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SPATIAL GAPS IN ARC VOLCANISM: THE EFFECT OF COLLISION OR SUBDUCTION OF OCEANIC PLATEAUS

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ABSTRACT

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One of the major processes in the formation and deformation of continental lithosphere is the process of arc volcanism. The plate-tectonic theory predicts that a continuous chain of arc volcanoes lies parallel to any continuous subduction zone. However, the map pattern of active volcanoes shows at least 24 areas where there are major spatial gaps in the volcanic chains (> 200 km). A significant proportion (~ 30%) of oceanic crust is subducted at these gaps. All but three of these gaps coincide with the collision or subduction of a large aseismic plateau or ridge.

The idea that the collision of such features may have a major tectonic impact on the arc lithosphere, including cessation of volcanism, is not new. However, it is not clear how the collision or subduction of an oceanic plateau perturbs the system to the extent of inhibiting arc volcanism. Three main factors necessary for arc volcanism are (1) source materials for the volcanics—either volatiles or melt from the subducting slab and/or melt from the overlying asthenospheric wedge, (2) a heat source, either for the dehydration or the melting of the slab, or the melting within the asthenosphere and (3) a favorable state of stress in the overlying lithosphere. The absence of any one of these features may cause a volcanic gap to form.

There are several ways in which the collision or subduction of an oceanic plateau may affect arc volcanism. The clearest and most common cases considered are those where the feature completely resists subduction, causing local plate boundaries to reorganize. This includes the formation of new plate-bounding transform faults or a flip in subduction polarity. In these cases, subduction has slowed down or stopped and the lack of source material has created a volcanic gap.

There are a few cases, most notably in Peru, Chile, and the Nankai trough, where the dip of subduction is so shallow that effectively no asthenospheric wedge exists to produce source material for volcanism. The shallow dip of the slab may be a buoyant effect of the plateau imbedded in the oceanic lithosphere.

The cases which are the most enigmatic are those where subduction is continuous, the oceanic plateau is subducted along with the slab, and the dip of the slab is clearly steep enough to allow arc volcanism;

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yet a volcanic gap exists. In these areas, the subducted plateau may have a fundamental effect on the physical process of arc volcanism itself. The presence of a large topographic feature on the subducting plate may affect the stress state in the arc by increasing the amount of decoupling between the two plates. Alternatively, the subduction of the plateau may change the chemical processes at depth if either the water-rich top of the plateau with accompanying sediments are scraped off during subduction or if the ridge is compositionally different.

INTRODUCTION

One of the key elements of plate tectonics is the link between the process of subduction and the formation of linear volcanic chains. It is expected that, in general, a continuous chain of arc volcanoes should lie parallel to any continuous zone of subduction. Calculations of chemical mass balance often assume such a relationship. However, an analysis of the map pattern of active volcanoes shows at least 24 significant spatial gaps in the volcanic chains. In fact, about 30% of oceanic crust is subducted where no volcanoes are active. All but three of these gaps coincide with the collision or subduction of a large aseismic plateau or ridge.

This paper discusses the hypothesis that the collision or subduction of these oceanic plateaus is responsible for the formation of volcanic gaps. The idea that the collision of a large buoyant feature can have a major tectonic impact on the overlying plate of a subduction zone is not new. Vogt (1973) and Vogt et al. (1976) first attributed various modifications of the subduction process to the presence of large bathymetric features on the subducting oceanic seafloor. They suggested that the greater buoyancy of a plateau, provided by the greater thickness of relatively lighter crust beneath the plateaus, may cause the plateau to resist subduction. The subsequent effects may include the formation of cusps or irregular indentations in the trace of the subduction zone and the inhibition of back-arc spreading, arc polarity reversals, anomalous seismicity, slab and volcanic arc segmentation, and the shearing-off of seamounts at the trench. Kelleher and McCann (1976, 1977) further correlated the subduction of oceanic plateaus with zones where there were fewer great earthquakes and shorter rupture zones, gaps in intermediate depth earthquakes and also gaps or offsets in volcanic chains. They roughly calculated the relative buoyancy of oceanic lithosphere containing a 70-km-thick oceanic plateau of intermediate density and concluded that there may well be enough density contrast for the lithosphere to resist subduction. Various other authors have investigated specific effects of the collision or subduction of a buoyant feature, including low-angle subduction and Laramide type tectonics (Pilger, 1981), anomalous seismicity and intermediate depth earthquake gaps (Chung, 1978), volcanic gaps in South America (Nur and Ben-Avraham, 1981) and rotation of an arc upon collision (McCabe and Uyeda, 1983; McCabe et al., 1982). The possibility that the buoyant feature may also accrete to the convergent margin upon collision has been discussed by Nur and Ben-Avraham (1981).

The spatial association between volcanic gaps and subducting oceanic plateaus is striking. However, the specific nature of the interaction at the subduction zone which inhibits arc volcanism is not as clear. In this paper, we first discuss the process of arc volcanism as it is tied to subduction, in order to outline the major factors required for arc volcanism. We then attempt to describe and classify the 24 volcanic gaps and discuss ways in which the subduction or collision of an oceanic plateau can perturb the process of arc volcanism. We find in fact that there are several distinct ways in which the collision or subduction of an aseismic ridge or oceanic plateau may create a volcanic gap.

ARC VOLCANISM

Vital to the understanding of how the subduction of an oceanic plateau could create a volcanic gap is the understanding of the process of arc volcanism itself. Although it is clear that subduction and arc volcanism are intimately linked, the specific details of the physical and chemical processes which produce arc volcanoes are fairly controversial. It is generally agreed that the genesis of arc magmas is a complicated process which involves source material from the subducted oceanic lithosphere and/or the asthenospheric wedge, and occasionally also the overlying arc lithosphere (Ringwood, 1977; Delany and Helgeson, 1978; Anderson et al., 1978, 1980; Wyllie, 1979, 1982; Gill, 1981). However, the nature and proportion of each contribution is still unclear.

Figure 1 shows a schematic model of the genesis of arc magmas. During subduction, the cold oceanic lithosphere is transported into the deep mantle beneath the volcanic arc. This lithospheric slab includes crustal basalt and gabbro formed at the oceanic spreading center, hydrothermally altered to an unknown depth and degree, overlain by pelagic and trench sediments and accompanied by serpentinite bodies and the peridotite of the depleted mantle. As the oceanic slab subducts, it is subjected to increasing pressures and temperatures which lead to progressive metamorphism and dehydration of the crust and overlying sediments (Anderson et al., 1976) below 70 km depth. This process absorbs heat and releases water and other volatile components. It also creates extra pore space which might allow fluid to circulate through the slab and to escape to the overlying mantle (Delany and Helgeson, 1978). The motion of the relatively cold slab also sets up a counterflow in the asthenosphere above (Andrews and Sleep, 1974), which provides a constant supply of heat and new mantle, and may cause frictional melting of the uppermost slab if the temperatures and asthenospheric viscosity are high enough (Toksöz et al., 1971).

