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The Oman Ophiolite as a Cretaceous Arc-Basin Complex: Evidence and Implications

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The Oman ophiolite as a Cretaceous arc–basin complex: evidence and implications

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[Plates 1–3]

Geological and geochemical evidence suggest that the Oman ophiolite is a fragment of a submarine arc–basin complex formed above a short-lived subduction zone in the mid-Cretaceous. Detailed studies of the lava stratigraphy and the intrusive relationships of dykes, sills and high-level plutons provide further evidence for the magmatic and tectonic development of the complex in question. Four consecutive events can be recognized to have taken place before emplacement: (1) eruption of basalts of island arc affinity onto pre-existing (Triassic) oceanic crust; (2) creation of new oceanic crust by back-arc spreading; (3) intrusion of magma into this back-arc oceanic crust accompanied by eruption of basalts and andesites from discrete volcanic centres; (4) further intrusion of magma accompanied by uplift and eruption of basalts and rhyolites in submarine graben. A combined structural and geochemical analysis of the dyke swarm indicates that extension took place in approximately a N–S (ridge) and an ESE–WNW (leaky transform) direction relative to an inferred direction of subduction to the NE, and that a small but significant proportion of the sheeted dykes were injected during the 'arc' rather than the earlier 'back-arc spreading' episode. These various observations can be explained in terms of the progressive response of a non-isotropic lithosphere to the stresses induced during subduction.

Introduction

The precise tectonic setting of ophiolite complexes is still the subject of considerable debate. However, it is probable that at least some of the best preserved examples formed in back-arc basins. If such complexes can be identified, they may provide some of the answers to the many puzzling questions surrounding the extensional tectonics and magma genesis in currently and recently active basins. In this paper we provide geological and geochemical evidence to show that one of the largest and best exposed ophiolites in the world, the Semail thrust sheet of Oman, represents a section through an incipient island arc and oceanic crust of an underlying back-arc basin, and attempt to reconstruct the magmatic and tectonic evolution of the basin in question.

Geological evidence for a back-arc origin

The geology of the Oman mountains is already well documented (see, for example, Lees 1928; Wilson 1969; Reinhardt 1969; Allemann & Peters 1972; Glennie *et al.* 1973, 1974). Glennie's classic map, summarized in figure 1*a* shows the important features: an autochthonous carbonate platform developed on the Arabian basement and now exposed in tectonic windows; a stack of thrust sheets composing the Hawasina and Haybi complexes; the Semail thrust sheet, a fully developed ophiolite complex and the main subject of this paper; and a post-orogenic sedimentary cover. The schematic cross section presented in figure 1*b* summarizes the composition, ages and tectonic relationships of these various units.

Although there is no consensus on the precise details, the general sequence of events that produced this structure is also well understood (Gealey 1977; Welland & Mitchell 1977; Searle

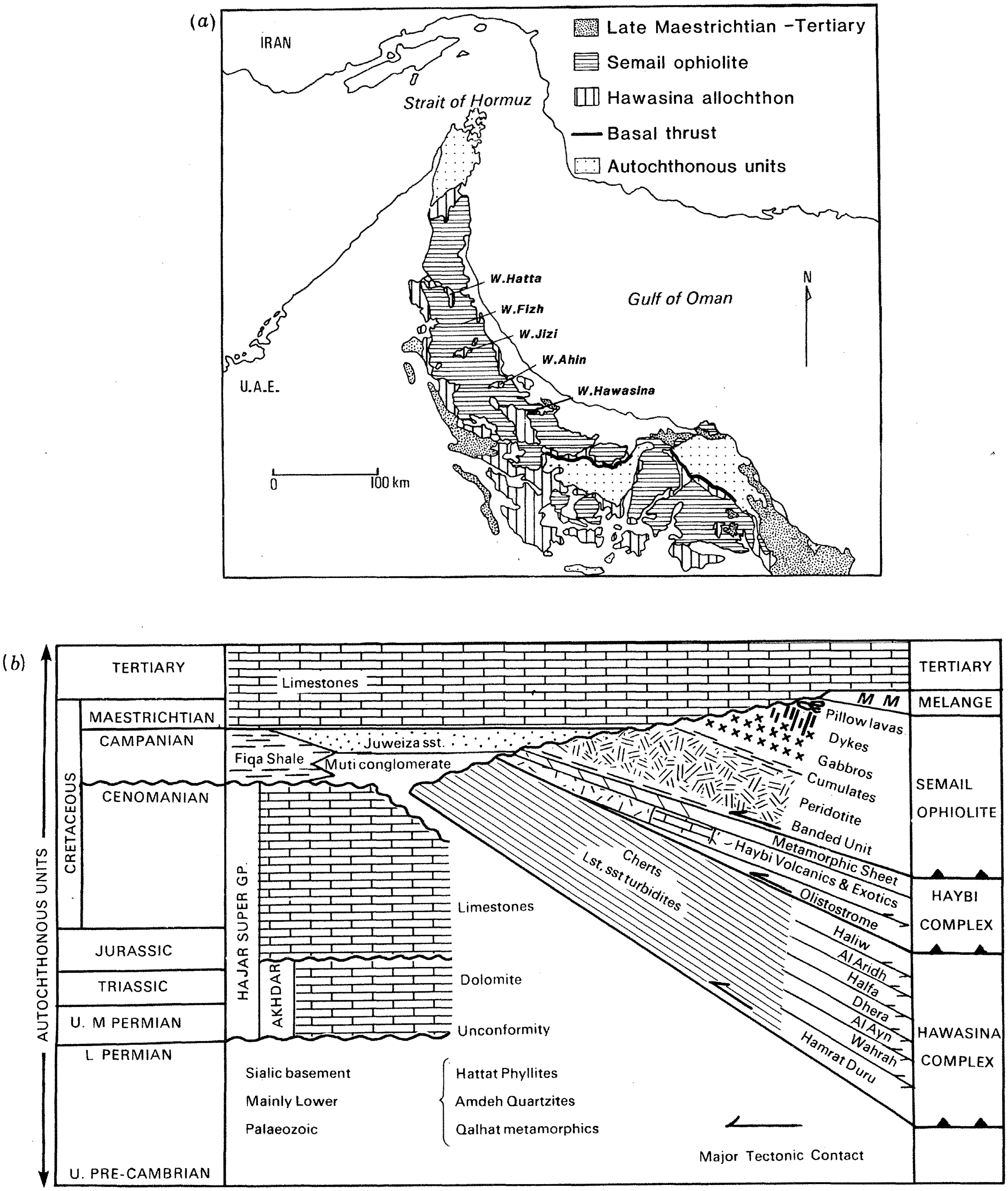

FIGURE 1. Map of the Oman mountains (after Glennie *et al.* 1974 and schematic section (from Searle *et al.* (1980), after Gealey (1977)).

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et al. 1980; Graham 1980). Tectonic activity began by continental rifting and alkali volcanism in the early Mesozoic, followed by sea floor spreading in the Late Triassic and perhaps also the Jurassic to form one of the small ocean basins that seem to have characterized western Tethys at that time. The platform carbonates and deeper water (Hawasina) sediments developed on the passive NE-facing continental margin of this ocean between the Upper Triassic and mid-Cretaceous. Emergence of the offshore rise, deposition of olistostromes and debris flows and resumption of volcanism, characterized initiation of subduction in the mid-Cretaceous (Searle *et al.* 1980). This was followed by formation of the ophiolite complex itself in the Cenomanian – earliest Turonian (Glennie *et al.* 1973; Allemann & Peters 1972; Tippit & Pessagno 1979). Subsequent interaction of the Arabian margin with a NE-dipping subduction zone, resulting in the imbrication of the Hawasina and Haybi sequences, the formation of the Haybi melange and emplacement of the ophiolite complex, was well under way by the Santonian (the age of metamorphism at the base of the ophiolite) and was complete by the lower Maestrichtian (the age of the oldest post-tectonic sediment) (Allemann & Peters 1972). Shallow marine sedimentation followed by Oligocene to Recent uplift and minor deformation completes the tectonic picture.

We can be almost certain from these ages that northeastwards subduction of oceanic crust was taking place at the time that the Oman ophiolite formed, but did the ophiolite form in front of or behind this subduction zone? The most convincing arguments in favour of the subduction-related origin were originally made by Gealey (1977), who proposed that the Oman mountains developed as a result of continent–island arc collision. He wrote: 'In the collision process, as exemplified by the geology of Oman, the passive continental margin rides down the subduction zone and underthrusts the island arc edifice, which consists of the plutonic arc proper and a frontal fore-arc limb floored by oceanic crust incorporated during formation of the arc structure. The obducted ophiolite slab represents the fore-arc limb emplaced on the underthrusting continental margin.'

Gealey's arguments were based on uniformitarian comparisons between Oman and more recent examples of continental-arc collision in different stages of development. As analogues he used the Aleutians (pre-collision stage), the Banda Sea (collision between platform carbonates and a fore-arc limb), Timor (underthrusting of continental crust) and New Caledonia (ophiolite emplacement). Gealey derived his evolutionary model from sedimentary and structural considerations without any recorded island arc volcanism or plutonism to support his hypothesis. In the following sections we present the evidence which shows his hypothesis to be correct in its most important aspects.

Evidence from the lava stratigraphy

The results of a study on the lava stratigraphy of the Oman ophiolites were reported by Alabaster & Pearce (1979) and Alabaster *et al.* (1980). This work consisted of detailed mapping of a small area around the Lasail ore deposit (figure 2) coupled with a regional survey to relate this map to the 300 km (along strike) of pillow lava exposure in the ophiolite complex as a whole.

In the field, the volcanic stratigraphy can be divided into three distinct lava units. The basal '*Geotimes* unit', named after the front cover of *Geotimes* on which one of its type localities was featured (Coleman 1975), directly overlies the sheeted dyke swarm virtually throughout the ophiolite, a transition zone of 500–100 m separating 100 % dykes from almost 100 % lava. The

