The influence of regional stress and magmatic input on styles of monogenetic and polygenetic volcanism

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Abstract. A new model is proposed to relate the development of monogenetic and polygenetic volcanoes to magmatic input and regional stress in the lithosphere. Output-stress diagrams relate magmatic output rate and crustal deformation rate (or strain rate). The magmatic output rates of polygenetic volcanoes and monogenetic volcano fields, including lava fields and central (axial) volcanoes on the mid-oceanic ridge systems, are estimated. Crustal deformation rates obtained from the literature are used as indicators of differential stress ($\Delta\sigma$). These data are normalized over time periods of $10^4$ years, and over areas of $10^5$ km$^2$. The ratio of output rate to input rate, inferred from ophiolite sections, and from crustal deformation around volcanoes, is about 0.1 to 0.3. Based on the average ratio, an input-stress diagram may be obtained at some volcanoes. Polygenetic volcano fields are plotted in the region of both larger output rate and smaller $\Delta\sigma$. Monogenetic volcano fields are plotted on a rift trend, or in the region of a small output ($\leq 1$ km$^3/10^4$ yr / $10^3$ km$^2$). In some regions, polygenetic volcanoes and monogenetic volcanoes coexist, for example, during 0.1 m.y. periods, and within the area of 30 km x 30 km. The output-stress diagram indicates that coexistence of these volcano types results from variation in differential stress, or variation in production rate of magma in the mantle at different levels over periods of time less than 0.1 m.y. This is supported using extrusive volumes and crustal deformation data from the eastern volcanic zone, Iceland, Taupo Volcanic Zone, New Zealand, TransMexican Volcanic Belt, Mexico, and the volcanic field on and around the Izu peninsula, Japan. The observed relationship between deformation rate, magmatic output rate, and volcano type supports the crack interaction theory of magma-filled cracks. The output-stress diagram also indicates that the balance between local stress induced by magma accumulation and regional stress, and stress relaxation, govern the structure of a volcano.

Introduction

Volcanoes are most simply classified into monogenetic and polygenetic types [e.g., Macdonald, 1972; Nakamura, 1975, 1986; Williams and McBirney, 1979; Moriya, 1983]. Monogenetic volcanoes are defined as volcanoes that erupt only once over a brief period of time; polygenetic volcanoes are defined as volcanoes that erupt repeatedly from the same general vent (summit or main crater) [Nakamura, 1975, 1986]. Monogenetic volcanoes include lava fields which erupt from different fissures on continental rifts. Central (axial) volcanoes on oceanic ridges [e.g., Macdonald, 1989] are composed of aggregates of lava fields which erupt at different fissures, as evidenced in parallel dikes which develop under central volcanoes as observed in ophiolite sections [Nicolas, 1989], and so are considered to be monogenetic volcanoes. Although monogenetic and polygenetic volcanoes are often classified simply on the basis of volcano morphology, the differences among these volcanoes may be controlled by tectonic setting. For example, drawing on thermodynamic arguments, Fedotov [1981] proposed that magma supply rate governs the variation among volcanoes: a low magma supply rate leads to the formation of monogenetic volcano fields, and high supply rates maintain the heat and mass flow necessary to sustain polygenetic volcanoes. Hildreth [1981] discussed variation in volcanism in terms of magma supply, but he also incorporated magma percolation, or magma extrusion/intrusion rate, as an important parameter regulating differences in the style and development of volcanism. However, there is ample field evidence to indicate that monogenetic volcano fields are found predominantly in extensional tectonic settings [e.g., Nakamura, 1986]. This led Nakamura [1977, 1986] to conclude that differential stress in the lithosphere plays an important role in the development of polygenetic or monogenetic volcanism. Takada [1994a] has proposed a model of crack interaction that integrates these ideas, in an effort to explain tectonic controls on the style and development of basaltic volcanoes. Crack interaction theory shows that dikes are more likely to coalesce given a relatively high magma input rate and a relatively low regional differential stress. The increased likelihood of crack coalescence may result in polygenetic volcanism. Alternatively, low magma supply rate and high regional differential stress both work to prevent crack coalescence, and this may result in the development of monogenetic volcano fields.

The ideal test of this hypothesis would be a comparison of magma supply rate and regional differential stress in areas of polygenetic and monogenetic volcanism. Although these data are impossible to obtain, a comparison of magma output and crustal deformation between volcanic regions is possible. In
this paper these data are compiled and compared, using an output-stress diagram introduced here. Variation between monogenetic and polygenetic volcanism is discussed using the output-stress diagram and results are interpreted in light of crack interaction theory [Takada, 1994a, b].

Estimating Magma Output and Stress From Field Data

The long-term magmatic output rate and the crustal deformation rate (or strain rate) must be estimated from field observations, and are used to make the output-stress diagram. Magmatic output rate is normalized to periods of $10^4$ years, over an area of $10^3$ km$^2$ in order to compare different areas. The normalized time range is far smaller than that required for the evolution of a polygenetic volcano, which may be 0.2-0.3 m.y. [Moriya, 1983]. Lava fields in Iceland, Great Rift, and Higashi Izu Monogenetic volcano field have the same resolution as a polygenetic volcano (Table 1). However, geochronological methods used in most monogenetic volcanic fields have low resolution, of the order of $10^5$ years, compared with those of polygenetic volcanoes. Normalized output to $10^4$ years is done for comparative purposes. For large polygenetic volcanoes, the normalized area will include the entire volcanic edifice. Most monogenetic volcanic fields are far larger than the area of 30 km x 30 km. In that case, the output of a monogenetic volcanic field is normalized to the area of 30 km x 30 km, (the total output) x (the total area) / ($10^3$ km$^2$).