Magma will form when the temperature exceeds the solidus of the source material. This condition has been proposed to be achieved within a subduction/arc system by frictional melting of the slab, by movement of material into higher temperature regions through either convection within the wedge or diapiric uprise,

TABLE 1

Data of the 24 volcanic gaps discussed in this paper

Gap location	Plateau	Gap size (km)	S [a] (km)	Dip [b] (°)	Change in dip (°) [c]	Convergence rate [d] (cm/yr)	Seafloor age [e] (m.y.)	Age of ridge [f] (m.y.)	T (km) [g]	L (km) [h]	D (km) [i]
A. Collision / Reorg.											
1. Costa Rica	Cocos Ridge	200	28.5	25 (1)	40	9.6	11	14 (21)	18 (25)	140	80
2. Colombia	Panama arc	400	57.8	21 (2)	15	2.2	70	70 (2)	26 (26)	500	210
3. Mindoro	Palawan	300	–	90 (3)	−45	7.69	27 (14)	L. Pal. (14)	> 15 (27)	300	286
4. Taiwan	Asia Margin	500	–	85 (3)	−40	7.04	37 (14)	–	30	250	200
5. Java	Australia	550	37.4	60 (4)	17	7.00	150	Pre-C (15)	32–34(15)	750	650
6. Yap	Caroline R.	500	–	–	–	1.60	Olig.	32 (16)	–	–	–
7. S. Mariana	Caroline R.	350	81.1	50 (5)	40	5.35	80/170	32 (16)	–	350	200
B. Shallow dip											
8. Peru	Nazca Ridge	1600	59.7	0–10 (6)	20–30	8.4	40	38 (17)	18 (28)	720	200
9. Chile	Juan Fernandez R.	700	60.0	0–10 (7)	20–30	8.7	36	36 (17)	–	770	190
10. S. Chile	Chile Rise	1000	185.7	–	–	2.24	4 (18)	0 (18)	–	–	–
11. Nankai	Palau–Kyushu	700	63.5	10 (8)	–	4.54	24 (19)	38 (22)	15 (29)	200	70
12. New Guinea	Euripik Rise	800	–	25 (9)	–	9.3	35	26 (23)	18 (30)	475	200
C. Enigmas											
13. Tonga	Louisville R.	550	87.2	53 (10)	25	8.0	82–118		–	800	550
14. N. Tonga	Capricorn Smt.	300	87.2	40 (10)	25	9.3	82–118		–	900	450
15. S. New Hebrides	Norfolk Ridge	350	143.6	65 (11)	5	6.6	35	–	23 (31)	260	200
16. N. New Hebrides	Torres Rise	375	143.6	75 (11)	−5	9.9	44			440	375
17. Solomons	Rennel Island	375	117.4	45 (12)	15	4.4	–	–	–	120	100
18. N. Marianas	Magellan Smt.	450	81.1	55 (5)	35	7.59	170	100 (16)	–	420	300

19. Izu-Bonin	Ogasawara Smt.	350	65.0	45 (5)	10	8.49	155	100 (16)	–	800	520
20. Ryukyus	Oki-Daito/Daito	550	63.5	35 (5)	10	5.58	60	50 (22)	16 (32)	500	300
21. Middle America	Tehuantepec	250	28.5	55 (1)	15	7.6	10–35 (20)	10 (20)	13 (20)	480	200
D. No ridge											
22. N. Macquerie	–	200	–	–	–	2.12	–	–	–	–	–
23. N Kermadec	–	400	66.2	43 (10)	–	7.46	–	–	–	950	660
24. S. Kermadec	–	600	66.2	62 (10)	–	6[illegible]	–	–	–	600	500
E. No gap											
25. New Hebrides	D'Entrecasteaux	–	143.6	75 (11)	–	9.6	–	–	–	275	200
26. Kamchatka	Meiji Seamount	–	36.9	50 (13)	–	8.7	–	72 (24)	18	–	–

[a] Average spacing (S) between volcanoes. Data from Shimozuro and Kubo (1981).

[b] Dip angle of the seismic zone at depths beneath volcanic gaps.

[c] The change in dip angle between a volcanically active zone and volcanic gap. A positive value indicates that the dip beneath the gap is shallower. Sources as for (b).

[d] Relative convergence rates calculated using the poles of Minster and Jordan (1978), Ranken et al. (1984), or Seno (1977). Rate calculated at the point of intersection with plateau.

[e] Age of the seafloor being subducted to either side of the oceanic plateau mostly from magnetic anomalies shown on the Plate-Tectonic Map of the Circum-Pacific Region (1981) unless otherwise stated.

[f] Age of the oceanic plateau where known or reasonably inferred.

[g] Thickness (T) of the oceanic plateau from refraction or gravity data.

[h] Length (L) of the subducting slab beneath the volcanic gap. Sources as for (b).

[i] Maximum depth (D) of the Benioff zone. Sources as for (b).

References: (1) Burbach et al. (1984); (2) Pennington (1981); (3) Hamburger et al. (1983); (4) Fitch (1970); (5) Katsumata and Sykes (1969); (6) Barazangi and Isacks (1979); Hasagawa and Sacks (1981); (7) Barazangi and Isacks (1976); (8) Kanamori (1972); (9) Denham (1969); (10) Sykes (1966); (11) Pascal et al. (1976); (12) Pascal (1979); (13) Veith (1974); (14) Taylor and Hayes (1980); (15) Jacobson et al. (1979); (16) Vogt et al. (1976); (17) Pilger (1981); (18) Herron et al. (1981); (19) Kobayashi and Nakada (1978); (20) Couch and Woodcock (1981); (21) Heath and Van Andel (1973); (22) Klein and Kobayashi (1980); (23) Winterer et al. (1971); (24) Scholl and Creager (1973); (25) Barday (1974); (26) Briceno-Guarupe (1978); (27) Murauchi et al. (1973); (28) Couch and Whitsett (1981); (29) Ludwig et al. (1973); (30) Den et al. (1971); (31) Collot and Missegue (1977); (32) Murauchi et al. (1968).

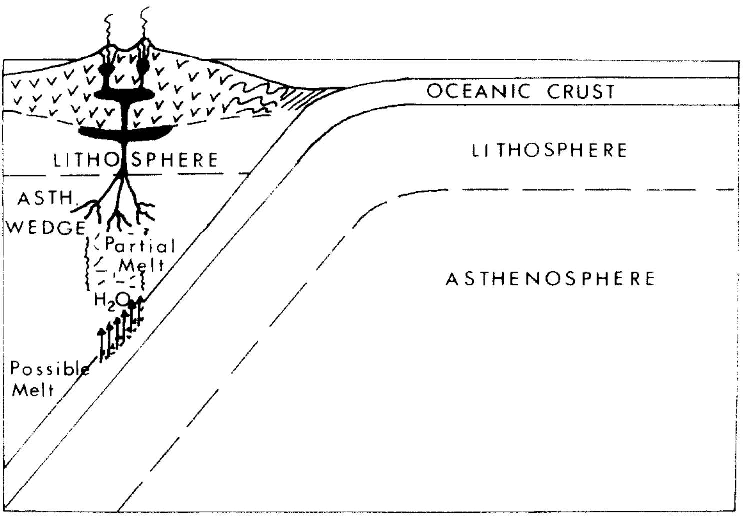


Fig. 1. Schematic cross-section of subduction system showing general model of arc magma genesis.

and by the lowering of the melting temperature by the introduction of volatiles into the system. The actual process may involve various combinations of these depending on initial conditions. The resultant basaltic melt ascends, collecting into magma reservoir systems within or at the base of the crust. The crust itself acts as a filter, retarding the rise of the magma and promoting differentiation of the magma into a more silica-rich melt (Gill, 1981).

The most important components to producing andesitic magma seem to be the presence of a moving oceanic slab, an asthenospheric wedge, water, and either a compressional stress within the arc or a 25-km-thick crust (Gill, 1981). The moving slab, although itself cold, provides the heat for melting by causing convection within the wedge and also possibly by frictional heating along the slab–wedge interface. The slab also contains water and other volatiles which, if they escape into the overlying mantle, may allow for a lower melting temperature within the wedge. It is also possible that the water necessary for partial melting is already contained in the mantle. The asthenospheric wedge is essential in order to permit convection of hot material and is also necessary as a source area to explain the volume and geochemistry of the volcanics. Finally, a mild horizontal compressive stress or a 25 km thick crust within the arc is necessary to produce orogenic andesite by fractionation of the basaltic magma.