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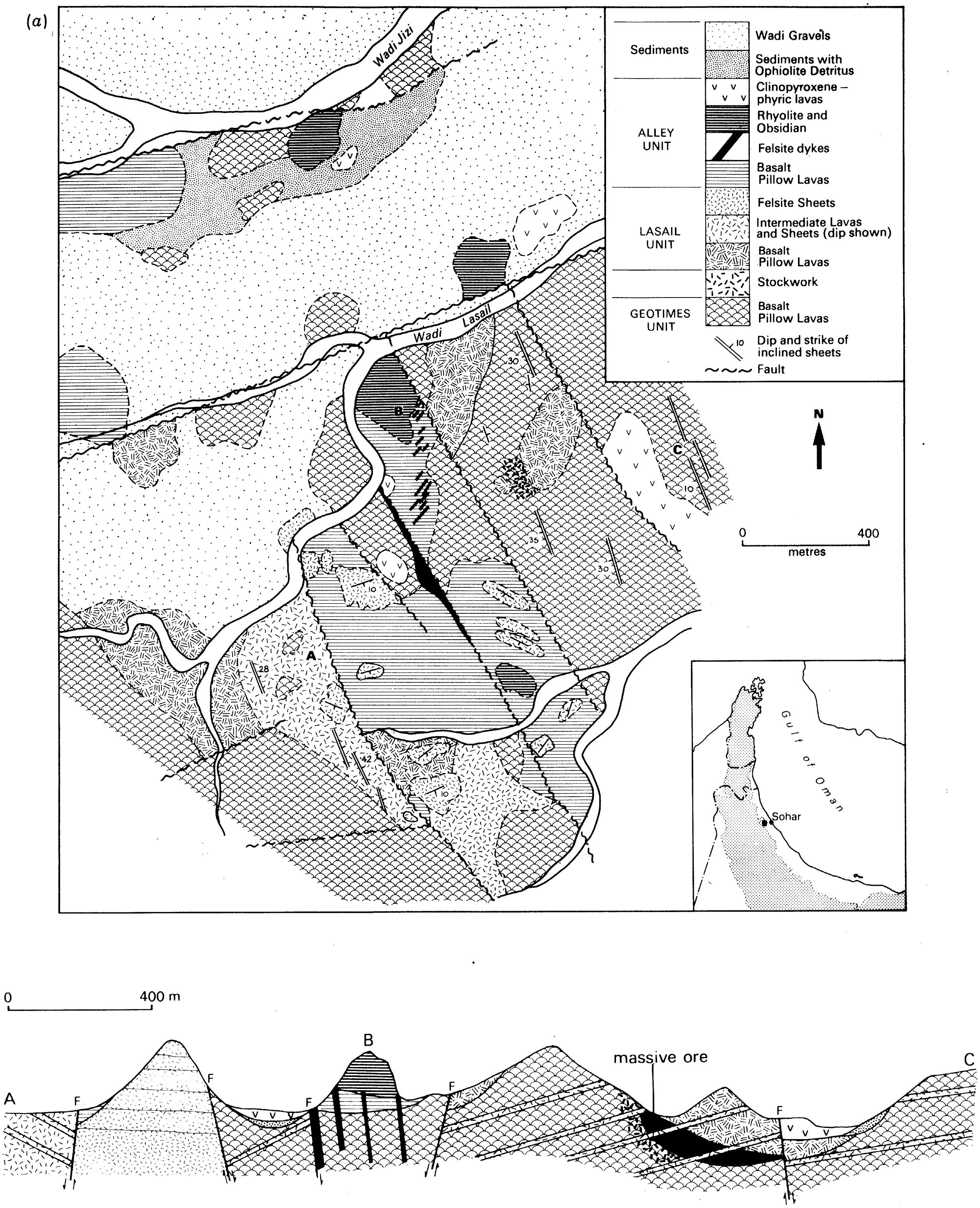

FIGURE 2. Map of the volcanic stratigraphy of a small area of the Oman ophiolite just south of Wadi Jizi (from Alabaster *et al.* 1980).

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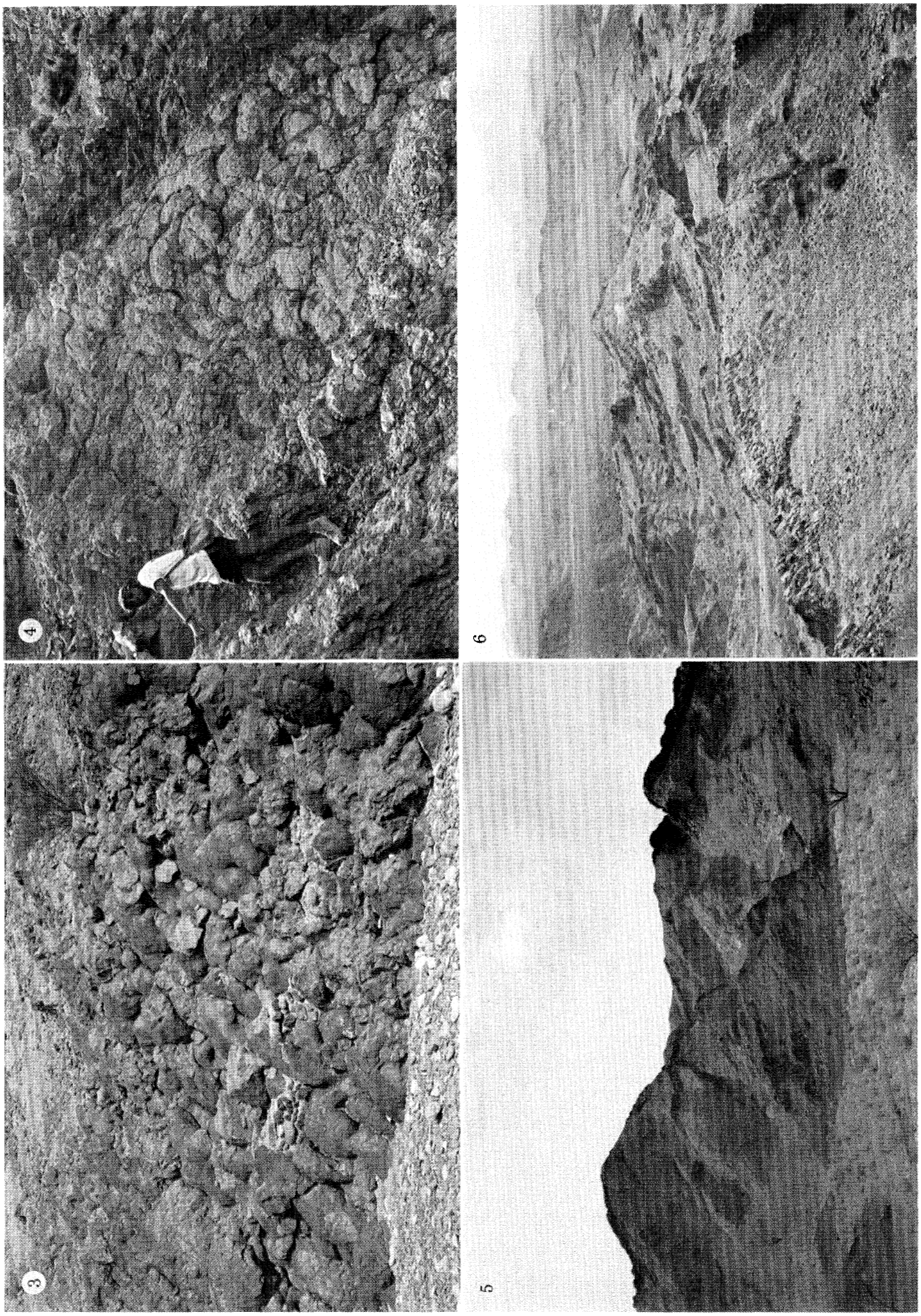

FIGURES 3–6 For description see opposite plate 2.

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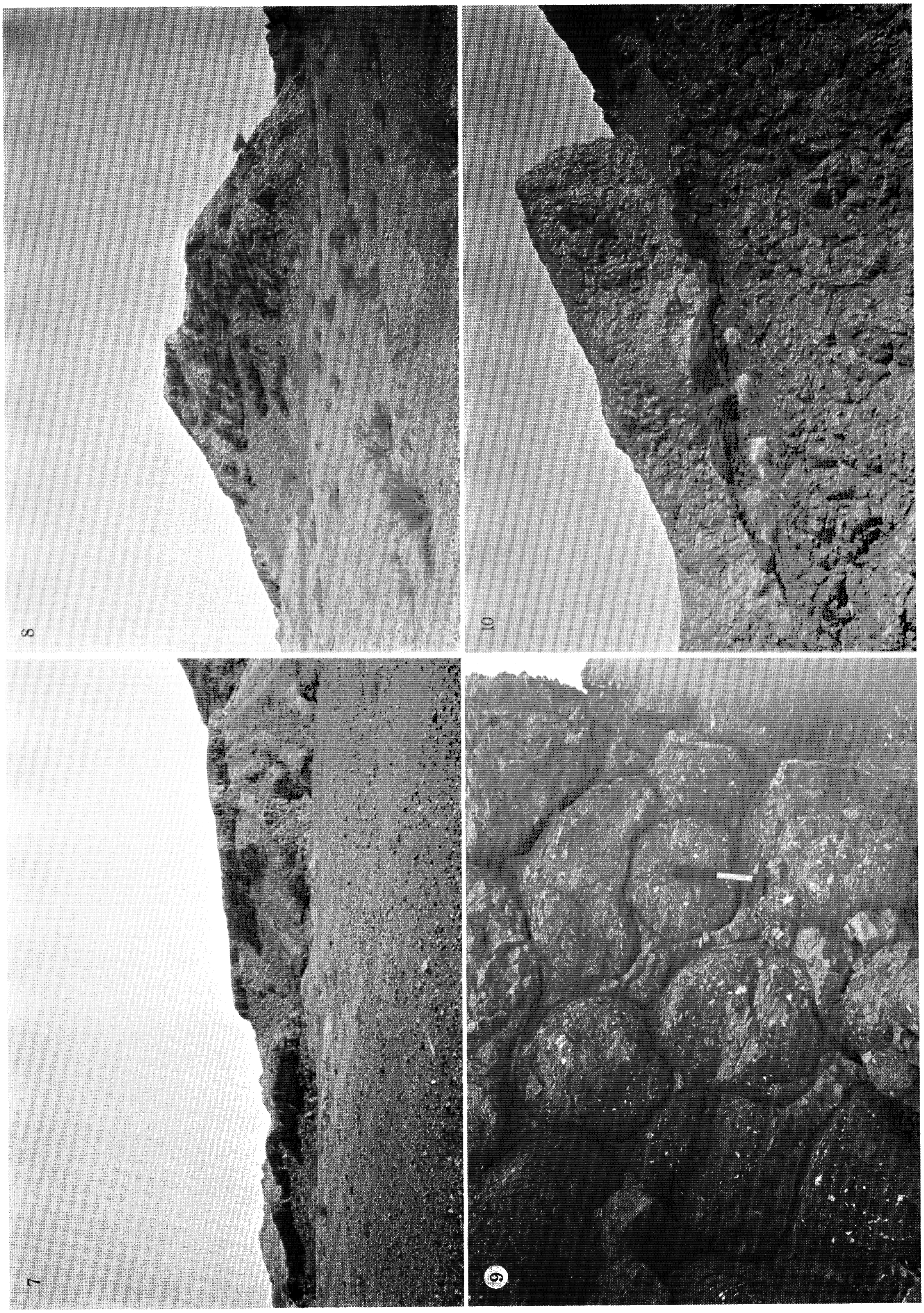

FIGURES 7–10. For description see opposite.

