggested that hydroclastic fragmentation is favoured by magma of higher viscosity which, for magmas of constant composition, is determined by temperature (see also Shimozuru, 1986); fragmentation is thus favoured by low rates of discharge from the magma chamber through vents and surface flows.

The complex forms of the fluidal magma wisps indicate slumping of and subsequent redeposition of

the intruded sediment mass while the magma was still hot and ductile. Lorenz (1984) described mud–magma mixtures in which extremely fluidal lava forms were deformed by sediment remobilization. Cohesive sediments were involved in that case, which may have limited the amount of flow dilation and mixing relative to the tuff breccia described here.

FOSSILS AND SEDIMENT DERIVATION

All body fossils observed at Punta Cono occur within the coarse-grained tuffs, lapilli tuffs and tuff breccia, suggesting that they were carried to their present location by high-concentration turbidity currents. Branching, horizontal burrows (approximately 10 mm thick) also occur within the coarse-grained and lapilli tuffs. Body fossils tend to occur in groups of the same type, perhaps indicating that individual flows tapped areas dominated by different organisms. Disarticulated 'oyster' shells, for instance, were probably deposited by flows generated in shallow water. In one locality dozens of ammonites litter the surfaces of two beds. Elsewhere a few beds are rich in belemnite debris, and others contain abundant disarticulated, broken and whole scallop-like bivalve shells. These shells occur only in association with block-sized lithic fragments deposited in groups within lapilli tuff, indicating that they were transported to their present location along with the lithic clasts.

Large, thick, broken and whole oyster-like shells originated in shallow water, high-energy environments, probably along the coast of an island volcano. Rare wood fragments also point to subaerial derivation for some sediment components. A single occurrence of rounded clast conglomerate suggests fluvial or shoreline reworking of some material, but most debris (volcanic rock fragments, crystals and pumice) reached the site of deposition unabraded.

ENVIRONMENT OF DEPOSITION

Pyroclastic rocks may be deposited directly from surges, or from subaerial or subaqueous pyroclastic flows or fall-out; alternatively, they may result from remobilization and redeposition of pyroclastic debris by streams, debris flows, or submarine sediment gravity flows (Fisher & Schmincke, 1984). The absence of such 'fingerprints' of submarine eruption as doubly graded tephra sequences (Fiske & Matsuda, 1964),

combined with other features considered below, indicate that the bulk of the pyroclastic rocks exposed at Punta Cono were erupted from subaerial volcanos. Fine-grained tuffs in the Punta Cono section were probably produced by ash emissions from nearby volcanos, largely redeposited or reworked by turbidity currents. The polylithologic nature of the clasts in the lapilli tuffs and tuff breccias indicates remobilization of the coarse pyroclastic debris as well, although its largely unabraded nature suggests this was limited. This remobilization and redeposition must have occurred relatively close to the vent region, on the flanks of an active volcano where fissures or surface flows (*cf.* Fuller, 1931) fed basaltic intrusions penecontemporaneous with sedimentation.

Thick wedges of coarse-grained, texturally and compositionally immature sediment may prograde into marine environments from adjacent highlands (fans-deltas; Wescott & Etheridge, 1984) or on the flanks of island arc volcanoes (arc aprons; Hickey, 1984). The Punta Cono section represents the deep marine part of an arc apron, although a non-marine to littoral source for the detritus is inferred. The preservation potential of subaerial parts of arc aprons, or of fan-deltas along faulted coastlines, is low, but marine fan-delta facies are common in the geologic record (Hickey, 1984; Surlyk, 1984; Busby-Spera & Keller, 1985; Busby-Spera & Boles, 1986). On modern arc aprons of the Lesser Antilles, pyroclastic flows, lahars and post-eruptive streamfloods carry tephra to the coastline to be deposited largely as sediment gravity flows on marine parts of the arc apron (Sigurdsson *et al.*, 1980). Ancient deep marine arc apron sections like the Punta Cono section commonly have lava flows and shallow level intrusions (Hayward & Ballance, 1982; Hickey, 1984; Busby-Spera, 1987, indicating proximity to central vents or related fissures; this requires steep topographic gradients and a very narrow (or non-existent) shelf.