Since the process of arc volcanism is not well understood, an analysis of perturbations of the process is necessarily speculative. However, regardless of the specific models for the process, one can safely argue that there are three separate factors necessary for the formation of arc volcanoes: (1) source materials for the volcanics—melt and volatiles from the subducting slab or from the overlying

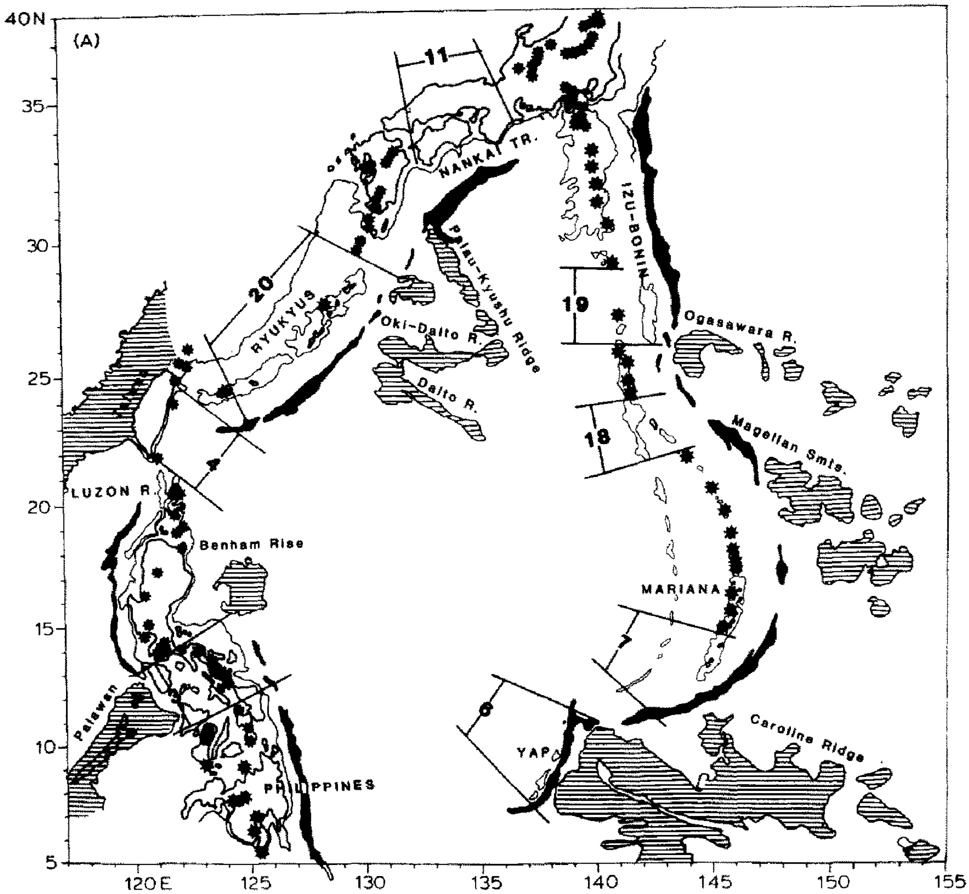


Fig. 2. Maps showing locations of volcanic gaps, trenches, and bathymetric rises. The gaps are numbered as in Table 1. A. The Philippine Sea region. B. The Solomon, New Hebrides, Tonga and Kermadec arcs. C. South America. D. Central America. E. Northern margin of Australia and New Guinea. (To be continued on pp. 202–204.)

asthenospheric wedge; (2) a heat source, either for the dehydration or for the melting of the wedge or the slab and (3) a favorable state of stress in the overlying lithosphere. The absence of any one of these components may cause a volcanic gap to form.

VOLCANIC GAPS

The 24 volcanic gaps discussed in this paper are listed in Table 1 and shown in Fig. 2A–E. These maps also include the location of trenches and oceanic plateaus. The volcano data base is from the "List of the World Active Volcanoes" edited by Katsui (1971) and also the Plate-Tectonic Map of the Circum-Pacific Region (1981)

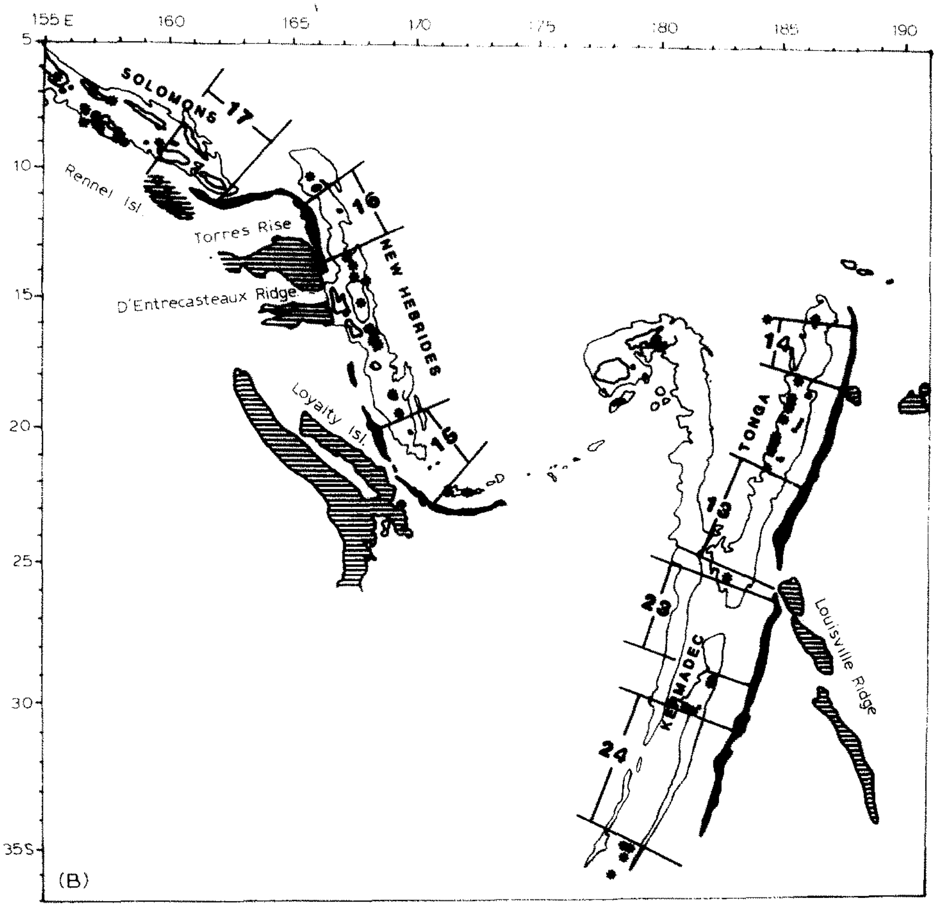
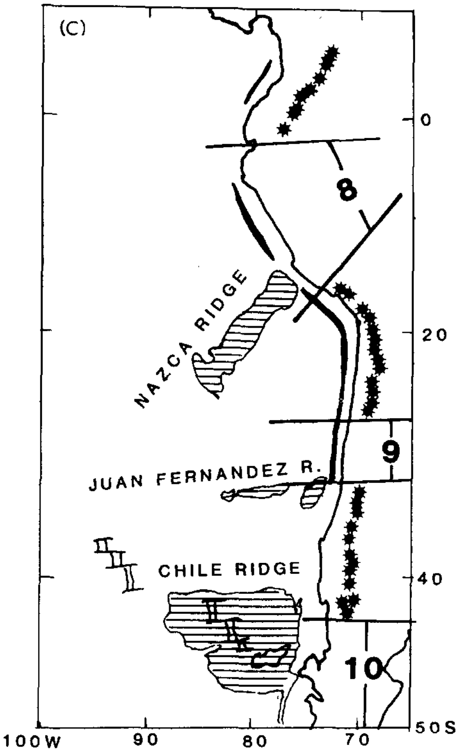
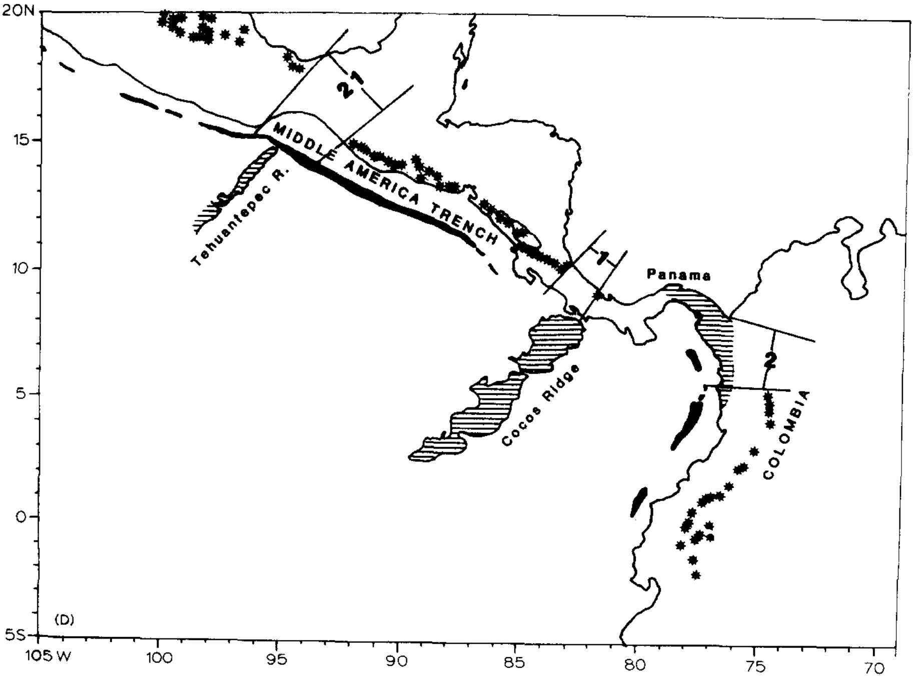