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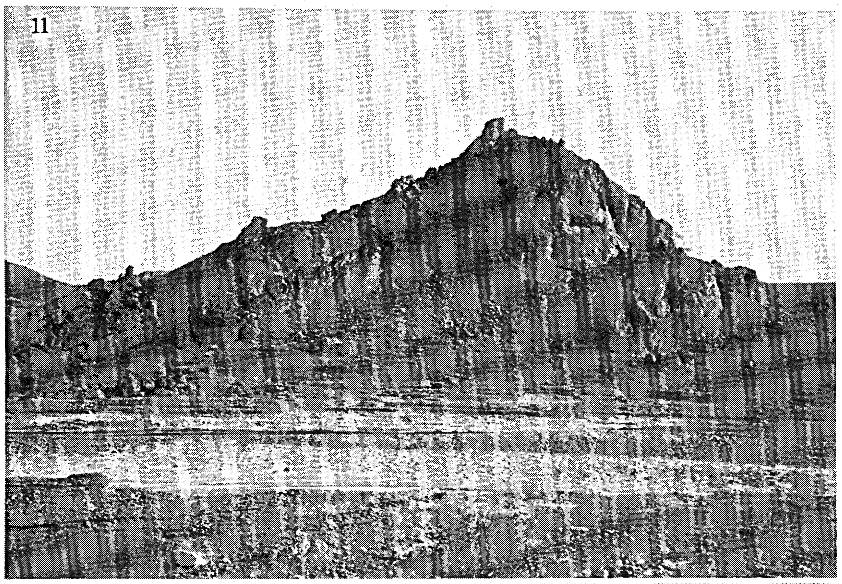

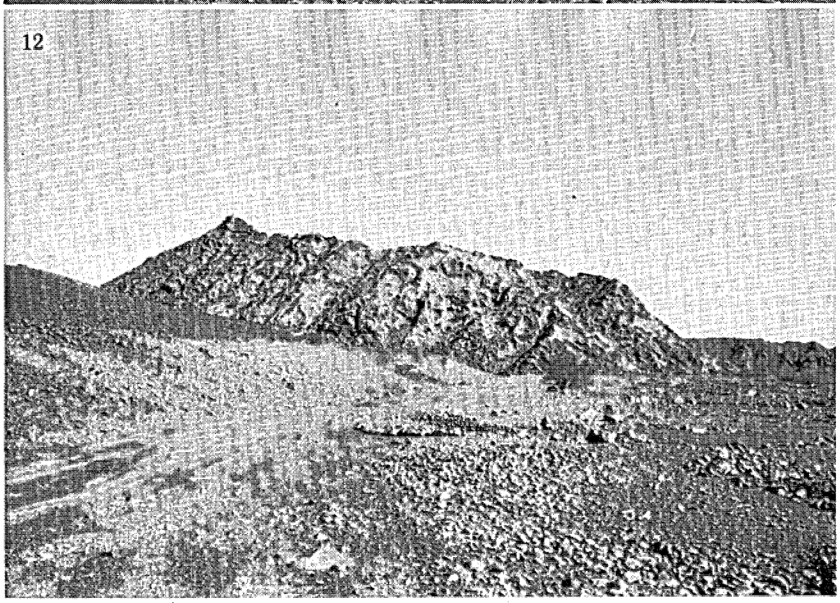

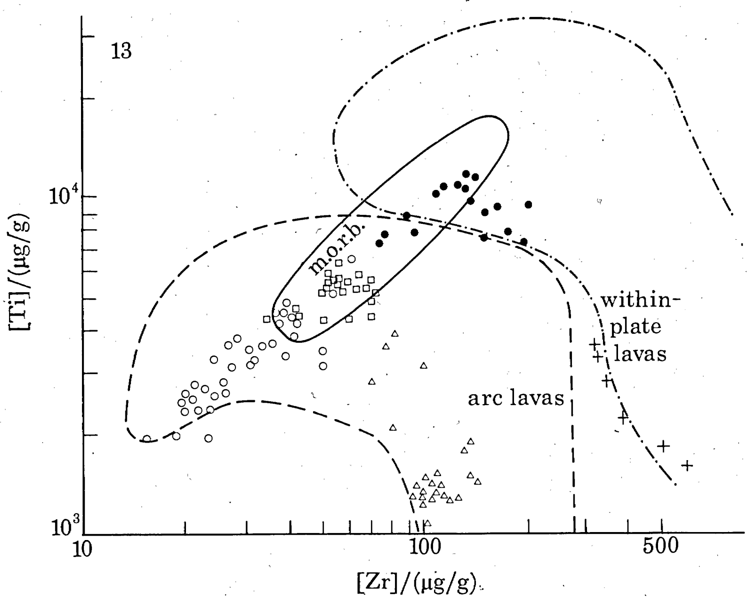

FIGURES 11–13. For description see opposite plate 2.

Description of plate 1

Figure 3. Typical exposure of the basal lavas, the *Geotimes* lava group. Inferred eruptive setting at a back-arc spreading axis. Pillow lavas are metabasalts and are red. Field of view about 12 m across.

Figure 4. Typical exposure of the first group of upper lavas, the Lasail lavas unit. Inferred eruptive setting in an island arc seamount. Pillow lavas are metabasalts and are green. Field of view about 6 m across.

Figure 5. Contact between *Geotimes* and Lasail lava units, possibly marking the contact between back-arc ocean crust and an island arc seamount. Contact stands out because of the dark red – pale green contrast. Lasail unit contains inclined sheets. Field of view at contact about 50 m across.

Figure 6. Lasail lava unit intruded by at least two intersecting sets of inclined sheets.

Description of plate 2

Figure 7. Columnar jointed andesite lava flow capping andesites and basalts of the Lasail unit. Thickness of flow about 20 m.

Figure 8. Subhorizontal felsite (trondhjemite) sheets representing the final stage of fractionation of the Lasail unit. The outcrop is about 100 m across.

Figure 9. Typical exposure of the vesicular Alley unit lavas. Inferred eruptive setting in graben between the island arc seamounts. Lavas are metabasalts and are grey–brown.

Figure 10. Contact between the *Geotimes* lava unit (foreground) and overlying Alley lava units (background), marked by about $\frac{1}{3}$ m of umbiferous sediment (the time-equivalent of the Lasail unit).

Description of plate 3

Figure 11. Volcanic centre of the Alley unit developed within one of the seamount areas. The core of the hill is a dacite plug, the top is an obsidian lava. The hill is about 100 m across at its base.

Figure 12. Ridge of rhyolite lavas of the Alley unit. The ridge is elongated along the strike of the dyke swarm of the underlying oceanic crust. The ridge is about 150 m across.

Figure 13. Ti–Zr covariation diagram for the lavas of the Oman ophiolite, showing the distinctive character of the lower 'back-arc spreading' unit (solid symbols) compared with the upper 'arc' units (open symbols). The fields occupied by the various magma types are taken from Pearce (1980); spreading axis plagiogranite analyses (crosses) used for comparison with 'arc' acid lavas are taken from Aldiss (1978). Lasail and Alley units: △, acid; □, intermediate; ○, basic. *Geotimes* units: ●, basic and basic–intermediate.

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unit reaches $1\frac{1}{2}$ km in thickness and typically consists of large, reddish-brown, non-vesicular, almost aphyric pillow lavas (figure 3, plate 1) interbedded with rare breccia flows and columnar jointed massive flows with pillowed tops. This sequence forms the volcanic 'basement' on which successive lava units were erupted.

The overlying lavas were divided into two groups, the Lasail and Alley units. The type area for the Lasail unit is the Lasail area itself (figure 2). It comprises a fractionation sequence from basalt through andesite to felsite and can reach 1 km in total thickness. The basaltic member consists of small, grey-green, non-vesicular, pillow lavas (figure 4) which directly overlie the lower (*Geotimes*) lava unit. The contact between the units can be seen in figure 5, standing out because of the red–green colour contrast between the lava units; the contact forms an irregular and sinuous outline with no intercalated sediment. The pillow lavas are invariably associated with swarms of thin intrusive inclined sheets of basaltic andesite and andesite composition that bifurcate in places and frequently transgress up the lava pile (figure 6). The extrusive equivalents are massive, columnar jointed andesite lava flows up to 40 m thick (figure 7, plate 2). The orientation of the inclined sheets indicates that they emanated from magma chambers now marked by acidic centres (see the section in figure 2). These centres occur as high-level trondhjemite plugs with a peripheral stacking of felsite sheets or massive flows (figure 8). The Lasail unit does not occur throughout the ophiolite but is restricted to centres spaced at intervals along the N–S strike of the complex (see figure 22). It is therefore considered to have formed seamounts erupted onto ocean crust; there is no evidence on whether or not some of the intermediate and acid flows were erupted above or below sea level.

The Alley unit has its type locality in the fault-bounded structure north of the Lasail area known as the 'Alley' (Smewing *et al.* 1977). It represents a second basalt–rhyolite fractionation sequence and reaches a maximum thickness of $\frac{1}{2}$ km. It may overlie the extrusive members of the Lasail unit, in which case a thin layer of ferromanganoan sediment separates the units. Alternatively, where no Lasail unit lavas were deposited, the 'Alley' lavas directly overlie the *Geotimes* unit, in which case a much thicker horizon of umbiferous sediment marks the volcanic unconformity (figure 10). The basaltic rocks form a series of large brownish-green, often highly fragmentary, vesicular pillowed flows (figure 9). The intermediate member of the Alley unit is represented by columnar jointed flows up to 20 m thick which are overlain by rhyolite (including obsidian) pillow lavas and massive flows (figures 11 and 12, plate 3). These acid centres occur only in the 'seamount' areas; in the 'inter-seamount' areas, marked by faulted depressions such as the 'Alley', volcanism is almost entirely basic in character. The final eruptive episode of the Alley unit is volumetrically small and consists of clinopyroxene–phyric and olivine–clinopyroxene–phyric pillow lavas and fragmentary flows erupted along certain fractures (see figure 2).

It is therefore apparent that this lava stratigraphy bears little resemblance to that found during deep-sea drilling of normal oceanic crust. The lower (*Geotimes*) unit, most of the underlying dyke swarm and the main layered gabbro sequence were seemingly formed at an oceanic constructive margin, but the later lavas and the intrusive rocks that feed them are clearly related to the development of an overlying seamount chain or volcanic ridge. A volcanic edifice of this type could have developed either in a within plate or an island arc setting. However, the abundance of basaltic andesites and andesites in the Lasail unit, with green diopsidic augite (or occasionally hornblende) as the dominant phenocryst, argues strongly for an island arc origin. The absence of any significant sediment between these lavas and the underlying *Geotimes* lavas

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further suggests that this island arc was built on new ('back-arc') oceanic crust rather than old ('normal') oceanic crust.

These interpretations are also supported by evidence from the plutonic rocks. Whereas the main gabbro sequence can be modelled as a number of large sub-ridge crest magma chambers (e.g. Hopson *et al.* 1979; Smewing 1980), this is not true of the large high-level intrusions of gabbro, tonalite and trondhjemite (in the seamount areas) and of peridotite and gabbro (in the 'inter-seamount' areas), which feed the later lava units. A preliminary examination of the chemistry of these units (figure 13) leaves no doubt that the lavas and the intrusives can be