The output rate of the mid-oceanic ridge, $M_o$, is given as

$$ M_o = h_p a_{sp} l $$

where $h_p$ is the thickness of pillow lava; $a_{sp}$ is full spreading rate; $l$ is the length along the ridge axis in the normalized area. Layer 2 in the seismic section of the oceanic crust is composed of pillow lava and a sheeted dike complex [Nicolas, 1989]. The thickness of the pillow lava is assumed to be the half thickness of layer 2, so that the output rate of the East Pacific Rise is estimated to be 30-48 km$^3$/10$^4$ yr/10$^3$ km$^2$, using (1) for $h_p = 1$ km, $a_{sp}$=100-160 mm/yr, and $l = 30$ km [Orcutt et al. 1976], or 15-38 km$^3$/10$^4$ yr/10$^3$ km$^2$ for $h_p$ = 0.5-0.8 km, $a_{sp}$=100-160 mm/yr, and $l = 30$ km [Harding et al., 1989].

The output rate of the Mid-Atlantic Ridge is estimated to be 1.8-3.6 km$^3$/10$^4$ yr/10$^3$ km$^2$ for $a_{sp}$ = 20 mm/yr, and $h_p$ = 0.4-0.6 km (thickness of layer 2A) [Talwani et al., 1971], 2.4 km$^3$/10$^4$ yr/10$^3$ km$^2$ for $a_{sp}$ = 20 mm/yr, and $h_p$ = 0.4 km (thickness of layer 2A) [Moore et al., 1974], or 4.8 km$^3$/10$^4$ yr/10$^3$ km$^2$ for $a_{sp}$ = 20 mm/yr, and $h_p$ = 0.8 km (pillow lava) [Fowler, 1976; Nisbet and Fowler, 1978].

Values of differential stress are needed as magma ascends through the lithosphere. Differential stress controls short-term volcanic events such as dike intrusion. It is difficult to estimate regional differential stress directly, but it can also be estimated from crustal displacement. The relationship between crustal strain and regional differential stress is defined on two timescales. The formulas of transformation into differential stress are

$$ \Delta \sigma = k \epsilon $$

$$ \Delta \sigma = \eta \delta e/\delta t $$

for short term and long term, respectively, where $\Delta \sigma$, $\epsilon$, $\eta$, and $\delta$ are differential stress, crustal strain, elastic modulus of the crust, crustal viscosity, and time, respectively. Differential stress estimated using (1) may be different from that estimated using (3).

Differential stress will change through time. Therefore, (3) is more valid than (2) because (3) can incorporate average values over long time periods and large areas. Strain rate governs the long-term volcanic activity, that is, development of a volcano including stress relaxation. In order to evaluate the relationship between tectonic stresses and volcanism in the long-term, strain rate is available as a good indicator of regional stress [Takahashi, 1990]. Thus crustal displacement rate including spreading rate, as well as strain rate in the normalized area, are used, because these can be estimated using geologic or geophysical data. Crustal displacement is used roughly in the same sense as cumulative finite strain.

Crustal deformation rates are estimated roughly. Some are order of magnitude estimations, but they are sufficient to discuss in the logarithmic form (Table 1; Figure 1). Parson and Thompson [1991] pointed out that dike intrusions can accommodate stress as easily as normal faulting. Therefore, deformation rate in volcanic fields may be higher than that estimated by fault slip and uplift alone. Spreading rate includes both the extension rate accomplished by dike intrusion and the slip rate due to normal faulting [Parson and Thompson, 1991].

Iceland, a ridge-centered hot spot, has four volcanic zones. For example, the eastern volcanic zone is a 60-km-wide rift zone. Lava fields derived from areal fissure eruptions occur in the Veidivötn-Laki area of the volcanic zone. The present full spreading rate in the rift zone is available for this work because the rift system may be stable in postglacial time [Schilling et al., 1982]. On the other hand, the output from the area including Hekla volcano is estimated only for a short period (1693-1970) [Wadge, 1982]. Moreover, there are no data available on the spreading rate around Hekla. Considering that Hekla is located at the northwestern margin of the volcanic zone, the spreading rate around Hekla is assumed to be of the order of 1-10 mm/yr, smaller than that of the central part of the rift zone (20-30 mm/yr).

The Rhine graben, 30-km-wide, is part of the European Cenozoic rift system. As a typical monogenetic volcanic field related to the rift, the Eifel volcanic field is located within an area of 20 km x 50 km in and around the Rhine graben. The tephra volume reported by Schmincke et al. [1990] in Table 1 is not a dense rock equivalent and results in an overestimate of output. However, lava volume has not been estimated and this likely causes an underestimate of output rate. The slip rate due to strike-slip faulting or normal faulting is not available. Consequently, crustal deformation rate is estimated from uplift or subsidence rate [Prodehl et al., 1992].