Fig. 2 (continued).

and includes volcanoes of Holocene age. These compilations generally show just the locations of stratovolcanoes or volcanic edifices which are exposed above sea level, and may therefore be biased in favor of andesitic volcanoes or volcanoes exposed on continental margins and biased against basaltic volcanoes, especially those in island arcs. More detailed surveys, especially in the southwest Pacific may reveal several sites of submerged volcanic activity not included in current compilations. Therefore, a gap in a chain of volcanoes does not necessarily mean a complete lack of volcanism of any kind, but rather a lack of the differentiated, more silica-rich magmas typical of "arc volcanism". It is this specific kind of volcanism that the subduction or collision of a plateau seems to affect.

The gaps were chosen by visual inspection of the map patterns using three criteria. (1) There must be either a complete spatial gap or an abrupt change in the average spacing or orientation of the volcanic chain. (2) The gap must be signifi-


(C)
0
8
20
NAZCA RIDGE
9
JUAN FERNANDEZ R.
CHILE RIDGE
40
10
50S
100W
90
80
70


20N
15
10
5
0
5S
(D)
105 W
100
95
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75
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21
MIDDLE AMERICA TRENCH
Tehuantepec R.
1
Panama
Cocos Ridge
2
COLOMBIA

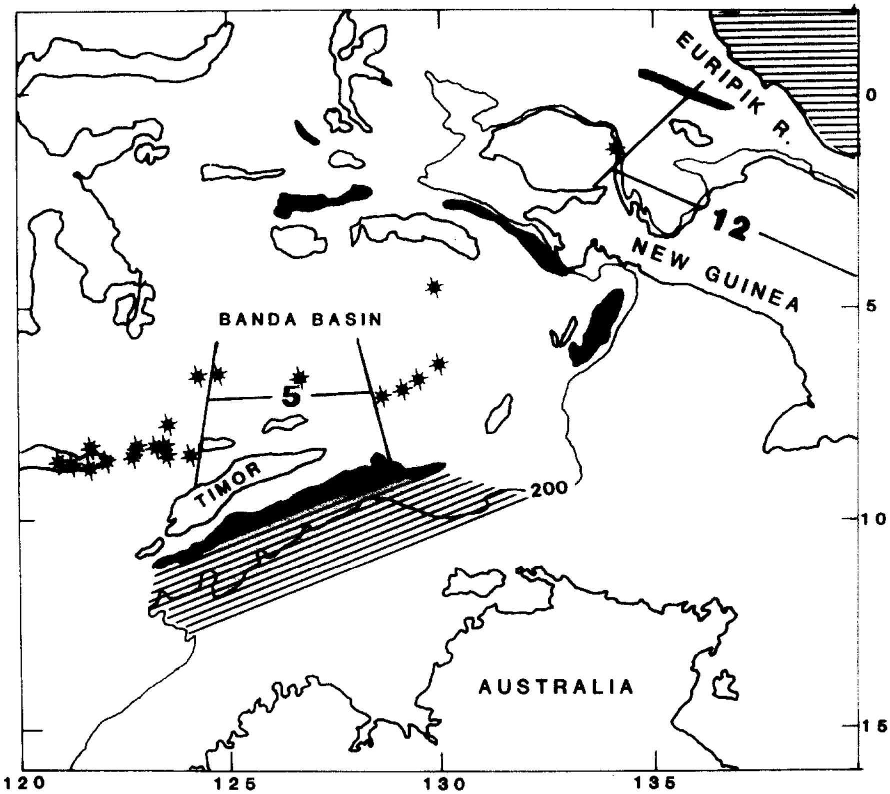

Fig. 2 (continued).

cantly larger than the average volcano spacing within that particular arc. Although the average spacing between volcanoes within active volcanic chains varies a great deal from arc to arc, the volcano spacing within each arc tends to be fairly consistent. In general, the gaps chosen are at least four times as large as the average volcano spacing defined by Shimozuru and Kubo (1983). Exceptions occur in the New Hebrides and Solomon arcs where the calculation of the average spacing included the gaps and therefore produced anomalously large spacing values. (3) The gap must be larger than 200 km in length which is larger than the maximum average volcano spacing of 185.7 km in southern Chile (Shimozuru and Kubo, 1983). The overall average volcano spacing varies according to author but ranges from 55–70 km (Vogt, 1974) to 100 km (Lingenfelter and Schubert, 1976). The variation in volcano spacing between arcs may be related to differences in lithospheric thickness (Vogt, 1974), differences in convergence rates (Shimozuru and Kubo, 1983), inherited fractures within the basement of the arc (Kienle et al., 1983), or segmentation within the convergent margin (Carr et al., 1979). Although the size of the gaps defined by such criteria may be affected by the possible sampling bias towards andesitic volcanoes mentioned above, several factors suggest that the locations of the gaps are not just a function of the sampling. The locations of the gaps concur with other anomalous tectonic features (Kelleher and McCann, 1976; Vogt et al., 1976). Within arcs where other factors such as convergence rate or the age of the plate remain fairly constant along strike, such as the Tonga arc, Peru, or Chile, the volcanoes still seem to cluster or be completely absent.

Figures 2A–E show clearly the close spatial association between the location of these volcanic gaps and the zones of collision or subduction of oceanic plateaus. The consequent question then is in what ways may these plateaus perturb the process of arc volcanism. The effect of a plateau at a given subduction zone may be partially dependent on characteristics inherent to the plateau such as relative buoyancy, crustal thickness, age, and size, and also partially dependent on characteristics of the convergent system itself such as the rate of convergence, age of the plate, dip, length, and depth of the slab, relative coupling at the interface, and nearness of plate boundaries. Therefore, we examined various characteristics for each volcanic gap and oceanic plateau in an effort to classify the gaps and to aid in understanding their origin. Table 1 includes the most important data for the 24 gaps. Figure 3 shows histograms of each characteristic.

Although no single pattern stands out (Fig. 3), the gaps can be divided into four groups based on combinations of characteristics: (1) cases where the collision of a buoyant oceanic plateau has caused subduction to stop or slow down and plate boundaries have had to reorganize to accommodate continuing convergence; (2) cases where the subduction of a buoyant feature has caused the slab to dip at an anomalously shallow angle; (3) cases where the subduction of an oceanic plateau doesn't drastically change the subduction geometry but somehow inhibits the process of arc volcanism at depth; and (4) cases where no unusual bathymetric feature is currently subducting. Within each group, the volcanic gaps seem to have formed by basically the similar process.

Case A: Collision / plate boundary reorganization

The clearest and probably most common way in which a volcanic gap may form is when a feature embedded within the oceanic crust is so buoyant that it completely resists subduction, in effect, colliding with the arc and causing local plate boundaries to reorganize (Fig. 4a). There are 7 cases (listed in Table 1) ranging in size and thickness from an oceanic plateau (Caroline R./Yap), to an island arc (Panama/ Colombia) to a continental margin (Asia/Taiwan). The two most important factors seem to be the relative buoyancy of the feature and the nearness of the collision to an adjustable plate boundary.

