JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 99, NO. B7, PAGES 13,563–13,573, JULY 10, 1994

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 10^3 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 volcanoes 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 (≤ 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.

Intoduction

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

Paper number 94JB00494.
0148-0227/94/94JB-00494$05.00

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 volcano fields have low resolution, of the order of 10^6 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 volcano fields are far larger than the area of 30 km x 30 km. In that case, the output of a monogenetic volcano 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_0, is given as

$$M_0 = h_p\, a_{sp}\, l \tag{1}$$

where is h_p 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

$$\Delta\sigma = k\,\varepsilon \tag{2}$$

$$\Delta\sigma = \eta\; d\varepsilon/dt \tag{3}$$

for short term and long term, respectively, where $\Delta\sigma$, ε, k, η, and t 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 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 accommodated 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 volcano 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 Rèunion in the Indian Ocean [*Stieltjes and Moutou*, 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*

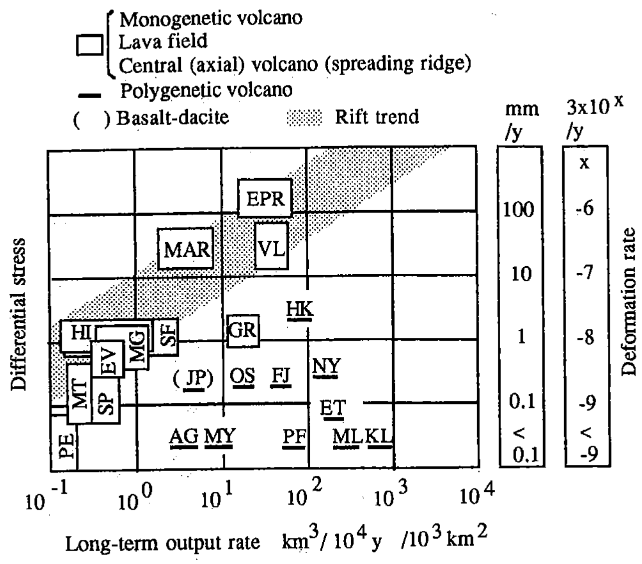
Table 1. Normalized Output Rate and Crustal Deformation (Strain, Spreading) Rate

Volcano or Volcanic Field	Type of Volcano	Total Output, km^3	Estimated Area, km^2	Duration of Output, years	Normalized Output Rate, km^3 $/10^4$ yr $/10^3$ km^2	Deformation Rate, mm/yr or Strain Rate, /yr	Duration of Deformation, years	Normalized Crustal Deformation, mm/yr	References
East Pacific Rise	C				15-48	Sp 100-160	present	100-160	*Harding et al.* [1989]
Mid-Atlantic Ridge	C				1.8-4.8	Sp 20	present	20	*Moore et al.* [1974]
Veidivö'tn-Laki Iceland	L	Lv144	3200	10^4	45	Sp 20-30	present	20-30	*Jakobsson* [1972]
Hekla	P			1693-1970	90	< 20-30	present	1-10	*Wadge* [1982]
Eifel VF, Rhine graben	M	Tv40	1000	$7x10^5$	0.6	Sb 0.5-0.7	Quaternary	0.5	*Schmincke et al.* [1990], *Prodehl et al.* [1992]
Etna	P			1600-1980	90-210	Up 0.071-0.12 Up 0.01-0.2	$3x10^5$(0.7-0.4Ma) $1.7x10^5$(0.17-0Ma)	≤ 0.1	*Wadge* [1982], *Brogan et al.* [1975]
Piton de la Fournaise	P			1930-1985	90	hydrostatic	present	< 0.1	*Stieltjes and Moutou* [1989]
Nyamuragira	P			1900-1980	105	Sp < 0.5, <1-2	present	≤ 0.5	*Wadge* [1982], *Baker and Wohlenberg* [1971] *Wohlenberg* [1976]
Kilauea	P			1920-1983	250-1000	hydrostatic	present	< 0.1	*Shaw* [1987]
Mauna Loa	P			1840-1980	250	hydrostatic	present	< 0.1	*Shaw* [1987]
Posterosional volcanism of Hawaii	M	≤ 1	1000	$> 10^6$	< 0.1	hydrostatic	present	< 0.1	*Macdonald et al.* [1983], *Nakamura* [1986]
San Francisco VF	M	0.75-14	4800	$4.75x10^6$ (5-0.25Ma)	0.16-3	SlEq 0.38-1* SlEq 0.23-7.4† St 10^{-16}-10^{-17}†	Quaternary Quaternary present	1	*Tanaka et al.* [1986], *Eddington et al.* [1987]
		1.8	4800	$2.5x10^5$ (0.25-0Ma)	0.38				
Mt. Taylor VF	M	100	1000	$4x10^6$ (4-0Ma)	0.25	weak extension	Quaternary	< 1	*Perry et al.* [1990]
Springerville VF	M	300	3000	$1.8x10^6$ (2.1-0.3Ma)	0.6	extension -hydrostatic	Quaternary	< 1	*Condit et al.* [1989], *Connor et al.*[1992]
Great Rift	L	30	1600	$1.5x10^4$	12.5	0.08-2 St $3x10^{-16}$, 10^{-17}	Quaternary Present	1	*Kuntz et al.* [1986], *Eddington et al.* [1987]
Michoacán	M	31	15000	$4x10^4$	0.5	Sl 1	Late Quaternary	1	*Hasenaka and Carmichael* [1985]
-Guanajuato VF		0.12	1250	10^3	1				*Suter et al.* [1992]
Higashi Izu	M		1000	$1.5x10^5$	0.1	Sl 1	Quaternary	1	*Koyama and Umino* [1991]
Monogenetic VF				$1x10^4$	2				*Hayakawa and Koyama* [1992], Ref.1
Izu Oshima	P			$2.5x10^4$	18-30	Sl 0.1	Quaternary	0.1	*Nakamura* [1964], *Tsukui et al.* [1986], Ref.1
Fuji	P			$8x10^4$	50	Sl 0.1	Quaternary	0.1	*Tsukui et al.* [1986], *Miyaji* [1988], Ref.1
Miyakejima	P			$3x10^3$	8	Sl < 0.1	Quaternary	< 0.1	*Isshiki* [1977]
Aogashima	P			$3x10^3$	3	Sl < 0.1	Quaternary	< 0.1	*Takada et al.* [1992]
Other volcanoes, Japan	P			$1x10^5$	2-50	Sl 0.01-1	Quaternary	0.01-1	*Moriya* [1983], *Tsukui et al.* [1986], Ref.1

