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# How do deep-water volcanoes grow?

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## **Abstract**

Deep-water volcanoes are emplaced in water depths >1.0 km and are widespread along continental margins and in ocean basins. Whilst the external morphology of deep-water volcanoes can be mapped using bathymetric surveys, their internal structure and true volume remain enigmatic. It is thus difficult to determine how deep-water volcanoes grow. We investigate 13 Late Miocene-to-Quaternary, deep-water volcanoes that are imaged in 3D by seismic reflection data from the northern South China Sea, which allow us to quantify their external morphology and examine their internal structure. These deep-water volcanoes were emplaced in water depths >1.5 km, are relatively small (<3.0 km diameter, <0.56 km tall, and <0.92 km<sup>3</sup> in volume), and have steep slopes (up to 42°). Most of the volcanoes have erosional, 'crater-like' bases, infilled with sub-horizontal seismic reflections. These crater-like bases are overlain by downward-converging, conical seismic reflections delineating the classical volcano morphology. We suggest the crater-like bases formed by excavation of cold, wet, and poorly consolidated near-seabed

31 sediment during expulsion of hydrothermal fluid, and not by explosive magmatic eruptions or  
32 gravitational subsidence. Erupted igneous material infilled the precursor craters with the observed  
33 sub-horizontal layers, likely comprising hyaloclastites. After this initial phase of volcanism, the  
34 buildup of volcanic material produced layers that are now represented by the flank-parallel or  
35 downward-converging, conical seismic reflections. We suggest high hydrostatic pressures of >15  
36 MPa, which are typical of water depths >1.5 km, inhibited degassing and fragmentation of  
37 ascending magma and thus erupted lava. This lack of degassing and fragmentation permitted  
38 effusive eruptions during the latter stages of volcanism. Our models for volcano growth in the  
39 deep submarine realm demonstrate the power of using 3D seismic data when investigating the  
40 internal structure and total volume of deep-water volcanoes.

41

42 **Keywords:** deep-water volcanoes, volcanism, extrusion dynamics, growth mechanism, erosion,  
43 South China Sea

44

## 45 **1. Introduction**

46 Volcanoes occur in a variety of plate boundary and intra-plate settings across Earth's surface.  
47 Determining how volcanoes grow is not only critical to predicting and mitigating volcanic  
48 hazards, but this understanding can also provide information on the underlying plumbing system  
49 structure (e.g., feeder and reservoir locations) and magma dynamics (e.g., composition and supply  
50 rate) (e.g., [Moore and Clague, 1992](#); [Arnulf et al., 2014](#); [Clague et al., 2018](#)). Volcanoes emplaced  
51 on land are typically well-studied and, by comparing their external morphology to similar  
52 neighboring edifices, we can infer they are broadly built through fluctuations between so-called  
53 summit- and diameter-prone growth (e.g., [Moore and Clague, 1992](#); [Rossi, 1996](#); [Grosse et al.,](#)  
54 [2009](#); [Karlstrom et al., 2018](#)). Yet without direct access to volcano interiors, it is difficult to test  
55 growth models predicted from their external morphology alone. Examining ancient, (partially)  
56 eroded volcanoes provides some insight into how volcanoes are constructed, but modification of  
57 their original shape means we cannot assess relationship between internal structure and external  
58 morphology (e.g., [Goto and Tomiya, 2019](#)). By using traditional remote-sensing and/or field-  
59 based techniques, we can therefore either quantify the external morphology of uneroded

60 volcanoes, but not know their internal structure, or study the interiors of eroded volcanoes where  
61 information on their original edifice shape has been lost. Our ability to only constrain either the  
62 external morphology *or* internal structure of onshore volcanoes, but not both, limits our  
63 understanding of how volcanoes grow.

64 Remote sensing data and lithostratigraphic analysis of well cores allow us to constrain the  
65 external geometry and internal structure of the evolution of shallow- and deep-water volcanoes  
66 (e.g., [Smith, 1988](#); [Magee et al., 2013](#); [Cocchi et al., 2016](#); [Buchs et al. 2018](#)). In particular,  
67 seismic reflection imaging of volcanoes provides a unique opportunity to resolve uncertainties  
68 regarding volcano growth, given these data can image both the external morphology *and* internal  
69 structure of volcanoes (e.g., [Gatliff et al., 1984](#); [Calves et al., 2011](#); [Magee et al., 2013](#); [Reynolds  
70 et al., 2018](#); [Sun et al., 2019](#)). For example, by using 2D seismic reflection data offshore southern  
71 Australia, [Magee et al. \(2013\)](#) showed trends in the external morphology of buried, shallow-water  
72 shield volcanoes were consistent with growth via summit eruptions and a proportionate increase  
73 in summit height and volcano diameters. Interpretation of reflections within the volcanoes reveal  
74 the majority of volcanoes did indeed grow by a proportionate increase in summit height and basal  
75 diameter (i.e. the layers were parallel to the volcano flanks) ([Magee et al., 2013](#); see also [Reynolds  
76 et al., 2018](#)). A similar seismic-based study of shallow-water volcanoes (water depth <200 m),  
77 emplaced along the western Indian rifted margin, reveal they preferentially grew via increases in  
78 diameter without a commensurate increase in summit height ([Calves et al., 2011](#)). Whilst seismic  
79 reflection data have been used to unravel the growth of shallow-water volcanoes, few studies  
80 have utilized these data to study the internal structure of deep-water (>1 km) volcanoes (e.g.,  
81 [Gatliff et al., 1984](#); [Sun et al., 2019](#)).

82 Discerning how deep-water volcanoes erupt and grow is critical for: (1) accurate assessment  
83 of deep-water volcanic hazards (e.g. submarine landslides and associated tsunami; e.g. [Staudigel  
84 and Clague, 2010](#)); (2) calculation of accurate eruptive and total volume estimates, which  
85 contribute to understanding melting conditions in the underlying crust and/or mantle (e.g. [Buchs  
86 et al., 2018](#); [Sun et al. 2019](#)); and (3) determining the role of volcanoes in gas venting and  
87 hydrothermal circulation (e.g. [Planke et al., 2005](#)). Importantly, high hydrostatic pressures in  
88 deep-water settings, which can inhibit degassing, ascent rate, and fragmentation of magma, mean

89 the extrusion dynamics of deep-water volcanoes may fundamentally differ from their onshore and  
90 shallow-water counterparts (e.g. [Gregg and Fornari, 1998](#); [Cas and Simmons, 2018](#); [Carey et al.,](#)  
91 [2018](#); [Manga et al., 2018](#); [Sun et al., 2019](#)). These differences in eruption style and underlying  
92 controls suggest we may not be able to simply apply our knowledge of volcanism in other,  
93 subaerial or shallow water settings, to understand how deep-water volcanoes grow (e.g. [Gregg](#)  
94 [and Fornari, 1998](#); [Manga et al., 2018](#)). It is therefore necessary to image the internal structure of  
95 deep-water volcanoes to reveal their growth history.

96 Here, we use 3D seismic reflection data to investigate 13 deep-water volcanoes located along  
97 the continental margin of the northern South China Sea. These Late Miocene-Present volcanoes  
98 were emplaced close to the Continent-Ocean Boundary (COB) in water depths >1.5 km. Our 3D  
99 seismic reflection data allow us to map the external morphologies and internal structures of these  
100 volcanoes in unprecedented detail. From our seismic reflection imaging, we propose the majority  
101 of studied deep-water volcanoes grew through an initial phase of crater formation driven by  
102 escape of hydrothermal fluids, which became infilled. Volcanic cones developed on top of these  
103 infilled craters, or in two cases directly on undisturbed seabed sediment, primarily grew by  
104 proportionate increases in summit height and basal diameter, thereby maintaining their slope  
105 angle; some volcanoes appear to have grown by preferential addition of material to summit  
106 regions. Although similar growth models have been proposed for volcanic cones in subaerial and  
107 shallow marine settings, we demonstrate the deep-water volcanoes we study are relatively smaller  
108 and have steeper slopes. We attribute the initial phase of crater formation and morphological  
109 differences between deep-water volcanoes and those in other settings, to the unique physical  
110 conditions under which deep-water volcanoes evolve. Our work shows seismic reflection data is  
111 a powerful tool for unravelling volcano growth.

112

## 113 **2. Geological setting**

114 The South China Sea is located in a complex tectonic region between the Eurasian, Pacific and  
115 India-Australia plates (e.g. [Briais et al., 1993](#); [Franke et al., 2014](#); [Li et al., 2014](#)) ([Fig. 1a](#)). The  
116 South China Sea evolved as a magma-poor rift, culminating in seafloor spreading, the onset of  
117 which varied across the region (e.g. [Clift et al., 2001](#); [Cullen et al., 2010](#); [Larsen et al., 2018](#)).

118 Seafloor spreading began in the East Sub-basin in the early Oligocene (~33 Ma; [Briais et al., 1993](#);  
119 [Li et al., 2014](#)), before the spreading center jumped to the Southwest Sub-basin in the late  
120 Oligocene (~25 Ma) (e.g. [Franke et al., 2014](#)). Spreading ceased sometime in the middle Miocene  
121 (~15.0-15.5 Ma; [Briais et al., 1993](#); [Li et al., 2014](#)). Since the late Miocene (~10.5 Ma), tectonic  
122 activity in the northeastern part of South China Sea has been mainly driven by its collision with  
123 the Philippine Sea Plate (i.e. the Dongsha Event; e.g. [Lüdmann and Wong, 1999](#)).

124 The study area is located to the south of the Dongsha Islands in the northern South China Sea  
125 ([Fig. 1a](#)). Geological and geophysical studies (e.g. borehole, gravity, magnetic, and 2D and 3D  
126 seismic reflection data) indicate widespread Cenozoic volcanism across the northern South China  
127 Sea (e.g. [Li and Liang, 1994](#); [Yan et al., 2006](#); [Zhao et al., 2016](#)). From the early Paleocene to  
128 earliest Oligocene, before the onset of seafloor spreading, intermediate-acidic volcanoes were  
129 emplaced in a subaerial setting ([Yan et al., 2006](#)). From the Oligocene to middle Miocene, there  
130 was a compositional and environmental transition to the emplacement of mafic-to-intermediate  
131 volcanoes in shallow-water (<200 m) and subaerial settings (e.g. [Yan et al., 2006](#); [Lester et al.,](#)  
132 [2014](#)) ([Fig. 1a](#)). Rapid post-emplacement subsidence led to these volcanoes being deeply buried  
133 (up to depths of 1.5 km) beneath the current seafloor (e.g. [Zhao et al., 2016](#)). Late Miocene and  
134 younger, intra-plate volcanoes ([Figs. 1b-c](#)) were emplaced close to the continent-ocean boundary  
135 ([Clift et al., 2001](#); [Sun et al., 2019](#)). Recent IODP Expeditions 349/367/368 drilled several of  
136 these deep-water volcanoes in the South China Sea, revealing they are primarily basaltic (e.g. [Li](#)  
137 [et al., 2014](#); [Larsen et al., 2018](#)), and that some were emplaced during continental breakup and  
138 directly covered by deep-water (>1.3 km) nanofossil-bearing clay sediments ([Larsen et al., 2018](#)).  
139 Many of the deep-water volcanoes, emplaced since the Late Miocene, feed long run-out lava  
140 flows that have irregular basal morphologies ([Sun et al., 2019](#)) ([Figs. 1b-c](#)). The volumes of these  
141 long run-out lavas appear equivalent to, or substantially greater than, that of the erupted material  
142 contained in the volcanoes themselves ([Sun et al., 2019](#)).