Mount Etna is a large shield volcano on the Island of Sicily. The long-term output rate of the volcano is not available. The crustal deformation rate around Mount Etna is estimated from regional uplift rates in northeastern Sicily and in southern Italy [Brogan et al., 1975].

Piton de la Fournaise is an active basaltic shield volcano on the island of la Réunion in the Indian Ocean [Stieltjes and Mouot, 1989]. The long-term output rate of the volcano is not available. If the growth rate of Piton de la Fournaise is almost constant during the last 10$^4$ years, except during a caldera formation event (the third and fourth phases [Gillot and
<table>
<thead>
<tr>
<th>Volcano or Volcanic Field</th>
<th>Type of Volcano</th>
<th>Total Output, km³</th>
<th>Estimated Area, km²</th>
<th>Duration of Output, years</th>
<th>Normalized Output Rate, km³/10³ yr</th>
<th>Deformation Rate, mm/yr</th>
<th>Duration of Deformation, years</th>
<th>Normalized Crustal Defomation, mm/yr</th>
<th>References</th>
</tr>
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<tr>
<td>East Pacific Rise</td>
<td>C</td>
<td></td>
<td></td>
<td></td>
<td>15-48</td>
<td>Sp 100-160</td>
<td>present</td>
<td>100-160</td>
<td>Harding et al. [1989]</td>
</tr>
<tr>
<td>Mid-Atlantic Ridge</td>
<td>C</td>
<td></td>
<td></td>
<td></td>
<td>1.8-4.8</td>
<td>Sp 20</td>
<td>present</td>
<td>20</td>
<td>Moore et al. [1974]</td>
</tr>
<tr>
<td>Veiðivötn-Lakagígar, Iceland</td>
<td>L  Lv144</td>
<td>3200</td>
<td>10⁴</td>
<td></td>
<td>45</td>
<td>Sp 20</td>
<td>present</td>
<td>20</td>
<td>Jakobsson [1972]</td>
</tr>
<tr>
<td>Hekla</td>
<td>P</td>
<td></td>
<td></td>
<td></td>
<td>1693-1970</td>
<td>90</td>
<td>&lt; 20-30</td>
<td>present</td>
<td>Wadge [1982]</td>
</tr>
<tr>
<td>Eifel VF, Rhine graben</td>
<td>M</td>
<td>7x10⁵</td>
<td>0.6</td>
<td></td>
<td>1990-2010</td>
<td>3x10⁵(0.7-0.4Ma)</td>
<td>≤ 0.1</td>
<td>&lt; 0.1</td>
<td>Schmincke et al. [1990], Prodehl et al. [1992]</td>
</tr>
<tr>
<td>Etna</td>
<td>P</td>
<td></td>
<td>1600-1980</td>
<td></td>
<td>90</td>
<td>0.071</td>
<td>1.7x10⁵(0.17-0Ma)</td>
<td>&lt; 0.1</td>
<td>Wadge [1982], Brogan et al. [1975]</td>
</tr>
<tr>
<td>Piton de la Fournaise</td>
<td>P</td>
<td></td>
<td>1930-1985</td>
<td></td>
<td>90</td>
<td>hydrostatic</td>
<td>present</td>
<td>&lt; 0.1</td>
<td>Stieltjes and Moutou [1989]</td>
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<tr>
<td>Nyamuragira</td>
<td>P</td>
<td></td>
<td>1900-1980</td>
<td></td>
<td>105</td>
<td>sp &lt; 0.5</td>
<td>hydrostatic</td>
<td>present</td>
<td>Wadge [1982], Baker and Wohlenberg [1971]</td>
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<tr>
<td>Kilaea</td>
<td>P</td>
<td></td>
<td>1920-1983</td>
<td></td>
<td>250</td>
<td>hydrostatic</td>
<td>present</td>
<td>&lt; 0.1</td>
<td>Wohlenberg [1976]</td>
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<td>Mauna Loa</td>
<td>P</td>
<td></td>
<td>1840-1980</td>
<td></td>
<td>250</td>
<td>hydrostatic</td>
<td>present</td>
<td>&lt; 0.1</td>
<td>Shaw [1987]</td>
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<tr>
<td>Posterosional volcano of Hawaii</td>
<td>M 1</td>
<td>1000</td>
<td>&gt; 10⁶</td>
<td></td>
<td>0.1</td>
<td>hydrostatic</td>
<td>present</td>
<td>&lt; 0.1</td>
<td>Macdonald et al. [1983], Nakamura [1986]</td>
</tr>
<tr>
<td>San Francisco VF</td>
<td>M 0.75-14</td>
<td>4800</td>
<td>4.75x10⁶ (5-0.25Ma)</td>
<td></td>
<td>0.16-3</td>
<td>SIEQ 0.38-1</td>
<td>Quaternary</td>
<td>1</td>
<td>Tanaka et al. [1986], Eddington et al. [1987]</td>
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<td></td>
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<td></td>
<td></td>
<td>SIEQ 0.23-7.4</td>
<td>Quaternary</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>St 10⁻¹⁰,10⁻¹⁷</td>
<td>Present</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mt. Taylor VF</td>
<td>M 100</td>
<td>1000</td>
<td>4x10⁵ (4-0Ma)</td>
<td></td>
<td>0.25</td>
<td>weak extension</td>
<td>Quaternary</td>
<td>&lt; 1</td>
<td>Perry et al. [1990]</td>
</tr>
<tr>
<td>Springerville VF</td>
<td>M 300</td>
<td>3000</td>
<td>1.8x10⁶ (2.1-0.3Ma)</td>
<td></td>
<td>0.6</td>
<td>extension</td>
<td>Quaternary</td>
<td>&lt; 1</td>
<td>Condit et al. [1989], Connor et al.[1992]</td>
</tr>
<tr>
<td>Great Rift</td>
<td>L 30</td>
<td>1600</td>
<td>1.5x10⁴</td>
<td></td>
<td>12.5</td>
<td>0.08-2</td>
<td>Quaternary</td>
<td>1</td>
<td>Kuntz et al. [1986], Eddington et al. [1987]</td>
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<td>Michoacan</td>
<td>M 31</td>
<td>15000</td>
<td>4x10⁴</td>
<td></td>
<td>0.5</td>
<td>Sl 1</td>
<td>Late Quaternary</td>
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<td>Hasenaka and Carmichael [1985]</td>
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<td>-Guadalupe VF</td>
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<td>0.12</td>
<td>1250</td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td>Suter et al. [1992]</td>
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<tr>
<td>Higashi Izu</td>
<td>M 100</td>
<td></td>
<td>1.5x10⁵</td>
<td></td>
<td>0.1</td>
<td>Sl 1</td>
<td>Quaternary</td>
<td>1</td>
<td>Koyama and Umino [1991]</td>
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<td>Monogenetic VF</td>
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<td>0.12</td>
<td>1250</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Hayakawa and Koyama [1992], Ref.1</td>
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<tr>
<td>Izu Oshima</td>
<td>P</td>
<td></td>
<td>1x10⁴</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Nakanura [1964], Tsukui et al. [1986], Ref.1</td>
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<td>Fuji</td>
<td>P</td>
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<td>2.5x10⁴</td>
<td></td>
<td>18-30</td>
<td>Sl 0.1</td>
<td>Quaternary</td>
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<td>Tsukui et al. [1986], Miyaji [1988], Ref.1</td>
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<td>Miyakejima</td>
<td>P</td>
<td></td>
<td>8x10⁴</td>
<td></td>
<td>50</td>
<td>Sl 0.1</td>
<td>Quaternary</td>
<td>0.1</td>
<td>Ishiki [1977]</td>
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<td>Aogashima</td>
<td>P</td>
<td></td>
<td>3x10⁴</td>
<td></td>
<td>8</td>
<td>Sl &lt; 0.1</td>
<td>Quaternary</td>
<td>&lt; 0.1</td>
<td>Takada et al. [1992]</td>
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<tr>
<td>Other volcanoes, Japan</td>
<td>P</td>
<td></td>
<td>1x10⁵</td>
<td></td>
<td>3</td>
<td>Sl &lt; 0.1</td>
<td>Quaternary</td>
<td>0.1</td>
<td>Moriya [1983], Tsukui et al. [1986], Ref.1</td>
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</table>