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, *Kaizuka and Imaizumi* [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.



AG, Aogashima, Japan
EPR, East Pacific Rise
ET, Etna, Italy
EV, Eifel VF, Rhine graben, Germany
FJ, Fuji, Japan
GR, Great Rift, Idaho
HI, Higashi-Izu VF, Japan
HK, Hekla, Iceland
(JP, Poly. volc., Japan)
KL, Kilauea, Hawaii
MAR, Mid-Atlantic Ridge
MG, Michoacàn Guanajuato VF, Mexico
ML, Mauna Loa, Hawaii
MT, Mt. Taylor VF, New Mexico
MY, Miyakejima, Japan
NY, Nyamuragira, Zaire
OS, Izu Oshima, Japan
PE, Posterosional volcanism, Hawaii
PF, Piton de la Fournaise, Reunion
SF, San Francisco VF, Arizona
SP, Springerville VF, Arizona
VL, Veidivötn-Laki, Iceland

Figure 1. Output-stress diagram. The data of each volcano or volcanic field are shown in Table 1. Deformation rate includes strain rate and spreading rate.

Nativel, 1989]), the output rate of the volcano for 1930-1985, estimated by *Stieltjes 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/ 10^4 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 Luedke*, 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 [*Eddington 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 [*Eddington et al.*, 1987] are useful to estimate the deformation rate near the Great Rift.

The Michoacàn-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 on 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 *Kaizuka and Imaizumi* [1984] and *the Research Group for Active Faults of Japan (RGAF)* [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/ 10^4 yr/ 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, b; *Rubin and Pollard*, 1987; *Takada*, 1989; *Lister and Kerr*, 1991]. Vesiculation around the level of neutral buoyancy gives magma the additional buoyancy for eruption [e. g., *Rubin and Pollard*, 1987; *Ida*, 1990; *Kazahaya and Shinohara*, 1990]. According to this model, the output volume is controlled by vesiculation 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 ophiolite 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 Kilauea is estimated to be 0.06 km^3/ yr [*Shaw*, 1987]. The supply rate of the volcano is 0.1 km^3/ yr in 1969 [*Swanson*, 1972], 0.086 km^3/ yr during 1956-1983 [*Shaw*, 1987], or 0.13 km^3/ 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 Kilauea 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