The reorganization of plate boundaries, which is essential if subduction is to stop at the collision location, may occur in several ways. For example, the collision of the Cocos Ridge at the Middle America Trench has been held responsible for previous

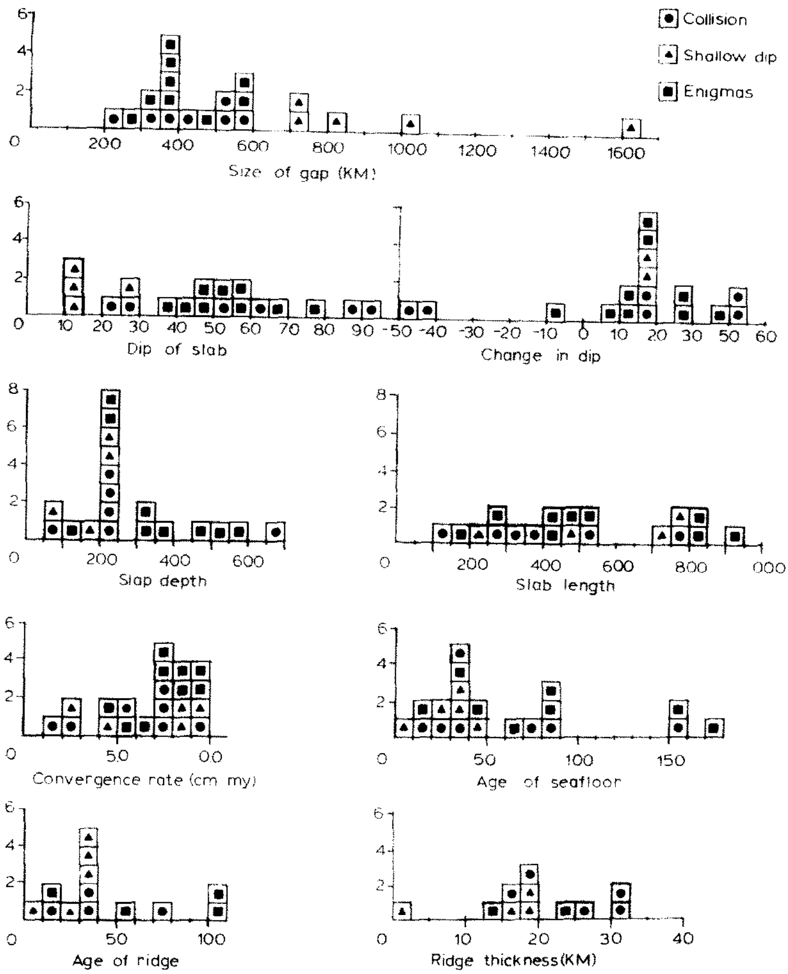

Fig. 3. Histograms of the most important characteristics of volcanic gaps and aseismic ridges.

plate reorganizations in the Galapagos area and is now thought to be blocking the system in Costa Rica. Besides the presence of a volcanic gap, the trench has disappeared, the Benioff Zone is 40° shallower in dip and shorter in length than to the west of the collision and the coastal region is experiencing rapid uplift. Pennington (1981) has suggested from earthquake studies that a new plate-bounding transform fault is forming west of the Cocos collision zone. This would replace the Panama Fracture Zone and would allow subduction to continue along the rest of the Middle America Trench.

The subduction system may also reverse polarity as inferred for Taiwan where the Asia continental margin is unable to subduct beneath the volcanic arc of the North

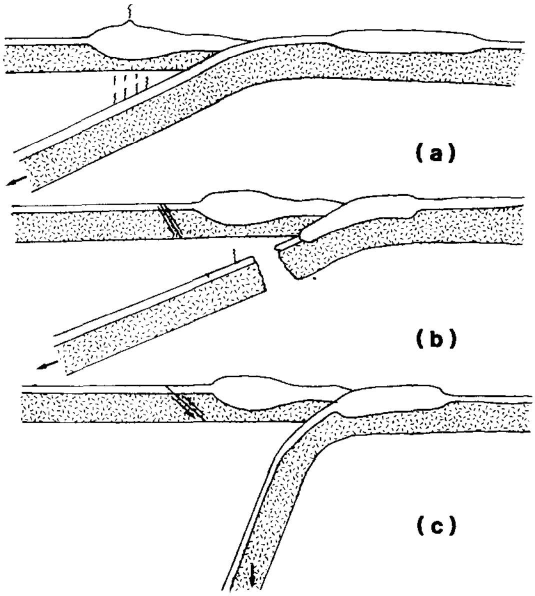

Fig. 4. Schematic model of the collision of an aseismic ridge. a. Before collision. b. After collision. Slab breaks off. c. After collision has stopped subduction. Slab steepens as it sinks.

Luzon Ridge (Hamburger et al., 1983) and for Timor where the Australian continental margin is colliding with the Java arc (Hamilton, 1979). In both cases, the oceanic crust which is currently in the back-arc position with respect to the arc appears to underthrust the arc, possibly starting a new subduction zone. Older examples of such polarity reversals may have been caused by the collision of the Ontong–Java Plateau with the Solomon arc (Halunen and Von Herzen, 1973) or the collision of the Benham Rise with the Philippine trench (Hamburger et al., 1983).

The collision of the Caroline Ridge with the Yap–Mariana system has caused the back arc of Yap to thrust over the arc (Hawkins and Batiza, 1977) and the South Marianas to rotate during back-arc spreading into an oblique position with respect to subduction (Larson et al., 1975; McCabe et al., 1982). Both the Panama/Colombia collision and the Palawan/Mindoro collision are occurring within convergent systems complicated by strike-slip and thrust fault tectonics. These systems are therefore able to adjust to the collisions through non-rigid tectonics.

In all these cases, subduction has slowed down or stopped. Without a moving slab, the supply of water or melt from the slab will disappear, there will be no frictional heating along the interface nor will there be a driving force for convection within the wedge which also provides heat. With a reduced magma source (factor 1) and no source of heat (factor 2) volcanism will cease.

An interesting problem is what happens to the slab when the buoyant feature collides. In six of the seven cases listed in Table 1, a Benioff zone still exists. In four cases the dip is shallow where the plateau is subducting but in at least two cases the dip is nearly 90. Although the dip might initially become shallower as the plateau attempts to subduct, a full-scale collision would slow or stop convergence at the trench. With a limited horizontal component of convergence, the vertical pull of gravity would presumably take precedence, and the slab would either detach from the anchored surface block (Fig. 4b) or would become steeper (Fig. 4c). A mechanism for such a detachment has been discussed by Price et al. (1983). At both the Palawan and Taiwan collision areas, the dip of the Benioff zone appears to increase from 50°–60° to nearly 90°. In Colombia, Panama seems to have been accreting to the margin of the continent even as the slab to which it was attached subducts at a shallow angle. Whether the slab is broken at the surface or the Panama arc is scraped off the top can't be ascertained from available seismic data.

Case B: Plateau subduction with shallow dip

A second way in which a volcanic gap may form is where the oceanic plateau subducts rather than collides, but the dip of the subducting slab becomes so shallow that a sufficient asthenospheric wedge does not exist to produce source material or heat for volcanism (Fig. 5). Five cases are listed in Table 1. The best studied cases are the volcanic gaps in Peru and Chile which correspond to regions of anomalously shallow subduction (< 15). Pilger (1981) has attributed the low-angle subduction in South America to the presence of the relatively buoyant Nazca and Juan Fernandez Ridges within the subducting slab. The other three cases are at the Nankai Trough, the intersection and south of the Nazca–Antarctica spreading ridge in southern Chile, and the Euripik Rise/northwestern New Guinea boundary.

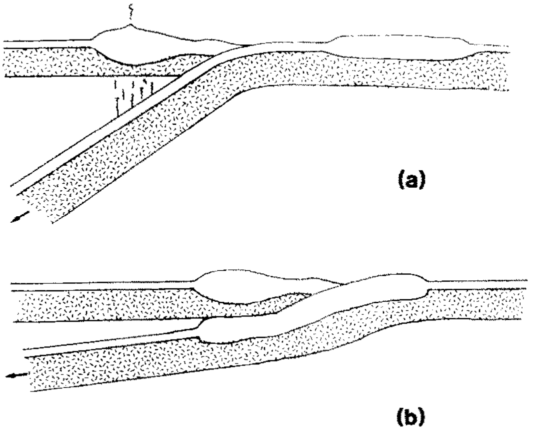

Fig. 5. Schematic model of subduction of quite buoyant ridge. a. Before subduction of ridge. b. Ridge is subducting and dip of slab is so shallow that no mantle wedge exists.

