Volc. Seis. 14(4), 1993, pp. 411-417
Reprints available directly from the publisher
Photocopying permitted by license only

Printed in United States of America

EFFECT OF CRUSTAL THICKNESS AND SPREADING RATE ON THE EVOLUTION OF VOLCANIC AND HYDROTHERMAL ACTIVITY

V. L. LEONOV
Institute of Volcanology, Far East Division, Russian Academy of Sciences, Petropavlovsk-Kamchatskiy, 683006 Russia

(Received January 9, 1992)

A comparative study has been carried out of the evolution of volcanic and hydrothermal activities in various geothermal areas of the world. The data available in the literature have been systematized and tabulated. They indicate that there is a close relationship between the periods of hydrothermal activity, eruption periodicity, and the composition of volcanic products, on the one hand, and the thickness of the crust, on the other. Three types of geothermal areas are identified: on oceanic, transitional and continental crust. Each type is characterized by its own pattern of volcanic and hydrothermal activity. The prospects of geothermal fields in Kamchatka are discussed.

The geotectonic distribution pattern of volcanic events and hydrothermal activity is still not clearly understood. As noted by Milanovskiy [4], volcanism is known to occur in all basic types of tectonic environment, but its character and products are different. Moreover, volcanic eruptions do not occur simultaneously, or vary in intensity, or may not take place at all even within geotectonic structures of the same type. At the same time, the problem of classification and correlation of volcanic and geothermal areas is urgent in geothermal investigations when the type of a geothermal deposit and its heat conditions have to be predicted before starting exploration. Attempts to classify geothermal areas in terms of plate tectonics and group them into spreading-, subduction- or intraplate-type areas [17], [19], etc., are not always encouraging, because areas that have little in common happen to be included in one group, e.g. Yellowstone and the Hawaiis.

The classification of geothermal areas attempted here is based on the type of crustal structure. It includes three types of areas associated with oceanic, transitional and continental crust (Table 1). According to available data, they differ considerably in the periodicity of volcanic events, in the products of eruptions, and in the duration of hydrothermal phases. So this approach seems to be justified and worth offering it for discussion.

It should be noted that this classification was based on abundant literary sources, the number of which is too large to be included in the list of references. So the reader is referred to basic works that

Table 1 Types of geothermal areas (based on crustal thickness).

Type of geothermal area	*Type and thickness of crust, km*	*Eruptive periodicity, years*	*Dominant rock type*	*Time of hydrothermal activity, years*	*Examples of hydrothermal areas (figures are crustal thickness, km)*		*Note*
I	Oceanic (up to 20)	10^2—10^4	Basalt	1—10^2	Assal (Djibouti)	5	Spreading rate > 2 cm/year
					East Pacific Rise	5—7	
				1—10^3	Puna, Loichi (Hawaii)	12—17	Spreading rate < 2 cm/year
					TAG (Atlantic)	10	
					Danakil (Ethiopia)	12	
					Krafla, Reykjanes (Iceland)	15—20	
				10^3—10^4	Cerro Prieto (Mexico)	17—20	
II	Transitional (20-35)	10^4—10^5	Andesite		Olkaria (Kenya)	20	–
					Geysers (USA)	20	
					Kamojang, Dieng (Indonesia)	20—25	
					Palinpinon, Tongonan (Philippines)	25	
					Albani, Sabatini (Italy)	25	
					Hohi, Kurikoma (Japan)	25—27	
					Paushetka, Mutnovskiy (Kamchatka)	26—28	
			Rhiolite	10^4—10^5	Wairakei (New Zealand)	36	–
				$n \times 10^6$	Emi-Kussi (Tchad)	45	–
III	Continental (up to 80)	$n \times 10^6$			Yellowstone, Long Valley (USA)	40—50	
					El-Tatio, Calabosos (Chile)	60	
			–	–	Fang, San-Kampjeng (Thailand)	45	Nonvolcanic
					Puga-Chumatang, Parabati (India)	60	
					Yangbajan (China)	70	

Note. Data on eruption periodicity, periods of hydrothermal activity and spreading rates are after [1], [5], [13], [16], [18], [22] and other authors listed in the References.

adduce a large number of absolute ages, and most of them have been determined with sufficient certainty and exactness. Data of crustal thickness are controversial in some areas, e.g. in Iceland. Such areas were classified using the bulk of data available, including the composition of volcanic products and periodicity of volcanic events and hydrothermal phases.