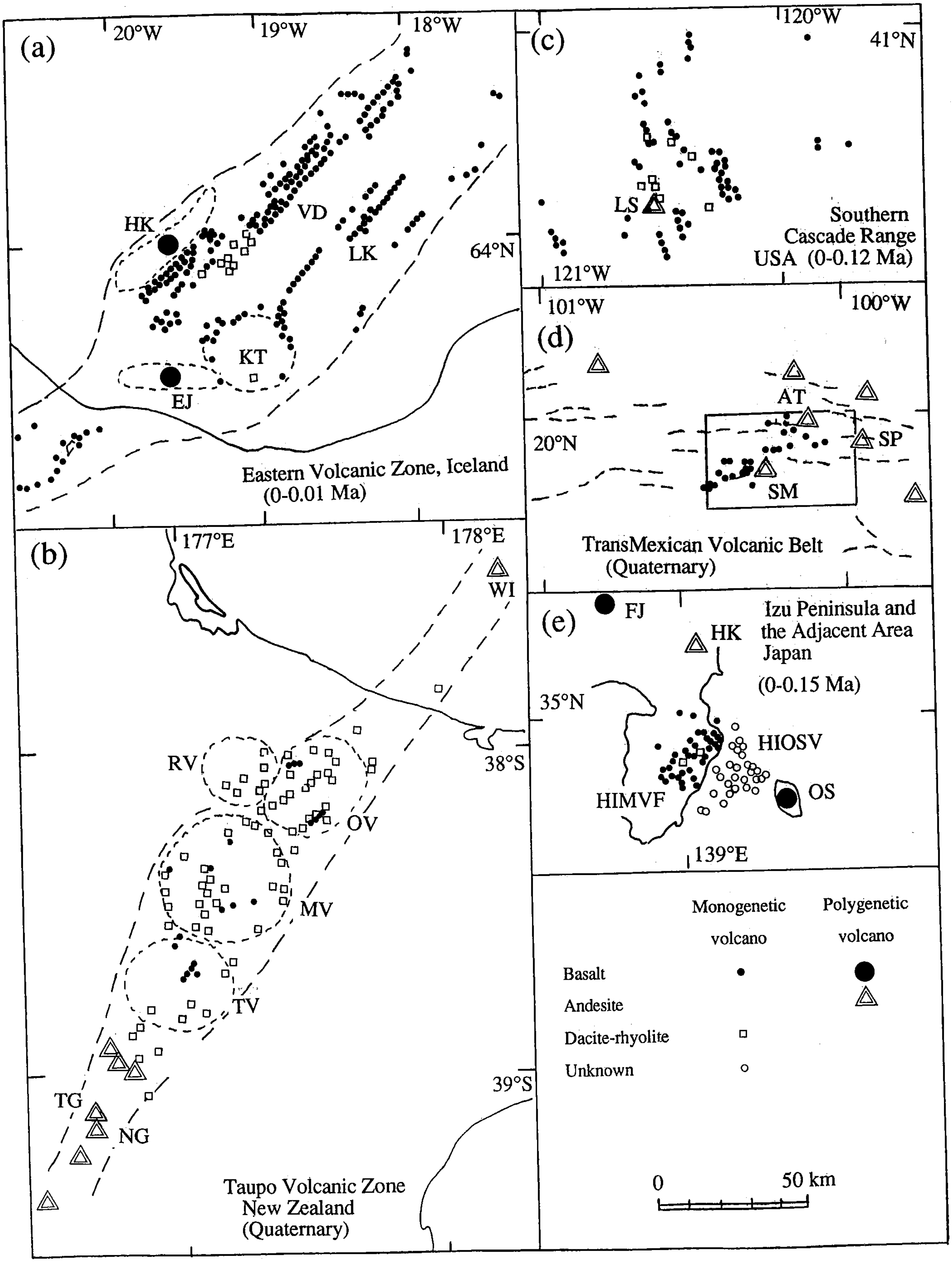
In the Eastern Volcanic Zone (60 km wide), Iceland, monogenetic and basaltic polygenetic volcanoes coexist (Figure 2a): fissure eruptions without polygenetic cones occurred in the Veidivötn-Laki area; Hekla volcano develops in the northwestern margin [e.g., *Jakobsson*, 1972, 1979]. A similar coexistence 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); andesitic 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 [*Guffanti and Weaver*, 1988; *Guffanti 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 [*Hasenaka and Carmichel*, 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°30'-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) (0-0.15 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 Imaizumi* [1984] and *RGAF* [1991], variation of differential stress exist in this area. Moreover, there is a variation of input rates among those volcanoes.


(a)
20°W
19°W
18°W
64°N
HK
VD
LK
KT
EJ
Eastern Volcanic Zone, Iceland
(0-0.01 Ma)
(b)
177°E
178°E
WI
38°S
RV
OV
MV
TV
TG
NG
39°S
Taupo Volcanic Zone
New Zealand
(Quaternary)
(c)
120°W
41°N
121°W
LS
Southern
Cascade Range
USA (0-0.12 Ma)
(d)
101°W
100°W
20°N
AT
SP
SM
TransMexican Volcanic Belt
(Quaternary)
(e)
FJ
HK
Izu Peninsula and
the Adjacent Area
Japan
(0-0.15 Ma)
35°N
HIOSV
HIMVF
OS
139°E
Monogenetic volcano
Polygenetic volcano
Basalt
Andesite
Dacite-rhyolite
Unknown
0
50 km

