# Understanding active volcanoes: the case of Usu Volcano, Japan, with emphasis on the 1977 summit eruption

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## Resumen

Comprender el comportamiento de un volcán activo requiere un análisis multidisciplinario de sus actividades pasadas y presentes. En este trabajo se pretenden explicar los mecanismos eruptivos del volcán Usu, uno de los volcanes activos más pequeños del mundo, pero no por ello menos peligroso, ya que ha producido diversos tipos de erupciones como eyecciones de pómez, cenizas y piroclastos y la formación de domos de lava en la cumbre y en su base. La actividad pasada del volcán, particularmente las erupciones piroclásticas, se discute brevemente basado en los artículos y mapas geológicos publicados. Usu es un volcán dacítico que ha producido erupciones desde la cumbre y desde bocas parásitas en su base, como las de 1910 y 1943. La primera fue freática y la segunda produjo un domo de lava. La erupción de 1977 inició con una erupción pumítica en la cumbre y continuó con erupciones freáticas por más de 5 años. La erupción de 2000 fue freato-magmática, desde la base NW. Las erupciones de este volcán desde el siglo 17 muestran un tiempo de recurrencia media de 57 años. La erupción de 1977 es una de las más grandes y fue un caso típico de las erupciones de la cumbre. El objetivo de esta recopilación de las diversas observaciones geofísicas de esa erupción contribuye a la comprensión de su estructura interna y de los mecanismos de erupción del Usu.


Palabras clave: Erupciones desde la cumbre, erupción parásita, modelo de fractura por cizalla, movimiento de bloque, estructura del volcán, depósito de magma.


## Abstract

Understanding the behavior of an active volcano requires a multidisciplinary analysis of its past and present activities. The aim of the present study is to explain the eruption mechanisms of Usu Volcano, which is one of the smallest active volcanoes of the world, yet a hazardous one as it has produced various types of eruptive activity such as pumice, ash and pyroclastic ejections, and the formation of lava domes at the summit and at its base. The past activity of the volcano, particularly the pyroclastic eruptions, is briefly discussed on the basis of the published papers and geologic maps. Usu is a dacitic volcano that has produced summit and basal eruptions from parasitic vents, as those of 1910 and 1943. The former was phreatic and the latter resulted in a lava dome formation. The 1977 eruption began with a pumice eruption at the summit, and phreatic eruptions continued for over 5 years. The 2000 eruption was phreato-magmatic with the outbreak at the NW base. The eruptions of this volcano since the 17$^{th}$ century show a 57-year mean recurrence time. The 1977 eruption is one of the largest and was a typical case of the summit eruptions. This compilation of the various geophysical observations of this eruption will contribute in the understanding of its internal structure and of the eruption mechanisms of this volcano.


Key words: Summit eruption, parasitic eruption, shear fracture model, block movement, volcano structure, magma reservoir.

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## 1. Introduction

Usu Volcano is located in the S part of Hokkaido Island (Figure 1) and was formed about 30,000 YBP. This volcano is one of the world's smallest active stratovolcanoes, measuring 6 km in basal diameter and 0.5 km in relative height. Usu has produced various types of eruptive activity including ejection of pumice, ash and pyroclasts, and the formation of lava domes at the summit and at the base. It is a dacitic volcano that has erupted about every 57 years, in average, since the 17th century. The present study of the different types of volcanic activity at Usu, and of the multiple geophysical observations on it is aimed to improve our understanding of the eruption dynamics of Usu and of active volcanoes in general.

Historical records in Hokkaido are short, only after the 17th century. The first recorded eruption of Usu Volcano was at the summit in 1663 and subsequently, the volcano repeated eruptions at the summit and the base. In the last summit eruption in August 1977, coinciding with the timely development of the activity, the new volcano observatory belonging to the University of Hokkaido began its observation routine. The volcanic activity continued till around 1982. In the present paper, these activities are simply named the 1977 eruption. And in 2000, a parasitic eruption occurred at the NW base.

During these periods, many papers were published after each respective eruption. However, there still remain many unsolved problems. To discuss some of them, the present authors would like to address some questions about the 1977-1982 eruption of Usu Volcano.