143

### 144 **3. Datasets and methods**

145 We use a time-migrated 3D seismic reflection dataset covering ~1150 km<sup>2</sup> to study the external  
146 morphology and internal structure of deep-water volcanoes in the South China Sea ([Fig. 1a](#)). The

147 data were acquired in 2012 using eight tuned air source guns, each with a volume of  $8 \times 20 \text{ in}^3$ ,  
148 to produce a total shot volume of  $8 \times 160 \text{ in}^3$ . Six 3000 m-long, 240-channel streamers with a  
149 spacing of 12.5 m were used to tow the hydrophones. The data are zero-phase processed with  
150 ordinary processing procedures (e.g. digital filtering, deconvolution, dynamic/static correction,  
151 offset stack, etc.), and displayed with Society of Exploration Geophysicists (SEG) normal polarity.  
152 A downward increase in acoustic impedance therefore corresponds to a positive reflection event  
153 (red on displayed seismic profiles) and a downward decrease in acoustic impedance corresponds  
154 to a negative reflection (black on displayed seismic profiles) (e.g. [Brown, 2004](#)).

155 The dominant frequency in the interval of interest (i.e. 0–400 m below the seabed) is  $\sim 40 \text{ Hz}$ .  
156 The estimated limit of separability within the deep-water strata (i.e. nanofossil-bearing clay)  
157 encasing the volcanoes is  $\sim 14 \text{ m}$ , based on a seismic velocity of  $2.2 \text{ km/s}$  for the sedimentary  
158 rocks; this velocity is derived from nearby seismic refraction profiles ([Yan et al., 2001](#); [Wei et al.,](#)  
159 [2011](#)) ([Fig. 1a](#)). There are no seismic velocity data available for the studied deep-water volcanoes,  
160 but we assume they have an interval velocity of the  $4.0 \pm 1.0 \text{ km/s}$  based on: (i) measured seismic  
161 velocities of  $\sim 3.0$ – $5.0 \text{ km/s}$  for basaltic rocks (lava flows, volcanoclastic breccias and pyroclastics)  
162 intersected by nearby boreholes (e.g. BY7-1 and IODP 1431) ([Li et al., 2014](#); [Zhao et al., 2016](#));  
163 (ii) velocities of  $\sim 3.0$ – $5.5 \text{ km/s}$  obtained from seismic refraction profiles that cover other deep-  
164 water volcanoes within the basin ([Yan et al., 2001](#); [Wei et al., 2011](#)); (iii) typical seismic velocities  
165 calculated from boreholes penetrating basaltic submarine volcanoes ( $\sim 3.3$ – $5.5 \text{ km/s}$ ) elsewhere  
166 ([Calvès et al., 2011](#)); and (iv) the size of observed seismic velocity anomaly-induced ‘pull-ups’  
167 beneath the studied volcanoes (V7, V11, and V13, [Fig. S1](#)), caused by acoustic waves travelling  
168 faster through hard, crystalline igneous rocks than the surrounding sedimentary strata ([Jackson,](#)  
169 [2012](#); [Magee et al., 2013](#); [Reynolds et al., 2018](#)). With regard to the latter point, we calculate  
170 interval velocities of  $3.2$ – $4.1 \text{ km/s}$  for the three volcanoes (V7, V11, and V13), derived from the  
171 magnitude of velocity pull-up artifacts ( $\sim 82.7 \text{ ms}$  -  $161.3 \text{ ms}$  TWT high) present in underlying  
172 seismic reflections ([Fig. S1](#)):

$$173 \quad V_{pi} = \frac{T_s \times V_{ps}}{T_i}$$

174 where  $V_{ps}$  ( $V_{ps} = 2.2 \text{ km/s}$ ) and  $V_{pi}$  are the seismic velocities of encasing rocks/sediments and  
175 igneous rocks;  $T_s$  and  $T_i$  are the seismic wave travel time in the encasing rocks/sediments and

176 igneous rocks, respectively (Fig. 2a).

177 Given a dominant frequency of  $\sim 40$  Hz and an interval velocity of the  $4.0 \pm 1.0$  km/s, the  
178 estimated limits of separability and visibility of layers within the volcanoes is  $25 \pm 6$  m ( $\lambda/4$ ) and  
179  $3.5 \pm 0.5$  m ( $\lambda/30$ ), respectively (Sun et al., 2019). When the volcanic structures are thicker than  
180 the estimated limit of separability, their top and base reflections can be distinguished. However,  
181 if their thickness lies between the limits of separability and visibility, they will appear as tuned  
182 reflection packages; i.e. reflections from their top and base interfere on their return to the surface  
183 and cannot be distinguished (e.g. Brown, 2004). Volcanic structures thinner than the limit of  
184 visibility will likely not be distinguishable from noise within the seismic data (Eide et al., 2017).

185 The volcanoes we study comprise two distinct components, involving a volcanic edifice and  
186 an underlying infilled crater-like base (Fig. 2b). We mapped three key seismic horizons: TV (top  
187 of volcano), BV (base of volcano), and the seabed (Figs. 3a-c). From these mapped horizons, we  
188 measured key geomorphologic parameters of the volcanoes, including diameter and height/depth  
189 of the edifices and crater-like bases (Fig. 2b). We define volcano thickness, which we also use to  
190 calculate volume, as the difference in height between TV and BV (Fig. 2b); volume estimates also  
191 take into account the irregular morphologies of TV and BV. Because the observed volcano flanks  
192 are rugged, we calculated average flank dips as  $\text{height}/(\text{diameter}/2)$  (Fig. 2b). In places, where  
193 volcanoes appear to merge, we constrain the plan-view extent of each edifice by distinguishing  
194 the location of minimum thickness between them (Fig. 3d). Errors in height, depth, volume, and  
195 flank dip measurements largely arise from uncertainties in the seismic velocities ( $4.0 \pm 1.0$  km/s)  
196 used to undertake the depth conversion rather than measurement errors. The collected edifice and  
197 crater dimensions data allow us to better understand how much volcanic material may be  
198 underestimated by surficial remote-sensing techniques, and thus unaccounted for when  
199 calculating volumes of magma production. We also compare the geomorphologic characteristics  
200 of the volcanic edifices to volcanoes emplaced in different environments with varying  
201 composition, such as ocean basins (Basalt; Smith, 1988), subaerial volcanic arcs (Basalt - andesite;  
202 Grosse et al., 2009), submarine volcanic arcs (dacite, basalt-andesite; Wright et al., 2006) and  
203 shallow water (Basalt; Magee et al., 2013).

204

## 205 **4. Characteristics of the deep-water volcanoes**

### 206 **4.1. Seismic expression**

207 We mapped the top and bases, and thus constrained the thickness and volume of 13 volcanoes  
208 (Figs. 3b-c; Table 1). In our seismic data, several volcanoes appear to have merged to form a  
209 single, large edifice defined by multiple distinct summits (i.e. V4-V6 and V11-V12; Figs. 3b-c).  
210 All the volcanoes are at least partly buried by a thin layer (<300 m) of Late Miocene- Quaternary  
211 strata (Figs. 4-5; Table 1), with the tips of edifices (i.e. V6, V7, V9, V11, V12 and V13) breaching  
212 the seabed (Figs. 3a, 4a-b, 4d, 4f). Except for V1, all volcanoes are encircled by moats that are  
213 up to 75 m deep, and which, depending on their stratigraphic occurrence, are unfilled (i.e. moats  
214 expressed at seabed), partly infilled, or fully filled (i.e. buried moats) (e.g. V6-V9; Figs. 3a, 4a-b,  
215 4d-f, 5). The volcanoes are typically characterized by continuous-to-moderately continuous, high-  
216 amplitude top reflections (i.e. TV), and discontinuous, primarily low-amplitude base reflections  
217 (i.e. BV) (Figs. 4-5; Fig. S1). Occasionally BV is continuous and high-amplitude (e.g. V6; Fig.  
218 4f). Projected boundaries dividing the crater-like bases and edifices of individual volcanoes occur  
219 at different stratigraphic levels (Fig. 4). For example, the edifice-crater boundary for V12 is  
220 coincident with the modern seabed, whilst for V9 the edifice-crater boundary is located ~50–100  
221 ms TWT (~40-80 m) beneath the current seafloor (Figs 4a-b).

222 We identify two types of volcano bases: (i) crater-like bases that truncate underlying seismic  
223 reflections (Figs. 4a-d); and (ii) relatively flat bases that are conformable with underlying strata  
224 (e.g. V6 and V8; Figs. 4e-f). Based on these differences in basal geometry, we sub-divide the  
225 volcanoes into two groups: (i) GP1 (11 volcanoes), which have crater-like bases (Figs. 4a-d); and  
226 (ii) GP2 (2 volcanoes), which have strata-concordant bases (Figs. 4e-f). The bases of both groups  
227 of volcanoes are located at various stratigraphic horizons, up to ~264 m beneath the seabed.  
228 Seismic reflections directly beneath the volcanoes, as well as those below lavas emanating from  
229 the volcanic edifices, have very low-amplitude compared to their typical seismic character away  
230 from the overlying volcanoes (Fig. 5). These reflections beneath the volcanoes are also typically  
231 disturbed and occasionally appear to be deflected upward relative to their regional dip (e.g. those  
232 beneath V9 in Fig. 5).