VF, volcanic (Volcano) field; P, polygenetic volcano; M, monogenetic volcano; C, central (axial) volcano on the oceanic ridge; L, lava field on the continental rift; Lv, lava volume; Tv, tephra volume; Sp, spreading rate; Sl, slip rate; Sb, subsidence rate; Up, uplift rate; Eq, earthquake; St, strain rate; Ref.1, Kaitadza and Inami [1984], and RGAF [1991].

* Central Utah.

† Southern Utah. If the output rate which has been estimated is available for the purpose of this paper, the output volume is not listed in this table.
Nativel, 1989), the output rate of the volcano for 1930-1985, estimated by Stietjes and Moutou (1989), is roughly valid for $10^4$ years. Piton de la Fournaise is growing in an intraplate environment, so that the regional stress condition is nearly hydrostatic.

Nyamuragira volcano is growing in the western rift of the East African Rift. The long-term output rate of the volcano is not available. On the assumption that the volcano has grown constantly, the output rate is estimated from historical data [Wadge, 1982]. In the eastern rift of the East African Rift, which is more active than the western rift, crustal deformation rates of 0.5 mm/yr [Baker and Wohlenberg, 1971] and 1-2 mm/yr [Wohlenberg, 1976] have been reported. Thus, the deformation rate of the western rift is of the order of, or smaller than, 0.5 mm/yr.

Kilauea and Mauna Loa are large shield volcanoes on the Hawaiian-Emperor Chain. The growth rate of the volcano on this volcanic chain for the 74 Ma period, or 1 Ma period is in the range of 100 to 1000 km$^3$/yr, and is of the same order as the short-term output rate of Kilauea (250-1000 km$^3$/yr) or Mauna Loa (250 km$^3$/yr)[Shaw, 1987]. The short-term output rate is available for normalization for $10^4$ years. However, the output rate during posterosional volcanism of Hawaii is smaller than 1 km$^3$/yr/10$^3$ km$^2$, based on the data of Macdonald et al. [1983]. Hawaiian volcanoes are growing in an intraplate environment, so that the regional stress condition is nearly hydrostatic except the volcanic edifice effects, including rift zones.