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 σ_3 (the minimum compressional principal stress)[*Anderson*, 1951]. According to the crack theory [*Anderson*, 1951], σ_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 σ_3 is horizontal and σ_1 (the maximum compressional principal stress) is vertical, which includes the formation of normal faults; the other is that both σ_3 and σ_1 are in the horizontal plane, which includes the formation of strike-slip faults. The application of crack interaction to developments of a monogenetic volcano 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 two 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 volcano 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 volcano 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_e/dz , composed of both the density difference between magma and the host rocks, and the stress gradient [*Takada*, 1989] (Figure 4) ; dP_e/dz is

$$dP_e/dz = -(\Delta\rho + d\sigma_h/dz) \tag{4}$$

where P_e, $\Delta\rho$, z, and σ_h 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 $dP_e/dz=0$ (the level of neutral buoyancy), the crack cannot propagate upward [*Ryan*, 1987a,b; *Rubin and Pollard*, 1987;

Figure 2. Distributions of monogenetic volcanoes and polygenetic volcanoes. One symbol is equivalent to one vent of a monogenetic volcano, including large fissures. "Basalt" includes basaltic andesite. (a) Eastern Volcanic Zone, Iceland, in postglacial time (simplified after *Jakobsson* [1972, 1979]). HK, Hekla; VD, Veidivötn; LK, Lakagigar; KT, Katra; EJ, Eyjafjöll. Vent sites of parasitic cones on the flanks of HK and EJ are omitted. (b) Taupo Volcanic Zone, New Zealand (simplified after *Cole and Nairn* [1975], and *Cole* [1990]). TV, Taupo volcanic center; MV, Maroa volcanic center; RV, Rotorua volcanic center; OV, Okataina volcanic center; WI, White Island; TG, Tongariro; NG, Ngauruhoe. TV, MV, RV, and OV are felsic volcanic centers associated with caldera formation. (c) Southern Cascade Range, United States, for 0-0.12 Ma (simplified after *Guffanti et al.* [1990]). LS, Lassen. (d) TransMexican Volcanic Field (Quaternary) (simplified after *Suter et al.* [1992]). SM, San Miguel; AT, Altamirano; SP, San Pedra. Cinder cones and stratovolcanoes are assumed to be monogenetic volcanoes of basalt and polygenetic volcanoes of andesite, respectively, based on the descriptions of *Suter et al.* [1992]. The data of monogenetic volcanoes are available only the rectangle area. (e) Izu Peninsula and the adjacent area, Japan, including Higashi Izu monogenetic volcano field for 0-0.15 Ma (simplified after *Koyama and Umino* [1991]). HIMVF, Higashi Izu monogenetic volcano field on the Izu Peninsula; HIOSV, Higashi Izu oki submarine volcanoes between the Izu Peninsula and Izu Oshima; FJ, Fuji; HK, Hakone; OS, Izu Oshima.

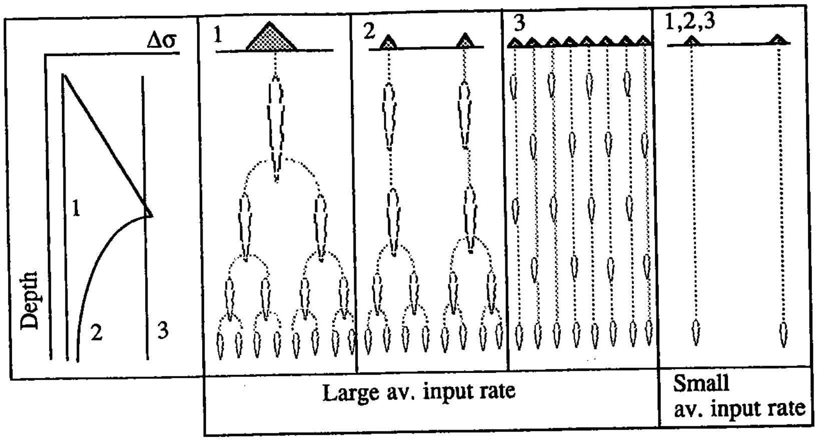


Figure 3. Schematic diagram of magmatic system by crack interaction. (Left) the regional differential stress ($\Delta\sigma$)-depth diagram. 1: case of small $\Delta\sigma$. 2: case that $\Delta\sigma$ varies with depth. 3: case of large $\Delta\sigma$.

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 $dP_e/dz = 0$ 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 ($dP_e/dz > 0$; 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, σ, is

$$\sigma = -\sigma_R + \sigma_L - \sigma_X \qquad (5)$$

where σ_R, σ_L, and σ_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 Kilauea 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 Rèunion, and Kilauea, 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].