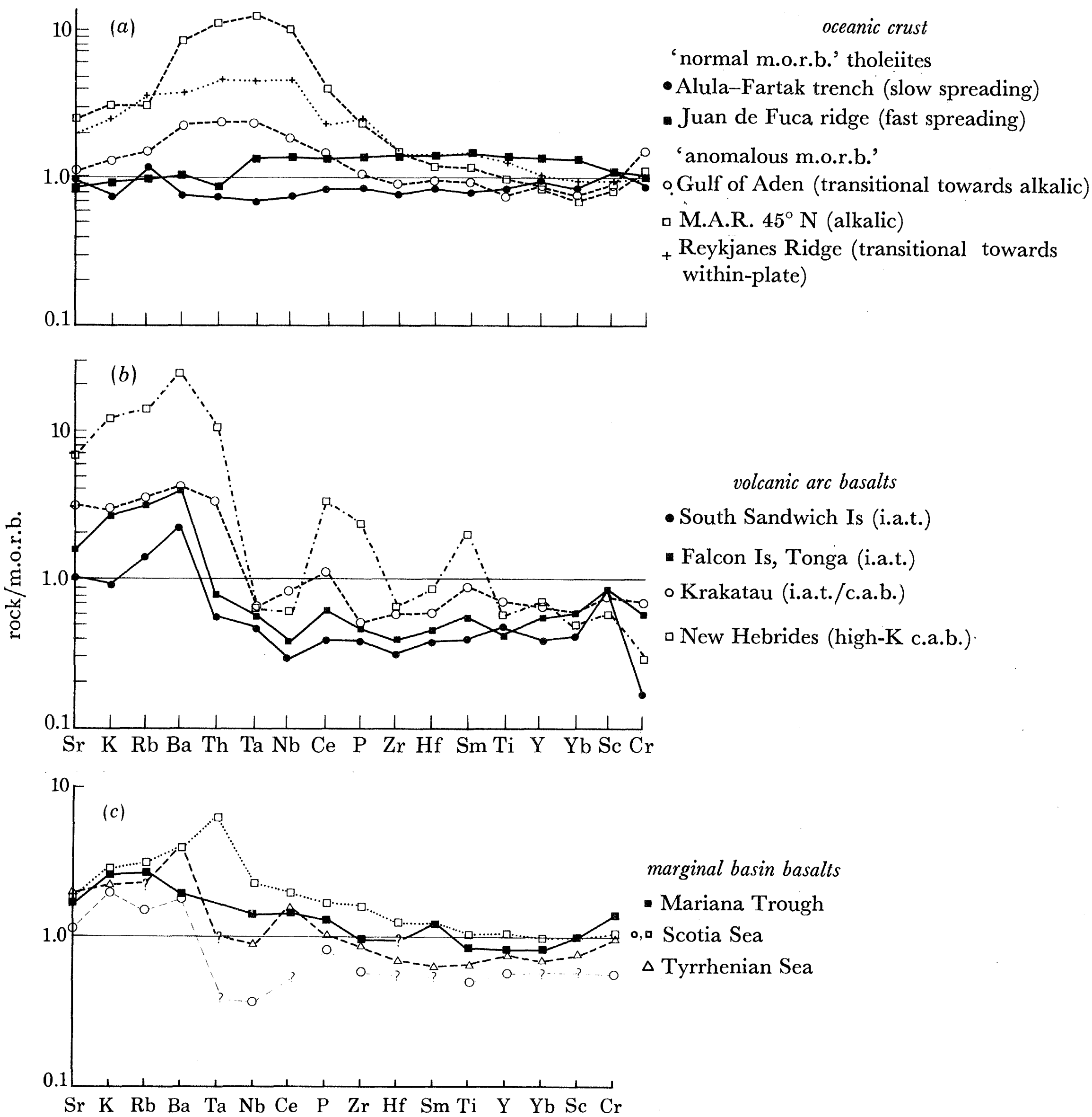

FIGURE 14. Geochemical patterns for (*a*) m.o.r.b., (*b*) island arc basalts and (*c*) some back-arc basin basalts. Analyses by J. A. Pearce except M.A.R. 45° N (Wood *et al.* 1979*b*), New Hebrides (Gorton 1977), Scotia Sea (Saunders & Tarney 1979), and Tyrrhenian Sea (Dietrich *et al.* 1977). Normalizing factors (micrograms per gram unless percentages indicated): Sr, 120; K_2O, 0.15 %; Rb, 2.0; Ba, 20; Th, 0.2; Ta, 0.18; Nb, 3.5; Ce, 10.0; P_2O_5, 0.12 %; Zr, 90; Hf, 2.4; Sm, 3.3; TiO_2, 1.5 %; Y, 30; Yb, 3.4; Sc, 40; Cr, 250.

subdivided into two sets: a constructive margin 'axis' set to which the *Geotimes* lavas belong, and an 'arc' set to which the Lasail and Alley units belong. The Lasail and Alley units can be subclassified as 'arc seamount' and 'arc rifting' respectively, on the basis of their field relations. A more detailed examination of the geochemistry in the next section provides further evidence for this conclusion.

One further aspect of lava stratigraphy that is important in interpreting the settling of the Oman ophiolite concerns the Haybi volcanics beneath the Semail Nappe. This sequence, described and defined by Searle *et al.* (1980) and Searle & Malpas (1979), has usually suffered disruption and metamorphism during emplacement. Where intact sequences do occur, the lower part is made up of mid–late Triassic alkali basalt (and ankaramite) lavas and pyroclastics, the upper part of tholeiitic pillow lavas. At the stratigraphic top of the tholeiitic sequence, the lavas are interbedded with red radiolarian cherts dated as early to middle Cretaceous. Searle *et al.* (1980) relate the Triassic lavas to the early stages of ocean formation and the uppermost tholeiites to the first episode of arc volcanism. The geochemical studies that follow include data from both these lava types, and from amphibolites in the metamorphic sheet, for comparison with the ophiolite itself.

Geochemical evidence for a back-arc origin

Island arc lavas erupted at the present day can readily be distinguished from mid-ocean ridge basalts. There are three main features to look for, two of which are highlighted in the geochemical patterns shown in figure 14*a*, *b*:

1. Island arc lavas show an enrichment in elements of low ionic potential (Z/r) (such as K, Rb, Sr, Ba and to a lesser extent Th and Ce) relative to elements of high ionic potential (such as Ta, Nb, Hf, Zr, Ti, Y and Yb) (see, for example, Gill 1974; Saunders & Tarney 1979). This feature is thought to be caused by selective enrichment of the island arc source region in elements that are mobile in the aqueous fluids driven off subducted oceanic crust.

2. At a given Cr content, the elements of high ionic potential are depleted in island arc lavas (Pearce 1975; Garcia 1978; Sharaskin *et al.*, this symposium). The explanation of this feature is more controversial. High degrees of partial melting (see, for example, Pearce & Norry 1979), remelting of some already depleted material (see, for example, Green 1973) and stability of minor oxide phases in the mantle residue (see, for example, Dixon & Batiza 1979; Saunders *et al.* 1980) have all been proposed. All these explanations require subducted water in the mantle source region, in the first two cases to lower the mantle solidus, and in the third to increase oxygen fugacity.

3. Island arc lavas exhibit an increase in their $^{87}Sr/^{86}Sr$ ratio relative to their $^{143}Nd/^{144}Nd$ ratio (Hawkesworth *et al.* 1977; DePaolo & Johnson 1979). This feature is explained by the high Sr/Nd ratio and the high $^{87}Sr/^{86}Sr$ ratio in the seawater-derived fluids driven off the subduction zone.

Most basalts from well established back-arc basins do not exhibit this subduction zone component (Dietrich *et al.* 1978; Saunders *et al* 1980; Tarney *et al.*, this symposium), but there are exceptions, such as part of the Scotia Sea (Saunders & Tarney 1979), the Bransfield Strait (Weaver *et al.* 1979) and the Tyrrhenian Sea (Dietrich *et al.* 1977). It is likely that the geochemical influence of the subduction zone was most commonly felt during the early stages of back-arc spreading, in areas not yet well sampled.

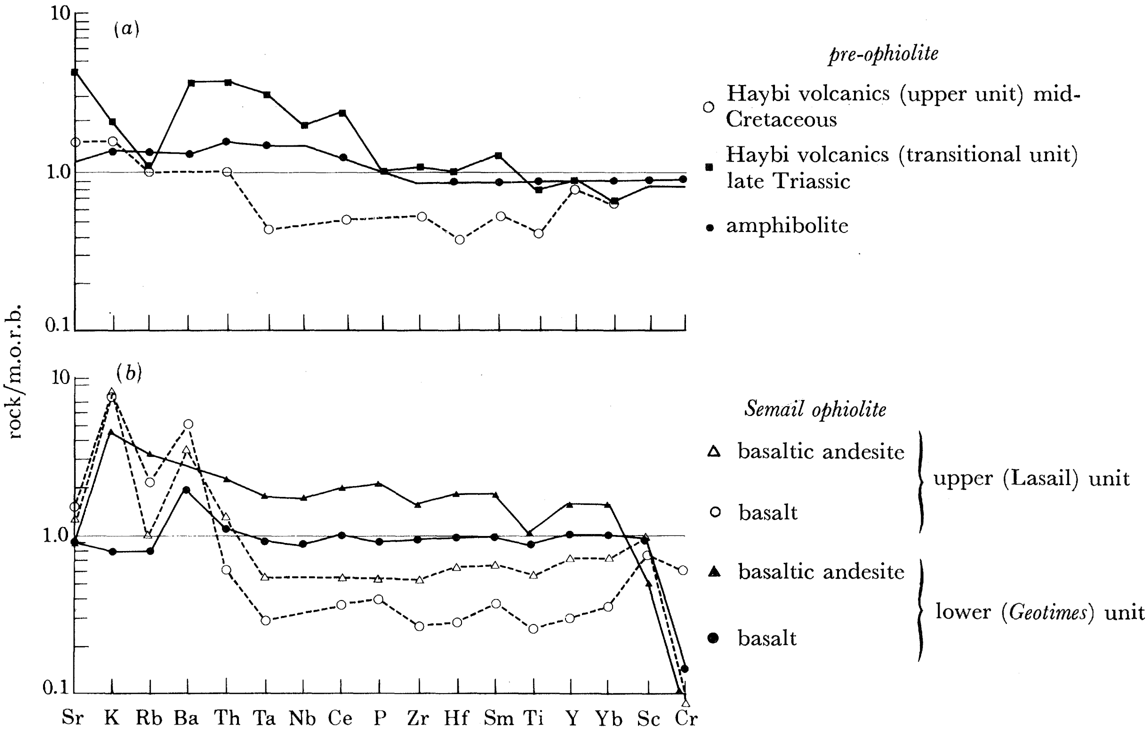

FIGURE 15. Geochemical patterns for (*a*) pre-ophiolite and (*b*) ophiolite basic volcanic rocks. Normalizing factors as in figure 14.

We now attempt to detect these features in both the late Triassic – early Jurassic volcanics, now structurally beneath the ophiolite, and in the Cretaceous ophiolite complex itself.

(*a*) *Geochemical patterns*

The patterns shown in figure 15 fall into three types. The late Triassic Haybi volcanics and the late Triassic or possibly Jurassic amphibolites show patterns that range from the flat pattern of normal mid-ocean ridge basalts (m.o.r.b.) with its ratio close to unity, to a 'humped' pattern characteristic of transitional m.o.r.b. (cf. figure 14*a*). Clearly, these patterns are consistent with an inferred origin during the early stages of continental separation, i.e. in a small Red Sea-type ocean unrelated to subduction. The lower–mid-Cretaceous lava from the Haybi unit and the Upper lava from the ophiolite complex show patterns that resemble the island arc tholeiite patterns in figure 14*b*, both in their high relative abundances of K, Rb, Sr, Ba and Th and in their low absolute abundances of the elements of high ionic potential, including Cr. The enrichment in elements such as K can also be caused by submarine alteration and therefore cannot be regarded as completely diagnostic. However, the degree of enrichment observed is greater than we would expect for greenschist facies alteration and a subduction zone component is therefore probable. Finally the ophiolite lower lava has a more transitional pattern. All incompatible elements have abundances close to typical m.o.r.b., but Cr is extremely low, lower than in any basalt lava dredged from the major ocean basin. The following diagrams examine these features more closely.

(*b*) Cr–Y *covariations* (*figure* 16)

These two elements highlight the second feature of island arc lavas described earlier. M.o.r.b. and island arc lavas are almost completely separated on this diagram, as they are on a similar diagram of Ti against Cr (Pearce 1975). The position of the data points confirm the m.o.r.b. characteristics of the late Triassic–Jurassic lavas and the island arc character of the Cretaceous lavas. They suggest further that the ophiolite lower lavas are strongly fractionated for basaltic rocks (Cr is accommodated in olivine, spinel and clinopyroxene and therefore rapidly depleted during fractionation) but they can be related to island arc basalts more closely than to m.o.r.b.