Active volcanic aprons have been likened by Hickey (1984) to alluvial fans because of their relatively limited size, steep surface slopes (1–10°), coarse average grain size, and tendency toward catastrophic deposition. The similarity of deposits from the two environments is underscored by the evolution of thought concerning the Miocene Piha Formation of New Zealand. Hayward & Ballance (1982) originally interpreted the Piha Formation as an alluvial fan, deposited 'from streams, flash floods, and debris flows on poorly vegetated fans at the foot of volcanic cones ...' The Piha Formation has subsequently been interpreted as a marine deposit on the basis of pillow

lavas and fossils (Ballance & Hayward, 1983), but the authors emphasize that the Piha 'resembles known montane alluvial facies more closely than known deep-water conglomerate facies'. The present study provides further evidence that bedload deposition of coarse-grained sediment produces similar depositional geometries and facies relationships in streams (high-concentration floodflows) and high-concentration turbidity currents (populations 2 and 3 of Lowe, 1982).

Like alluvial fan deposits, the internal organization of fan-delta or arc apron deposits is poor relative to deep-sea fan or deltaic deposits because sedimentation is controlled by random, external tectonic or volcanic processes, rather than by predictable autocyclic processes, such as channel migration or lobe switching. For example, the Middle Jurassic Gran Canon Formation on Cedros Island (Baja California) represents a largely pyroclastic arc apron that prograded into a back-arc rift during the growth of an island arc volcano (Busby-Spera & Keller, 1985; Busby-Spera, 1987). Internal organization of the apron is poor, because sedimentation was the direct result of acyclic eruptive activity. The Lower Cretaceous Asuncion Formation on the Vizcaino Peninsula (Baja California) preserves the marine part of a very coarse-grained fan-delta system that accumulated along fault scarps in a forearc region (Busby-Spera & Boles, 1986). Sedimentation was triggered by random seismic or flood events, resulting in a lack of internal organization, although an overall upward fining of the formation probably reflects erosional retreat of scarps after faulting ceased. The lack of vertical organization of the vertical sequence at Punta Cono probably reflects the unpredictability of such processes as eruptions, winds, seismic shaking and storms as agents of transport of tephra.

IGNEOUS ROCKS

At least three generations of intrusive activity have affected the Punta Cono section. An early, syndepositional group of basaltic dikes and sills are present. A later set of basaltic dikes, straight sided with a predominant NNW/SSE orientation, cut the section after lithification. All but the largest of these straight-sided basaltic dikes have been omitted from the coastal strip map (Fig. 3). Lastly, a group of silicic sills with pronounced flow banding or, rarely, columnar joints, cuts the basaltic dikes.

Syndepositional igneous rocks

Both extrusive and shallow-level intrusive syndeposi-
tional igneous rocks occur within a well-exposed
hyaloclastite-peperite complex at the northern end
(the base) of the Punta Cono section described in this
paper. The complex comprises a mixture of bedded
volcaniclastic rocks, intrusive and extrusive fragmen-
tal basalt, and non-fragmental basalt (Fig. 11). Figure
12a is a cross section showing a generally tabular
basalt intrusion within stratified volcaniclastic rocks.
Note the presence, from base to top, of (1) tuff breccias
and tuffs containing no basalt fragments; (2) overlying
tuff breccia, containing hydroclastic basalt blocks
(blocky fragments formed by spalling, chilling and/or
steam explosion), jigsaw-puzzle hyaloclastite basalt
blocks, and breccia-enveloping fluidal basalt sheets;
(3) massive jigsaw-puzzle hyaloclastite, containing
some rafts of (1) and (2); (4) a second section of
bedded coarse-grained tuffs and lapilli tuffs without
hydroclasts; (5) peperite, a magma–sediment mixture
that includes all of the above rock types plus blocks of
bedded calcareous tuff and fossiliferous tuff set in a
tuff matrix; and (6) the tabular intrusive body, cutting
through the section at a low angle and containing rafts
of bedded tuff.