233

## 234 4.2. External volcano morphology and dimensions

### 235 4.2.1. Volcano edifices

236 The volcanic edifices have circular to elliptical basal sections, with diameters of  $\sim 0.6\text{--}3.0$  km  
237 (average of  $\sim 1.3$  km), covering areas of  $\sim 0.25\text{--}7.15$  km<sup>2</sup> (Table 1). Edifice height ranges from  
238  $\sim 79\pm 20$  to  $560\pm 140$  m (Figs. 3b, 6a; Table 1). There is a very weak (i.e.  $R^2 = \sim 0.21$ ), positive  
239 correlation between edifice diameter and height, with an average height:diameter ratio of 0.25  
240 (Fig. 6a). Flanks are linear, convex-upward, or convex-downward, and are moderate-to-steep,  
241 with dips of up to  $42^\circ$  (average dip of  $\sim 26^\circ$ ) (Figs. 4-5; Figs. S1-S2); most (nine) of the volcanoes  
242 have slopes  $>20^\circ$  (Table 1). Flank dip is weakly, negatively correlated with edifice diameter ( $R^2$   
243  $= \sim 0.12$ ; Fig. 6b) and weakly, positively correlated with height ( $R^2 = \sim 0.39$ ; Fig. 6c). Overall,  
244 edifice volumes range from  $\sim 0.0160\pm 0.0040$  to  $0.9213\pm 0.2303$  km<sup>3</sup> and show a strong, positive  
245 correlation to diameter ( $R^2 = \sim 0.94$ ; Fig. 6d), but a weak correlation to height ( $R^2 = \sim 0.25$ ; Fig.  
246 6e) and no correlation with flank dip and volume ( $R^2 = \sim 0.06$ ) (Fig. 6f).

247

### 248 4.2.2. Crater-like bases

249 The depth and diameter ranges of the crater-like bases are  $\sim 87\pm 22$  to  $517\pm 129$  m and  $\sim 0.8$  to  
250  $4.6$  km, respectively, and only weakly, positively correlated ( $R^2 = \sim 0.20$ ) (Fig. 7a; Table 1); their  
251 volumes range from  $0.0082\pm 0.0021$  to  $0.8144\pm 0.2036$  km<sup>3</sup> (Table 1). The dips ( $5^\circ\text{--}32^\circ$ ) of the  
252 basal crater flanks are only weakly, negatively correlated to crater diameter ( $R^2 = \sim 0.29$ ; Fig. 7b)  
253 and very weakly, negatively correlated with depth ( $R^2 = \sim 0.17$ ; Fig. 7c). Crater volume is  
254 moderately-to-strongly, positively correlated ( $R^2 = \sim 0.65$ ) with crater diameter but only weakly  
255 correlated to crater depth ( $R^2 = \sim 0.22$ ) and very weakly, negatively correlated with crater flank  
256 dip ( $R^2 = \sim 0.16$ ) (Figs. 7d-f).

257

### 258 4.2.3. Total volcano morphometrics

259 The heights of volcano edifices and depths of crater-like bases are weakly, positively correlated  
260 ( $R^2 = \sim 0.36$ ; Fig. 7g), whilst their diameters are moderately-to-strongly, positively correlated ( $R^2$   
261  $= \sim 0.65$ ; Fig. 7h). We note the diameters of the crater-like bases are typically greater than (e.g.  
262 V5 and V11) or equal to (e.g. V10 and V13) those of their overlying edifices (Fig. 7h). These

263 differences in diameter mean that the *volumes* of crater-like bases are typically larger than those  
264 of the overlying edifices (Fig. 7i), e.g. by more than five times (e.g. V5; Table 1); the volumes of  
265 the crater-like bases and edifices are weakly, positively correlated ( $R^2 = \sim 0.23$ ; Fig. 7i).

266 The average diameters, combining that of the edifices and crater-like bases, of individual  
267 volcanoes show a weak ( $R^2 = \sim 0.22$ ), positive correlation to volcano thickness (Fig. 7j). The total  
268 volumes of volcanoes range from  $0.0277 \pm 0.0070$  to  $1.2669 \pm 0.3167$  km<sup>3</sup>. Volcano thickness is  
269 only weakly ( $R^2 = \sim 0.28$ ) positively correlated with total volcano volume (Fig. 7k). However,  
270 there is a strong ( $R^2 = \sim 0.88$ ), positive correlation between total volcano volume and average  
271 diameter (Fig. 7j).

272

### 273 **4.3. Internal architecture and seismic facies**

274 We define two principal intra-volcano seismic facies (Fig. 4). Seismic facies 1 (SF1) is bound  
275 at its base by BV and predominantly comprises discontinuous, short, parallel to sub-parallel,  
276 moderate- to high-amplitude seismic reflections within the crater-like bases of GP1 (Figs. 4a-d).  
277 Reflections within SF1 appear broadly parallel with those of the surrounding sedimentary layers.  
278 Some outwardly dipping seismic reflections, which define broadly conical structures, are locally  
279 observed within the cores of SF1 (Figs. 4a-d). Overlying SF1, seismic facies 2 (SF2) constitutes  
280 the upper parts of all GP1 and GP2 volcanoes, broadly comprising stacked, continuous-to-  
281 moderately continuous, moderate-amplitude reflections that downlap onto SF1 or BV (Fig. 4). In  
282 most instances the internal SF2 reflections, where clearly observed, broadly parallel the outer  
283 margins of the volcanic edifices (e.g. Figs 4a, e-f); in some edifices the SF2 reflections converge  
284 down-dip (e.g. Fig. 4b).

285

## 286 **5. Discussion**

287

### 288 **5.1 Age and environment of volcanism**

289 Biostratigraphic data from nearby boreholes constrain the age of the nanofossil-bearing  
290 sedimentary sequences encasing the 13 mapped volcanoes, which have edifice bases that mark

291 the syn-eruptive seabed located at different stratigraphic levels (Figs. 4-5), and indicate volcanism  
292 occurred periodically between the Late Miocene (e.g. ~6.3 Ma of V1; Sun et al., 2019) and  
293 Quaternary. Analysis of ODP ( Site 1146) and IODP (Site U1501) data reveal that, at least since  
294 the Early Miocene (~23 Ma) and throughout this prolonged period of intermittent volcanism, the  
295 study area was a deep-water (>1.0 km) environment (e.g. Clift et al., 2001; Li et al., 2014; Larsen  
296 et al., 2018). Because the mapped volcanoes are within an area characterized by a present water  
297 depth of >1.3 km, and the subsidence-corrected, syn-emplacement Miocene-Quaternary sea level  
298 was ~200 m higher than it is today (Xu et al., 1995), we consider it likely that eruptions occurred  
299 in water depths >1.5 km; these water depths correspond to overlying hydrostatic pressures of >15  
300 MPa.

301

## 302 **5.2. Volcano formation and growth**

303 Most of the thirteen mapped volcanoes (11 of 13) can be sub-divided into an edifice and a  
304 crater-like base (Figs. 4a-d, 5). These crater-like bases truncate the underlying stratigraphic  
305 reflections and are infilled by sub-horizontal reflections, onto which a conical edifice is developed  
306 (Figs. 4a-d, 5). Here we discuss how each of these features relates to the initiation and growth of  
307 these deep-water volcanoes.

308

### 309 **5.2.1. Formation of crater-like bases**

310 Crater-like bases have been observed beneath volcanoes and hydrothermal vents in subaerial  
311 and shallow-water settings, and their formation has primarily been attributed to disaggregation  
312 and material removal during explosive eruptions (e.g. Planke et al., 2005; Wright et al., 2006;  
313 Geyer and Martí, 2008). Alternatively, crater-like bases could form by the collapse of subsurface  
314 conduits following magma extraction and subsidence of overlying material (e.g. Walker, 1993;  
315 Geyer and Martí, 2008) and/or post-emplacement gravitational subsidence in response to volcano  
316 loading (e.g. Moore and Clague, 1992; de Silva and Lindsay, 2015); depressions generated by  
317 these subsidence processes are expected to host inward-dipping layers (e.g. de Silva and Lindsay,  
318 2015). We consider it unlikely that the crater-like bases documented here (e.g., Figs 4a-d) formed  
319 by subsidence because: (i) gravitational subsidence driven by volcano loading could not produce

320 craters volumetrically larger than, and which occasionally extend beyond the footprint of, the  
321 overlying edifices (Moore and Clague, 1992; de Silva and Lindsay, 2015) (Fig. 7i; Table 1); (ii)  
322 volcano loading will cause underlying reflections to sag and will *not* produce craters that truncate  
323 and erode underlying strata (Figs. 4a-d); and (iii) the sub-horizontal reflections observed within  
324 the crater-like bases are inconsistent with collapse of pre-existing strata into evacuated magma  
325 conduits or deformation imposed by the volcano load (Figs. 4a-d).

326 By ruling out subsidence as a mechanism for driving crater formation, our observed truncation  
327 and erosion of underlying strata by the craters may suggest they formed via an initial phase of  
328 explosive activity (e.g. Planke et al., 2005; Wright et al., 2006; Geyer and Martí, 2008). However,  
329 whilst evidence for explosive volcanism (e.g. pyroclast occurrence) has been documented in  
330 silicic, volatile-rich deep-water settings, high hydrostatic pressures caused by large water  
331 columns (e.g. >1.0 km) are expected to prevent substantial exsolution of volatiles from magma  
332 and thereby inhibit explosive eruptions (e.g. Walker, 1993; de Silva and Lindsay, 2015; Carey et  
333 al., 2018; Cas and Simmons, 2018). Although we lack the well data required to test whether the  
334 crater-infilling-material was generated by explosive volcanic activity, we consider it plausible  
335 that the deep-water emplacement (>1.5 km) and the basaltic, inferred volatile-poor nature of  
336 magma extruded from these volcanoes may have restricted a namely “explosive” eruption style.  
337 In particular, the interplay of the deep-water setting and magma composition may have led to  
338 primarily effusive eruptions or rapid magma extrusion into the water column in a “non-explosive”  
339 manner, e.g. such as suggested for high mass eruption rates by Manga et al. (2018). If explosive  
340 activity was inhibited, a different mechanism for producing the observed erosive craters is  
341 required.

342 Given the sub-horizontal seismic reflections (i.e. SF1) infilling the craters and truncation of  
343 underlying strata by the basal surface, we suggest the crater-like bases could have formed in  
344 response to the escape of magma-related hydrothermal fluids (e.g. fluids from magma and/or  
345 heated pore fluids from the surrounding sediments). We propose that fluid escape disaggregated  
346 and excavated the weak, near-seabed sediments via a similar process to that inferred for ancient,  
347 seismically-imaged hydrothermal vents (e.g. Planke et al., 2005; Buarque et al., 2016).  
348 Considering expelled, fine-grained sediments are likely to be removed by bottom currents (e.g.