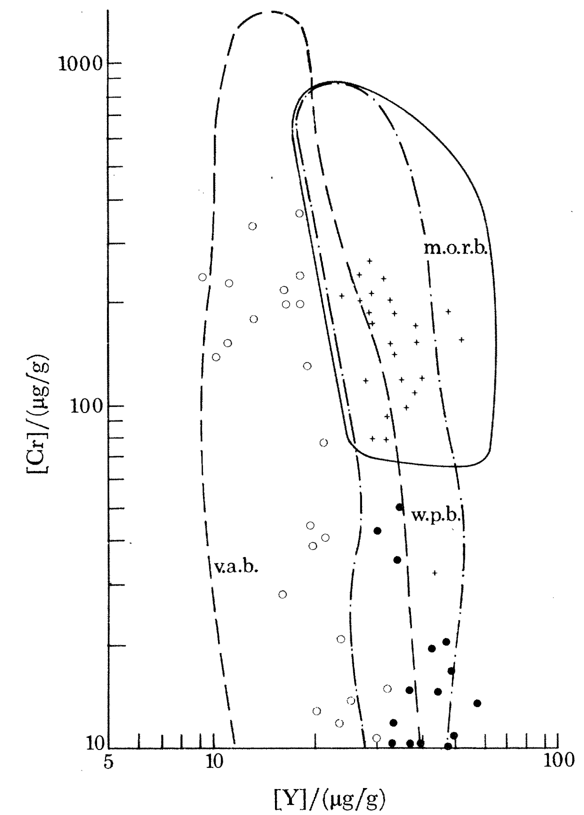

FIGURE 16. Cr–Y covariation diagram (Pearce 1980), showing the distinction between the mainly late Triassic pre-ophiolite lavas (Haybi volcanics and amphibolites, +) and the late Cretaceous ophiolite lavas (*Geotimes* (●) and Lasail and Alley (○) units).

(*c*) Th/Yb–Ta/Yb *covariations* (*figure* 17)

These axes highlight the first feature of island arc tholeiites described earlier. Th is used here as the element of low ionic potential and Ta as an element of high ionic potential, following the work of Wood *et al.* (1979*a*). The advantage of using Th, which shows much less enrichment than K, Rb, Sr or Ba in island arc basalts, is that it does not appear to have been mobile during sub-seafloor alteration in the Oman ophiolite. As figure 17 shows, the late Triassic – early Jurassic lavas plot within the field area occupied by m.o.r.b. and within plate basalts, in which Ta and Th are strongly intercorrelated. By contrast the Upper Cretaceous lavas all show the displacement to higher Th abundances characteristic of island arc lavas. This displacement is much more marked in the Haybi and ophiolite upper lavas, adding further evidence for their arc origin. The ophiolite lower lavas show a displacement which is smaller and only just significant; the geochemical influence of the subduction zone is therefore relatively small during the genesis of these lavas.

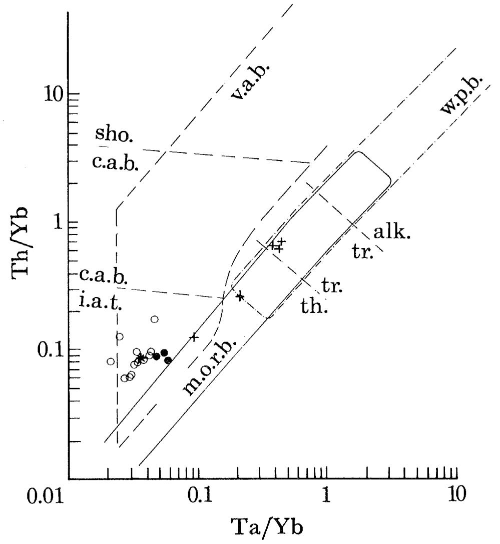

FIGURE 17. Th/Yb–Ta/Yb covariation diagram based on the work of Wood *et al.* (1979*a*), showing the Th shift of the Cretaceous lavas, which indicates an origin above a subduction zone. Key as figure 16, except ∗, Uppermost Haybi lava; m.o.r.b., mid-ocean ridge basalt; w.p.b., within-plate basalt; v.a.b., volcanic arc basalt; th., tholeiitic; tr., transitional; alk., alkalic; c.a., calc-alkalic; sho., shoshonitic; i.a.t., island arc tholeiite.

(*d*) $^{87}Sr/^{86}Sr$–$^{143}Nd/^{144}Nd$ *covariation* (*figure* 18)

Whereas the previous diagram depended on a Th shift from a m.o.r.b.-within plate correlation line to identify a subduction zone component, this diagram depends on a $^{87}Sr/^{86}Sr$ shift. It is therefore even more sensitive to alteration and, in the Oman ophiolite, can only be applied with confidence if unaltered clinopyroxene separates are used. Only two analyses are available from the northern Oman ophiolite at present: a clinopyroxene separate from a porphyritic dyke belonging to the Alley unit (Pearce (1980), from an analysis by C. J. Hawkesworth) and a clinopyroxene separate from a gabbro in the Wadi Fizh area (McCulloch *et al.* 1980). Although McCulloch *et al.* have not classified their gabbro as 'axis' or 'arc', an 'axis' origin can be tentatively inferred from their absolute Nd abundances.

As McCulloch *et al.* pointed out, the Wadi Fizh analysis plots in the ambiguous field occupied by both island arc and m.o.r.b. lavas and cannot be uniquely characterized. Our analyses of the Alley pyroxene, when expressed as ϵ values (DePaolo & Johnson 1979), plots unambiguously in the island arc field. This conclusion is consistent with the previous diagram, where the upper lavas contained a much greater subduction zone component than the lower lavas, but until further work has been carried out these results must be regarded as preliminary.

The geochemical studies can therefore be summarized as follows.

1. The ocean crust that existed before the formation of the Oman ophiolite formed in a small (mainly Triassic) ocean unrelated to any subduction zone.

2. The first lavas erupted onto this crust in the middle Cretaceous had a strong subduction zone component in their chemistry, consistent with an island arc origin.

3. The lower lavas in the Oman ophiolite show a slight but significant subduction zone input; they are therefore consistent with formation during the early stages of back-arc spreading.

4. The upper lava units in the Oman ophiolite have a strong subduction zone component in their chemistry, consistent with an island arc origin.

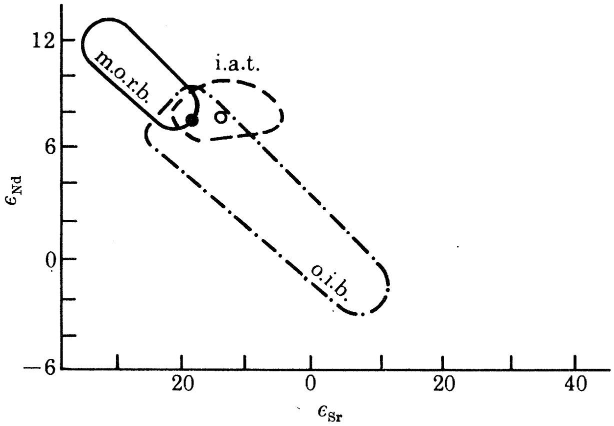

FIGURE 18. ϵ_{Nd}–ϵ_{Sr} covariation diagram (McCulloch *et al.* 1980), showing the two available pyroxene analyses from the northern Oman ophiolite. Key: i.a.t., island arc tholeiite; m.o.r.b., mid-ocean ridge basalt; o.i.b., ocean island basalt; ●, lower (*Geotimes*) lava unit; ○, upper (Alley) lava unit.

EVIDENCE FROM THE DYKE SWARM FOR BACK-ARC EXTENSIONAL TECTONICS

Because of the fractures developed during emplacement, dyke trends are a better guide than fault trends to the directions of minimum stress in the Oman back-arc basin. The sheeted dyke swarm on the Troodos Massif of Cyprus was also used in this way to establish a sea floor spreading origin for the ophiolite complex, to determine the spreading direction and to establish the Arakapas fault belt as a transform fault (Gass 1968, Moores & Vine 1971; Simonian & Gass 1978).

The sampling programme for this study was carried out by one of the authors (A.W.S.) as part of a palaeomagnetic study. The first step was to establish which dykes were produced by the back-arc spreading event and which were produced by the later arc events. To do this, some representative samples were analysed for Ti, Zr and Y at Cambridge University (Nisbet *et al.* 1979) and the analyses compared with the lava compositions. As figure 19 shows, of 50 samples, 31 plotted in the field of the 'axis lavas' whereas 19 plotted in the field of 'arc' lavas on this basis. The analyses and their classification are presented on figure 20.

It is therefore apparent that a significant part of the dyke swarm can be related to the arc rather than to the spreading event. The actual proportion is less than figure 20 suggests, however, because of the sampling bias towards the fresher and less fragmented later dykes. Nevertheless, many of the dykes classifying as 'arc' were taken from apparently homogeneous sheeted units and must therefore denote complete extension locally.

The second step was to correlate dyke type with dyke direction, and a stereogram has been plotted for this purpose in figure 21. The various data points have been annotated according to whether they classify as 'axis' or 'arc' and also according to geographic location. The latter subdivision was carried out as a precaution against the possibility that different segments of the ophiolite had suffered relative rotation during emplacement. Each of the 'blocks' defined in figure 21 is thought, on field evidence, to have behaved throughout as an internally coherent unit. As figure 21 shows, however, there is no obvious variation in dyke direction between the various blocks, thereby suggesting that the distribution of data points does reflect pre-emplacement variations.

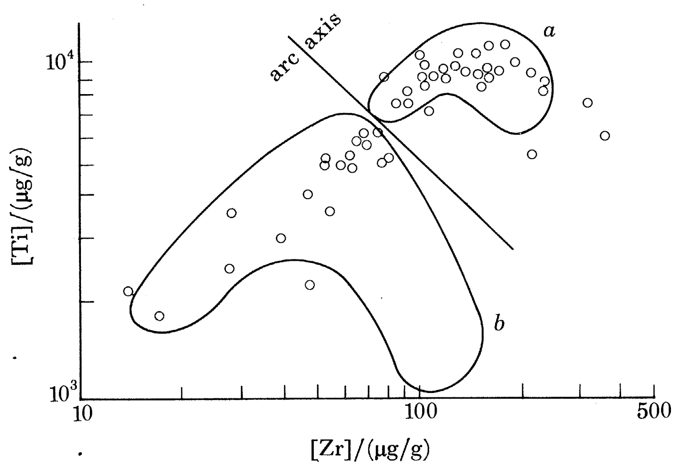

FIGURE 19. Classification of analysed samples from the sheeted dyke swarm into 'arc' and 'axis' by using the Ti–Zr covariation diagram. Lava fields are taken from figure 13: (*a*) field of lower (*Geotimes*) unit; (*b*) field of upper (Lasail and Alley) units.

The stereogram reveals: (1) that there is a distinctly bimodal distribution to the strike of the dyke sets, one at about 175°, the other at about 115°; (2) both 'axis' and 'arc' dyke sets have this bimodal distribution; and (3) the 115° set predominates in the north, between Wadi Hatta and Wadi Fizh, whereas the 175° set predominates throughout the rest of the area studied.

There are major fault trends within the ophiolite, which correspond to both dyke directions. The 115° direction is followed by a number of faults that cut, displace and deform the ophiolite (one lies along Wadi Hatta itself) and which were interpreted as transform faults by Smewing (1980). The 175° direction is followed by graben faults, such as those bounding the 'Alley' (Smewing *et al.* 1977; Fleet & Robertson 1980), that were also active on the sea floor and must lie parallel to an original rifting direction. Early hypotheses for the variability in the dyke directions were: intrusion of magma with a sigmoidal stress field at ridge transform intersections (Smewing 1980), and superimposition of a subradial 'arc' dyke set onto a 175° 'spreading axis' trend (Alabaster & Pearce 1979). Neither hypothesis, however, fits the bimodality of the data, nor the field observation that dyke sets of both orientations can be found within a few hundred metres of each other.

A more likely possibility is that the main back-arc spreading direction was 175° (relative to present-day north) with an oblique transform direction of 115° and that, in the Wadi Hatta – Wadi Fizh block, spreading took place in the transform direction itself. Since subduction took place towards the northeast, both the ridge and the transform direction must have made an angle of about 60° with the subduction direction and, initially at least, sea floor spreading could have taken place in either direction.