Figure 12b–e illustrates the sequence of events
inferred to have produced the complex. The lowest
unit exposed at Peperite Point (Fig. 11) is massive tuff
breccia, overlain by 4 m of medium bedded tuff and
lapilli tuff (1). Overlying the bedded tuffs is a massive
'mixed' tuff breccia (2) that contains, in addition to
the usual volcanic lithic fragments, abundant blocky
basalt fragments, 'packets' of jigsaw-puzzle hyaloclas-
tite (rock formed by hydroclastic fragmentation of
lava with little or no subsequent movement of clasts
relative to one another), as well as rare sheets of basalt,
approximately a metre thick and exposed over tens of
square metres on bedding-plane surfaces. Clasts
identical to those in the tuff breccia, as well as small
inclusions of tuff breccia and tuff, are locally present
within the basalt sheets. The sheets' margins are
irregular, un-pillowed, and tend to 'detour' around
large clasts.

We infer that the mixed tuff breccia (2) was
produced by shallow-level intrusion of basalt into
unconsolidated sediments, with concomitant remobil-
ization and redeposition of the resulting mixture (Fig.
12b). The presence of unfragmented lava sheets
indicates a nearby source of low-viscosity magma (the
vent area is not exposed at Punta Cono). Enveloped
tuff breccia clasts and bits of tuff breccia further
support an intrusive rather than an extrusive origin,
as the sediments would be more easily trapped by a
magma moving through them than by a flow passing
overhead. The jigsaw-puzzle hyaloclastite packets and
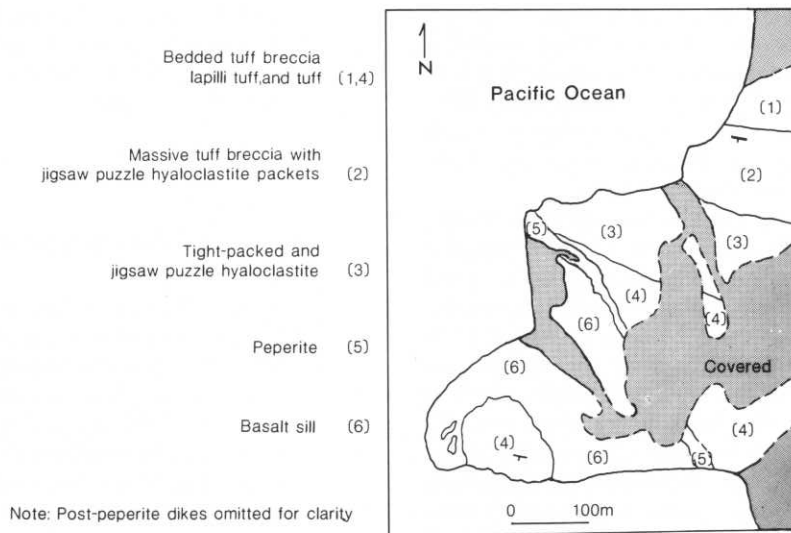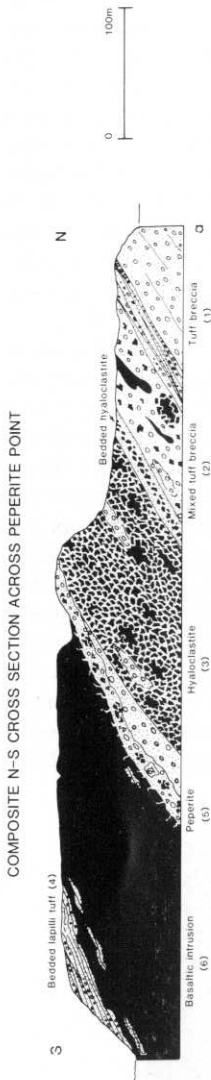
MAP OF PEPERITE POINT

Bedded tuff breccia
lapilli tuff,and tuff   (1,4)