349 [Judd and Hovland, 2007](#)), we suggest the crater may have been infilled by sub-horizontal  
350 packages composed of either: (i) erupted dense material that settles out of the water column (i.e.  
351 it is not affected by bottom currents), perhaps forming layers of hyaloclastites; and/or (ii) material  
352 eroded from the depression flanks and deposited within the crater. The process we infer for the  
353 formation of the crater-like bases is similar to the generation of deep-water pockmarks, which are  
354 usually of kilometer-scale and infilled by sub-horizontal sedimentary strata (e.g. [Judd and](#)  
355 [Hovland, 2007](#)). Further exploration of the material filling these basin-like structures by drilling  
356 and coring would help clarify the style and nature of the eruptive activity and crater formation.

357       Regardless of the process(es) driving crater formation, the absence of crater-like bases beneath  
358 GP2 edifices indicates volcanism did not always involve near-seabed excavation and was site  
359 specific ([Figs. 4e-f](#)). It is difficult to determine exactly what factors (e.g. seabed cohesivity and  
360 porosity, water depth, mass eruption rate, magma composition and volatile content) controlled  
361 the initial emplacement styles of the GP1 and GP2 volcanoes solely from the geophysical data  
362 we use here.

363

#### 364 5.2.2. Model of volcano growth

365       Volcano geometry is influenced by the interplay of constructive (e.g. dyke intrusion and  
366 stacking of lava flows) and destructive processes (e.g. flank collapse and erosion) (e.g. [Annen et](#)  
367 [al., 2001](#); [Kervyn et al., 2009](#); [Magee et al., 2013](#)). To evaluate edifice growth, field- and remote  
368 sensing-based studies broadly rely on the assumption that, in any given volcanic field or setting,  
369 small volcanoes develop into large volcanoes (e.g. [Walker, 1993](#); [de Silva and Lindsay, 2015](#)).  
370 Patterns in volcano morphometry have therefore been used to infer growth models (e.g. [Rossi,](#)  
371 [1996](#); [Calvès et al., 2011](#); [Magee et al., 2013](#); [Karlstrom et al., 2018](#)). From these morphometric  
372 data, the following growth models for various volcanoes from subaerial and shallow marine  
373 settings have been proposed ([Fig. 8](#)): (i) proportional increase in summit height and basal diameter,  
374 maintaining flank dip ([Magee et al., 2013](#); [Figs 8b, f](#)); (ii) preferential addition of material to the  
375 summit area and upper volcano flanks, whilst the diameter remains consistent and flank dip  
376 increases with time ([Magee et al., 2013](#); [Figs 8c, f](#)); (iii) lateral progradation of the edifice flanks  
377 while summit height is fixed, such that flank dip decreases with time ([Calvès et al., 2011](#); [Figs](#)

378 8d, f); and (iv) maintenance of a proportional increase in summit height and basal diameter with  
379 time, interrupted by a short-stage of lateral progradation of the edifice flanks (Rossi, 1996; Figs  
380 8e, f). However, these growth models derived from volcano morphometry data are difficult to test  
381 because we cannot easily access and evaluate the internal 3D structure of volcanoes. Seismic  
382 reflection data uniquely allows us to image volcano interiors in 3D, meaning we can interrogate  
383 how edifices build up through time by mapping internal layers (Magee et al. 2013; Sun et al.,  
384 2019). Here, we compare the internal architecture of our deep-water volcanoes to growth models  
385 from those emplaced in subaerial and shallow marine conditions (Fig. 8), and discuss potential  
386 environmental controls on the differences we recognize in their geometry and evolution.

387 The internal seismic facies variations we recognize within our GP1 volcanoes differ from the  
388 seismic facies observed within monogenetic volcanoes that are mainly characterized by  
389 homogeneous seismic reflections (e.g. Reynolds et al., 2018). These facies differences suggest  
390 the GP1 volcanoes were instead formed through multiple eruptive events (i.e. they are polygenetic)  
391 and we propose they likely developed in three stages (Fig. 9): (Stage 1) during the first stage,  
392 crater-like bases formed through the explosive expulsion of hydrothermal fluids (see Section  
393 5.2.1); (Stage 2) crater infilling through eruption of material and/or mass wasting of crater flanks,  
394 forming the aggradational SF1 facies; (Stage 3) construction of a broadly conical edifice on a  
395 relatively flat surface, following crater infilling, through summit eruptions that promoted vertical  
396 and lateral growth as evidenced by the positive correlation between volcano height and basal  
397 diameter, and the parallelism between their external morphology and internal SF2 reflections (cf.  
398 Figs. 4, 6a, 8a-b, f). Down-dip convergence of internal SF2 reflections in some volcanoes (e.g.  
399 V12; Fig. 4b) suggests that, for some edifices, vertical aggradation through accumulation of  
400 erupted material at the summit may have outpaced lateral expansion of the basal diameter (e.g.  
401 Figs. 8a, c, f) (Vail et al. 1977; Magee et al. 2013); i.e. in this scenario, little erupted material  
402 reached the base flanks of the volcano, perhaps because eruption rate was low and episodic. The  
403 growth of some volcanoes by vertical aggradation may explain why flank dips of the GP1  
404 population correlate moderately positively with volcano height, but not basal diameter (Figs. 6b-  
405 c). During Stage 2 or Stage 3, after full or partial infilling of the crater-like base, intrusions feeding  
406 summit eruptions may have modified the core of SF1 to form the conical structures that are locally

407 observed (Figs. 4a-d, 9).

408 GP2 volcanoes lack crater-like bases but otherwise appear similar to GP1 (Figs 4 and 6); i.e.  
409 the volcanic materials contained in the GP2 volcanoes were expelled directly onto the paleo-  
410 seabed, feeding a volcano that grew both vertically and laterally (Stage 3; Fig. 9). The narrow,  
411 low-amplitude zone directly beneath the volcanoes and associated deformations (e.g. deflected-  
412 upward seismic reflections) suggest that dykes or faults may have served as magma ascent  
413 pathways (MP) (Fig. 5). However, these upward-deflected seismic reflections may also be  
414 possibly interpreted as seismic artefacts (i.e. velocity pull-ups) that are caused by the overlying  
415 thick, high-density volcanic rocks.

416

### 417 5.2.3 Controls on edifice morphology

418 Compared to subaerial and shallow-water basaltic volcanoes, as well as moderate- to deep-  
419 water (0.9-3 km) andesitic-basaltic volcanoes, the deep-water basaltic volcanoes we study: (i) are  
420 ~41–427 times (in volumes) smaller than basaltic and basaltic-andesitic volcanoes from  
421 elsewhere (Fig. 6a, d-e; Table S1); (ii) display similar positive correlations between height,  
422 diameter, and volume, implying volcano growth broadly involved a proportionate increase in  
423 summit height and basal diameter (Fig. 8-9) (e.g., Magee et al., 2013); but (iii) have steeper flanks  
424 (most of them  $>20^\circ$ ), with some volcanoes evidently growing via preferential vertical aggradation  
425 (Fig. 6b-c). We tentatively suggest that the small size and steeper flanks we observe likely reflect  
426 differences in the environment of emplacement (e.g., water depth) and seabed lithology. Below  
427 we consider how mass eruption rate and magma volatile content may control the  
428 geomorphological characteristics of volcanoes in this study.

429 The magnitude, duration, and steadiness of eruption rate influence the distribution of extruded  
430 material (e.g. de Silva and Lindsay, 2015; White et al., 2015). For example, low eruption rates  
431 drive lava to move slowly over short distances (e.g. Rossi, 1996) and, thus, erupted materials are  
432 more likely to accumulate around the vents/upper flanks and form high-angle slopes; i.e. growth  
433 is via preferential vertical aggradation. Low eruption rates could explain the steep slopes of the  
434 volcanoes we study, and may relate to the limited magma supply during post-rift volcanism (e.g.  
435 Yan et al., 2006; Li et al., 2014) and/or volatile undersaturation in the basaltic parental magma.

436 Episodic, shorter-duration emplacement of lava (as opposed to a single event) would also build  
437 notably steeper flanked volcanoes, as demonstrated in experiments by [Fink et al. \(1993\)](#). However,  
438 we note that the presence of long run-out lava flows flanking the volcano edifices (>9.0 km long)  
439 likely indicates eruption rates varied significantly through time, with intermittent periods of short-  
440 lived, high eruption rates of, possibly, volatile-enriched magma feeding the longest run-out flows  
441 ([Sun et al., 2019](#)) ([Fig. 5](#)). In addition to the low mass eruption rates, volatile-undersaturated lavas  
442 (as primarily inferred here) have higher cooling rates, higher glass transition temperatures and  
443 higher viscosities, and thus, lava may quench and build-up more proximal to the eruptive source  
444 ([Del Gaudio et al., 2007](#)).

445 Because of high hydrostatic pressure, wet, cold, and unconsolidated sediments, and the overall  
446 magma-deficient (low eruption rate and magma supply), post-rift setting during the Late Miocene,  
447 the deep-water volcanoes documented here geomorphologically and genetically differ to their  
448 subaerial and shallow-water counterparts in other tectonic environments ([Fig. 6](#)). In future,  
449 physical and geochemical studies of eruptive products, particularly within GP1 volcanoes, may  
450 help resolve the unusual morphologies and eruptive mechanisms within this tectonic setting.

451

## 452 **6. Conclusion**

453 We use 3D seismic reflection data to investigate the three-dimensional structure of thirteen  
454 Late Miocene-to-Quaternary deep-water volcanoes. Two groups of volcanoes, one with (GP1)  
455 and one without (GP2) crater-like bases, are identified. Internally, these volcanoes comprise two  
456 dominant seismic facies types that document volcano growth processes. We are able to investigate  
457 the relationship between the external morphology and internal structure of deep-water volcanoes,  
458 and thereby build growth models for these hitherto poorly understood volcanic structures. The  
459 growth of most of the volcanoes is defined by two main stages: crater formation and infilling,  
460 likely initiated by the escape of hydrothermal fluids, and subsequent construction of an overlying  
461 conical edifice. Importantly, recognition of crater-like bases beneath the volcanoes implies the  
462 volume of modern deep-water volcanoes, which are typically quantified by bathymetric surveys,  
463 may be grossly underestimated as the volcanoes may not have a flat, seabed-parallel base. In this  
464 study, most of the deep-water volcanoes have edifice volumes less than the underlying craters.

465 Our growth models suggest the morphology of the studied deep-water volcanoes were primarily  
466 controlled by the high hydrostatic pressure occurring in the deep-water setting, the volatile-poor  
467 nature of the parent magma, and variable magma supply due to the post-rift tectonic setting. In  
468 particular, these factors led to erupted material primarily accumulating near the summit and on  
469 the upper flanks of the volcanoes, meaning they have relatively smaller sizes (basal diameters,  
470 heights and volumes) and are characterized by slopes steeper than that typically seen in their  
471 subaerial, shallow-water, and deep-water arc-related counterparts. This study adds a unique  
472 dataset to the global database of submarine volcano morphologies. Moreover, this study also  
473 highlights that 3D seismic surveys could help revise previous estimates of submarine volcano or  
474 seamount volumes and morphologies, and further our understanding of submarine volcanoes that  
475 are already relatively-well studied.