The first type includes geothermal areas underlain by oceanic crust. Rona [5] has recently systematized data on geothermal fields from oceanic spreading centers. He showed, for example, that high-temperature hydrothermal activity in the TAG fields (26 °N mid-Atlantic ridge with a < 2 cm/year spreading rate) occurred in cycles, the active phases being repeated every ~ 10 000 years and lasting from one to 100 years.

Shorter phases of high-intensity hydrothermal activity have been recorded in oceanic ridges of larger spreading rates (> 2 cm/year). Such phases last merely a few to a few tens of years in the 21 °N East Pacific Rise. Rona [5] associated hydrothermal activity in such areas with volcanic events (basalt intrusion). He noted that the recurrence of cyclic peaks of volcanic activity may be 10 to 100 times more frequent in spreading centers with medium and large spreading rates than in oceanic ridges with small spreading rate values.

Rona [5] systematized numerous data on Krafla Volcano, situated in a neovolcanic spreading center in northern Iceland. Although this area shows a small spreading rate, it clearly demonstrates a close relationship between recent hydrothermal activity and recent episodes of graben formation. 20 subcycles, lasting a few months each, have been recorded during a seven-year monitoring of the volcano since 1975. Each subcycle includes doming and subsidence in the caldera. The floor of the caldera may bulge to a height of one meter. Bulging is usually accompanied by seismic activity which is believed to be caused by the filling of the magma chamber with magma. Magma rises into the chamber until the critical pressure is attained. Magma heats the surrounding rocks and activates the hydrothermal system of the volcano. The subsequent subsidence of the caldera is usually accompanied by fracturing, dike injection, and volcanic eruptions. Seismic activity occurs as a result of magma injection from the chamber into the fracture zone. The resulting dikes, which may extend as far as 60 km from the center of the caldera, heal the fractures and destroy the preexisting hydrothermal system.

This pattern is important for correlating hydrothermal activities with the cycles of regional magmatism, volcanism, and tectonics in regions of active crustal extension. The rates and magnitudes of magmatic and hydrothermal processes in such regions may be high enough to compensate for the absence of a long-lasted process of magma differentiation. Many of the known geothermal systems are likely to operate as the systems of Iceland described above. Examples are the Cerro Prieto and Salton Sea systems in the axial trough of the Gulf of California spreading center, the Hawaiian Ridge, and the Danakil Depression in Afar. The periodical injection of basalts seem to be the most important source of heat in such systems.

The hydrothermal systems of island arcs are markedly different from spreading center systems. The largest, high-temperature hydrothermal systems are associated there with large volcanic centers whose eruptive products are essentially of an acidic composition. An example is the Taupo volcanic zone, North Island, New Zealand, where six caldera volcanoes discharged 300 to 1000 km^3 of lava each for the last 600 000 years. 97 percent of the lava volume are rhyolites [25]. There are 18 high-temperature and about 50 low-temperature geothermal systems in the area. They are believed to have been operating since approximately 200 000 years ago, the more active phases being interrupted by repose periods. Changes were recorded in their surface manifestations. For instance, the Waimangu geothermal field is believed to have been formed ~ 100 years ago, the age of the Tikorangi-Rotoma system was reported to be younger than 3000 years [9].