Tōya caldera was formed with pumice flows about 110,000 YBP, and presently, its rims are not clearly distinguishable. The original caldera rims may have been dissected and the caldera area expanded. In Figure 2, the thick broken contour connects triangulation points on peak lines, and does not outline the caldera rim. The red contours show the Bouguer gravity anomalies on Tōya caldera in mgal (Yokoyama, 1964), roughly corresponding to the caldera depression. Based on the gravity anomalies, the original Tōya caldera can be estimated to be about 8 km in diameter. A probable caldera boundary is shown by the thin black broken line in Figure 2. The post-caldera activities resulted in formation of Nakajima Island (andesitic rocks) at the center of the caldera, and Usu Volcano was initially formed about 30,000 YBP with basaltic rocks. Usu volcano is located on the rim or nearly outside of Tōya caldera. Such relationship is similar to that between Aira caldera and Sakurajima Volcano in Kyushu, and that between Krakatau and Anak Krakatau in the Sunda straits (Bemmelen, 1949). In general, during the post-caldera activities, it may be difficult for magma to develop new conduits in the "soft" caldera deposits and would prefer to select conduits along the boundaries between caldera deposits and basements.

In Section 2, the eruptive history of Usu Volcano is reviewed from a geological viewpoint. Section 3 is focused on summarizing past eruptions from the standpoints of associated physical manifestations such as explosions, seismicity, and deformations, with emphasis on the parasitic eruptions, and introducing a tectonic structural line (T.S.L.) at the N base of Usu Volcano, which is not directly related to the boundary of Tōya caldera. In Section 4

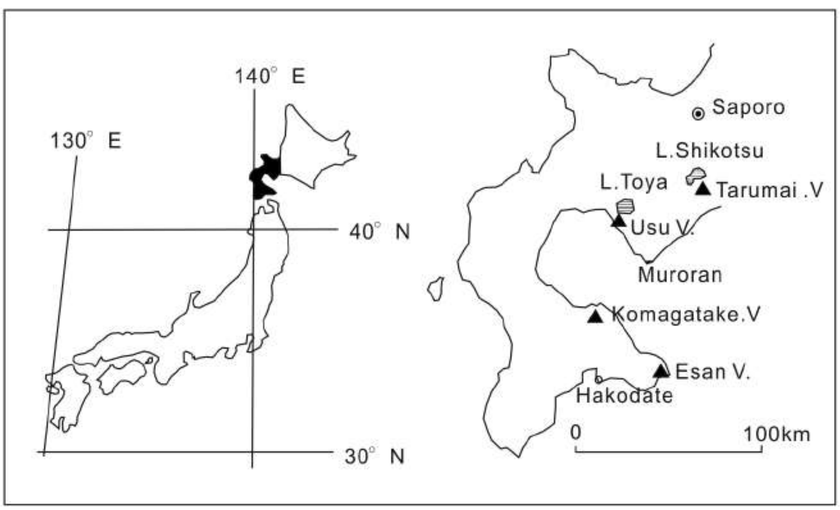



**Figure 1.** Location map of Usu Volcano.

the 1977 summit eruption is discussed and a hypothetical model of the block movements of the upper part of the volcano is presented. In Section 5, subsurface structure of the volcano from various geophysical observations such as P- and S-wave velocity structure, aquifer configuration, electrical resistivity and muographic outline of the summit are discussed. The five main results obtained in the present paper are finally detailed.

At this point a special mention is made of two technical terms to be used in this work:

Craterlet: The activity of Usu Volcano formed many explosion craters in the summit and at the base, usually shallow and with small diameters. In the present paper, to avoid confusion between the summit crater and the explosion craters, the latter shall be named "craterlet" after Ōmori (1911, 1913). These craterlets may be ephemeral since they are rather small in dimension.

Lava dome and mound: Once the 1910 hill or Meiji-Shinzan (new mountain formed in the Meiji era (ab. MS) was called cryptodome. Later, Tanaka and Yokoyama (2013) based on muography (cf. 4.2. Parasitic eruptions) proved it to be a mound without any hidden domes; it was forced up by magma intrusion. The magma intrusion front remains at a depth of about 70 m. In the present paper, cryptodomes and mounds are told apart. Such a structure would be formed by viscous dacitic magmas.

In the following, the discussion will be initiated with the volcano activities during historical time after 1663, and proceed to the 20th century, and finally concentrate on the 1977 eruption.

## 2. Activities of Usu Volcano in historical times

### *2.1 Historical eruptions of Usu Volcano*

Historical records of Hokkaido Island are rather short and eruptions of Usu Volcano have been recorded only after the 1663 eruption as summarized in Table 1.