476

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485

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622

## 623 **Figure Captions**

624

625 Figure 1: (a) Geological setting of the study area. Red polygonal line located to the south of  
626 Dongsha Islands is the 3D seismic survey. Top left: regional setting of the South China Sea. It is  
627 bounded by the Red River Strike-slip faults (RRFs) to the west and by the subduction trench  
628 (Manila Trench) to the east. Northern South China Sea is marked with black square. Igneous rocks

629 with ages from exploration wells and seamount dredges are marked with blue circles/rings (Jin,  
630 1989; Li and Liang, 1994; Zou et al., 1995). Crustal structure profiles (OBS1993 (Yan et al., 2001)  
631 and OBS2006-3 (Wei et al., 2011)) are marked with pink solid lines. ODP sites 1145 and 1146,  
632 IODP site U1501 and location of Figure 3 are also labeled. The base map is modified from Yang  
633 et al. (2015) and Sun et al. (2019); (b) and (c) Regional seismic strata of the study area. Volcanic  
634 materials are mainly located in the shallow level (0-300 m) of post-rifting strata. See location in  
635 (a).

636

637 Figure 2: (a) Schematic diagram of the calculation method for igneous velocity within the volcano  
638 and surrounding sediments; (b) Schematic diagram of geomorphic parameters measured in this  
639 study, using an example volcano with an identified crater-like base and overlying edifice. Note  
640 that the travel-time distances between the volcano summit and its base ( $T_s$ ), or the top of the  
641 velocity pull-up ( $T_i$ ), were measured *within* the volcano edifices.

642

643 Figure 3: (a) Present seabed morphologies of the study area, interpreted from the 3D seismic data.  
644 The landmark areal projections of buried or partly buried volcanoes are marked; (b) Thickness  
645 (and thus, height) of volcano edifices in the study area; (c) Thickness (and thus, depth) of volcano  
646 craters in the study area; (d) Total thickness of the volcanoes, calculated from the vertical addition  
647 of (b) and (c). The boundaries of merged volcanoes are marked. 100 ms (twt) = ~200 m.

648

649 Figure 4: (a) - (d): Seismic characteristics of volcanoes (GP1). (a) Volcano 9 (V9) and its line  
650 drawing; (b) Volcano 12 (V12) and its line drawing; (c) Volcano 1 (V1) and its line drawing; (d)  
651 Volcano 7 (V7) and its line drawing. (e) - (f): Seismic characteristics of volcanoes (GP2). (e)  
652 Volcano 8 (V8) and its line drawing; (f) Volcano 6 (V6) and its line drawing. 150 ms (twt) for  
653 volcano is equal to ~300 m. TV = top of volcano; PS = present seabed (solid pink line); PLS =  
654 paleo-seabed (solid green line); IS = inferred present seabed (dashed pink line); SF1 = seismic  
655 facies 1; SF2 = seismic facies 2; VE. = vertical exaggeration. See locations in Fig. 3a.

656

657 Figure 5: 3D seismic profile crosses through V8, V9 and V10. The igneous pathways underneath

658 the volcanoes are narrow, vertical structures (dashed arrows) and the surrounding strata are  
659 slightly pushed upward, suggesting them probably as dykes. VE. = vertical exaggeration. See  
660 location in Fig. 3a.

661

662 Figure 6: Geomorphologic parameters of deep-water volcanoes (gray solid circles; this study),  
663 shallow-water volcanoes (pink squares; [Magee et al., 2013](#)), subaerial arc volcanoes (green  
664 triangles; [Grosse et al., 2009](#)), submarine arc volcanoes (grey cross; [Wright et al., 2006](#)) and ocean  
665 volcanoes (blue rhombus; [Smith, 1988](#)). (a) Height vs diameter; (b) Dip vs diameter; (c) Dip vs  
666 height; (d) Diameter vs volume; (e) Height vs volume; (f) Dip vs volume; The deep-water  
667 volcanoes in this study have different trends (slopes) to other types. The errors of geomorphologic  
668 parameters of volcanoes in the study area are from the ranges of volcano velocities.

669

670 Figure 7: (a)-(f): Geomorphologic characteristics of the craters of deep-water volcanoes in this  
671 study. (a) Depth vs diameter (crater); (b) Dip vs diameter (crater); (c) Dip (crater) vs depth; (d)  
672 Depth vs volume (crater); (e) Diameter (crater) vs volume (crater); (f) Dip (crater) vs volume  
673 (crater). (g)-(i): Geomorphologic characteristics between the volcano edifices and craters of deep-  
674 water volcanoes. (g) Height vs depth; (h) Diameter (crater) vs diameter (edifice); (i) Volume  
675 (crater) vs volume (edifice); (j)-(l): Geomorphologic characteristics of the total volcanoes. (j)  
676 (Height + depth) vs average diameter; (k) (Height + depth) vs total volume; (l) Average diameter  
677 vs total volume.

678

679 Figure 8: Volcano growth models. (a) Model of deep-water volcano growth (GP1; purple dashed  
680 lines) with preferentially vertical aggradation of the edifice flanks; (b) Model of shallow-water  
681 volcano growth (red dashed lines) through a proportional increase in summit height and basal  
682 diameter (offshore southern Australia; [Magee et al., 2013](#)); (c) Model of shallow-water volcano  
683 growth (green dashed lines) where summit height increases, whilst basal diameter remains  
684 consistent (offshore southern Australia; [Magee et al., 2013](#)); (d) Model of shallow-water and  
685 subaerial pioneer cones of hyaloclastite mounds (dark blue dashed lines) where the basal diameter  
686 increases, whilst the summit height remains consistent (western Indian rifted margin; [Calvès et](#)

687 [al., 2011](#)); (e) Model of subaerial shield volcano growth (orange dashed lines) involving a  
688 proportional increase in summit height and basal diameter, which is disrupted by a short stage of  
689 preferentially lateral progradation of the edifice flanks (Iceland, [Rossi, 1996](#)). (f) The expected  
690 trends in summit or basal diameter plotted against volcano volume and average flank dip for all  
691 models (a-d).

692

693 Figure 9: Cartoon showing proposed three-stage evolution of GP1 volcanoes (see text for details).

694 GP2 volcano growth may be akin to Stage 3. SF1 = Seismic facies 1; SF2 = Seismic facies 2; MP  
695 = Possible magmatic intrusions.

696

697

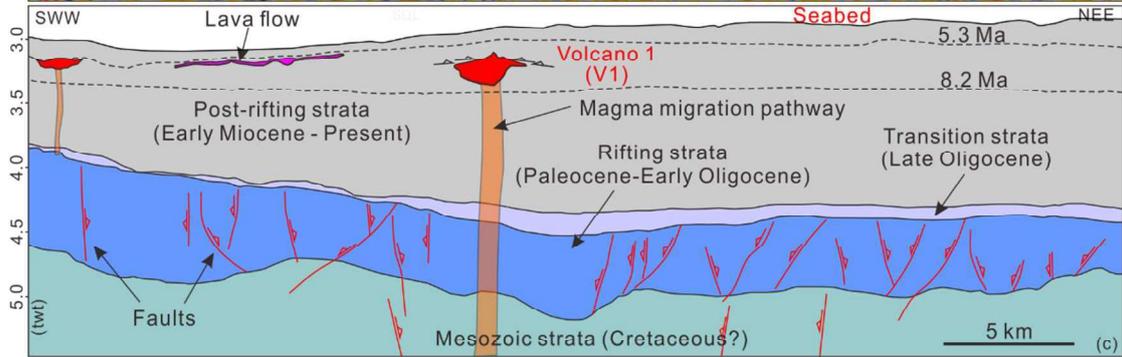
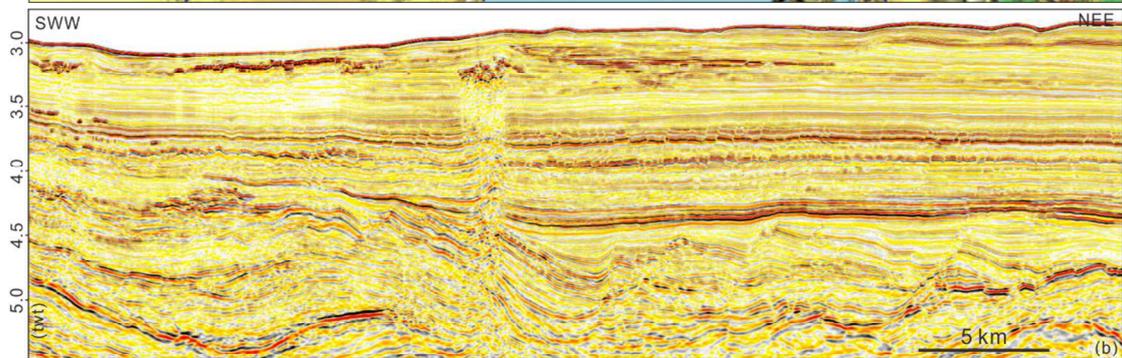
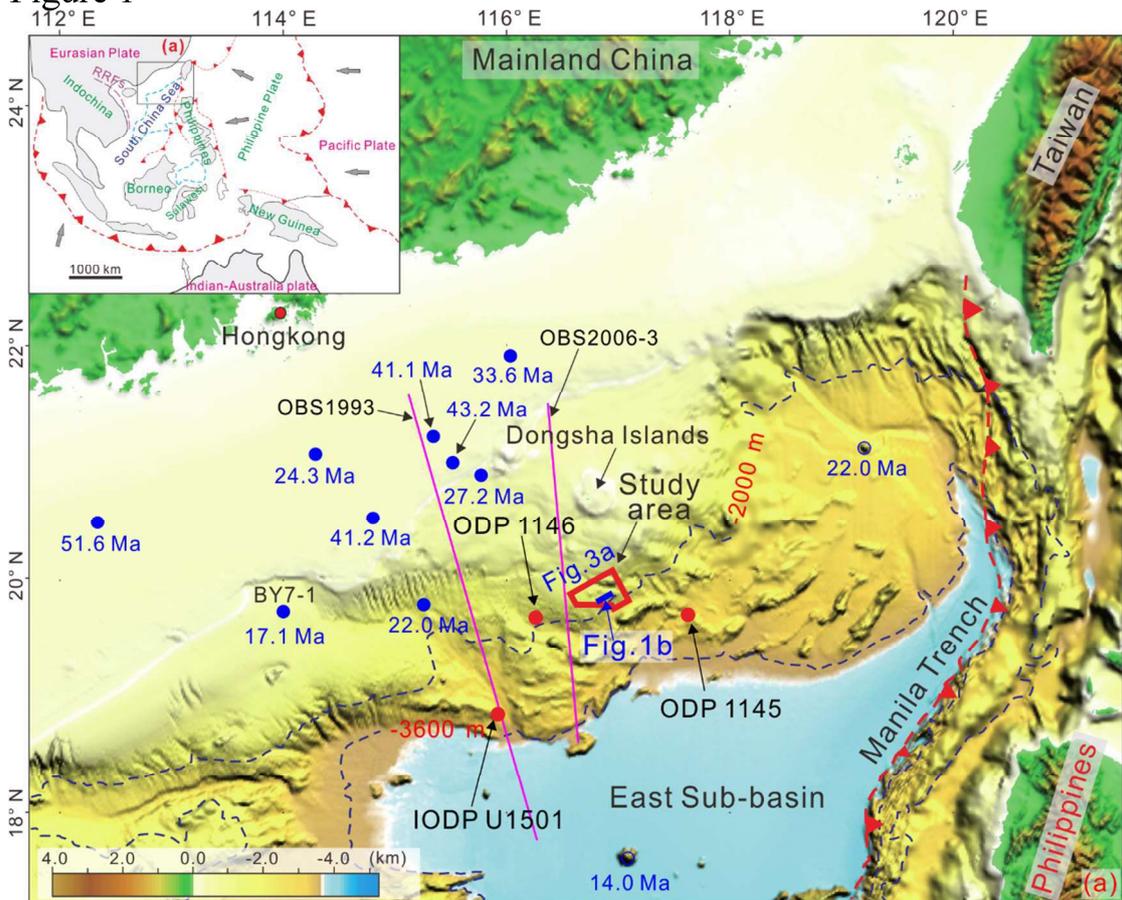
698 **Table Caption**