Massive tuff breccia with
jigsaw puzzle hyaloclastite packets   (2)

Tight-packed and
jigsaw puzzle hyaloclastite   (3)

Peperite   (5)

Basalt sill   (6)

Note: Post-peperite dikes omitted for clarity

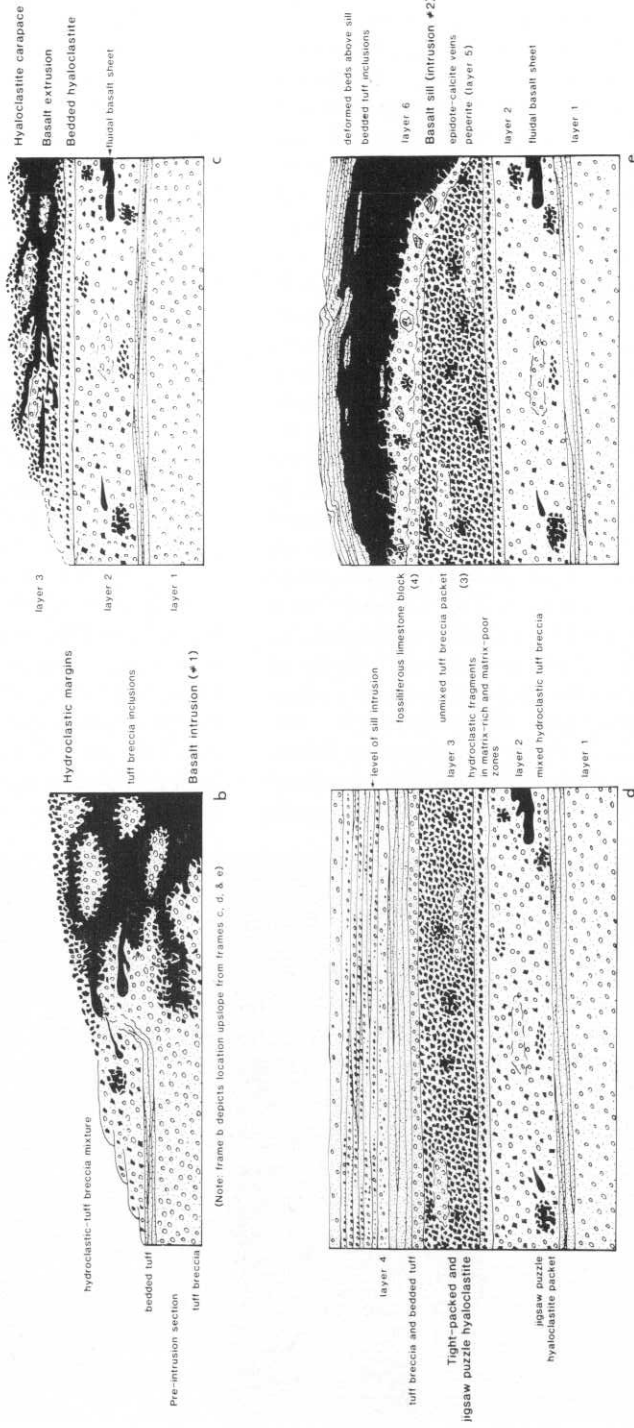**Fig. 11.** Geological map of Peperite Point (see Fig. 3).

Fig. 12. Composite cross section of Peperite Point (a), and schematic representation of sequence of events (frames (b–e), not to scale) inferred to have produced the hyaloclastite–peperite complex. (b) Coarse-grained and lapilli tuffs are intruded by magma. Magma chills and spalls (hydroclastic fragmentation) at its margins, and moves into and mixes with tuffs. Slumps and debris flows result from the disturbance, and carry away mixtures of hydroclastic fragments and host sediment. (c) Magma and associated hyaloclastite (jigsaw puzzle hydroclastically fragmented magma) carapace penetrate sediment cover and override previous deposits. Fragmenting magma advances within hyaloclastic carapace. Some re-sedimentation of hydroclastic fragments occurs ahead of advancing mass. (d) Magmatic activity ceases. Tuffs and lapilli tuffs, some with a significant shell fragment component, are deposited. (e) A second intrusion occurred after deposition of an unknown thickness of tuff. Peperite developed along lower margin of the roughly tabular body, while the top is marked by epidotized rafts of tuff incorporated by the magma. Margins of intrusion are increasingly fractured and veined near both upper and lower margins.

blocky hydroclastic fragments may have formed along the intrusion's margins where the magma was cooler and more viscous (Lonsdale & Batiza, 1980).