Important evidence has been obtained by Japanese investigators [22], [23] on the ages of geothermal systems, the duration of hydrothermal activity, and its relation to volcanism. They demonstrated that geothermal fields may contain areas of young hydrothermally altered rocks that continue to develop at the present time and areas with hydrothermal products of very old age, overlain by unaltered younger deposits. The rocks usually exhibit several periods of hydrothermal alteration. For instance, there are four generations of altered rocks in the Kuju geothermal system: older than 32 000, ~30 000, and 25 000 years and recent, and four generations in Tamagawa: older than 40 000, 27 000, and ~5000 years and recent.

Sumi and Takashima [22] proved that the periods of intensive hydrothermal activity were closely related in time to episodes of volcanic activity. This enabled them to distinguish different types of geothermal fields, including a particular type associated with middle/late Quaternary volcanism. Geothermal fields of this type are widespread in island arcs elsewhere, including all high-temperature fields of Kamchatka [3], [6].

The above evidence indicates that, as in spreading centers, hydrothermal activity in island arcs occurs in a cyclic manner, its active phases being in tune with active volcanic periods. It also suggests that active hydrothermal cycles last 3 to 5 thousand years there, a span one or two orders of magnitude larger than the periods of high-intensity hydrothermal activity in seafloor spreading centers [5]. This can probably be explained by different spreading rates.

According to Rona's data, the mid-oceanic ridges can be subdivided into those of high spreading rate (> 2 cm/year) and those of low spreading rate (< 2 cm/year). At the same time, a spreading rate of 7 mm/year was measured geodetically in the Taupo volcanic zone, New Zealand [21], and ~ 0.2 mm/year in the Kagosima volcanic area in southern Japan, which bears resemblance to the Taupo zone [26]. A similar value, 0.3 mm/year, was found for areas of recent faults in eastern Kamchatka [2]. Although there are scarce data on spreading rates in volcanic areas of island arcs, this seems to be an important factor which imposes a marked effect, as in seafloor spreading centers, on the evolution of volcanism and hydrothermal activity. It is probably owing to a comparatively higher spreading rate in the Taupo zone (~ 7 mm/year), that volcanic eruptions are more frequent in this area than in the other island arcs of the world (35 eruptions occurred there for the last 50 000 years, each of them having deposited a thick tephra cover). The Taupo area is also distinguished by the predominance of rhyolites and a great power of its high-temperature hydrothermal systems: 142 400 kcal/s at Waiotapu, 101 000 kcal/s at Wairakei, 81 800 kcal/s at Orakeikorako [8]. For comparison the most powerful geothermal fields of Kamchatka show the following values: 75 000 kcal/s in Bolshoi Semyachik, 70 000 kcal/s in the Valley of Geysers, and 64 000 kcal/s in the Uzon caldera [6]. This provides a basis for classifying the geothermal fields of the Taupo zone as a special subtype, intermediate between types II and III (Table 1).

The example of the Taupo volcanic zone demonstrates that there is a complex relationship between the periodicity of large magnitude volcanic events and the duration of active hydrothermal phases, on the one hand, and the spreading rate, on the other. So we may not be loggical enough when we classify geothermal areas into types I and II [3]. At the same time, for lack of data on spreading rates, this question cannot be resolved at the present time and hence remains open.