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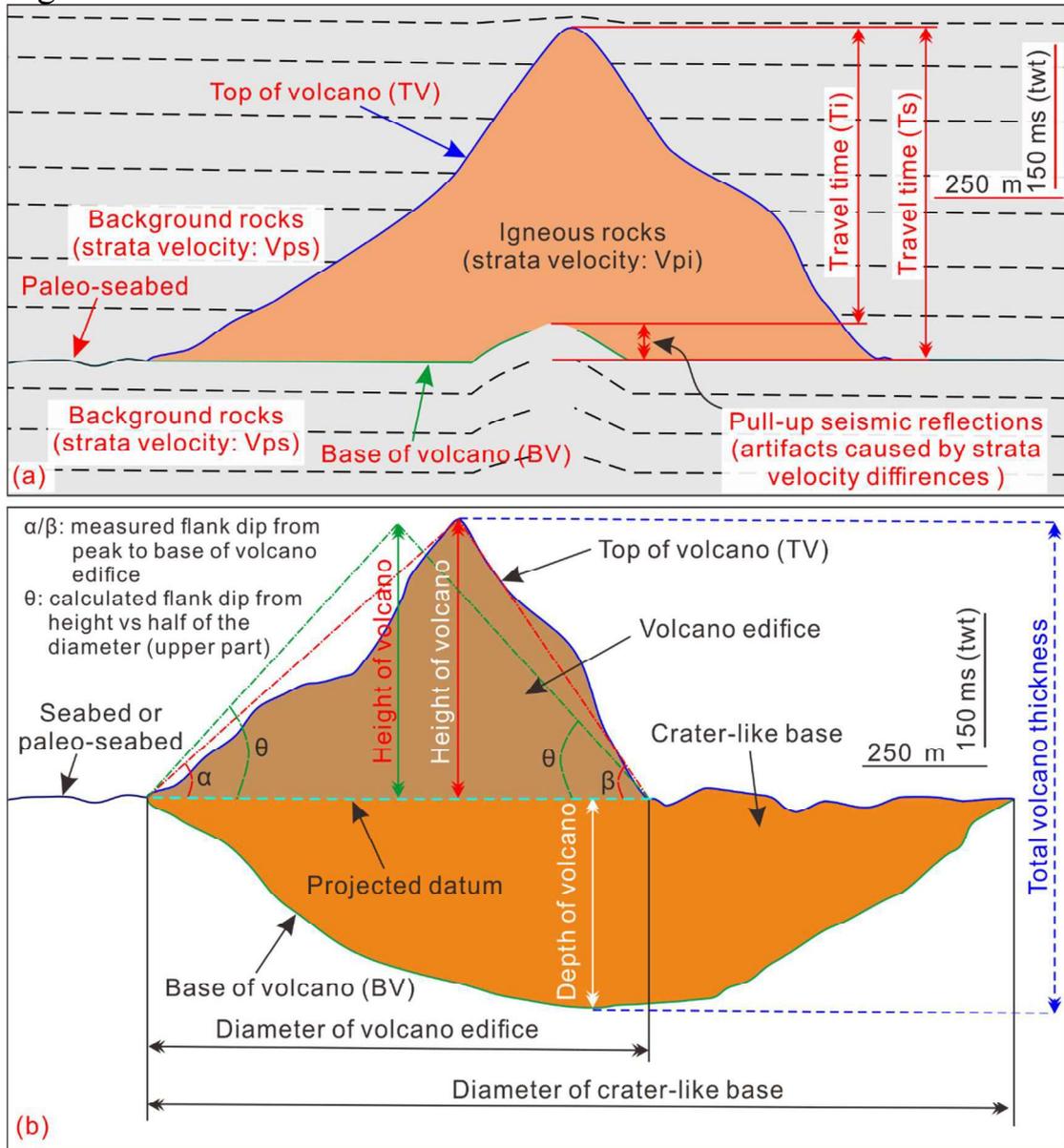
700 Table 1: Geometrical parameters of the edifices and craters of volcanoes. (1) = water depth of the  
701 seabed where the volcanoes emplace or emplace underneath it (W.D.); (2) = sediment thickness  
702 overlying the buried volcanoes (Th.).