Usu Volcano is considered small, measuring 6 km in the basal diameter and 500 m in relative height, making it one of the smallest active volcanoes of the world. It is also remarkable that eruption activity has occurred almost regularly in time since 1663, with 57 years as the longest interval between the eruptions, and that the eruption sites irregularly changed between the summit and the base. In historical times, it has not erupted simultaneously at the summit crater and the base. At the base of the volcano, we have observed three parasitic eruptions since 1910, and the 1977 eruption was the first summit eruption after 1853. At present, we cannot predict where the next eruption will occur.

Table 1 shows a wide range of eruption styles, including phreatic, pumice, pyroclastic eruptions, and mound and dome formation. Such sequence of varying volcanic activity is characteristic of this volcano. Matsumoto and Nakagawa (2010, Figure 6) discussed petrology of the volcanic rocks of Usu Volcano:

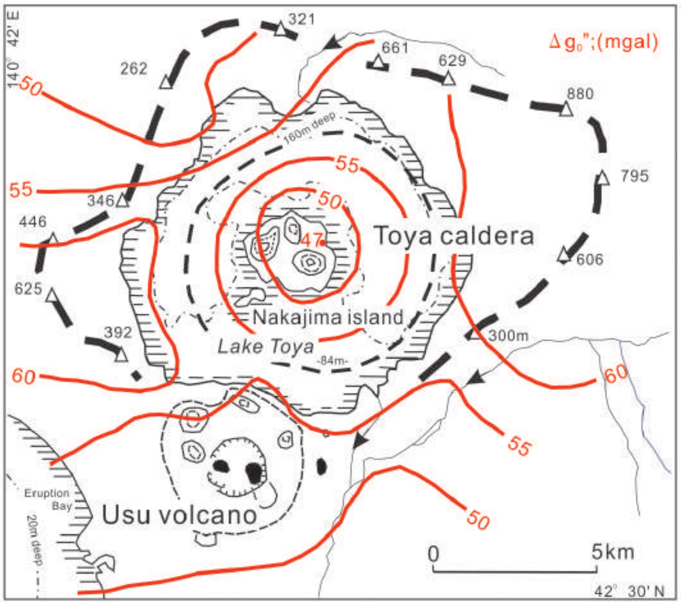



**Figure 2.** Tōya caldera and Usu Volcano. Thick broken line connects topographic peaks around the lake. Red contours show Bouguer gravity anomalies mainly on Tōya caldera (Yokoyama, 1964). 47 mgal at the island is the minimum value. A thin broken line outlines the probable caldera boundary.

**Table 1.** Eruptions of Usu Volcano in historical times.

| Period | Interval (years) | Eruption site | Type of activity | Type of ejecta | Volume of ejecta and lavas ($km^3$ DRE) |
|---|---|---|---|---|---|
| 1) 1663 | | summit | base-surge | pyroclastic flows (Us-b) | ~ 1 |
| 1’) | pre-1769 | | | | |
| 2) 1769 | | summit | Ko-Usu lava dome | ash | 0.03 |
| | 53 | | | | |
| 3) 1822 | | summit | Ogariyama mound | “Bunsei nue ardente” | 0.02 |
| | 31 | | | | |
| 4) 1853 | | summit | Oo-Usu lava dome | pyroclastic flows | 0.05 |
| | 57 | | | | |
| 5) 1910 | | N base | craterlets and mounds | ash | small |
| | 33 | | | | |
| 6) 1943 | | E base | Showa-Shinzan lava dome | ash | 0.05 |
| | 34 | | | | |
| 7) 1977 | | summit | craterlets and deformations | pumice and ash | 0.03 |
| | 23 | | | | |
| 8) 2000 | | NW base | craterlets | ash | 0.0001 |

1’) Between the 1663 and 1769 eruptions, Nakagawa *et al*. (2005) found the deposits of an unknown eruption. Underlined: Distributions of the ejecta are shown in Figure 3. Volume estimates of the ejecta are approximate.

One of their results shows that $SiO_2$ content of the whole-rock components of historical ejecta has decreased almost linearly from 75 % of 1663 to 70 % of 2000. This means that the Usu magmatic system has changed since 1663 from rhyolite to dacite, causing a decrease of the magma viscosity with time according to the experimental results of Goto (1997).