In all these cases, unlike the first group, the plateau actually subducts and subduction continues. Presumably, either the plateaus are not quite buoyant enough to completely stop subduction or the plate is large enough, with no nearby adjustable plate boundaries, that the pull of the slab on each side of the plateau is enough to overcome some of the buoyant effect, allowing subduction to continue, and preventing a collision between the plateau and the margin.

The buoyancy of the plateau containing slab is enhanced in the case of the Nankai Trough and southern Chile by the young age of the subducting, oceanic lithosphere and in Nankai and New Guinea by the inferred fairly recent onset of subduction. Subduction at the Nankai Trough is thought to be so recent that the slab has only reached 70 km into the mantle which is clearly too shallow for arc volcanism. However, the slab itself is at least 200 km long and would be long enough to produce arc volcanoes if the dip of the slab were steeper. In this case it is the buoyancy and shallow dip of the slab which prohibits arc volcanism rather than the inferred young age of subduction. In fact, recent work by Sacks (1983) suggests that the slab continues as an aseismic mass much farther beneath southwest Japan than can be seen from the Benioff earthquake zone. However, it is the dip of the slab which creates the gap. The slab itself is long enough to reach critical depths if the dip were steeper.

Case C: Plateau subduction with steep dip

The most enigmatic and possibly the most intriguing of the volcanic gaps are those where the oceanic plateaus subduct rather than collide and the dip of the slab remains fairly steep (Fig. 6a). There are nine cases shown in Table 1. In these cases,

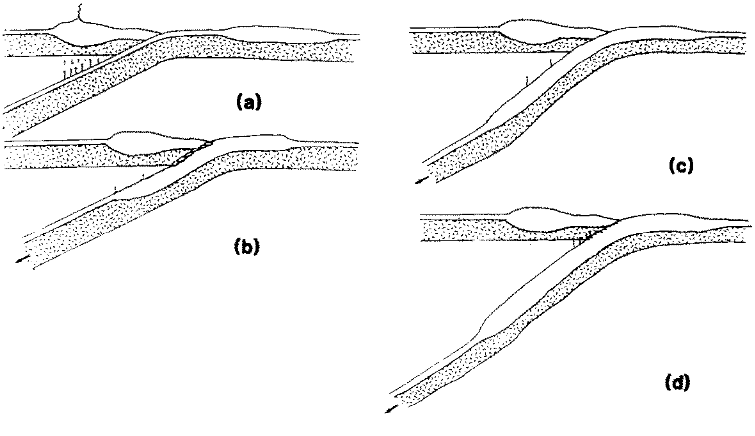

Fig. 6. Schematic model of subduction of only slightly buoyant ridge. a. Before subduction. b. Top of ridge is sheared off during subduction. c. Ridge contains too few hydrous minerals. d. Water is released from slab at depths too shallow to produce arc volcanism.

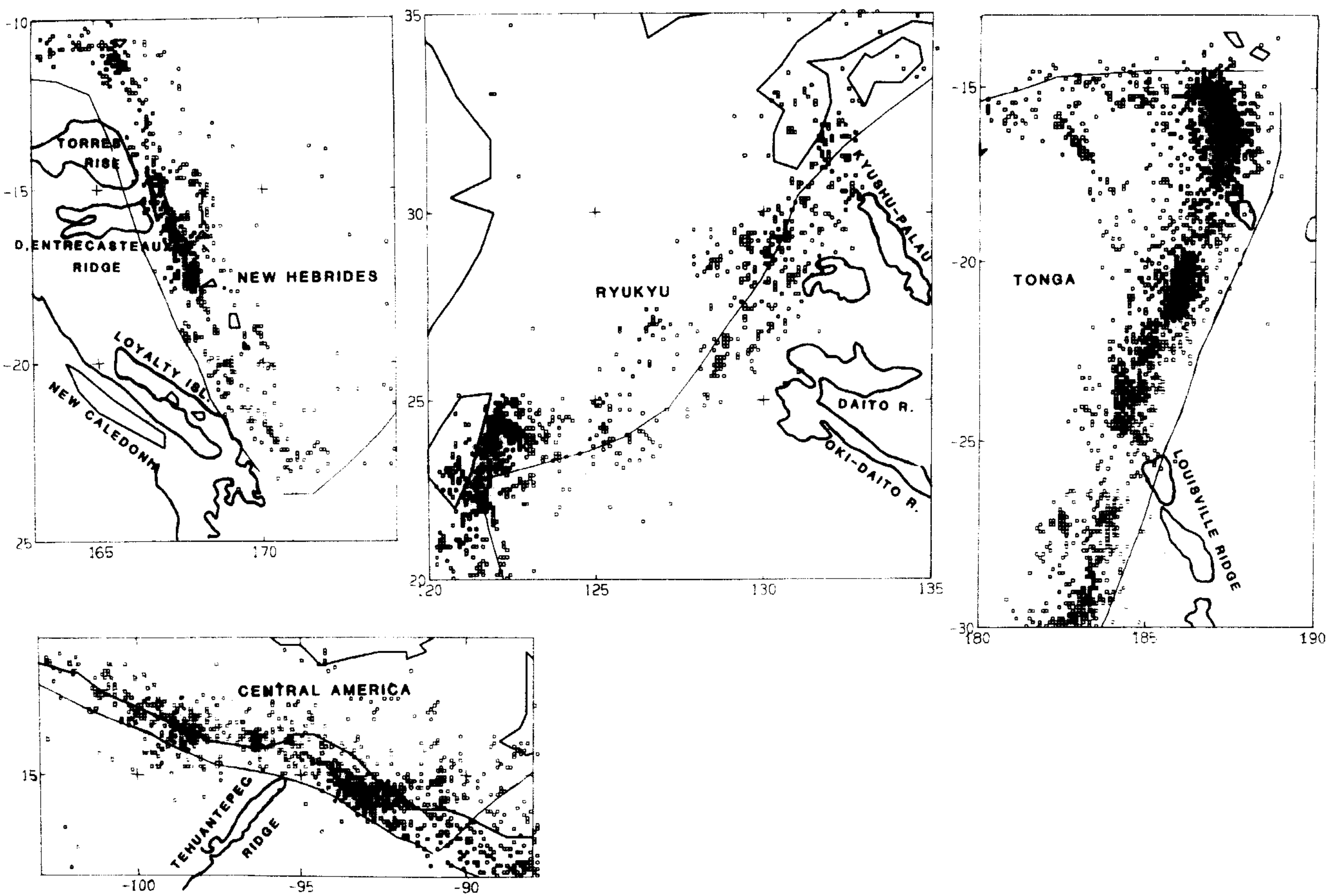

Fig. 7. Shallow seismicity is less frequent where a ridge subducts.

the relative buoyancy of the plateau may cause the dip of the slab to shallow locally by 5–15° (Table 1; Chung, 1978) but the resultant dip of the slab is still steep enough to allow the presence of an asthenospheric wedge between the plateau/slab and the overlying arc lithosphere. In these cases the subduction of the plateau may have a more fundamental effect on the process of arc volcanism, either by changing the state of stress in the arc (factor 3) or by perturbing the sources of magma, volatiles, or heat at depth (factors 1 or 2).

We examine first the possibility that the plateau changes the stress state of the arc. Intuitively, one might expect that although the plateau is not buoyant enough to stop subduction, it still has topographic relief and may therefore still act as a compressive force on the arc by increased coupling at the interface between the two plates. If this were the case, the shallow seismicity at the interface and within the arc should reflect such a lateral increase in stress. However, there is no such consistent change in shallow seismicity (Chung, 1978). In fact, seismicity data seem to suggest the opposite. Kelleher and McCann (1976, 1977) found reduced levels of seismic moment release where plateaus were subducted, with fewer great earthquakes and shorter rupture zones. An analysis of the map pattern of shallow earthquakes from the ISC (Fig. 7) shows a general decrease in the number of shallow earthquakes in the overlying plate where an oceanic plateau is subducted and an increase only where an active seismic fracture zone such as the D'Entrecasteaux Fracture Zone subducts (Chung and Kanamori, 1978).