We have considered volcanic and hydrothermal activities in areas of oceanic crust and in continent-ocean transition zones and found these areas to be different. Essential differences are revealed when we compare these environments with continental areas, especially with extensive regions of comprehensively studied volcanic and hydrothermal activities in the USA. Wilson *et al.* [25] compared the Taupo volcanic zone and the Yellowstone area. They found that with the comparable volumes of volcanic rocks and rates of magma transport, the Taupo zone was distinguished by shorter magmatic cycles, the absence of resurgence, and a younger fragmented crust. They concluded that these conditions precluded the formation of large magma chambers in the upper crust of New Zealand and did not favor the development of petrologic processes in magma reservoirs, as complex as in the caldera volcanoes of the western United States. The latter conclusion seems to be applicable not only to the western provinces of the USA, but to all regions with thick continental crust. For instance, three caldera-formation cycles with the ages of 14 to 11, 10 to 8, and 1 to zero m.y. have been reported by Newhall [18] from the Lake Atitlan area, western Guatemala. He found that mafic magmas had melted the crust and produced large volumes of rhyolitic magma during each cycle. Gardner *et al.* [13] reported that the Jemes volcanic field, New Mexico, had a very long volcanic history. Volcanic activity began there ~ 16.5 m.y. ago with the eruption of alkali basalts, which were succeeded by olivine tholeiites from 13 to 10 m.y., by andesites from 10 to 7 m.y. and by dacites from 7 to 4 m.y. ago. Calderas started to form as a result of explosive rhyolitic volcanism since 4 m.y. ago. Self *et al.* [20] considered these events to be the results of the evolution and periodical unplugging of one magma reservoir. Gardner *et al.* [13] reported that at least three episodes of hydrothermal activity occurred there, their ages being within 13 to 8 m.y., 7 to 6 m.y., and younger than 1 m.y.

We may note, for comparison, that numerous calderas were formed in Kuyshu, Japan in a cyclic manner, the intermissions between the cycles being as short as a few tens of thousand years: caldera eruptions occurred 120 000 and 70 000 years ago at Aso, 130 000 and 85 000 years ago at Ata, and 75 000 and 6300 years ago at Kikai [17]. Baker [7] compared the history of complex caldera volcanoes of the central Andes, the southwestern USA, and Mexico, on the one hand, and that of caldera volcanoes in Japan, Indonesia, and New Zealand, on the other hand. He noted that the former operated as ignimbrite centers with a small number of eruptions, and the latter erupted both lavas and pyroclastic material during numerous but relatively small eruptions. These differences obviously stem from the viscosity of magma, volatile content, the size of magma portions, and the speed of magma rise, which, in its turn, must depend on the crustal thickness [7].

The geothermal areas, located in regions of thick crust, are included in Table 1 into type III as a nonvolcanic subtype. The most probable source of heat in such geothermal systems are cooling granite intrusions. Some of them are believed to have been emplaced 20 m.y. ago [11].

It should be mentioned in conclusion that an important factor that need be considered in planning geothermal explorations or assessing various geothermal deposts is a tectonic environment, namely the thickness of the crust and the rate of its extension. Two extreme situations can be envisaged. One is a rapidly spreading thin oceanic crust, where heat, that generates geothermal systems, is provided by numerous, frequent basalt intrusions. Such systems are extremely dynamic, unstable and do not last long, even though they may accumulate large amounts of geothermal energy. The other situation comprises comparatively stable continental regions with a thick crust in which shallow magma reservoirs may be buried for millions of years and produce occasional eruptions of highly differentiated volcanic products, or may exist as buried bodies of hot rocks without any surface manifestations. Large calderas may be formed above such magma chambers with resurgent domes growing within them. Such heat sources produce powerful and long-lived geothermal systems.

The geothermal systems of Kamchatka are intermediate between these two types. The conditions beneath Kamchatka are not suitable for a rapid rise of large volumes of basaltic material into the upper crust in a manner, similar to the mechanisms operating in Iceland, mid-oceanic ridges, or, possibly, beneath the Cerro Prieto and Salton Sea geothermal systems [12], [15]. At the same time, there can hardly be any large buried magma chambers in Kamchatka, whose magma has never been discharged to the surface, similar to the chambers residing in the crust beneath the Socorro and Death Valley areas in the USA [10], [24]. The geothermal systems of Kamchatka are obviously different from those of New Zealand, too, the region of a peculiar geodynamic environment. These features should be considered when planning geothermal exploration, assessing geothermal fields, and modeling their heat supply.