*2.2 Pyroclastic ejecta in historical times*

Katsui (1973) studied pyroclastic deposits, falls and flows, on and around Usu Volcano. Three typical ash falls are depicted in Figure 3 where the explosion of the 1769 and the 1822 ash emissions are known as Ko-Usu dome and Ogari-yama mound, respectively. The origin of the 1663 ash surges is not clear. It is very important to determine the exact eruption center of each pyroclastic flow. In the figure, the location of the eruption centers are estimated, considering E-ward drifts as prevailing winds in this area. Probably, the 1769 pyroclastic falls may have occurred simultaneously with the growth of Ko-Usu dome. The 1853 pyroclastic flows may have been caused by collapses of the new Oo-Usu lava dome, considering that the dome partly collapsed prompted by earthquakes at the beginning of the 1977 eruption.

For any quantitative discussion, it is desirable and important to know the magnitudes of these eruptions that may be estimated by the volumes of their effusive ejecta. Distributions of ash deposits from the three summit eruptions, in 1663, 1769 and 1822, are shown in Figure 3 by Katsui (1973) and volume of some recent eruptions are already published. Approximate ejecta volumes of historical pyroclastic flows and lava domes in Figure 3 are graphically calculated with some assumptions made by the present authors as shown in Table 1. As for the recent activity, the 1977 eruption issued 0.030 $km^3$ DRE in total (Katsui *et al*., 1978) and the 2000 eruption produced 0.0001 $km^3$ DRE (Geological map, 2007). During these 350 years, Usu volcano eruptions have occurred almost regularly in time, but their ejected volumes have been variable. In Table 1, allowing for some errors, the summit eruptions ejected larger volumes than the parasitic eruptions.

## 3. Some characteristics of the eruptions, mainly in the 20th century

Modern Japanese development of natural sciences documenting actually started in 1860’s and the 20th century was a progressively advancing period; The 1853 eruption of Usu

Volcano was not monitored scientifically, but the 1910 eruption was monitored in a greater detail, even if the scientists arrived at the locations after the outbreaks. The activities of this volcano are characterized by dacitic magmas with high viscosity.

It is difficult to obtain precursory information on future eruption activity of a volcano from its static images. Eruption activity of Usu Volcano can be obtained from summarization of its eruption activity. In this Section, the eruption activities in the 20th century shall be studied at two localities, its base and the summit. At the former, numerous small parasitic vents formed in the 1910 and the 2000 eruptions, and lava dome in the 1943 eruption occurred. In the latter, only the 1977 eruption formed several vents and caused remarkable deformation of the summit. The latter is briefly referred in this section, but shall be discussed in detail in Section 5. The simplified distributions of the craterlets formed after the 1910 eruption is shown in Figure 4.

*3.1 Precursory earthquakes of the eruptions in historical times*

Precursory earthquakes of the Usu eruptions were counted by felt shocks in historical times and instrumentally observed after 1943. The leading manifestations of this volcano were usually explosive after precursory earthquakes of which duration usually ranged from 1 to 10 days, with one exception that lasted 6 months during the 1943 eruption: Deviation from 1 to 10 days may depend on magma paths from reservoirs to vents. Magma movements during the 6 months before the outbreaks of the 1943 eruption shall be discussed in Subsection 3.2.

In Table 2, it is remarkable that the 1977 eruption had the shortest precursory period of 32 hours and the smallest magnitude of precursory earthquakes detected was M 3.7. These characteristics shall be discussed in Subsection 4.1.