In the western part of San Francisco Volcanic Field (SFVF), there are older polygenetic volcanoes [Tanaka et al., 1986]. Their volumes are included in Table 1 so that the output rate of SFVF is overestimated. SFVF is located near the southern termination of the Sevier rift, a rift along the eastern margin of the Great Basin [Smith and Luecke, 1984; Tanaka et al., 1986]. Deformation rates in the central part of the Sevier rift, central Utah, and in the southern part of the Sevier rift, southern Utah [Edington et al., 1987], are available as the deformation rate near SFVF.

The Mount Taylor Volcanic Field and the Springerville volcanic field are located between the Sevier rift and the Rio Grande rift. In the Mount Taylor Volcanic Field, the stress condition is weak extension [Perry et al., 1990]. The deformation rates in these areas are likely smaller than those along the Sevier rift (1 mm/yr).

The Great Rift is a 2- to 8-km-wide volcanic rift zone in the Snake River Plain, Idaho. Basaltic lava fields derived from fissure eruptions were formed in and around the rift [Kuntz et al., 1986]. The slip rate and strain rate along faults in central Idaho [Edington et al., 1987] are useful to estimate the deformation rate near the Great Rift.

The Michoacan-Guanajuato Volcanic Field is located in the western TransMexican Volcanic Belt (TMVB), Mexico. There are no data available on the crustal deformation rate of the western TMVB. Only the slip rate of the Venta de Bravo fault in the central part of TMVB has been reported [Suter et al., 1992]. I adopt this value as the average deformation rate of TMVB. However, as the number of cinder cones relative to central volcanoes decreases with distance from the Middle American Trench, the stress condition may also vary in the TMVB [Connor, 1990].

Mount Fuji, the Higashi Izu Monogenetic Volcano Field (HIMVF), Izu Oshima, Miyakejima, and Aogashima have developed on the Izu-Bonin arc, Japan. Except for HIMVF and around the Izu Peninsula, they are basaltic polygenetic volcanoes. The Izu Peninsula is located on the northernmost tip of the Philippine Sea Plate. The stress condition varies on and around the Izu Peninsula. The differential stresses near Mount Fuji and Izu Oshima volcano are smaller than that of Higashi-Izu (1 mm/yr), judging from the compiled data of Katsuka and Imazumi [1984] and the Research Group for Active Faults of Japan (ROAF) [1991]. The differential stresses near Miyakejima and near Aogashima are smaller than those of Izu Oshima, judging from variations in focal mechanisms of shallow earthquakes [Nakamura, 1984] and from a low frequency of seismicity, respectively.

Output-Stress Diagram

Estimates of normalized output and differential stress (Table 1) for each region are compared on an output-stress diagram (Figure 1). Intraplate polygenetic volcanoes such as Kilauea and Mauna Loa are plotted in the region of both smaller deformation rate and larger output rate, that is, the lower right region of Figure 1. On the other hand, monogenetic volcanoes including lava fields and central (axial) volcanoes are plotted in the region of a small output rate, or on a rift trend, that is a trend with a positive gradient (stippled zone of Figure 1). The volcanic fields, defined as an independent group of monogenetic volcanoes [Nakamura, 1986], are plotted in the region of a small output rate (≤ 1 km$^3$/10$^3$ km$^2$). Monogenetic volcanoes in and around the Rhine graben are on the region of both a small output rate and an intermediate deformation rate. Lava fields, for example those of Iceland and
the Great Rift, are plotted on the region of both a large deformation rate and an intermediate output rate. Thus, monogenetic volcanoes of the Rhine graben, Iceland, and Great Rift are plotted on a rift trend. Monogenetic volcanoes in mid-oceanic ridges, for example, the East Pacific Rise and the Mid-Atlantic Ridge, are also on a rift trend.

Discussion

Input-Stress Diagram

For basaltic magma, the level of neutral buoyancy may exist in the shallow crust [e.g., Ryan, 1987a; Rubin and Pollard, 1987; Takada, 1989; Lister and Kerr, 1991]. Vesculation around the level of neutral buoyancy gives magma the additional buoyancy for eruption [e.g., Rubin and Pollard, 1987; Ida, 1990; Kazahaya and Shikohara, 1990]. According to this model, the output volume is controlled by vesculation processes above the level of neutral buoyancy. It is difficult to estimate the ratio of the output rate to input rate of magma (the output-input ratio) theoretically, but several observations may constrain this ratio.

The ratio of the thickness of the pillow lava layer to the thickness of the oceanic crust is equivalent to the output-input ratio. Based on ophiolitic sections of Moores and Jackson [1974] and Nicolas [1989], a ratio of about 0.1-0.3 is obtained. Based on seismic structures of the oceanic crust [Talwani et al., 1971; Moore et al., 1974; Christensen and Salisbury, 1975; Fowler, 1976; Nisbet and Fowler, 1978; Harding et al., 1989], a ratio of about 0.1-0.25 is obtained. In other areas, the supply rate of magma can be evaluated from surface deformations around volcanoes. The intrusive rate of Kiluaea is estimated to be 0.06 km$^2$/yr [Shaw, 1987]. The supply rate of the volcano is 0.1 km$^2$/yr in 1969 [Swanson, 1972], 0.086 km$^2$/yr during 1956-1983 [Shaw, 1987], or 0.13 km$^2$/yr during 1983-1985 [Shaw, 1987]. The growth rate of the volcano is 0.1 km$^3$/yr [Shaw, 1987]. Thus, the output-input ratio of Kiluaea is estimated to be about 0.3 [Shaw, 1987]. In Krafla, Iceland, the output-input ratio is estimated to be about 0.17 [Björnsson, 1985].