REFERENCES

1. *Vulkanicheskiy tsentr: stroenie, dinamika, veshchestvo (Karumskaya struktura)* (Volcanic center: structure, dynamics and material, with reference to Karumskiy Volcano)(Moscow: Nauka, 1980).
2. V. A. Legler and L. M. Parfenov, in: *Tektonicheskoe raionirovanie i strukturno-veshchestvennaya evolutsiya severo-vostoka Azii* (Tectonic provinces and evolution of northeast Asia)(Moscow: Nauka, 1979): 134-155.
3. V. L. Leonov, *Strukturnye usloviya lokalizatsii vysokotemperaturnykh gidroterm* (Structural conditions for localization of hydrothermal reservoirs)(Moscow: Nauka, 1989).
4. E. E. Milanovskiy, in: *Geodinamika, magmoobrazovanie i vulkanizm* (Geodynamics, magma generation and volcanism)(Petropavlovsk-Kamchatskiy, 1974): 32-50.
5. P. Rona, *Hydrothermal Mineralization at Seafloor Spreading Centers* (Amsterdam: Elsevier, 1983).
6. V. M. Sugrobov, in: *Teplovoi rezhim nedr SSSR* (Terrestrial heat flow in the USSR)(Moscow: Nauka, 1970): 181-199.
7. M. C. W. Baker, *J. Volcanol. Geotherm. Res.* **11**: 293-315 (1981).
8. C. J. Banwell, in: *New Zealand Volcanology. Central Volcanic Region.* Inform. Ser. 50 (1966): 161-189.
9. P. R. L. Browne and M. P. Hochstein, in: *Abstr. Internat. Volcanol. Congr. New Zealand* (1986).
10. C. E. Chapin, A. R. Sanford, and L. D. Brown, *EOS* **61**, N46 (1980).
11. P. Copeland, T. M. Harrison, and P. Le Fort, *J. Volcanol. Geotherm. Res.* N44: 33-50 (1990).
12. W. A. Ebders, D. K. Bird, A. E. Williams *et al.*, *Geothermics* **13**, N1/2: 27-47 (1984).
13. J. N. Gardner, F. Goff, S. Garcia, and R. C. Hagen, *J. Geophys. Res.* **91**, NB2: 1763-1778 (1986).
14. M. Hachimoto, *J. Phys. Earth.* **33**, N3: 191-226 (1985).
15. P. W. Kasameyer, L. W. Younker, and J. M. Hanson, *Bull. Geol. Soc. Amer.* **95**, N10: 1242-1252 (1984).
16. H. Machida and F. Akai, *Abstr. Kagoshima Intern. Conf. Volc.* (1988).
17. L. J. P. Muffler, in: *Proc. 2nd Sympos. Development and Use of Geothermal Resources* **1**: 499-505 (1976).
18. C. G. Newhall, *J. Volcanol. Geotherm. Res.* **33**: 23-55 (1987).
19. L. Rybach, in: *Geothermal Systems: Principles and Case Histories* (1981): 3-36.
20. S. Self, F. Goff, J. W. Wright *et al.*, *J. Geophys. Res.* **91**, NB2: 1779-1798.
21. K. B. Spörly, *Circum-Pacific Orogenic Belts and Evolution of the Pacific Ocean Basin. Geodynamics Ser.* **18**: 115-132 (1987).
22. K. Sumi and J. Takashima, in: *Proc. 2nd Sympos. Development and Use of Geothermal Resources* **1**: 525-534 (1976).
23. J. Takashima, in: *Abstr. IAVCEI Sympos. Arc Volcanism* (1981).
24. B. Voogd, L. Serpa, L. Brown *et al.*, *Geology* **14**: 64-67 (1986).
25. C. J. N. Wilson, A. M. Rogan, I. E. M. Smith *et al.*, *J. Geophys. Res.* **89**: 8463-8484 (1984).
26. I. Yokoyama and S. Onkawa, *J. Volcanol. Geotherm. Res.* **30**, N3/4: 253-282 (1986).