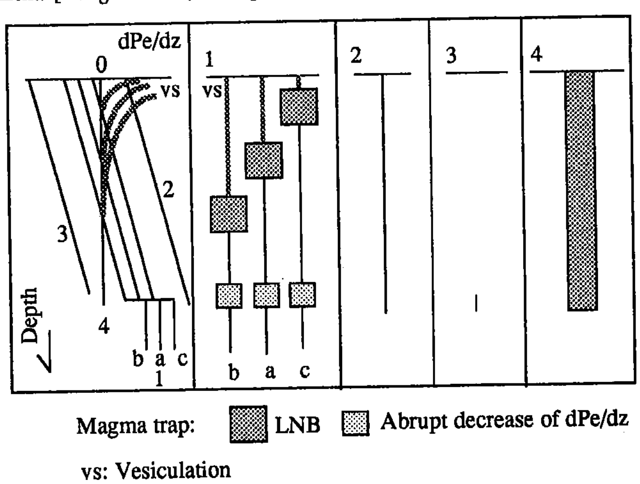


Figure 4. Schematic diagram of magmatic system by crack propagation: *dPe/dz*-depth diagram (left). Magma is trapped at the level of *dPe/dz* = 0 (level of neutral buoyancy (LNB)), or at the level where *dPe/dz* decreases abruptly [e.g.,*Takada*, 1989]. Vesiculation around LNB gives magma the additional 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; màgma 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 ($\sigma = -\sigma_R + \sigma_L - \sigma_X = 0$). 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 ophiolite 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 magma 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 (≤ 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 ophiolite 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.

Acknowledgments. The author thanks C. B. Connor of the Center for Nuclear Waste Regulatory Analyses, and J. P. Lockwood of USGS for valuable comments and critical readings of this manuscript. The author also appreciates constructive reviews by referees. The author benefited from discussions with M. Takahashi of Ibaraki University, and Y. Kobayashi of Tsukuba University.

References

Allan, J. F., Geology of the northern Colima and Zacoalco grabens, southwest Mexico: Late Cenozoic rifting in the Mexican volcanic belt, *Geol. Soc. Am. Bull.*, *97*, 473-485, 1986.

Anderson, E.M., *The Dynamics of Faulting and Dyke Formation*, 2nd ed., 206 pp., Oliver and Boyd, Edinburgh, 1951.

Baker, B. H., and J. Wohlenberg, Structure and evolution of the Kenya rift valley, *Nature*, *229*, 538-542, 1971.

Björnsson, A., Dynamics of crustal rifting in NE Iceland, *J. Geophys. Res.*, *90*, 10,151-10,162, 1985.

Borgia, A., L. Ferrari, and G. Pasquare, Importance of gravitational spreading in the tectonic and volcanic evolution of Mount Etna, *Nature*, *357*, 231-235, 1992.

Brogan, G. E., L. S. Cluff, and C. L. Taylar, Seismicity and uplift of southern Italy, *Tectonophysics*, *29*, 323-330, 1975.

Christensen, N. I., and M. H. Salisbury, Structure and constitution of the lower oceanic crust, *Rev. Geophys.*, *13*, 57-86, 1975.

Cole, J. W., Structural control and origin of volcanism in the Taupo volcanic zone, New Zealand, *Bull. Volcanol.*, *52*, 445-459, 1990.

Cole, J. W., and I. A. Nairn, *Catalogue of the Active Volcanoes of the World Including Solfatara Fields, Part 22: New Zealand*, 156 pp., International Association of Volcanology and Chemistry of the Earth's Interior, Naples, 1975.

Condit, C. D., L. S. Crumpler, J. C. Aubele, and W. E. Elston, Patterns of volcanism along the southern margin of the Colorado Plateau: The Springerville field, *J. Geophys. Res.*, *94*, 7975-7986, 1989.

Connor, C. B., Structure of the Michoacàn-Guanajuato volcanic field, Mexico, *J. Volcanol. Geotherm. Res.*, *33*, 191-200, 1987.

Connor, C. B., Cinder cone clustering in the TransMexican Volcanic Belt: Implication for structural and petrologic models, *J. Geophys. Res.*, *95*, 19,395-19,405, 1990.

Connor, C. B., C. D. Condit, L. S. Crumpler, and J. C. Aubele, Evidence of regional structural controls on vent distribution: Springville Volcanic Field, Arizona, *J. Geophys. Res.*, *97*, 12,349-12,359, 1992.

Dietrich, J. H., Growth and persistence of Hawaiian volcanic rift zones, *J. Geophys. Res.*, *93*, 4258-4270, 1988.