The mixed tuff breccia (2) is overlain by several metres of weakly bedded fine-grained hyaloclastite (aquagene tuff; Carlisle, 1963) and hydroclastic tuff breccia made up of blocky (hydroclastic) basalt fragments set in a fine-grained hyaloclastite matrix. The matrix of the hydroclastic tuff breccia is identical to that in hyaloclastite horizons lacking larger fragments, and the boundaries between matrix-rich and matrix-poor horizons are gradational. It is inferred that the sequence represents a re-sedimented, entirely hydroclastic, sequence. The hydroclastic fragments may have been derived from the same magma feeding the mixed breccia, as continued basalt eruption penetrated the sediment cover and exposed masses of hydroclastically fragmenting basalt at the surface (Fig. 12c). Kokelaar, Bevins & Roach (1985) illustrate similar synchronous intrusion, extrusion and reposition of hydroclastic debris associated with an Ordovician rhyolite. Tight-packed to jigsaw-puzzle hyaloclastite (Fig. 13; layer 3, Fig. 12) makes up a hyaloclastite layer of variable thickness that contains scattered inclusions of tuff breccias and tuffs (1). It conformably overlies the indistinctly bedded aquagene tuffs and hydroclastic tuff breccias (Fig. 12d). Also present within the hyaloclastite layer are blocks, up to several metres thick, of internally unfragmented lava (Fig. 12c). The boundaries of these blocks grade

into the surrounding, fully fragmented hyaloclastite, and represent bodies of basalt that escaped complete fragmentation. The simplest explanation for preservation of such blocks involves variable rates of magma supply. Large volumes of magma delivered in rapid pulses expose an initially small surface-area (per magma volume) to water (the hydrofracturing medium), and are thus less likely to be completely fractured than magma exposed to water in smaller increments. The hyaloclastite layer (3) may have been emplaced in the manner envisioned by Walker & Blake (1966), whereby magma moves forward through its own hyaloclastite debris, continually sending pulses of lava further into a growing clastic carapace (Fig. 12c).

After accumulation of an unknown thickness of section (Fig. 12d), a basalt sill (layer 5; Fig. 12e) was intruded, cutting one edge of the hyaloclastite body and entering a section of bedded tuffs, calcareous tuffs and tuff breccias. The base of the sill shows evidence of mixing with a wet, largely unlithified host. This evidence includes: (i) a 1–10 m thick 'mixed horizon' [(4), Fig. 12e] directly below the sill in which bedding is absent and blocks of the basalt are dispersed from the sill into a matrix of calcareous tuff breccia with occasional blocks of bedded tuff and fossiliferous tuffaceous limestone (Fig. 14). The basalt blocks were hot at the time of incorporation into this mixture, as indicated by development of jigsaw-puzzle hyaloclastite, and (ii) epidote–calcite skarn blocks, up to 2 m in
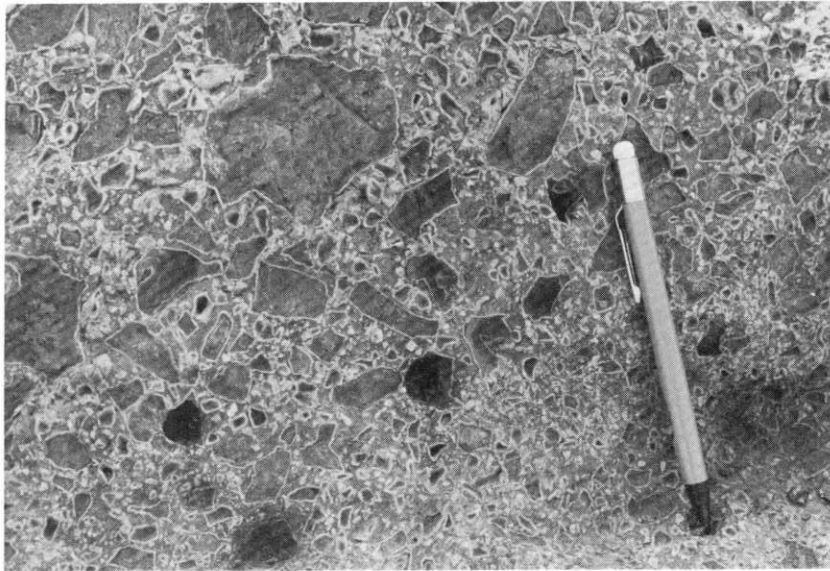