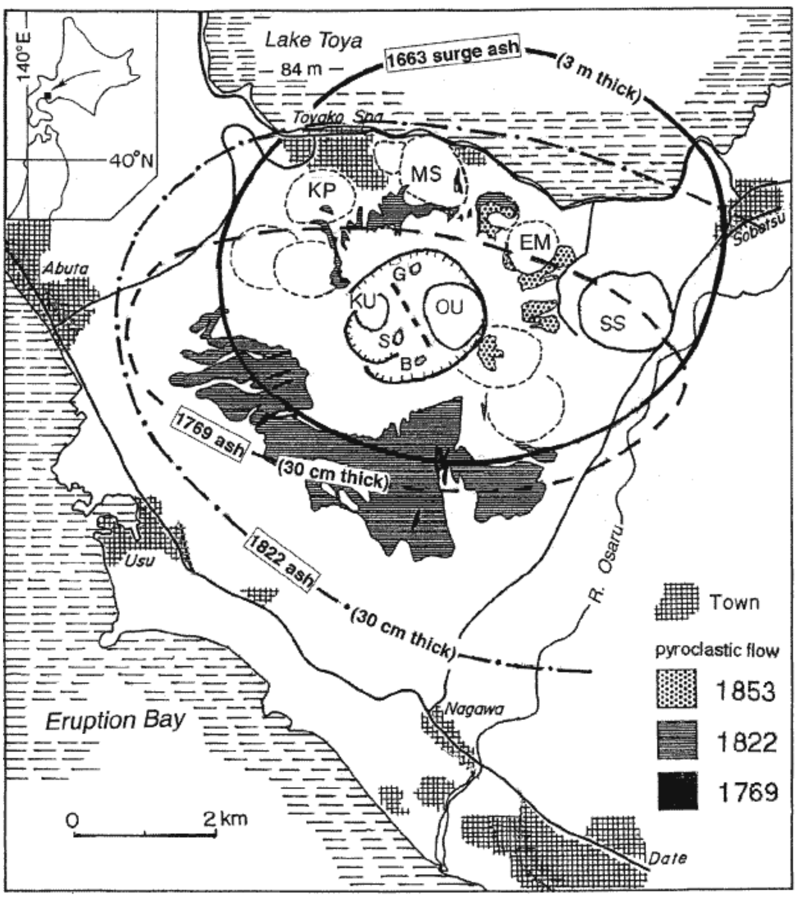



**Figure 3.** Distribution of pyroclastic ejecta from Usu Volcano copied and simplified from a geologic map prepared by Katsui (1973). G, S and B in the summit crater denote the ponds, respectively.

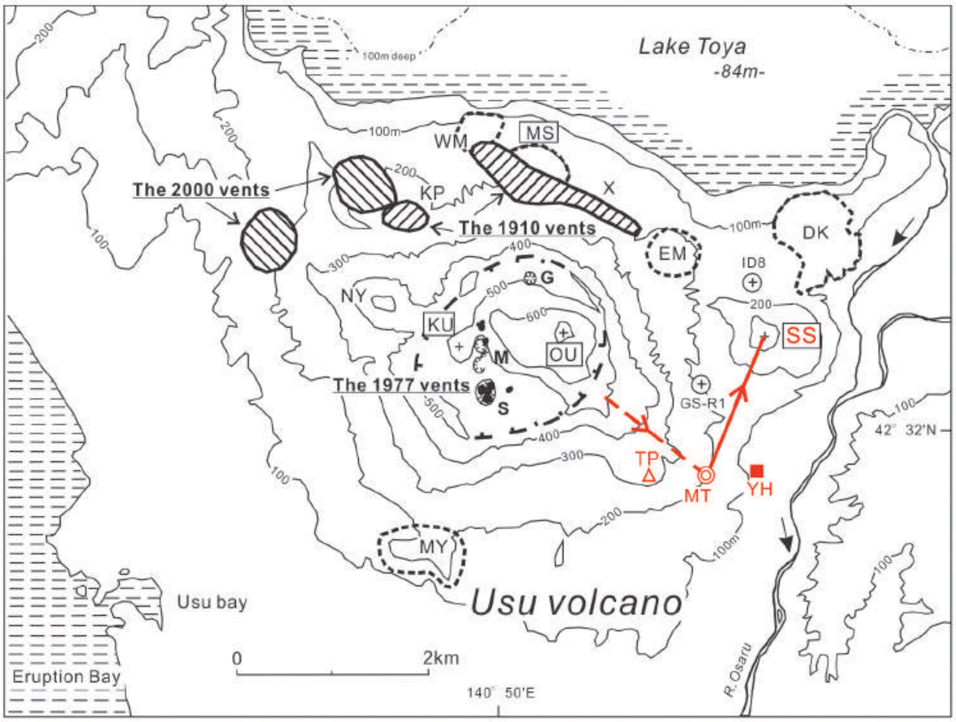



**Figure 4.** Domes, mounds, craterlets and parasites on Usu Volcano formed after 1910.
KU: the 1769 lava dome, OU: the 1853 lava dome, MS: Meiji-Shinzan (the 1910 mound), SS: Showa-Shinzan (the 1944 lava dome), WM: West maru-yama, KP: Kompira-yama, EM: East maru-yama (lava and scoria, basalt and andesite), DK: Donkoro-yama (scoria, basalt), NY: Nishi-yama, MY: Minami-yama, ■YH: Yanagi-hara station, △TP: Triangulation point, ◎MT: Assumed location of magma top (its depth is about 1 km), ⊕: Drilling sites, M: main craterlets (Nos. 1,2 and 3, and I) , G: "Golden Pond", S: "Silver Pond".