It is not clear physically how the oceanic plateau affects the shallow seismicity of either the arc or the plate interface. Kelleher and McCann (1976, 1977) suggest that

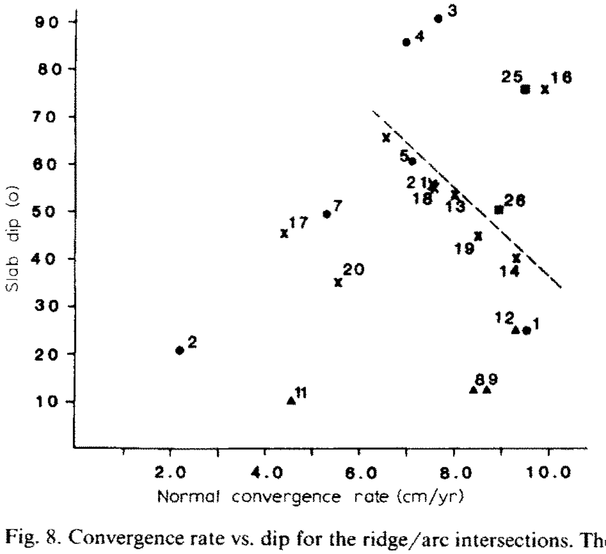

Fig. 8. Convergence rate vs. dip for the ridge/arc intersections. The black squares are the two areas where volcanism continues despite the subduction of a ridge.

the plateaus are so buoyant that they slow or stop subduction itself. A drop in the horizontal convergent motion would explain the drop in seismicity. However, we have shown at least nine areas where the plateaus subduct along with the slab. It is possible that the topographic relief of the plateau above the seafloor changes the nature of the contact surface between the two plates. An analysis of the rupture process of large earthquakes (Ruff and Kanamori, 1983) suggests that earthquake size is related to the asperity distribution along the plate interface, with large earthquakes associated with large asperities and small earthquakes associated with smaller scattered asperities. Ruff and Kanamori further correlate this seismic coupling with horizontal compressive stress between the plates, with stress proportional to the ratio of the summed asperity area to the total area of the interface. If the relief of the plateau reduces the actual contact area between the plates or modifies the contact by creating small scattered asperities, the coupling between the plates and in turn the horizontal compressive stress transmitted to the arc may be reduced. It is also possible that the different chemical composition and thickness of the aseismic plateau compared to normal oceanic crust may create small asperities by causing variations in the strength of the contact zone. Eissler and Kanamori (1982) attribute a large normal earthquake within the slab at the usually thrust-dominated Tonga plate interface to possible local decoupling caused by the subduction of the Louisville Ridge.

It is interesting to note that eight out of nine of these gaps, where an oceanic plateau subducts at a relatively steep dip, occur within convergent systems which correspond to the low-coupling, Mariana-type classification of Uyeda and Kanamori (1979). Each is at an island arc rather than a continental margin and has active or incipient back arc spreading. Perhaps these systems need to be only slightly perturbed to cause the arc stress state to inhibit formation of arc volcanoes. If the plateau locally decreases the coupling, this may lessen the amount of compressive stress transmitted to the arc. Gill (1981) states that one of the requirements necessary for the formation of andesitic arc volcanoes is either an arc crustal thickness of 25 km or more or a compressive stress in the arc. Most of the nine arcs are oceanic arcs. The plateau may then prevent the formation of volcanoes by not promoting the differentiation of the existent basaltic magmas into the more silica-rich magmas which characterize arc volcanism. The effect of subducting plateaus therefore may be a function of decreased coupling at the plate interface.

However, several of the subducting plateaus and their associated anomalous shallow seismicity are located at the edge of a volcanic gap rather than in the middle (for example, see Fig. 7, Tonga arc/Louisville Ridge), suggesting that the plateau may exert a deeper influence on the process. If we rule out surficial effects, we must assume that the bathymetric feature continues at depth and that the presence of this feature at depth affects the fundamental process of arc volcanism. Either the plateau changes the course of heat or the source of volcanic material.

It's possible that if a plateau maintains its 2–5 km relief with respect to the seafloor that it may modify the source of heat either by reducing the amount of frictional heat produced along the slab–mantle interface or by changing the nature of the asthenospheric counterflow. However, the details of these two processes are only inferred from models which are themselves poorly constrained and therefore such modifications would be difficult if not impossible to detect. It also seems likely that even if such modifications existed that there would still be sufficient heat to fuel the endothermic dehydration reactions within the slab and the partial melting within the wedge.

Therefore, the cause of volcanic gaps where plateaus are subducted may lie in a perturbation in the supply of volcanic materials. This perturbation is probably tied to the different geochemical compositions and crustal thicknesses of oceanic plateaus, compared to normal oceanic crust. Oceanic crust formed at a mid-ocean spreading center is quickly hydrothermally altered. Estimates of the depth and extent of this alteration vary a great deal, but it is thought to provide enough water through dehydration at depth to enhance partial melting in the asthenosphere and resultant arc volcanism (Gill, 1981). In contrast, an oceanic plateau may have formed originally as an island arc, a hot spot trace or a leaky transform. In that case, the underlying oceanic crust, if there is any, will be modified by the formation of the oceanic plateau, with the plateau itself having a different geochemistry and structure. In addition, subduction of sediment, especially turbiditic trench sediment, is likely to be limited where a wide topographic feature is subducted. The role of subducted sediments in arc volcanism is not yet clear. Although studies of various elements suggest that sediments may subduct and contribute distinctive chemical components to the magma, it is not known if sediments are essential to the process of arc volcanism.

Several possibilities exist for the effect of the plateau on the physico-chemical processes at depth:

(1) The top of the plateau may be sheared off or broken apart during subduction (Fig. 6b). If the top 2 or 3 km contain much of the water and/or hydrous minerals, this erosion may constitute a significant decrease in the quantity of water available deeper for melting. This erosion may also remove any sediments which may exist upon the surface of the plateau.

(2) The plateau remains intact through the subduction process, but doesn't initially contain the volume of hydrous minerals necessary to start the process of arc volcanism (Fig. 6c). During formation of the plateau, the high temperatures may bake the hydrous minerals out of the original oceanic crust. Although the surface of the plateau may be vesiculated and full of hydrous minerals, the crust of the plateau as a whole may contain less water than crust formed at a spreading ridge.

(3) A sufficient quantity of hydrous minerals exists in the plateau, but water may be released by dehydration reactions at a depth too shallow for melting within the mantle wedge ($<$ 70 km; Fig. 6d). Delong and Fox (1977) suggest that for spreading ridges the lack of a cap of sediments may allow water to escape too soon. A similar

TABLE 2

Amount of oceanic crust subducted at each volcanic gap

Gap location	Relative convergence poles				Coord. of gap at trench				Oceanic crust subducted	
	plate pair	lat.	long.	w(°/m.y.)	°N	°E to	°N	°E	km^2/yr	% of total
A. Collision										
1. Costa Rica	COCO–CARB	23.60	−115.55	1.543	9.0	−85.0	8.0	−83.0	0.0234	
2. Colombia	NAZC–SOAM	59.08	−94.75	0.835	8.0	−77.0	5.5	−78.0	0.0153	
3. Mindoro	EURA–PHIL	45.50	150.20	1.200	10.0	122.0	13.0	120.0	0.0157	
4. Taiwan	EURA–PHIL	45.50	150.20	1.200	20.0	120.0	23.0	122.0	0.0290	
5. Java	INDI–EURA	19.71	38.46	0.698	−11.0	124.0	−9.0	129.0	0.0341	
6. Yap	PCFC–PHIL	2.00	138.30	1.160	7.0	136.0	11.0	139.0	0.0064	
7. S. Marianas	PCFC–PHIL	2.00	138.30	1.160	11.0	139.0	13.0	146.0	0.0121	
									0.1360	5.5
B. Shallow dip										
8. Peru	NAZC–SOAM	59.08	−94.75	0.835	−2.0	−81.5	−17.0	−75.0	0.1528	
9. Chile	NAZC–SOAM	59.08	−94.75	0.835	−27.0	−72.0	−33.0	−70.0	0.0642	
10. S. Chile	ANTA–SOAM	87.69	75.20	0.302	−46.0	−76.0	−57.0	−70.0	0.0241	
11. Nankai	EURA–PHIL	45.50	150.20	1.200	30.0	132.5	31.0	133.0	0.0054	
12. New Guinea	INDI–PCFC	60.71	−5.79	1.246	0.0	135.0	−3.0	142.0	0.0706	
									0.3171	12.6