Fig. 13. Photograph of well-developed jigsaw puzzle hyaloclastite. Distinct chill(?) rims are recrystallized (Pencil is 0·14 m).

Fig. 14. Photograph of polymict peperite developed at western edge of intrusion. Intrusion just beyond photo at lower left. Some blocks of magmatic origin with solid outline, tuff blocks with dashed outline, fossiliferous limestone block with dotted outline (hammer is 0·32 m).

diameter, dispersed through the lower 5–10 m of the sill (Fig. 12e). Calcite/epidote veins, possibly developed at the same time, also cut the base of the basalt (Fig. 12e). The mixed zone formed by nearly complete rehomogenization of an originally bedded section. Only the beds of fossiliferous tuffaceous limestone or calcareous fossiliferous tuff remained coherent, possibly due to early cementation of the carbonate-rich beds; these beds were disrupted and dispersed to form blocks in the mixed horizon and skarn blocks within the intrusion. Homogenization of the unlithified host sediment and accompanying mixing of basalt hydroclasts through it may have been accomplished by: shearing at the base of the intrusion; thixotropic failure of the sediment as interstitial water was converted to steam or supercritical fluid; steam

explosions along the base of the intrusion; or some combination of the three. The absence of a mixed horizon at the top of the sill favours the interpretation that the sill (Fig. 12e) was emplaced very soon after the hyaloclastite body (Figs 12c and d), before much loading and de-watering of the bedded section took place by burial; thus, the sediments beneath the weight of the sill were more likely to fail than those above. Alternatively, the sill could represent part of a burrowed lava flow (Fuller, 1931), in which case the tuffs at the top of the sill may have been carried 'piggy-back,' while those beneath were mixed as discussed above.

SUMMARY AND CONCLUSIONS

Coarse and fine-grained volcaniclastic rocks were deposited in deep marine portions of a fan-delta complex (volcaniclastic apron) built onto the steep flanks of an andesitic stratovolcano. Evidence suggests limited subaerial reworking and transport of pyroclastic debris. Unreworked pyroclastic flow deposits are absent from the section. Multi-lateral channel geometries suggest deposition in a sedimentary system similar to those found in alluvial fans. Limited preservation of fine grained material and predominance of 'bedload' deposits from the bases of high concentration sediment gravity flows reinforce the analogy.

Syndepositional extrusion of mafic magma on the submarine flanks of the island arc volcano formed hyaloclastite deposits. Most syndepositional magma formed shallow sills, mixing with sediments. Predominance of shallow intrusions over extrusions in areas of water saturated sediment was predicted by Mc-Birney (1963) on the basis of relative densities of magma and wet sediment. Shallow 'intrusions' developed from flows burrowing beneath low-density sediments (Fuller, 1931) could produce similar field relationships and marginal features, but would imply initial extrusion in areas lacking an unlithified, low-density sediment cover.

Evidence for syndepositional intrusion includes peperite complexes and magma–sediment remobilization mixtures. Syndepositional shallow-level intrusions may constitute much of the section preserved in some cases. The intrusions can also drive sedimentation by causing sediment failure and remobilization.