### *3.2 Parasitic eruptions in the 20th century*

Historical eruptions of Usu volcano are clearly distinguished as basal and summit eruptions. On Usu Volcano, simultaneous eruptions at the base and the summit have never occurred and basal eruptions have been recorded only after 1910. The basal eruptions are necessarily parasitic.

(a) The 1910 parasitic eruption at the northern base

The activities of the 1910 eruption were reported in detail by Ōmori (1911) and Satō (1910). Precursory earthquakes large enough to be felt persisted for 6 days. About 45 vents of small and large diameters opened in random order, roughly along the contour of 200 m a.s.l. at the N base of the volcano, as shown in

**Table 2.** Precursory earthquakes observed in the historical times.

| Eruption | | Precursory period | Main type of activity | Hypocentral depths (km) | Maximum magnitude |
|---|---|---|---|---|---|
| 1663 | summit | 3 days | | | |
| ? | summit | ? | ? | | |
| 1769 | summit | a few days ? | Ko-Usu (KU dome) | | |
| 1822 | summit | 3 days | | | |
| 1853 | summit | 10 days | Oo-Usu (OU dome) | | |
| 1910 | N base | 6 days | phreatic, MS mound | | 5.1 |
| 1943 | E base | 6 months | Showa-Shinzan lava dome | 1 ~ 4 | 5.0 |
| 1977 | summit | 32 hours | pumice eruption | 0 ~ 5.5 | 3.7 |
| 2000 | NW base | 4 days | phreatic eruption | 5 ~ 7 | 4.6 |

Figure 5 (b); the distribution of the craterlets was originally described by Satō (1913) and the majority of those with small diameter have disappeared topographically years ago. The craterlets issued mudflows and a small quantity of lapilli and ash. Considering the random formation of small craterlets, we may assume a "tectonic structural line" (T.S.L.) for a series of the 1910 craters including Kompira-yama (KP), East Mound and the 1943 lava dome (SS-dome) in Figure 5 (b). Moreover the 2000 eruption formed many craterlets around the 200 m contour at the NW slope or at the extension of the above T.S.L.

As will be discussed in Subsection 4.3, T.S.L. is the surface edge of the summit block which tilted as volcanic activity developed during the 1977 eruption. At the 1910 craterlets in the region of point X in Figure 5 (b), magma was fed along the vertical plane of T.S.L Such a manner of magma supply may have caused the random occurrence of parasitic eruptions along T.S.L.

In the 1910 eruption, a mound Meiji-Shinzan (MS) of relative height about 80 m was formed. Tanaka and Yokoyama (2013) studied the structure of MS mound (Figures 5 and 6). Hitherto this mountain has been called a "cryptodome" describing a lava dome covered by a domelike hill. Muography revealed that actually it is a mound that was lifted by magma intrusion and the magma top remains about 70 m below the surface. The magmas exploded in the shallower parts forming three craterlets A, B and C at the surface as shown in Figure 6.

(b) The 1943 parasitic eruption at the eastern base:

The sequence of this eruption was reported by Minakami, Ishikawa and Yagi (1951) from synthetic application of geophysics and geology. Later Nemoto *et al*. (1957) studied the geothermal field on and around the lava dome applying multiple techniques.