If the average output-input ratio of 0.1-0.3 is adopted in the ridge volcanism and Hawaiian volcanoes, we can convert the output-stress diagram into the input-stress diagram. However, the application to the other volcanoes is difficult due to lack of field observations. If the output-input ratio is smaller than 0.1, the relative position of each data point may change considerably after transformation from the output-stress diagram to the input-stress diagram, and the relative positions of monogenetic volcanoes and polygenetic volcanoes may not be maintained. If the output-input ratio is in the range of 0.1-1, the relative position of each data point will not change appreciably. In the input-stress diagram obtained by this operation, polygenetic volcanoes are plotted in the region of both smaller deformation rate and larger input rate. Monogenetic volcanoes are plotted in the region of a small input rate, or on a rift trend, where the balance between input rate and deformation rate is maintained.

When we convert crustal deformation rate into differential stress on the output-stress diagram or the input-stress diagram, we should note that all volcanic fields are shifted to lower stress states in Figure 1 because the dikes associated with these volcanoes work to accommodate stress the same way faults do [Parson and Thompson, 1991].

Coexistence of Monogenetic and Polygenetic Volcanoes

In the Eastern Volcanic Zone (60 km wide), Iceland, monogenetic and basaltic polygenetic volcanoes coexist (Figure 2a): fissure eruptions without polygenetic cones occurred in the Veldivötn-Laki area; Hekla volcano develops in the northwestern margin [e.g., Jakobsson, 1972, 1979]. A similar coexistence of volcano types exists in the Taupo Volcanic Zone, New Zealand (200-250 km long) (Figure 2b): basalt fissure eruptions without polygenetic cones occurred in the central rift (50 km wide); andesite polygenetic volcanoes develop in the northeastern and the southwestern ends of the rift (20 km wide) [e.g., Cole and Nairn, 1975; Cole, 1990]. The output rates from volcanoes or volcanic fields of Iceland are almost constant (Figure 1). In Iceland and New Zealand, it seems reasonable that these across-rift and/or along-rift variations in volcano type can be explained by variation in the differential stress with a constant input rate.

Monogenetic volcanoes and polygenetic volcanoes have coexisted since 2 Ma in the Cascade Range [Guiffaut and Weaver, 1988; Guiffaut et al., 1990]. For example, in the southern Cascade Range, the two types of volcanoes (basaltic to dacitic) develop together in an area of 30 km x 30 km during the last 0.12 m.y. (Figure 2c). There are no data available on variation of differential stress in this region. If differential stress is constant in this region, variation of input rate, that is, variation of production rate of magma in the asthenosphere, is needed to explain variation among volcanoes.

In the Colima graben and the Tepic-Zacoalco graben, the westernmost part of the TMVB, several andesitic polygenetic volcanoes have developed inside grabens [Allan, 1986]. For example, the output rates of Colima and Ceboruco volcanoes were estimated to be 27 and 60 km$^3$/10$^4$ yr/10$^3$ km$^2$, respectively [Luhr and Carmichael, 1980, 1982; Nelson, 1980]. The output rates of these polygenetic volcanoes are far larger than that of the Michoacán-Guanajuato volcanic field, in the western part of the TMVB, composed of monogenetic volcanoes [Hasenak and Carmichael, 1985]. Variation of input rate may exist in the mantle.

In the central part of the TMVB, monogenetic volcanoes and polygenetic volcanoes coexist in the Quaternary period; there is a regional shift from polygenetic volcanism in the north to monogenetic volcanism in the south; polygenetic volcanoes are generally more deeply eroded than monogenetic volcanoes [e.g., Connor, 1987, 1990]. For the lack of dating, the coexistence in the short time range is not discussed. For example, there is a tendency for polygenetic volcanoes to exist at the margin of the graben, or outside the graben, based on the distribution map (100°20'-100° W) of Suter et al. [1992] (Figure 2d). Thus, variation of differential stress as well as that of input rate may exist in this area.

In the Izu Peninsula and the adjacent area (60 km x 60 km), the Higashi Izu monogenetic volcano field (group) (0.0-0.15 Ma), the Higashi Izu oki submarine (monogenetic) volcanoes, and polygenetic volcanoes such as Izu Oshima volcano (0-0.025 Ma), Hakone volcano (0-0.25 Ma), and Fuji volcano (0-0.08 Ma) develop (Figure 2e) [e.g., Nakamura, 1964; Tsukui et al. 1986; Koyama and Umino, 1991]. According to the crustal deformation data of Kaizuka and Imaiatsu [1984] and RGAP [1991], variation of differential stress exist in this area. Moreover, there is a variation of input rates among these volcanoes.
Crack Interaction

The output-stress diagram provides empirical evidence that crack interaction theory can explain variation in volcano morphology. The crack plane is perpendicular to $\sigma_3$ (the minimum compressional principal stress) [Anderson, 1951]. According to the crack theory [Anderson, 1951], $\sigma_3$ is in the horizontal plane for ascent of a magma-filled crack. There are two arrangements of the principal stress axes satisfying this condition: one is that $\sigma_3$ is horizontal and $\sigma_1$ (the maximum compressional principal stress) is vertical, which includes the formation of normal faults; the other is that both $\sigma_3$ and $\sigma_1$ are in the horizontal plane, which includes the formation of strike-slip faults. The application of crack interaction to developments of a monogenetic volcanic field and a polygenetic volcano is discussed in light of these two stress field geometries.