703

Figure 1



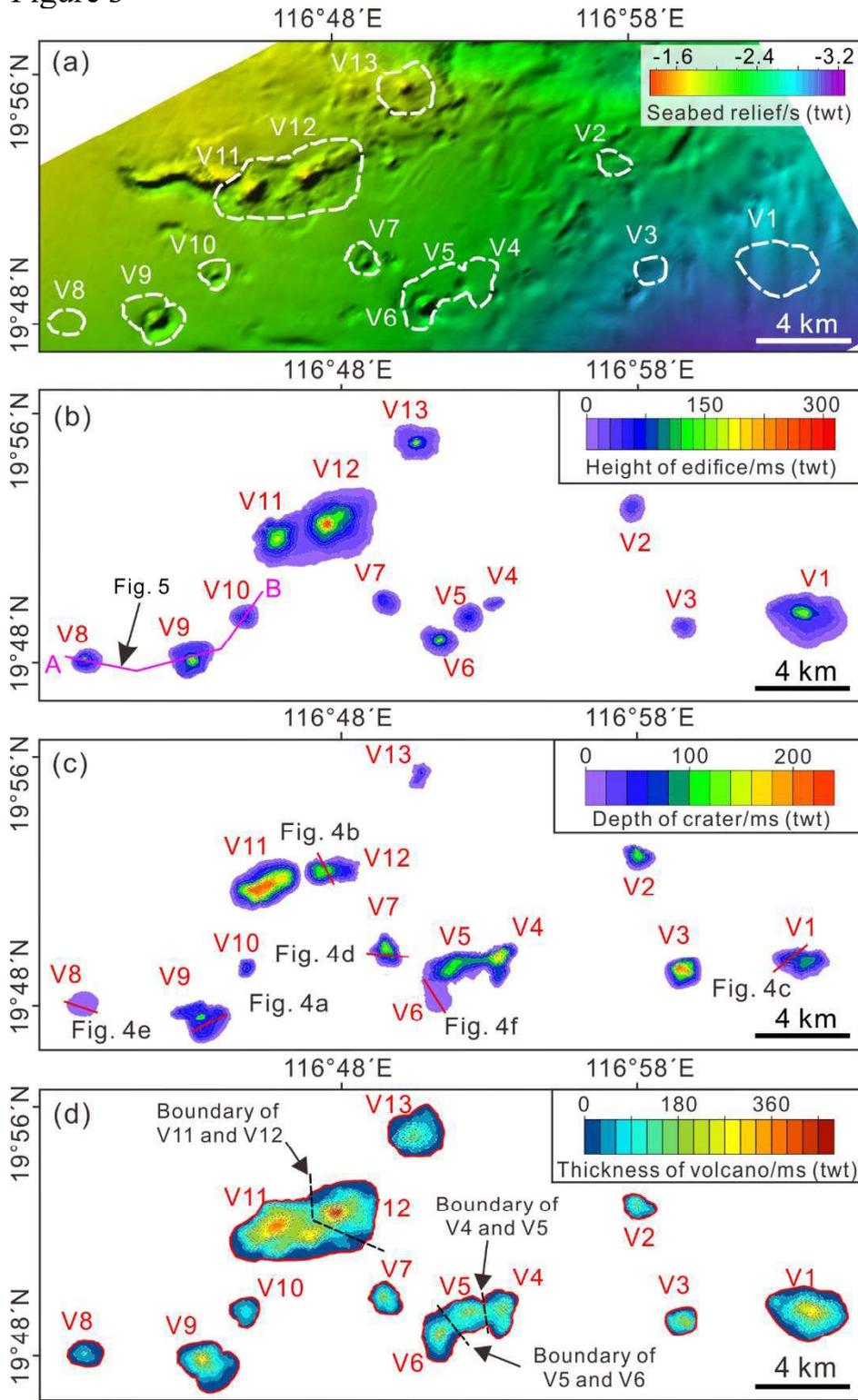
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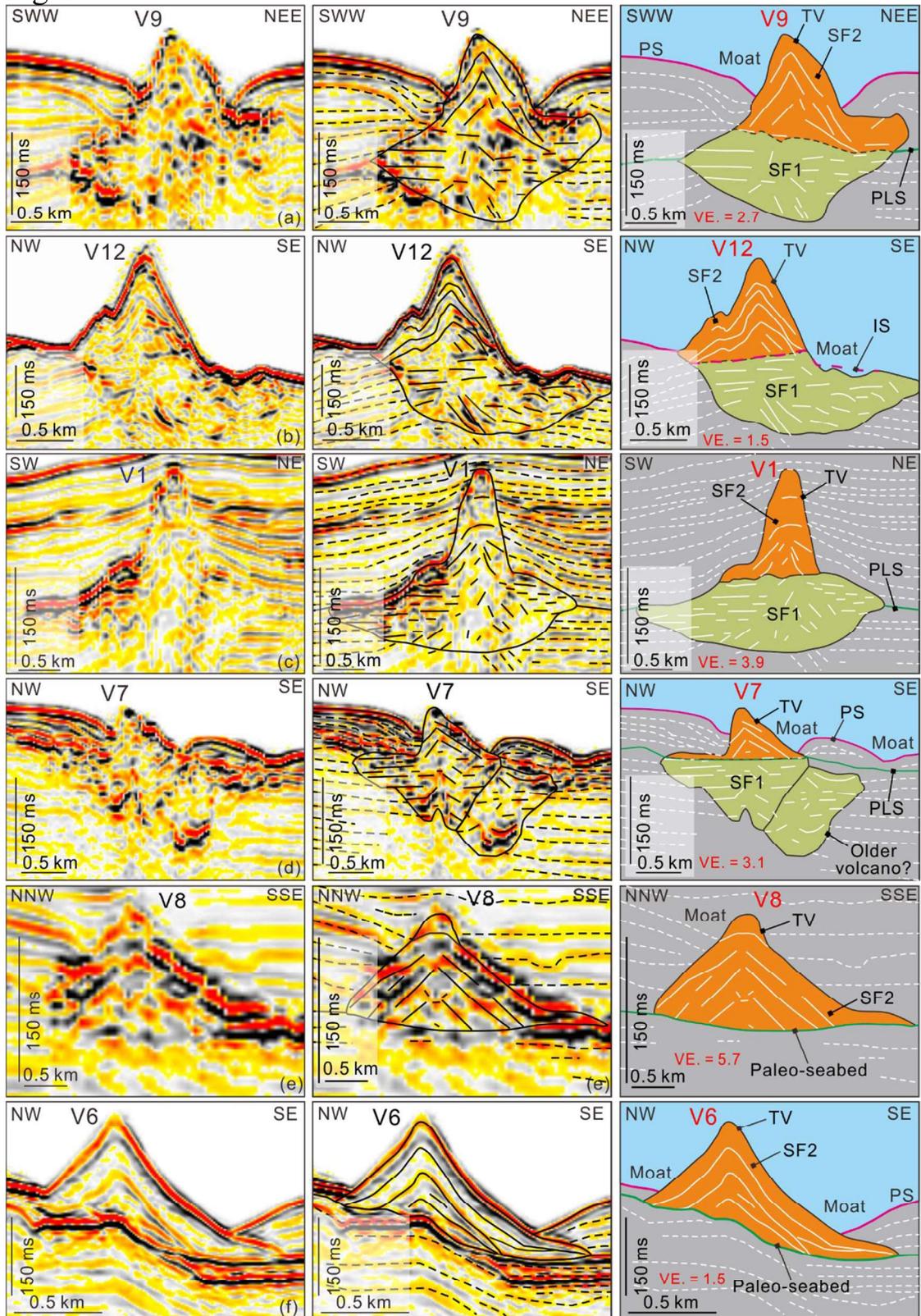
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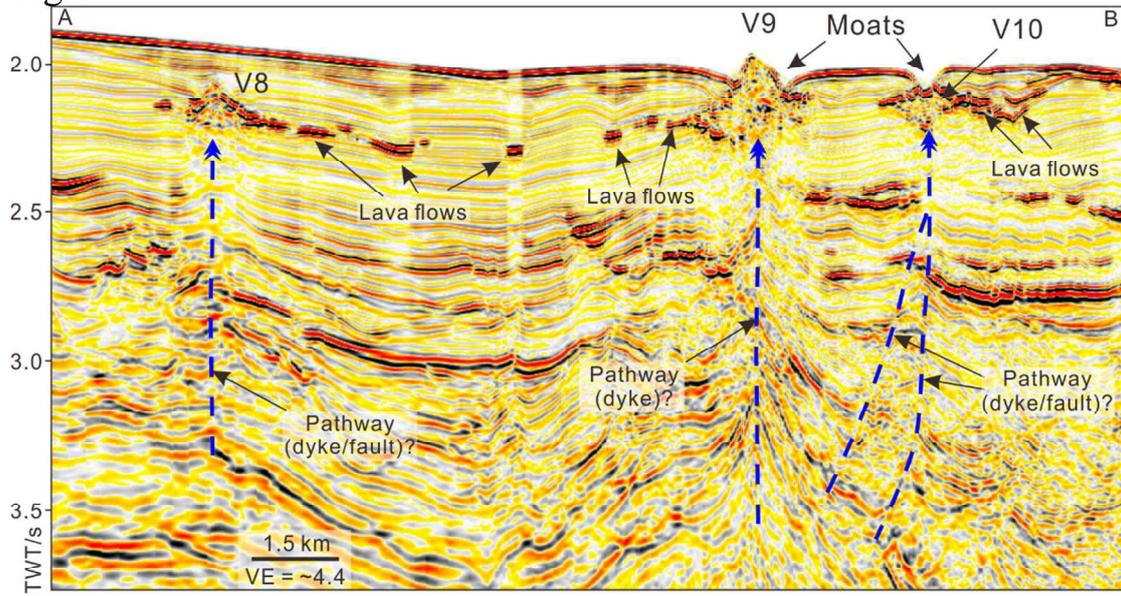
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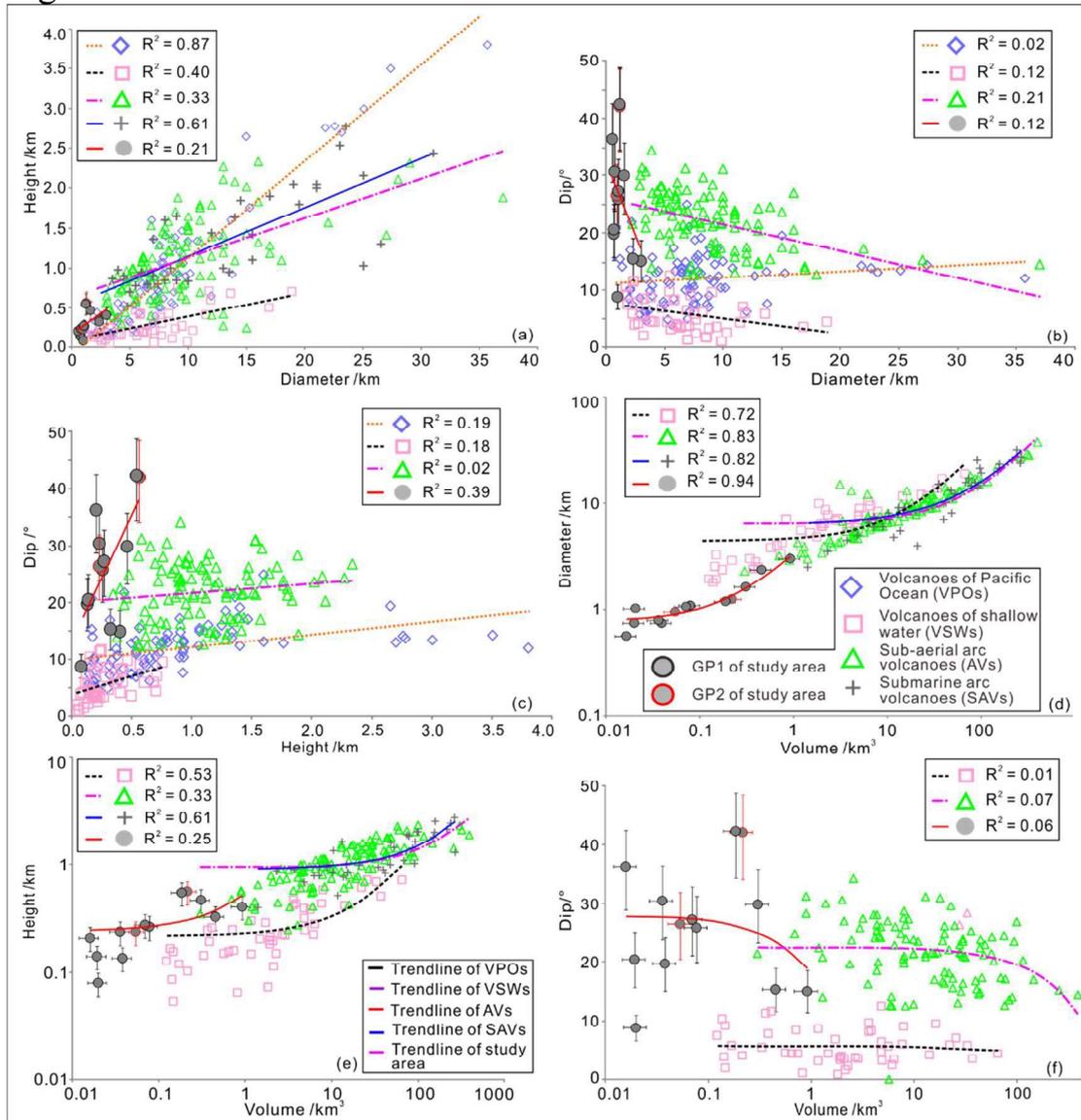
712 Figure 4