These two trends also exerted the main tectonic control on magmatism when 'arc' volcanic activity resumed. The 'seamount' event, with its 40 km spacing between major centres (figure 22) and its strong association with the transform direction, can be satisfactorily explained by considering the response of a non-isotropic lithosphere to stress (Vogt 1974; Mohr & Wood 1976; Bahat 1979). If the lithosphere is put under stress, by mantle diapirism or a related mechanism, it responds by propagation of fractures from the surface downwards. The first fractures to open will be those that transect the whole lithosphere – the transform faults – and the seamounts will therefore tend to be localized in these settings. The spacing of seamounts will, according to Vogt,

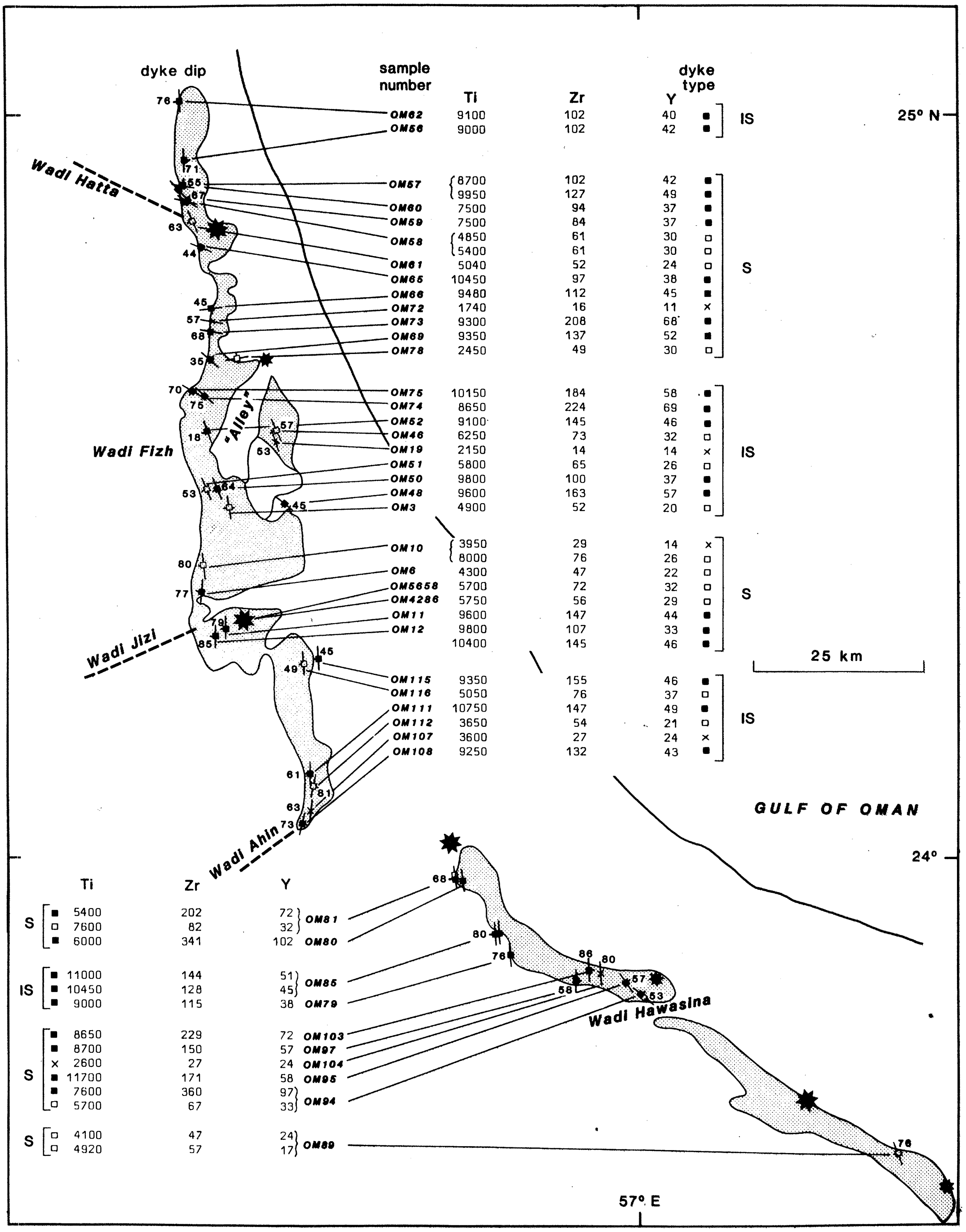

FIGURE 20. Sheeted dyke analyses (micrograms per gram) from the Oman ophiolite showing relations between strike and dip of dyke, geographical position and geochemical classification. Stippled area denotes lava and dyke outcrop. Solid squares denote classification as 'axis'; open squares denote classification as 'arc'; crosses denote classification as clinopyroxene ± olivine phyric 'arc'; asterisks denote arc seamount intrusives (large asterisks denote major intrusive centres); S, seamount areas; IS, inter-seamount area.

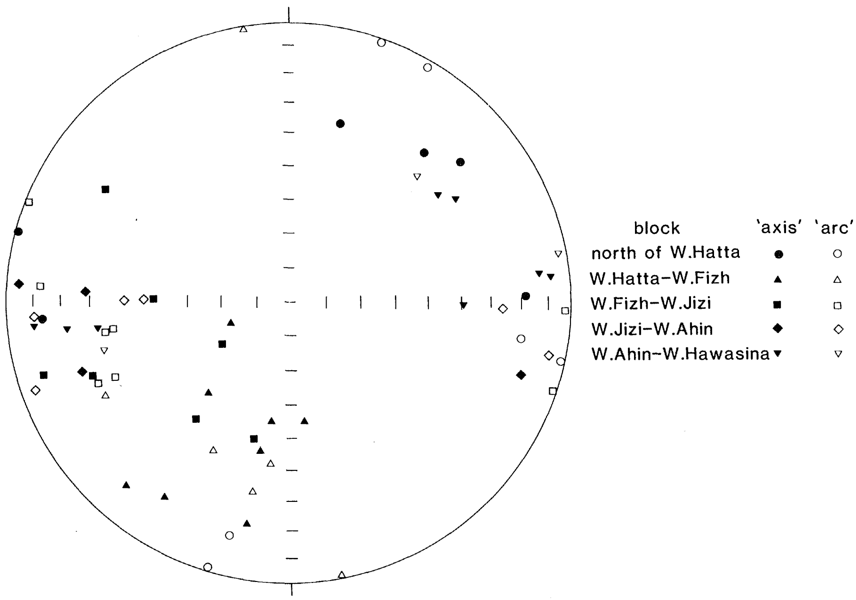

FIGURE 21. Stereographic plot of poles to dyke orientations, also showing relation to geographical position and geochemical classification.

be equal to the lithospheric thickness. In this case the estimated maximum thickness of 40 km is consistent with the concept of a young back-arc lithosphere (Abe & Kanamori 1970). To explain the 'rifting' event, we assume that further stress will cause domal uplift until new fractures propagate to intersect the zone of partial melting. The direction of propagation is now parallel to the original dyke swarm, the other plane of weakness, and the result is fissure eruptions and development of graben parallel to the original spreading axis.

Whatever the precise mechanism however, it is apparent that tectonic controls on magmatism are the same in the Oman back-arc basin as they are in within plate volcanic chains at present; the only major difference is that the stress at the base of the lithosphere was caused by subduction rather than by 'hot spot' activity.

SETTING AND EVOLUTION OF THE OMAN BACK-ARC BASIN

Our tentative reconstruction of the Oman back-arc basin just before continent – island-arc collision is shown in figure 22. Here, we briefly consider the evidence for this model, first for its geometry and then for the timing of events.

Geometry

In making the regional reconstruction, we can start with the African–Arabian and Eurasian plates whose relative positions and relative motions during the Cretaceous are known with reasonable certainty (see, for example, Smith 1971; Dewey *et al.* 1973). During the Cenomanian, the Arabian margin was moving approximately northeastwards towards Eurasia about a rotation pole in the North Atlantic. We have assumed in figure 22 that this direction corresponds with the direction of subduction beneath Oman, and the attitudes of thrust planes related to collision support this (Graham 1980). The position of India relative to Africa is also known but no attempt has been made here to tackle the difficult task of locating the various microcontinents.

One necessary assumption for the regional reconstruction is that all late Cretaceous (100–80 Ma) ophiolite complexes in this part of Tethys (Nowroozi 1972; Stoneley 1975; Stocklin 1977; Takin 1972) formed behind subduction zones by back-arc spreading and arc volcanism. This assumption is consistent with Cyprus, Hatay and Oman, the only complexes of this group so