Injecting dikes may interact with other liquid-filled cracks (ILL), as when dikes are intruded simultaneously, or with already cooled and solidified dikes, solid-filled cracks (ILS) [Takada, 1994b]. Numerical experiments indicate that only ILL is important for magma accumulation. Several conditions enhance crack coalescence. Buoyancy-driven cracks coalesce more easily than pressurized cracks; large differential stress prevents liquid-filled cracks from coalescing, whereas coalescence occurs easily at small differential stress; in addition, magma supply rate must be large enough for interaction of magma-filled dikes to be viable [Takada, 1994a] (Figure 3).

Individual polygenetic volcanoes occur in regions of smaller differential stress than lava fields on the continental rift and mid-oceanic ridges, and have larger magmatic output rates than most monogenetic volcanic fields (Figure 1). According to crack interaction theory, when differential stress in the vertical plane is smaller, cracks tend to coalesce easily, given a magmatic input rate large enough for cracks to interact (Figure 3, case 1). When input rate becomes larger, the tensile stress derived from regional differential stress cannot cancel the local compressional stress induced by magma accumulation. As a result, magma concentrates in space in its ascent process, so that a stable magma path may be formed. The above two cases provide the condition for the formation of polygenetic volcanoes. The theoretical model is consistent with the result of the output-stress (input-stress) diagram. In the natural system, differential stress may vary with depth (Figure 3, case 2). In this case, magma-filled cracks can coalesce in the lower lithosphere.

Monogenetic volcanoes are plotted on a rift trend. Applying crack interaction theory, at larger differential stress, cracks do not coalesce easily. This makes the formation of monogenetic volcanoes more likely (Figure 3, case 3). This condition is equivalent to the condition of tensile stress proposed by Nakamura [1986]. Nakamura [1986] proposed that monogenetic volcanism occurs in regions with crustal stress fields that are more tensional. The tensional stress field is one with large differential stress. On a rift trend, the balance between stresses produced by magma-filled cracks and stresses produced by crustal extension is maintained. Consequently, the local compressional stress does not increase with time. Large monogenetic volcanic fields also occur in areas with small magma supply rate [Fedotov, 1981]. According to crack interaction theory, at a small input rate, monogenetic volcanoes will also be formed whether differential stress is large or small (Figure 3). This is because the interaction of two liquid-filled cracks is unlikely at a small magmatic input rate. This condition is equivalent to that for the formation of monogenetic volcanoes of Fedotov [1981], based on thermodynamic arguments. The results of the output-stress (input-stress) diagram support the theoretical model on the formation of monogenetic volcanoes. Moreover, it removes the discrepancy between the condition proposed by Nakamura [1986] and that by Fedotov [1981].

In the above discussion, it is assumed that magma can rise without pause in the lithosphere. However, basaltic magma is commonly trapped on the way to the surface [Ryan, 1987a,b; Rubin and Pollard, 1987; Takada, 1989; Lister and Kerr, 1991]. The vertical direction of crack propagation is controlled by the driving force, $dp/dz$, composed of both the density difference between magma and the host rocks, and the stress gradient [Takada, 1989] (Figure 4); $dp/dz$ is

$$\frac{dp}{dz} = -(\Delta p + d\sigma/dz) \tag{4}$$

where $P_o$, $\Delta p$, $z$, and $\sigma_b$ are driving pressure of a liquid-filled crack, density differences between magma and the host rocks, depth, and compressional stress perpendicular to the vertical crack, respectively. At the depth satisfying the condition $dp/dz = 0$ (the level of neutral buoyancy), the crack cannot propagate upward [Ryan, 1987a,b; Rubin and Pollard, 1987;]
Takada, 1989, 1990; Lister and Kerr, 1991) (case 1 in Figure 4). In this instance, a magma chamber composed of a dike-sheet complex will be formed at depth. The development of parallel dike swarms under the mid-oceanic ridge suggests that the condition of $dP/dz = 0$ at any depth is satisfied for some magma (case 4 in Figure 4). On the other hand, when a stress gradient is superimposed, or when the average density in a crack decreases owing to vesiculation of volatile component ($dP/dz > 0$; case 2 of Figure 4), magma-filled cracks even at a small input rate can ascend directly toward the surface [Takada, 1989]. This leads to the formation of monogenetic volcanoes, and the transport of primary magma to the surface.

Stress Balance and Stress Relaxation

Dike intrusion deforms the host rocks around a dike. However, the stress resulting from this deformation will be relaxed in the long-term. The process of stress relaxation can weaken the effect of crack interaction. There is a difference in the stress relaxation process between polygenetic and monogenetic volcanoes.