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REFERENCES

BALLANCE, P.F. & HAYWARD, B.F. (1983) Piha Formation volcanic conglomerates (Waitakere Group, Lower Miocene, Auckland) – a sedimentological conundrum. *Misc. Pub. geol. Soc. N.Z.*, **30A**.

BEGGS, J.M. (1984) *Stratigraphy, petrology, and tectonic setting of the Alisitos Group, Baja, California, Mexico.* Unpublished PhD Thesis, University of California, Santa Barbara, 201 pp.

BUSBY-SPERA, C.J. (1987) Lithofacies of deep marine basalts emplaced on a Jurassic backarc apron, Baja California (Mexico). *J. Geol.*, 25 pp.

BUSBY-SPERA, C.J. (1987) Evolution of a Middle Jurassic backarc basin, Cedros Island, Baja California: Evidence from a marine volcaniclastic apron. *Bull. geol. Soc. Am.*, 35 pp.

BUSBY-SPERA, C.J. & BOLES, J.R. (1986) Sedimentation and subsidence styles in a Cretaceous forearc basin, southern Vizcaino Peninsula, Baja California, Mexico. In: *Cretaceous Stratigraphy, Western North America* (Ed by P. L. Abbott), pp. 79–90. Soc. econ. Paleont. Miner. Pac. Section.

BUSBY-SPERA, C.J. & KELLER, B. (1985) The Gran Canon Formation: a Middle Jurassic marine volcaniclastic wedge on Cedros Island, Baja California. *Abstr. geol. Soc. Am. with Prog. Cordilleran Section Mtg.*

BUSBY-SPERA, C.J. & WHITE, J.D.L. (1986) Interaction of magma and wet sediment in the Cretaceous Alisitos Group, Baja California. *Abstr. geol. Soc. Am. with Prog.*, **18(2)**, 91.

BUSBY-SPERA, C.J. & WHITE, J.D.L. (1987) Interaction of magma and wet sediment at the type locality of the Cretaceous Alisitos Group, Baja California, Mexico. *Geology*, in press.

CARLISLE, D. (1963) Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia. *J. Geol.*, **71**, 48–71.

DIMROTH, E., COUSINEAU, P., LEDUC, M. & SANSCHAGRIN, Y. (1978) Structure and organization of Archaean subaqueous basalt flows, Rouyn–Noranda area, Quebec, Canada. *Can. J. Earth Sci.*, **15**, 902–918.

FISHER, R.V. (1984) Submarine volcaniclastic rocks. In: *Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins* (Ed. by B. P. Kokelaar and M. F. Howells). *Spec. Publ. geol Soc. London*, **16**, 5–28.

FISHER, R.V. & SCHMINCKE, H.-U. (1984) *Pyroclastic Rocks*, Springer–Verlag, New York.

FISKE, R.S. & MATSUDA, T. (1964) Submarine equivalents of ash flows in the Tokiwa Formation, Japan. *Am. J. Sci.*, **262**, 76–106.

FULLER, R.E. (1931) The aqueous chilling of basaltic lava on the Columbia River Plateau. *Am. J. Sci.*, **21**, 281–300.

GALLOWAY, W.E. (1981) Depositional architecture of Cenozoic Gulf Coastal Plain fluvial systems. In: *Recent and Ancient non-marine Depositional Environments: Models for Exploration* (Ed. by F. G. Ethridge and R. M. Flores). *Spec. Publs Soc. econ. Paleont. Miner.*, Tulsa, **31**, 127–156.

GASTIL, G., PHILLIPS, P.R. & ALLISON, E.C. (1975) Reconnaissance geology of the State of Baja California. *Mem. geol. Soc. Am.*, **140**, 170 pp.

HAMPTON, M.A. (1972) The role of subaqueous debris flow in generating turbidity currents. *J. sedim. Petrol.*, **42**, 775–793.