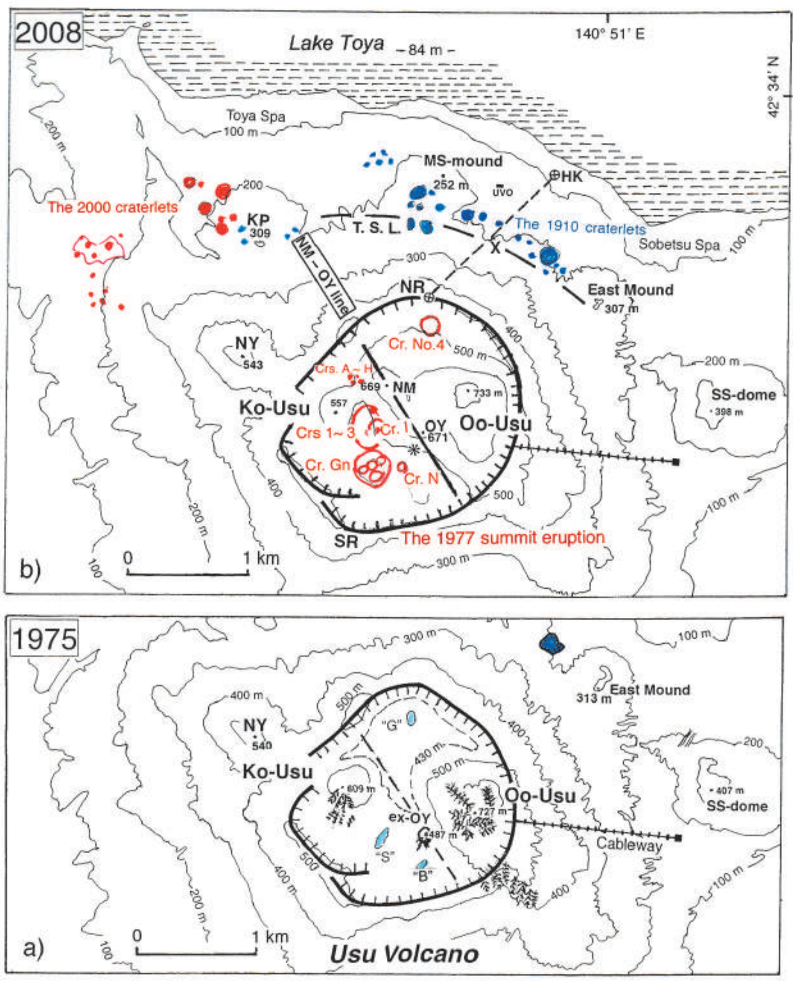



**Figure 5.** The volcanic features of Usu Volcano before and after the 1977 eruption. The topographic maps were surveyed in 2008 and 1975 respectively by the Geographic Survey Institute.
a) Summit part before the 1977 eruption. OY: Ogari-yama (a mound of unknown age).
b) Summit and base as of 2008, including volcanic structures formed in the 20$^{th}$ century.
NR – HK: GPS survey line, T.S.L. (tectonic structural line). Blue dots: the 1910 craterlets.
Red dots at the base: the 2000 craterlets.
Red colors in the summit crater: the craterlets formed in the 1977 eruption.
NR: north rim. SR: south rim. OY: Ogari-yama. NM: New Mountain. KP: Kompira-yama.
NY: Nishi-yama. SS: Showa-Shinzan lava dome formed in 1944. Asterisk: Site of a scarp shown in Photo 2.
NM –OY line corresponds to NM point in Figure 13.