At polygenetic volcanoes, which are plotted in the region of large input rate and small differential stress, the local compressional stress increases in the volcanic edifice with magma accumulation. Because of local increase in compressional stress, stress relaxation must occur in polygenetic volcanoes. The relaxation process can take place through the development of a dike swarm [Rubin, 1990; Takada, 1994b]. The long-term stress in the volcanic edifice, $\sigma$, is

$$\sigma = -\sigma_R + \sigma_L - \sigma_X$$

where $\sigma_R$, $\sigma_L$, and $\sigma_X$ are regional differential stress, local differential stress induced by cracks, and relaxed stress, respectively. The compressional stress is positive. If the local stress caused by magma accumulation in the volcanic edifice becomes larger than the regional stress, or if the relaxation process does not operate efficiently ($\sigma = -\sigma_R + \sigma_L - \sigma_X > 0$), radial dikes develop. However, efficient stress relaxation can cause the development of local rift zones in the volcanic edifice. For example, in Kiluaea and Mauna Loa, local balance is kept by (5), because the regional stress field around the Hawaiian volcanic chain is nearly hydrostatic ($\sigma = \sigma_L - \sigma_X = 0$; $\sigma_R = 0$). This local balance seems to be similar to the regional one on a rift trend in the output-stress (input-stress) diagram ($\sigma = -\sigma_R + \sigma_L = 0$; $\sigma_X = 0$). The dike emplacement in the Hawaiian rift zone occurs from slip of deep faults beneath the volcanic edifice [e.g., Nakamura, 1980; Dietrich, 1988]. Huge landslides in Piton de la Fournaise, La Reunion, and Kiluaea, Hawaii, are activated by a combination of stress due to the forceful emplacement of magma and the gravitational loading of the volcanic edifice [Duffield et al., 1982]. The gravitational spreading controls the growth of the volcanic edifice including a rift zone and a magma reservoir in Mount Etna [Borgia et al., 1992].

Magma trap: 

Abrupt decrease of $dP/dz$

vs: Vesiculation

Figure 4. Schematic diagram of magmatic system by crack propagation: $dP/dz$-depth diagram (left). Magma is trapped at the level of $dP/dz = 0$ (level of neutral buoyancy (LNB)), or at the level where $dP/dz$ decreases abruptly [e.g., Takada, 1989]. Vesiculation around LNB gives magma the additional buoyancy for eruption. 1a: case of hydrostatic stress condition; 1b and 1c: superposition of stress gradient on the hydrostatic condition can shift LNB; 2-4: cases of superposition of stress gradient; 2: case that the extensional stress increases upward; magma never stops in the lithosphere; 3: case that the compressional stress increases upward; magma never ascends in the lithosphere; 4: the extreme extensional stress condition that any level of the lithosphere is LNB.
The stress relaxation process is not important in a monogenetic volcano field. The local compressional stress does not increase with magma accumulation beneath monogenetic volcanoes plotted in the region of a small input, because the stress induced by dike intrusions at a low frequency is relaxed easily ($\sigma = -\sigma_R + \sigma_L \rightarrow \sigma_x = 0$). On a rift trend, the relaxation process is not needed ($\sigma = -\sigma_R + \sigma_L = 0$; $\sigma_x = 0$). A rift trend means that magmatic input rate and regional stress are closely linked together. To give a simple example, input rate increases on the mid-oceanic ridge as spreading rate becomes larger. This is because a rift trend has a positive gradient in the output-stress diagram. Magma intrudes into the vertical space in the horizontally extended crust to form dikes. The integrated system indicating this process is observed in ophiolitic sections. However, the uppermost oceanic crust composed of feeder dikes and pillow lavas may be deformed by dike intrusions and normal faulting. The surface stress varies with spreading rate [Parson and Thompson, 1991].

Conclusions
An output-stress diagram provides a means of explaining the change in volcano morphology with changes in tectonic setting and magmatic production rates. Crustal deformation rate in a normalized area of 30 km x 30 km, which is converted into differential stress ($\Delta \sigma$), is selected as the stress axis. Magmatic output rate is normalized for $10^4$ years and over an area of $10^3$ km$^2$.

Polygenetic volcanoes are located in regions of both larger output rate and smaller $\Delta \sigma$. Monogenetic volcanoes are plotted on a trend, or in the region of a small output ($\leq 1$ km$^3$) $10^4$ yr/10$^3$ km$^2$.

In the mid-oceanic ridges and Hawaiian volcanoes, an input-stress diagram may be obtained, based on the data of the output-input ratio, which is estimated from ophiolitic sections and crustal deformation around volcanoes.

The coexistence of polygenetic and monogenetic volcanoes in areas of 30 km x 30 km and during 0.1 m.y. periods indicates either variation of differential stress or variation of magmatic output (input) rate.

The output-stress diagram supports the hypothesis that variation among monogenetic and polygenetic volcanoes can be explained by a crack interaction model, governed by the input rate of magma-filled cracks and differential stress.

There is a difference in stress relaxation process between polygenetic and monogenetic volcanoes. The output-stress diagram indicates that the stress balance between the local stress induced by magma accumulation and the regional stress, and the stress relaxation, govern the structure of a volcano.

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References

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Nisbet, E. G., and C. M. R. Fowler, The Mid-Atlantic Ridge at 37 and


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