C. Enigmas										
13. Tonga	INDI–PCFC	60.71	−5.79	1.246	−25.0	185.0	−22.0	187.0	0.0310	
14. N. Tonga	INDI–PCFC	60.71	−5.79	1.246	−18.0	188.0	−15.0	188.0	0.0324	
15. S. New Hebr.	INDI–PCFC	60.71	−5.79	1.246	−22.5	171.0	−20.0	168.0	0.0262	
16. N. New Hebr.	INDI–PCFC	60.71	−5.79	1.246	−13.0	168.0	−11.0	167.0	0.0240	
17. Solomons	INDI–PCFC	60.71	−5.79	1.246	−11.0	163.0	−10.0	159.0	0.0186	
18. N. Marianas	PCFC–PHIL	2.00	138.30	1.160	20.0	147.5	22.0	146.0	0.0063	
19. Izu–Bonin	PCFC–PHIL	2.00	138.30	1.160	24.0	143.0	27.0	144.0	0.0182	
20. Ryukyus	EURA–PHIL	45.50	150.20	1.200	14.0	127.0	28.0	132.0	0.0375	
21. Middle America	COCO–CARB	23.60	−115.55	1.543	15.0	−95.0	14.0	−93.0	0.0158	
									0.2100	8.4
D. No ridge										
22. N. Macquerie	INDI–PCFC	60.71	−5.79	1.246	−60.0	155.0	−46.0	164.0	0.0216	
23. N. Kermadec	INDI–PCFC	60.71	−5.79	1.246	−28.0	185.0	−26.0	185.0	0.0172	
24. S. Kermadec	INDI–PCFC	60.71	−5.79	1.246	−36.0	182.0	−31.0	183.0	0.0363	
									0.0751	3.1
										29.6

situation may exist for a subducted oceanic plateau with significant topographic relief and intermediate geochemistry.

The problem is therefore to ascertain to what extent and depth the oceanic plateau may be hydrothermally altered and how this anomalous feature reacts chemically and rheologically upon subduction. At present, available information is too scanty to constrain speculation. The three listed possibilities all assume that the addition of volatiles from the subducted plate into the overlying mantle is required for partial melting in the wedge. If there is already enough water in the mantle to cause partial melting upon convection into higher temperatures, then the subduction of a ridge cannot significantly affect the deeper chemical processes.

Case D: Volcanic gaps with no subducting plateaus

Three volcanic gaps are included in Table 1 which cannot be attributed to the collision or subduction of a plateau. In the case of the northern Macquerie Ridge, one volcano is known just south of New Zealand but the rest of the plate margin is devoid of volcanoes. This could be a result of the low rate and oblique angle of convergence (Table 1). Even at the location of the solitary volcano, the perpendicular rate is only 2.1 cm/yr. The two gaps in the Kermadec chain are strikingly apparent and harder to explain. Kelleher and McCann (1976, 1977) show a strong spatial correlation in the Kermadec island arc between the occurrence of great earthquakes and the location of volcanoes. They suggest that the zone without either volcanoes or great earthquakes is caused by subduction of non-smooth oceanic crust. So, although a large bathymetric feature is absent, it may still be the rough topography of the seafloor which affects volcanism.

DISCUSSION AND CONCLUSIONS

It becomes clear from the previous data compilation and discussion that the subduction or collision of oceanic plateaus can have major tectonic impact on the process of arc volcanism on the overlying plate of a subduction system. If over 30% of the oceanic crust which is subducted in a year is subducted at volcanic gaps, then the process which creates the gaps is significantly modifying the "normal" orogenic processes at convergent margins. Table 2 shows the amount of oceanic crust subducted at each volcanic gap.

Volcanic gaps are relevant to the analysis of geochemical cycles of the elements which are concentrated in the earth's crust. It is often assumed that subduction implies a proportionate amount of arc volcanism and that whatever volatile and incompatible elements, such as water and potassium, that get subducted are automatically recycled to the surface in the arc volcanoes. Volcanic gaps, however, act as additional sinks for elements subducted with the slab and need to be included in any mass balance calculation. In particular, automatic recycling by island arc volcanoes cannot be presumed in a mass balance for the composition of the subducted oceanic crust. For example, alteration of the bulk of the oceanic crust to 3% hydration by weight implies a cycle time of less than 1 billion years for water. The presence of volcanic gaps further implies a massive loss of water throughout geological time and a decrease in ratio of water to chlorine, which proportionally is much less removed by subduction than water. As these variations are not evident, it is possible that only a kilometer or so of the oceanic crust is extensively altered and that the present cycle time for water is a few billion years. (See Ito et al., 1983 for more on the water cycle.)

It also becomes clear that there is no single cause of volcanic gaps. For the volcanic gaps where subduction has ceased or reoriented (case A) or where the dip of the subducting slab is extremely shallow (case B), it is easy to understand how the process of arc volcanism may be perturbed to the extent of creating volcanic gaps. Arc volcanism may be shut off without either a moving slab or an asthenospheric wedge to provide magmatic source material or the heat for dehydration and melting.

The volcanic gaps where subduction is continuous and the dip is steep enough (case C) are harder to explain. Clearly one or more of the three factors necessary for arc volcanism are being affected by the subduction of the plateaus. The pattern of shallow seismicity at the plate interface and within the arc suggests that the subducting plateau may alter the mechanism of coupling between the plates by reducing the contact area. This may in turn reduce the amount of compressive stress in the arc. Since a minimum amount of horizontal compressive stress is required for an island arc to produce andesitic arc volcanoes (Gill, 1981) and since most of the gaps of case 3 are within arcs that are already of the low-stress, low-coupling classification (Uyeda and Kanamori, 1979), it may only require a minimal reduction of compressive stress to produce a significant change in the nature of volcanism within the arc. The less differentiated basaltic magmas might form extensive, low-lying flows rather than volcano edifices or they might never even reach the surface.

Alternatively, the subducting plateau may be affecting the source of magma at depth. If the hydrous top of the plateau is sheared off during subduction or if the plateau crust contains lesser volumes of hydrous minerals, the small amount of melt within the wedge may inhibit the formation of arc volcanoes or may abruptly increase the spacing between volcanoes. Either of the proposed mechanisms seem to be possible but speculative given current knowledge.

One of the surprising results of this study is that buoyancy does not seem to be the most important characteristic of an oceanic plateau in its interaction at a subduction zone. The previously noted effects of plateau collision or subduction have been uniformly attributed to their greater crustal thickness and relative buoyancy compared to normal oceanic crust. Yet two-thirds of the plateaus do not collide and accrete; rather, they subduct. Of those that do subduct more than two-thirds do so without significantly changing the geometry of the subduction system and slab. In these cases, it is other characteristics of these features, namely their topographic relief and their different compositions compared to oceanic crust, that require a modification of the basic models of oceanic subduction.

If the collision or subduction of an oceanic plateau can have such pronounced effects on the process of arc volcanism, then the two areas where large bathymetric features are being subducted without volcanic gaps (Case E) are anomalous. Figure 8 shows a plot of convergence rate versus dip for each of the volcanic gaps listed in Table 1 and also the Kamchatka/Meiji Seamount and New Hebrides- D'Entrecasteaux intersection areas. The two areas where volcanism continues uninterrupted are areas with both high convergence rates and steep dips. We have suggested in this paper that the presence of a plateau on the subducting plate in some way reduces the amount of water or magma available from the slab. Of course, higher dips and greater convergence rates increases the amount of plateau/slab to pass through the pressure–temperature window where volcanism starts. Kamchatka and the D'Entrecasteaux areas may therefore exceed the critical amount necessary for arc volcanism.

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