714 Figure 5

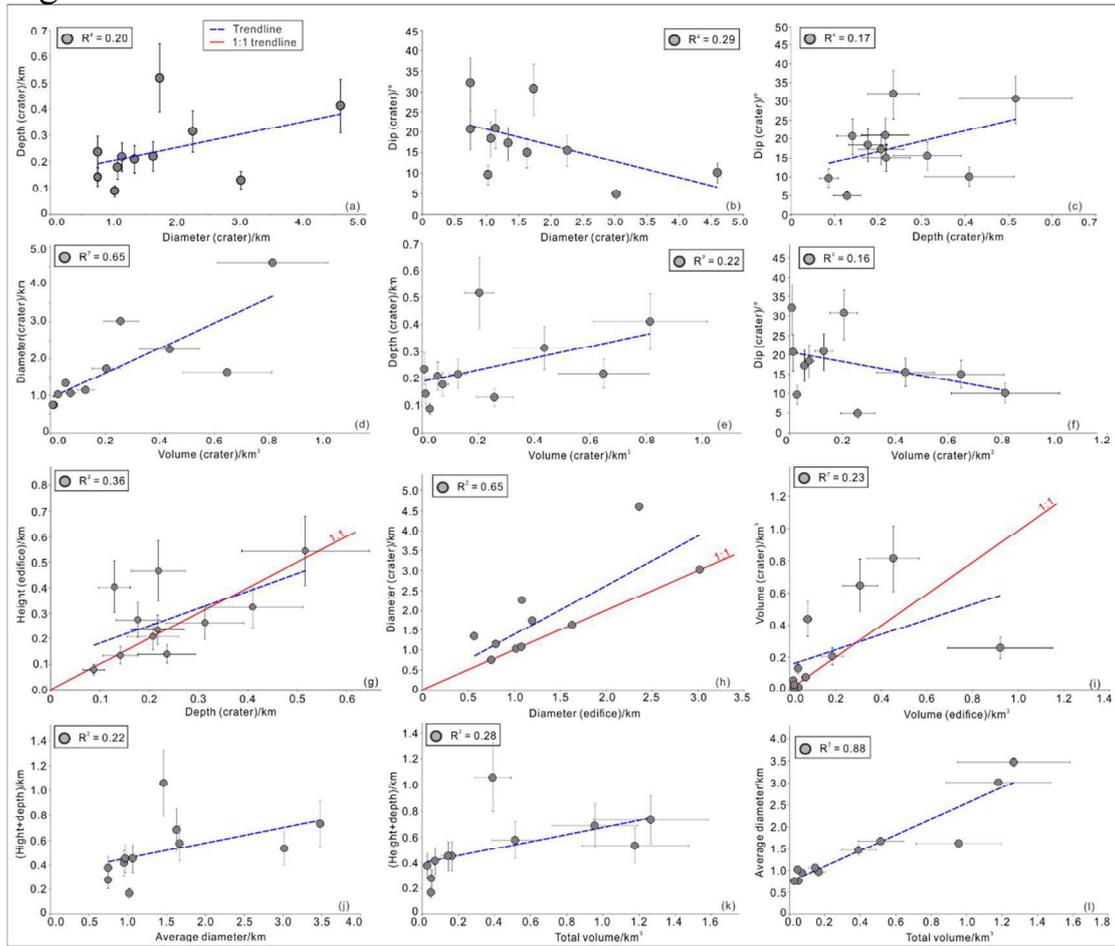


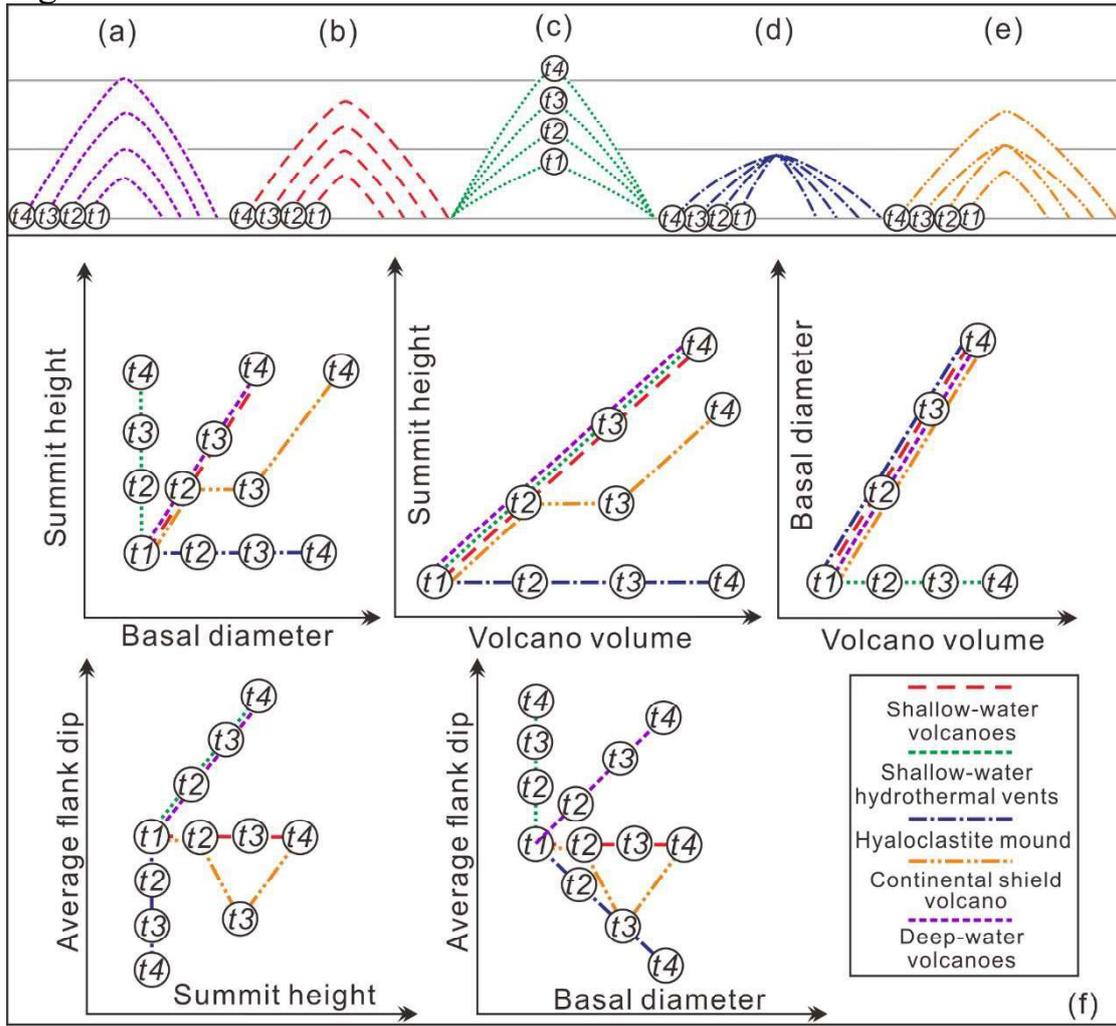
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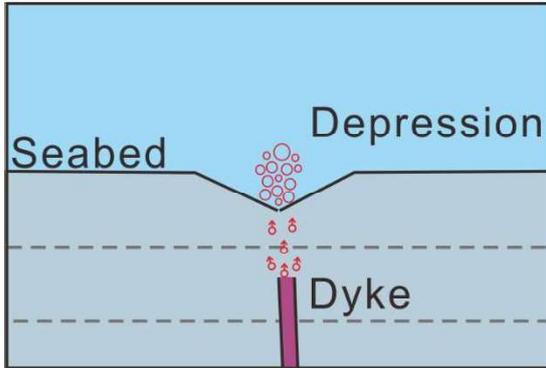
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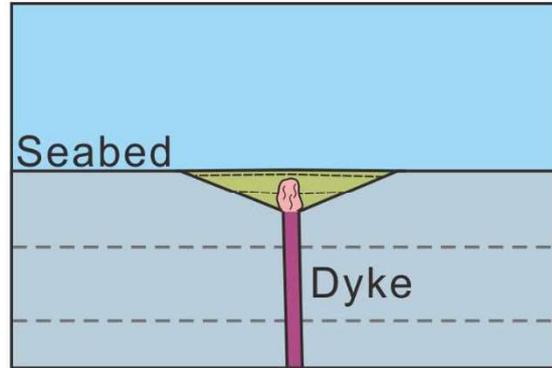


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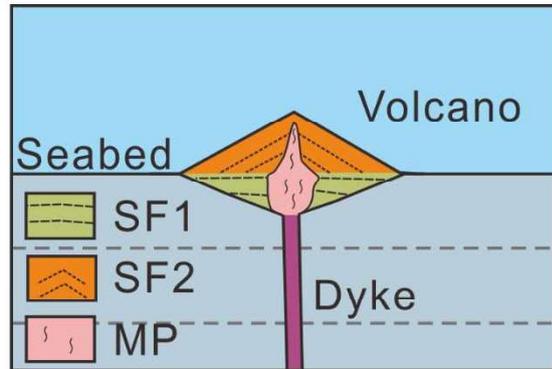
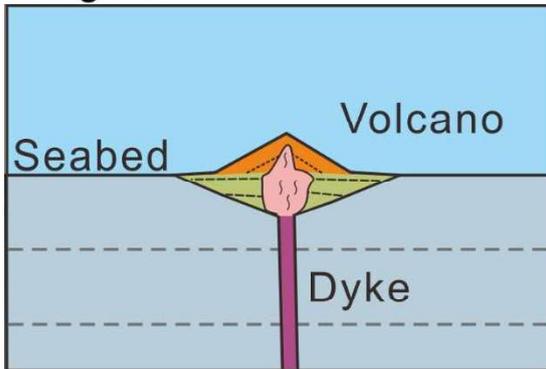
Figure 9  
Stage 1



Stage 2



Stage 3



726

727

728

1 Table 1: Geometrical parameters of the edifices and craters of volcanoes. (1) = water depth of the seabed where the volcanoes emplace or emplace underneath it (W.D.);  
 2 (2) = sediment thickness overlying the buried volcanoes (Th.).

3

Volcanoes No.	General parameters		Geometrical parameters of edifices					Geometrical parameters of craters			
	W. D./ km <sup>(1)</sup>	Th./m <sup>(2)</sup>	Diameter/km	Area/km <sup>2</sup>	Height/m	Min. dip/ <sup>o</sup>	Max. dip/ <sup>o</sup>	Volume/km <sup>3</sup>	Diameter/km	Depth/m	Volume/km <sup>3</sup>
V1	2.25	264.0	3.0	7.15	404±101	11	19	0.92±0.23	3.0	129±32	0.26±0.06
V2	2.14	110.0	0.8	0.44	134±34	15	24	0.04±0.01	0.8	142±35	0.01
V3	2.21	88.0	0.8	0.44	140±35	16	25	0.02	0.8	235±59	0.01
V4	1.83	82.5	0.6	0.25	207±52	29	43	0.02	1.3	208±52	0.06±0.01
V5	1.80	165	1.1	0.92	262±66	20	31	0.08±0.02	2.3	313±78	0.44±0.11
V6	1.65	55.0	1.2	1.21	560±140	34	48	0.22±0.05	--	--	--
V7	1.58	104.5	0.8	0.50	235±59	24	36	0.04±0.01	1.1	217±54	0.13±0.03
V8	1.37	121.0	0.9	0.70	235±59	20	32	0.05±0.01	--	--	--
V9	1.49	247.5	1.6	2.08	467±117	23	36	0.31±0.08	1.6	218±55	0.65±0.16
V10	1.46	115.5	1.0	0.82	79±20	7	11	0.02	1.0	87±22	0.03±0.01
V11	1.37	33.0	2.4	4.37	323±81	12	19	0.45±0.11	4.6	410±103	0.81±0.20
V12	1.35	33.0	1.2	1.12	544±136	34	49	0.19±0.05	1.7	517±129	0.21±0.05
V13	1.36	22.0	1.1	0.89	274±69	21	33	0.07±0.02	1.1	177±44	0.07±0.02

4

5