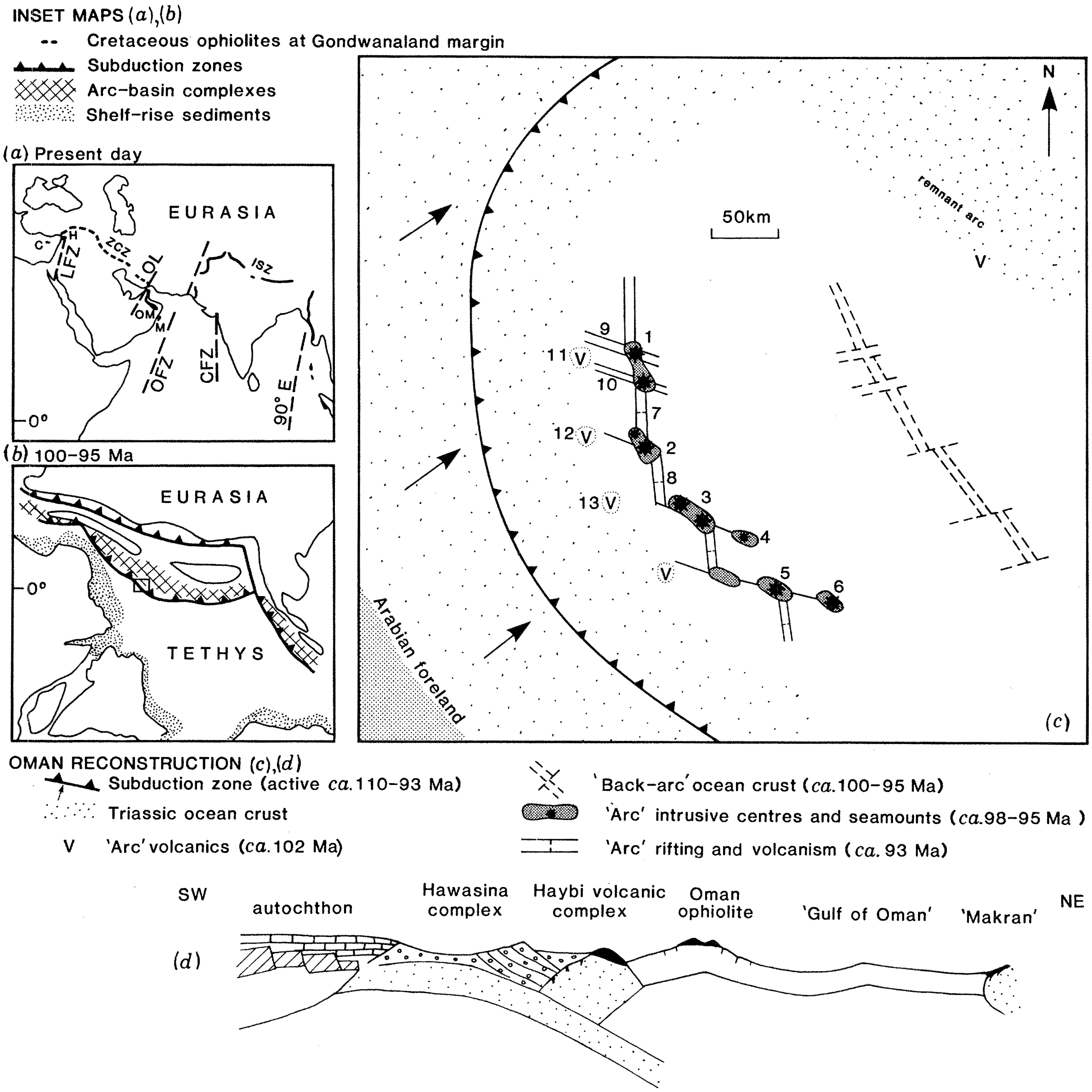

FIGURE 22. Location of Upper Cretaceous ophiolites, and important fracture zones in the present Alpine–Himalayan belt (*a*), regional setting of Oman in the mid-Cretaceous (*b*), and reconstruction of the Oman back-arc basin just prior to continental collision at about 93 Ma (*c*). The cross section (*d*) is based on the Lau Basin section of Barazangi & Isacks (1971). Key to (*a*): C, Cyprus; H, Hatay; OM, Oman; M, Masirah Is; ZCZ, Zagros Crush Zone; ISZ, Indus Suture Zone; LFZ, Levant Fracture Zone; OL, Oman Line; OFZ, Owen Fracture Zone; CFZ, Chagos Fracture Zone; 90° E, 90° E Ridge. Key to (*c*): 1, Hatta intrusive centre; 2, Jizi centre; 3, Mahab centre; 4, Ghuzahn centre; 5, Mabrah centre; 6, Daris centre; 7, the 'Alley'; 8, Jizi–Ahin zone; 9, Wadi Hatta fault zone; 10, Wadi Fizh–Wadi Rajmi fault zone; 11, Sumeini Window; 12, Asjudi Window; 13, Hawasina Window.

far studied geochemically (Pearce 1980), but no evidence is yet available for the others. A further assumption is that the arcuate locus, convex to the southeast, of the Oman mountain chain is the zone of greatest crustal thickening and therefore closely follows the configuration of the trench just before subduction. The locus of Indonesian ophiolites (Dickinson 1973) similarly locates the eastern part of the arc system, but the ophiolites around the northern margin of India (Stocklin 1980) and in the Zagros crush zone (Stocklin 1977; Takin 1972) may have moved northeastwards relative to Oman since they accreted to their respective continental margins.

For the detailed reconstruction of the Oman segment, we have assumed that the internal geometry of the Oman ophiolite has been retained (except around the domal structures), that lateral translation took place along a transform direction and that the locus of arc volcanism just before collision lay about 150 km from the trench. The positions of the arc seamounts (locations 1–6) and inter-seamount graben (7 and 8) can therefore be located from the geological evidence (Alabaster *et al.* 1980), as can some of the major faults (localities 9 and 10) that were active at the time of seamount volcanism. The sites where earliest arc volcanism (localities 11–13) developed on Triassic crust must be located between the seamount chain and the trench, the fore-arc region in the period just before collision. An interesting consequence of such a reconstruction is that the suture dividing the ophiolite complex proper from the underlying thrust sheets is located at the boundary between Triassic and Cretaceous oceanic crust.

The back-arc basin itself is difficult to reconstruct without palaeomagnetic information on whether the 175° (and 115°) spreading direction reflects the original spreading direction or whether rotation took place in the short period between initial back arc spreading and arc volcanism. The model depicted in figure 22 is based on an analogy with the Mariana trough (Karig *et al.* 1978), in which the trench appears to be migrating towards the converging Pacific plate during back-arc spreading. The geometry of Oman suggests that a similar migration might have taken place about a rotation pole in the north, but this part of the model is completely speculative at this stage.

Timing

Figure 23 summarizes the age data so far published on Oman. There is no direct evidence for the time at which subduction began. It must pre-date the first igneous event by at least 3 Ma (the period necessary to subduct several hundred kilometres of oceanic crust at a fast spreading rate) and perhaps by as much as 10 Ma; an age of 115–105 Ma is a reasonable estimate on this basis. Further refinement can be made by assuming that the development of new subduction zones in this part of Tethys must be balanced by new rifts or by increased spreading rates elsewhere. There are two main possibilities. Dewey *et al.* (1973) and Hays & Pitman (1973) pointed out that a sudden threefold increase in spreading rate in the World's rift system took place at about 110 Ma, continuing until about 80 Ma. Rabinowitz & LaBrecque (1979) and Johnson *et al.* (1976) have timed the splitting of South America from Africa and India from Antarctica at about 130 Ma. This latter date appears to be too early; 110 Ma is used here, therefore, to mark the start of subduction.

If this estimate is correct, the Cretaceous Haybi volcanic event must lie between 107 and 100 Ma. Unfortunately, no published dates are available, although Searle (1980) quotes lower–middle Cretaceous radiolarian ages from unpublished work by P. Tippit. For the ophiolite itself, radiometric ages of 98–95 ± 1 Ma have been determined on plagiogranites by Tilton *et al.* (1979). Assuming that the plagiogranites studied were of 'axis' rather than of 'arc' type, these ages correspond well with Tippit & Pessagno's (1979) early Cenomanian radiolarian age from sediments within the 'axis' lavas. Tilton *et al.* (1979) also identified a systematic geographical variation from 98 ± 1 Ma in the NW to 95 ± 1 Ma in the SE of the ophiolite, and deduced from this eastward younging a spreading rate of 2–5 cm a^{-1}, assuming a normal spreading axis. If a back-arc model is involved, with all the inherent complications in opening mechanism (Karig *et al.* 1978), these spreading rate calculations must be viewed as oversimplified. The concept of a fast spreading ridge is however consistent with the smooth topography of the back-arc basin (Smewing *et al.* 1977) and with an estimate of a 5–10 cm a^{-1} half-rate based on geochemical criteria (Pearce 1980).

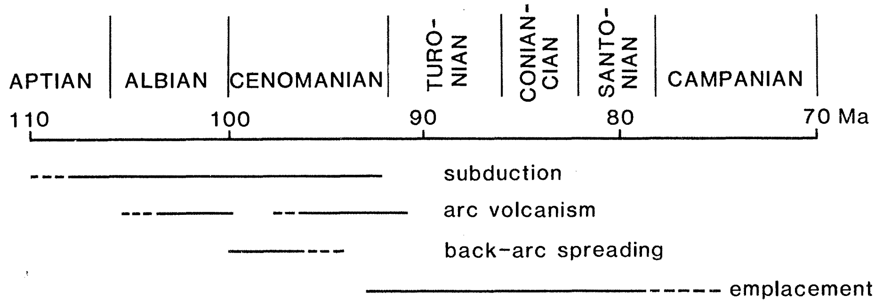

Figure 23. Summary of our model for the timing of events in the Oman back-arc basin. See text for discussion and sources of data. Cretaceous time scale from van Hinte (1976).

The curvature of the arc means that the back-arc basin must have been narrower in the north than the south, but an average of 3 Ma spreading at a 5 cm a^{-1} half rate giving an ocean 300 km wide may be a reasonable minimum estimate.

The time gap between axis and 'arc seamount' volcanism cannot have been very great because no significant thickness of sediment is found between the two (figure 5), but it must have been great enough to ensure that the two magma sources did not mix. The time gap between 'arc seamount' and the 'arc rifting' volcanism may be considered greater on sedimentological grounds (figure 10), but this may be a misleading comparison because hydrothermal activity was very intense at this time (Alabaster *et al.* 1980). Tippit & Pessagno (1980) dated the final (arc) volcanism as early Turonian, using the radiolaria present in pelagic sediments interbedded with and overlying the uppermost lavas.

At about this time convergence between Arabia and the Oman 'arc' was taken up by crustal thickening rather than subduction, and arc volcanism ceased. At 92.5 ± 2 Ma, alkali peridotite sills were intruded into the Haybi complex (Searle *et al.* 1980) and these rocks contain no record of any subduction-zone component. Probably subduction had ceased by this time and this and subsequent events can be related to emplacement tectonics (Searle & Malpas 1979).

It must finally be emphasized that this model applies only to the northern part of the Oman ophiolite. Coleman's (1979) synthesis of the area south of the Semail Gap indicates that that part of the ophiolite resembles more closely 'normal' oceanic crust. Moreover, the lavas of Masirah Island (figure 21) (Abbots 1979) do not show the low Cr of the *Geotimes* lava unit and again more closely resemble typical m.o.r.b. in their chemistry. These sequences therefore represent either the normal Tethyan ocean, or a part of the back-arc basin that is further removed from the subduction zone. Further geological studies in these areas and in the Gulf of Oman (White & Ross 1979) should enable these alternatives to be distinguished.

Discussion

Geological and geochemical evidence suggests that the Oman ophiolite is made up of back-arc oceanic lithosphere cut by the products of arc magmatism. Because the whole arc–basin complex, of which the ophiolite is a part, developed in less than 10 Ma (magmatism presumably stopping as a result of continent–arc collision), it is difficult to make direct comparisons between Oman and the long-lived arc-basin systems fringing the Pacific at the present day. Nevertheless, the Oman ophiolite appears to contain a complete section through a submarine arc, a part of the oceanic crust not yet sampled by dredging or drilling, and therefore may contain clues that will help solve some of the current problems of back-arc magmatism and tectonics. In particular, our studies in Oman suggest that the positive topographic feature represented by island arcs may be due to updoming of the underlying oceanic crust as much as to the accumulated products of volcanism and plutonism. They further indicate that the eruption and subsequent splitting of arc volcanoes closely resemble in mechanism the eruption of volcanoes and development of rift valleys in within plate settings. In the Oman model, upwelling mantle opens and propagates fractures along planes of weakness in the back-arc lithosphere, initially using transform faults and subsequently using faults parallel to the inherited trend of sheeted dykes. The resulting magmatism is initially restricted to discrete volcanic centres but later extends along the whole length of the crust, mainly within graben structures. It is easy to envisage that, had magmatism continued, these graben would have developed into new back-arc spreading centres. Studies in progress will enable this model to be tested further.

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Figure 3. Typical exposure of the basal lavas, the *Geotimes* lava group. Inferred eruptive setting at a back-arc spreading axis. Pillow lavas are metabasalts and are red. Field of view about 12 m across.

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FIGURE 4. Typical exposure of the first group of upper lavas, the Lasail lavas unit. Inferred eruptive setting in an island arc seamount. Pillow lavas are metabasalts and are green. Field of view about 6 m across.

IGURE 5. Contact between *Geotimes* and Lasail lava units, possibly marking the contact between back-arc ocean crust and an island arc seamount. Contact stands out because of the dark red – pale green contrast. Lasail unit contains inclined sheets. Field of view at contact about 50 m across.

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FIGURE 6. Lasail lava unit intruded by at least two intersecting sets of inclined sheets.

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'IGURE 7. Columnar jointed andesite lava flow capping andesites and basalts of the Lasail unit. Thickness of flow about 20 m.