Duffield, W. A., L. Stieltjes, and J. Varet, Huge landslide blocks in the growth of Piton de la Fournaise, La Rèunion, and Kilauea volcano, Hawaii, *J. Volcanol. Geotherm. Res.*, *12*, 147-160, 1982.

Eddington, P. K., R. B. Smith, and C. Renggli, Kinematics of Basin and Range interplate extension, in Continental Extension Tectonics, edited by M. P. Coward, J. F. Dewey, and P. L. Hancock, *Geol. Soc. Spec. Publ.*, *28*, 371-392, 1987.

Fedotov, S.A., Magma rate in feeding conduits of different volcanic centers, *J. Volcanol. Geotherm. Res.*, *9*, 379-394, 1981.

Fowler, C. M. R., Crustal structure of the Mid-Atlantic ridge crest at 37°N, *Geophys. J. R. Astron. Soc.*, *47*, 459-491, 1976.

Gillot, P.-Y., and P. Nativel, Eruptive history of the Piton de la Fournaise volcano, Rèunion Island, Indian Ocean, *J. Volcanol. Geotherm. Res.*, *36*, 53-65, 1989.

Guffanti, M., and C. S. Weaver, Distribution of late Cenozoic volcanic vents in the Cascade Range: Volcanic arc segmentation and regional tectonic considerations, *J. Geophys. Res.*, *93*, 6513-6529, 1988.

Guffanti, M., M. A. Clynne, J. G. Smith, L. J. P. Muffler, and T. D. Bullen, Late Cenozoic volcanism, subduction, and extension in the Lassen region of California, southern Cascade Range, *J. Geophys. Res.*, *95*, 19,453-19,464, 1990.

Harding, A. J., J. A. Orcutt, M. E. Kappus, E. E. Vera, J. C. Mutter, P. Buhl, R. S. Detrick, and T. M. Brocher, Structure of young oceanic crust at 13°N on the East Pacific Rise from expanding spreading profiles, *J. Geophys. Res.*, *94*, 12,163-12,196, 1989.

Hasenaka, T., and I. S. E. Carmichael, The cinder cones of Michoacàn-Guanajuato, central Mexico: Their age, volume and distribution, and magma discharge rate rate, *J. Volcanol. Geotherm. Res.*, *25*, 105-124, 1985.

Hayakawa, Y., and M. Koyama, Eruptive history of the Higashi Izu Monogenetic volcano Field 1: 0-32 ka (in Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, *37*, 167-181, 1992.

Hildreth, W., Gradients in silicic magma chambers: Implications for lithospheric magmatism, *J. Geophys. Res.*, *86*, 10,153-10,192, 1981.

Ida, Y., Condition to start volcanic eruption (Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, *35*, 299-312, 1990.

Isshiki, N., Activity of Miyake-jima volcano during the last 3000 years (in Japanese), *Bull. Volcanol. Soc. Jpn.*, *22*, 290, 1977.

Jakobsson, S. P., Chemistry and distribution pattern of recent basaltic rocks in Iceland, *Lithos*, *5*, 365-386, 1972.

Jakobsson, S. P., Petrology of Recent basalts of the Eastern Volcanic Zone, Iceland, *Acta Nat. Isl.*, *26*, 1-103, 1979.

Kaizuka, S., and T. Imaizumi, Horizontal strain rates of the Japanese Islands estimated from Quaternary fault data, *Geogr. Rep. Tokyo Metrop. Univ.*, *19*, 43-65, 1984.

Kazahaya, K., and H. Shinohara, A simple model on long-term magmatic process in relation to growth of volcanoes, paper presented at meeting of Int. Assoc. of Volcanol. and Chem. of the Earth's Interior, Mainz, 1990.

Koyama, M., and S. Umino, Why does the Higashi-Izu monogenetic volcano group exist in the Izu-Peninsula?: Relationship between late Quaternary volcanism and tectonics in the northern tip of the Izu-Bonin arc, *J. Phys. Earth*, *39*, 391-420, 1991.

Kuntz, M. A., D. E. Champion, E. C. Spiker, and R. H. Lefebvre, Contrasting magma types and steady-state, volume-predictable, basaltic volcanism along the Great Rift, Idaho, *Geol. Soc. Am. Bull.*, *97*, 579-594, 1986.

Lister, J.R., and R. C. Kerr, Fluid-mechanical model of crack propagation and their application to magma transport in dykes, *J. Geophys. Res.*, *96*, 10,049-10,077, 1991.

Luhr, J. F., and I. S. E. Carmichael, The Colima Volcanic complex, Mexico, Part I, Post-caldera andesite from Volcà Colima, *Contrib. Mineral. Petrol.*, *71*, 343-372, 1980.