HAYWARD, B.S. & BALLANCE, P.F. (1982) Possible alluvial fans of basic to intermediate volcanic rudites with their feet in deep water: Piha Formation, Miocene, New Zealand (abstr). *I.A.S. 11th Int. Sedim. Congr. Hamilton, Ontario 1982*, **17**.

HICKEY, J.L. (1984) Stratigraphy and composition of a Jura-Cretaceous volcanic arc apron, Punta Eugenia, Baja California Sur, Mexico. In: *Geology of the Baja California Peninsula* (Ed. by V. A. Frizzell, Jr.). *Soc. econ. Paleont. Miner., Pacific Sect.*, **39**, 149–160.

KOKELAAR, B.P. (1982) Fluidization of wet sediments during the emplacement and cooling of various igneous bodies. *J. Geol. Soc. London*, **139**, 21–33.

KOKELAAR, B.P., BEVINS, R.E. & ROACH, R.A. (1985) Submarine silicic volcanism and associated sedimentary and tectonic processes, Ramsey Island, SW Wales. *J. Geol. Soc. London*, **142**, 591–613.

LONSDALE, P. & BATIZA, R. (1980) Hyaloclastite and lava flows on young seamounts examined with a submersible. *Bull. geol. Soc. Am.*, **91**, 545–554.

LORENZ, B.E. (1984) Mud-magma interactions in the Dunnage Melange, Newfoundland. In: *Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins* (Ed. by B. P. Kokelaar & M. F. Howells). *Spec. Publ. geol. Soc. London*, **16**, 271–277.

LOWE, D.R. (1982) Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. *J. sedim. Petrol.*, **52**, 279–297.

MCBIRNEY, A.R. (1963) Factors governing the nature of submarine volcanism. *Bull. Volcanol.*, **26**, 455–469.

MCCAVE, I.N. (1972) Transport and escape of fine-grained sediment from shelf areas. In: *Sedimentation in Submarine Canyons, Fans, and Trenches* (Ed. by D. J. P. Swift *et al.*), pp. 163–176. Dowden, Hutchinson and Ross, Stroudsburg, PA, USA.

MOORE, D.G. (1969) Reflection profiling studies of the California continental borderland: structure and Quaternary turbidite basins. *Spec. Pap. geol. Soc. Am.*, **107**, 142 pp.

SCHUMM, S.A. (1960) The effect of sediment type on the shape and stratification of some modern fluvial deposits. *Am. J. Sci.*, **258**, 177–184.

SCHUMM, S.A. (1977) *The Fluvial System.* John Wiley and Sons, New York, 338 pp.

SHIMOZURU, D. (1986) Physical condition of occurrence of phreatomagmatic eruptions. In: *Int. Volcanol. Congr.*, New Zealand, 1986, p. 120.

SIGURDSSON, H., SPARKS, R.S.J., CAREY, S.N. & HUANG, T.C. (1980) Volcanogenic sedimentation in the Lesser Antilles Arc. *J. Geol.*, **88**, 523–540.

SILVER, L.T. (1986) Observations on the peninsular ranges batholith, southern California and Mexico, in space and time. *Abstr. geol. Soc. Am. with Progr.*, **18(2)**, 184.

SURLYK, F. (1984) Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston Forland Group, East Greenland. In: *Sedimentology of Gravels and Conglomerates* (Ed. by E. H. Koster and R. J. Steel). *Mem. Can. Soc. Petrol. Geol.*, **10**, 359–382.

WALKER, G.P.L. & BLAKE, D.H. (1966) The formation of a palagonite breccia mass beneath a valley glacier in Iceland. *J. Geol. Soc. London*, **122**, 45–61.

WALKER, R.G. (1975) Nested submarine-fan channels in the Capistrano Formation, San Clemente, California. *Bull. geol. Soc. Am.*, **86**, 915–924.

WESCOTT, W.A. & ETHRIDGE, F.G. (1980) Fan-delta sedimentology and tectonic settings – Yallahs fan delta, southeast Jamaica, *Bull. Am. Ass. Petrol. Geol.*, **64**, 374–399.