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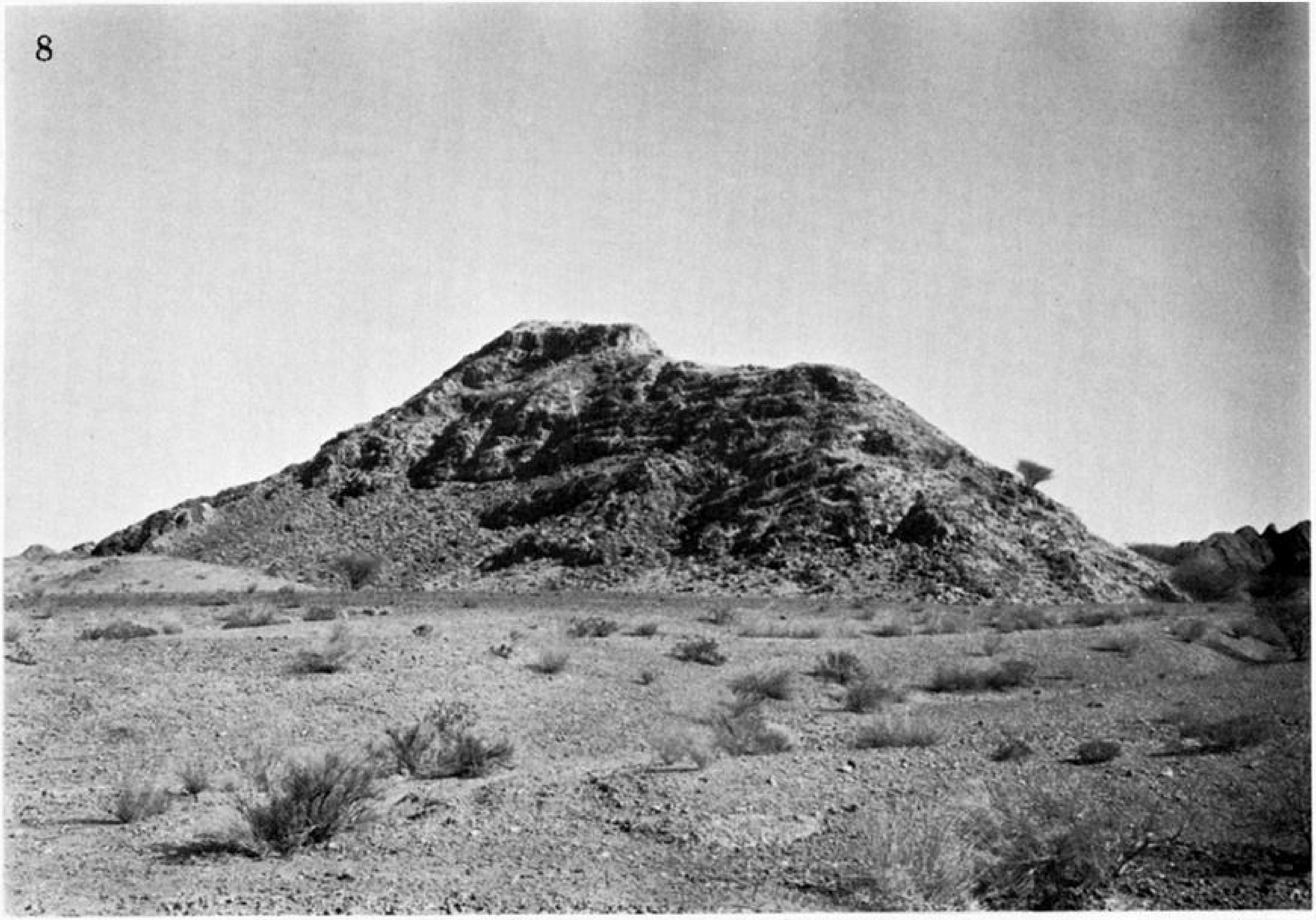

IGURE 8. Subhorizontal felsite (trondhjemite) sheets representing the final stage of fractionation of the Lasail unit. The outcrop is about 100 m across.

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Figure 9. Typical exposure of the vesicular Alley unit lavas. Inferred eruptive setting in graben between the island arc seamounts. Lavas are metabasalts and are grey–brown.

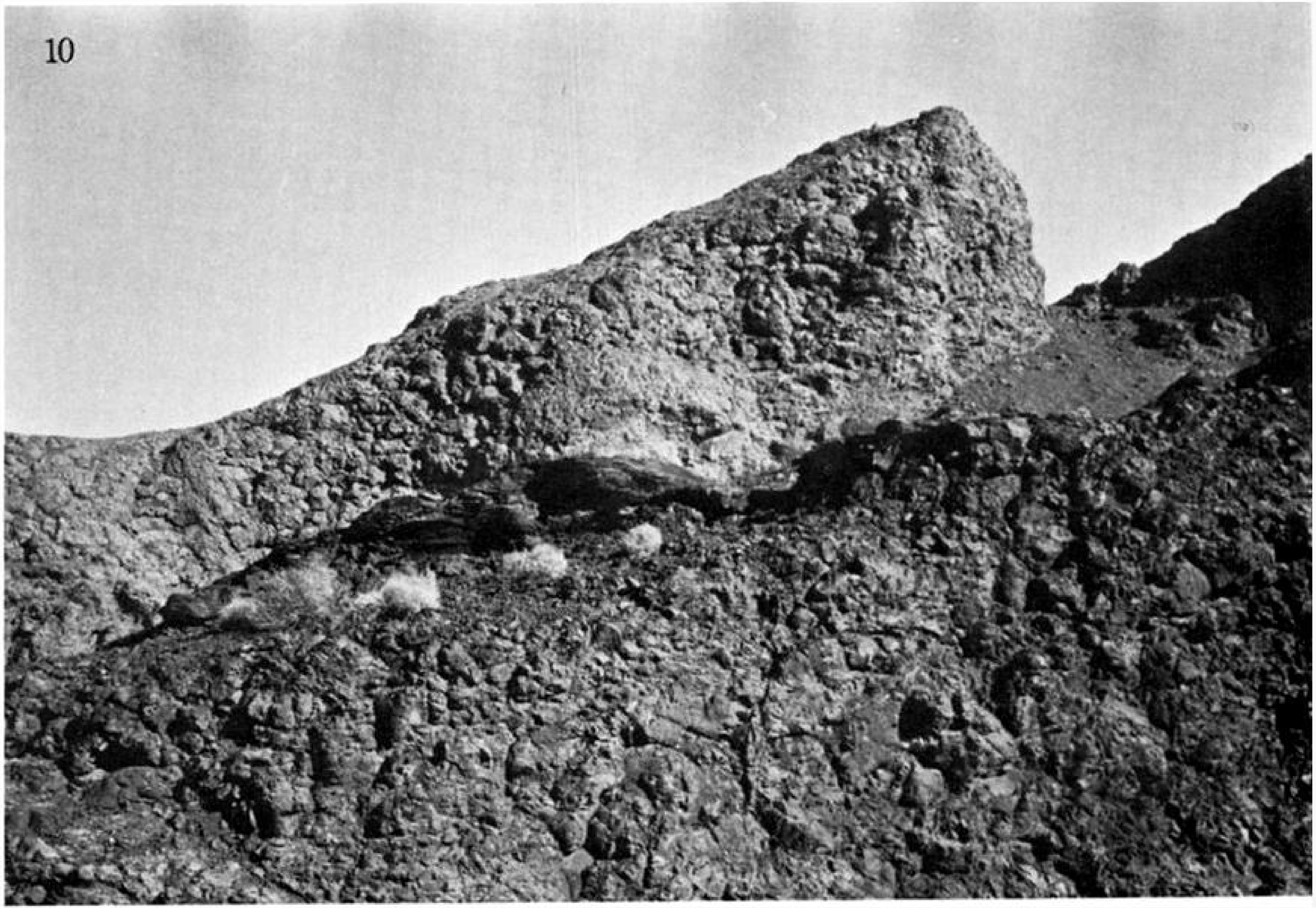

FIGURE 10. Contact between the *Geotimes* lava unit (foreground) and overlying Alley lava units (background), marked by about $\frac{1}{3}$ m of umbiferous sediment (the time-equivalent of the Lasail unit).

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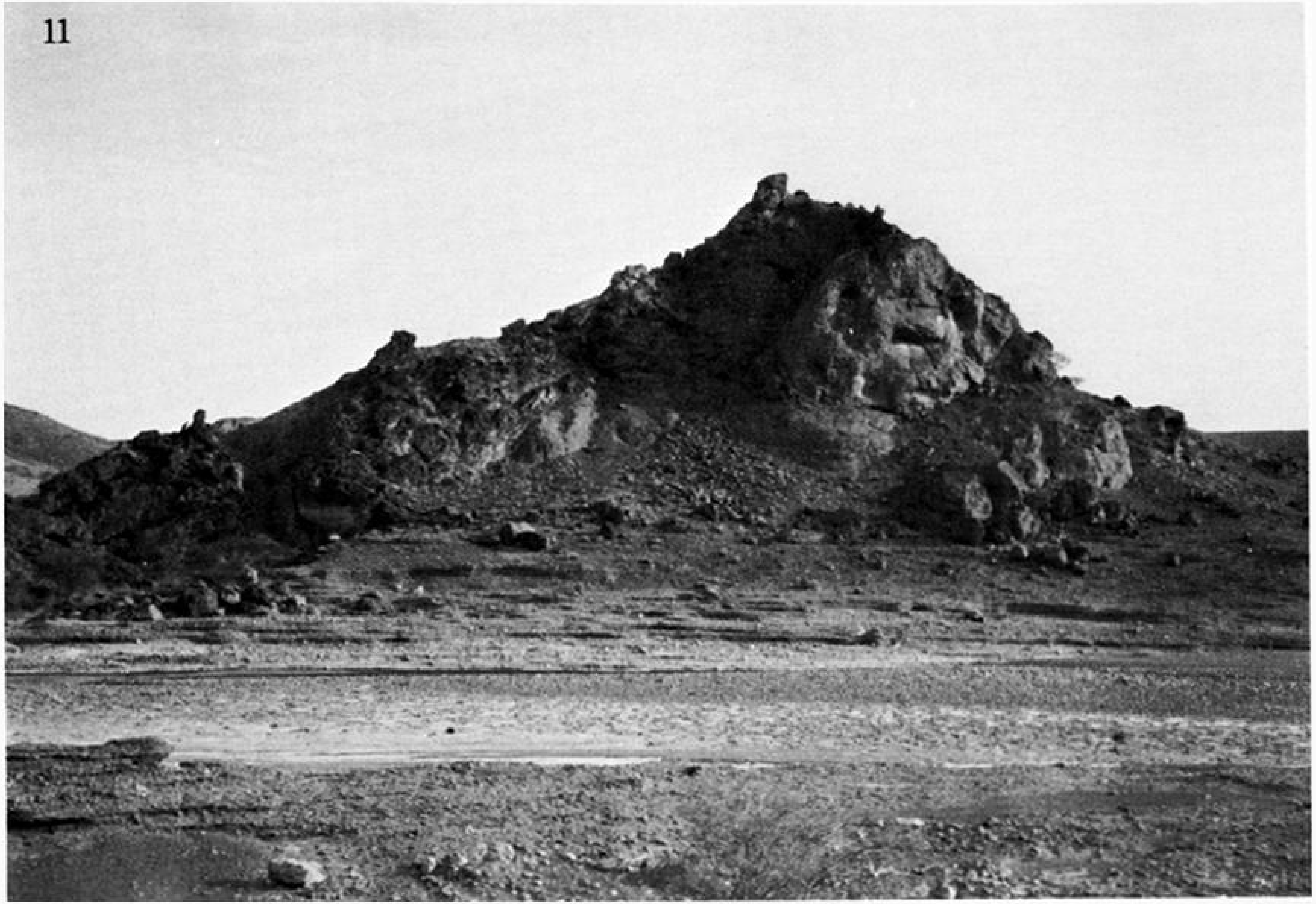

ɪɢᴜʀᴇ 11. Volcanic centre of the Alley unit developed within one of the seamount areas. The core of the hill is a dacite plug, the top is an obsidian lava. The hill is about 100 m across at its base.

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FIGURE 12. Ridge of rhyolite lavas of the Alley unit. The ridge is elongated along the strike of the dyke swarm of the underlying oceanic crust. The ridge is about 150 m across.