Luhr, J. F., and I. S. E. Carmichael, The Colima Volcanic complex, Mexico, Part III, Ash- and scoria-fall deposits from the upper slopes of Volcà Colima, *Contrib. Mineral. Petrol.*, *80*, 262-275, 1982.

Macdonald, G. A., *Volcanoes*, 510 pp., Prentice-Hall, Englewood Cliffs, N. J., 1972.

Macdonald, G. A., A. T. Abbott, and F. L. Peterson, *Volcanoes in the Sea*, 517 pp., University of Hawaii Press, Honolulu, 1983.

Macdonald, K. C., Tectonic and magmatic process on the East Pacific Rise, in *The Eastern Pacific Ocean and Hawaii*, *The Geology of North America*, *vol. N*, edited by E. L. Winterer, D. M. Hussong, and R. W. Decker, pp. 93-110, Geological Society of America, Boulder, Colo., 1989.

Miyaji, N., History of Younger Fuji volcano (in Japanese with English abstract), *J. Geol. Soc. Jpn.*, *94*, 433-452, 1988.

Moore, J. G., H. S. Fleming, and J. D. Phillips, Preliminary model for extrusion and rifting at the axis of the Mid-Atlantic Ridge, 36°48' North, *Geology*, *2*, 437-440, 1974.

Moores, E. M., and E. D. Jackson, Ophiolites and oceanic crust, *Nature*, *250*, 136-139, 1974.

Moriya, I., *Morphology of a Volcano in Japan* (in Japanese), 135 pp., University of Tokyo Press, 1983.

Nakamura, K., Volcano-stratigraphic study of Oshima volcano, Izu, *Bull. Earthquake Res. Inst. Univ. Tokyo*, *42*, 649-728, 1964.

Nakamura, K., Volcano structure and possible mechanical correlation between volcanic eruptions and earthquakes (in Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, *20*, *Spec. Issue*, S229-S240, 1975.

Nakamura, K. Volcanoes as possible indicators of tectonic stress orientation, *J. Volcanol. Geotherm. Res.*, *2*, 1-16, 1977.

Nakamura, K., Why do long rift zones develop in Hawaiian volcanoes- A possible role of thick oceanic sediments- (in Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, 25, 255-269, 1980.

Nakamura, K., Distribution of flank craters of Miyake-jima volcano and the nature of the ambient crustal stress field (in Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, 29, *Spec. Issue*, S16-S23, 1984.

Nakamura, K., Volcanology and the plate tectonics (in Japanese with English abstract), *Bull. Volcanol. Soc. Jpn.*, *30*, *Spec. Issue*, S1-S16, 1986.

Nelson, S. A., Geology and petrology of Volcà Ceboruco, Nayarit, Mexico, *Geol. Soc. Am. Bull.*, *Part II*, *91*, 2290-2431, 1980.

Nicolas, A., *Structures of Ophiolites and Dynamics of Oceanic Lithosphere*, 367 pp., Kluwer Academic, Norwell, Mass., 1989.

Nisbet, E. G., and C. M. R. Fowler, The Mid-Atlantic Ridge at 37 and 45°N: some geophysical and petrological constraints, *Geophys. J. R. Astron. Soc.*, *54*, 631-660, 1978.

Orcutt, J. A., B. L. N. Kennett, and L. M. Dorman, Structure of the East Pacific Rise from an ocean bottom seismometer survey, *Geophys. J. R. Astron. Soc.*, *45*, 305-320, 1976.

Parson, T., and G. A. Thompson, The role of magma overpressure in suppressing earthquakes and topography: Worldwide examples, *Science*, *253*, 1399-1402, 1991.

Perry, F. V., W. S. Baldridge, D. DePaolo, and M. Shafiqullah, Evolution of a magmatic system during continental extension: The Mount Tayor Volcanic Field, New Mexico, *J. Geophys. Res.*, *95*, 19,327-19,348, 1990.

Prodehl, C., St. Mueller, A. Glahn, M. Gutcher, and V. Haak, Lithospheric cross sections of the European Cenozoic rift system, *Tectonophysics*, *208*, 113-138, 1992.

Research Group for Active Faults of Japan (RGAF), *Active Faults in Japan, Sheet Maps and Inventories* (in Japanese with English abstract), rev. ed., 436 pp., University of Tokyo Press, 1991.

Rubin, A. M., A comparison of rift-zone tectonics in Iceland and Hawaii, *Bull. Volcanol.*, *52*, 302-319, 1990.

Rubin, A.M., and D. D. Pollard, Origins of blade-like dikes in volcanic rift zones, in Volcanism in Hawaii, edited by R. W. Decker, T. L. Wright, and P. Stauffer, *U. S. Geol. Surv. Prof. Pap.*, *1350*, 1449-1470, 1987.

Ryan, M. P., The elasticity and contractancy of Hawaiian olivine tholeiite, and its role in the stability and structural evolution of sub-caldera magma reservoirs and rift systems, in Volcanism in Hawaii, edited by R. W. Decker, T. L. Wright, and P. Stauffer, *U. S. Geol. Surv. Prof. Pap.*, *1350*, 1395-1447, 1987a.

Ryan, M. P., Neutral buoyancy and the mechanical evolution of magmatic systems, in Magmatic Process: Physicochemical Principles, edited by B.O. Mysen, *Geochem. Soc. Spec. Publ. 1*, 259-287, 1987b.

Schilling, J.-G., P. S. Meyer, and R. H. Kingsley, Evolution of the Icelandic hot spot, *Nature*, *296*, 313-320, 1982.

Schmincke, H.-U., P. v. d. Bogaard, and A. Freundt, *Quaternary Eifel Volcanism*, *Guidebook of Excursion 1AI*, *Workshop on explosive volcanism*, 188 pp., International Association of Volcanology and Chemistry of the Earth's Interior, Rome, Italy, 1990.

Shaw, H. R., Uniqueness of volcanic systems, in Volcanism in Hawaii, edited by R. W. Decker, T. L. Wright, and P. Stauffer, *U. S. Geol. Surv. Prof. Pap.*, *1350*, 1357-1394, 1987.

Smith, R. L., and R. G. Luedke, Potentially active volcanic lineaments and loci in western conterminous United States, in *Explosive Volcanism: Inception, Evolution, and Hazards*, edited by the Geophysics Study Committee, pp. 47-66, National Academy Press, Washington, D. C., 1984.

Stieltjes, L., and P. Moutou, A statistical and probabilitic study of the historical activity of Piton de la Fournaise, Reunion Island, Indian Ocean, *J. Volcanol. Geotherm. Res.*, *36*, 67-86, 1989.

Suter, M., O. Quintero, and C. A. Johnson, Active faults and state of stress in the central part of the Trans-Mexican volcanic belt, Mexico, 1, The Venta de Bravo fault, *J Geophys. Res.*, *97*, 11,983-11,993, 1992.

Swanson, D. A., Magma supply rate at Kilauea Volcano, 1952-1971, *Science*, *175*, 169-170, 1972.

Takada, A., Magma transport and reservoir formation by a system of propagating cracks, *Bull. Volcanol.*, *52*, 118-126, 1989.

Takada, A., Experimental study on propagation system of liquid-filled crack in gelatin: Shape and velocity in hydrostatic stress condition, *J. Geophys. Res.*, *95*, 8471-8481, 1990.

Takada, A., Accumulation of magma in space and time by crack interaction, in *Magmatic Systems*, edited by M. P. Ryan, Academic Press, San Diego, Calif., in press, 1994a.

Takada, A., Development of a subvolcanic structure by the interaction of liquid-filled cracks, *J. Volcanol. Geotherm. Res.*, in press, 1994b.

Takada, A., O. Oshima, S. Aramaki, K. Ono, T. Yoshida, and K. Kajima, Geology of Aogashima volcano, Izu Islands, Japan, *Bull. Volcanol. Soc. Jpn.*, *37*, 233-250, 1992.

Takahashi, M., Quaternary volcanism and regional tectonic stress field in Japanese islands, *Eos Trans. AGU*, *71*, 948, 1990.

Talwani, M., C. C. Windisch, and M. G. Langseth, Jr., Reykjanes ridge crest: A detailed geophyscal study, *J. Geophys. Res.*, *76*, 473-517, 1971.

Tanaka, K. L., E. M. Shoemaker, G. E. Ulrich, and E. W. Wolfe, Migration of volcanism in the San Francisco volcanic field, Arizona, *Geol. Soc. Am. Bull.*, *97*, 129-141, 1986.

Tsukui, M., M. Sakuyama, T. Koyaguchi, and K. Ozawa, Long-term eruption rates and dimensions of magma reservoirs beneath quaternary polygenetic volcanoes in Japan, *J. Volcanol. Geotherm. Res.*, *29*, 189-202, 1986.

Wadge, D., Steady state volcanism: Evidence from eruption histories of polygenetic volcanoes, *J. Geophys. Res.*, *87*, 4035-4049, 1982.

Williams, H., and A. R. McBirney, *Volcanology*, 397 pp., Freeman, Cooper, San Francisco, Calif., 1979.

Wohlenberg, J., The structure of the lithosphere beneath the East African Rift zones from interpretation of Bouger anomalies, in *Afar Between Continental and Oceanic Rifting*, edited by A. Pilger and A. Rösler, pp. 125-130, Schweizerbart, Stuttgart, 1976.

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(Received July 14, 1993; revised February 3, 1994; accepted February 15, 1994.)