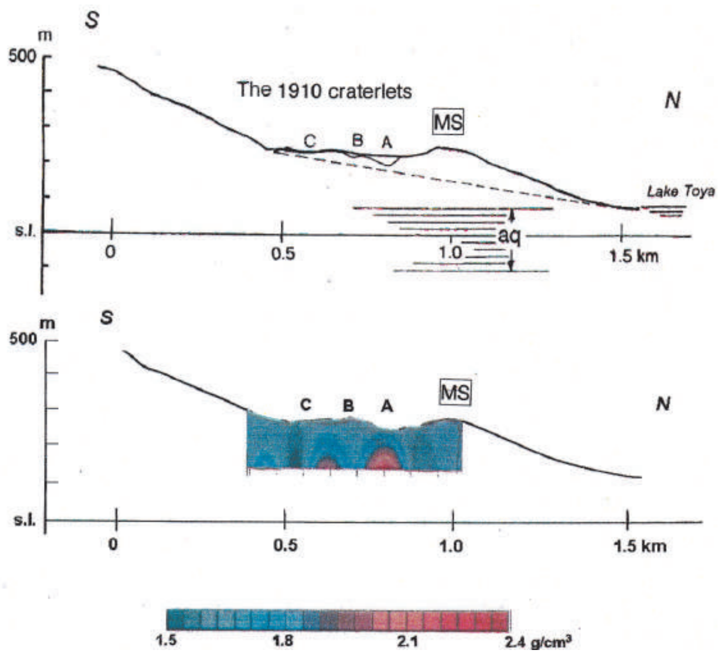



**Figure 6.** A muograph of Meiji-Shinzan (MS) mound after Tanaka and Yokoyama (2013).

Precursory earthquakes of the eruption began on Dec. 28, 1943. After 4 months, on May 1, upheaval of the railroad track at YH point (Figure 4) had reached 23 m at the maximum location within a distance of about 3 km (Inoue, 1948). The upheavals were surveyed along the railroad in the N-S direction but may have extended wider into the area covered with forests. The present authors suspect the center of upheaval may have been located a little W-ward from YH point in Figure 4. We newly found deformation of a triangulation point at about 1 km west (TP in Figure 4). This was 302.5 m a.s.l. in 1920 and 1932, and had changed to 316.8 m by a re-survey in 1955; namely this point had lifted about 15 m probably related with the 1943 activity. Now, we propose that the magma top may have been around MT point between YH and TP, or around the 200 m contour as shown in Figure 4.

Adopting the point-source model, the depth of the pressure source (or magma top) is approximately determined at about 1 km depth. In other words, viscous dacitic magma took 4 months to reach under the MT area from their origin, probably around 4 ~ 5 km b.s.l. Thereafter, the magma migrated N-ward and upward and reached SS point around June 1, 1944, with an apparent velocity of 1.5 km / month or 50 m/ day. On the other hand, in the period of dome upheaval near the ground surface discussed by Yokoyama (2002, Figure 2), its averaged velocity was about 1.8 m / day. Such considerable difference was surmised to be partly due to a remarkable increase of viscosity of the magma by cooling and dehydration even though their circumstances were rather different.

Minakami *et al*. (1951) carried out seismological observation using mechanical seismometers at five points during the period June to September 1944. Volcanic earthquakes of A-type during July 1944 were located at the S slope of the volcano or at the NW of MT area (Figure 4) and the hypocenter depths ranged mainly between 3 to 5 km. Duration of the precursory earthquakes was exceptionally long, lasting 6 months (Table 2). SS lava dome is peculiar in being located at the base while the other lava domes such as Ko-Usu and Oo-Usu are in the summit area. The viscous dacitic magma may have developed a separate route.

Why did the magma change its course towards the N at MT? There may exist a caldera boundary or an extension of the tectonic structural line (an extension of T.S.L) to obstruct its E-ward migration. The magma may have followed the weakest route with least resistance. During the migration, the magma top ascended to about 200 m b.s.l. (Yokoyama, 2002, Figure 2). At SS point, the rising magma had lifted the ground surface and contacted the aquifer on June 23, 1944 causing the first explosion.

The dome finally reached a relative height of about 400 m, and grew laterally into an onion-shaped dome and completed its development as the Showa-Shinzan (SS) lava dome. Yokoyama (2002) discussed growth mechanism of the lava dome, and Tanaka and Yokoyama (2008) carried out a muographic survey and clarified the structure of the upper part of the lava dome.

(c) The 1977 parasitic eruption in the summit crater: This is a unique summit eruption in the 20th century. This eruption shall be discussed in Section 4.

(d) The 2000 eruption at the NW base: Precursory earthquakes occurred for 4 days, similar to the 1910 eruption (6 days) and the magmas took the paths in similar structure, but in different directions. The eruption started on March 31 at the W foot of Nishi-yama (NY in Figure 4) around the 200 m (a.s.l.) contour and was magma-phreatic. The explosion column reached to about 3.2 km that was lower than that of the 1977 summit eruption (12 km). On the next day, new craterlets opened at Kompira-yama (KY) where a few craterlets of the 1910 eruption pre-existed. By the middle of April, the eruption had formed 65 small craterlets roughly on the extension of T.S.L. as shown in Figure 5b. At the W base of (NY), the central part finally had upheaved about 80 m by intrusion of magmas (Mori and Ui, 2000), as well as Meiji-Shinzan (MS). In Figure 5, a contour of 50 m upheaval is shown in red color. At the Kompira-yama (KP) area, it is not clear whether a few of the 2000 vents reoccupied the 1910 ones. At present, a precise determination of the 1910 vents on KP is difficult because the vents were very small and many of them have disappeared.

In contrast to main craters, parasitic ones usually don't repeat eruptions at the same vents probably because conduits of the latter are small in diameter and remain blocked with magmatic material of previous eruptions. However, if the previous conduits are not totally blocked, following magma-phreatic material may pass through the conduits and reach the previous vents to repeat the eruptions. If the previous conduits are totally blocked, usually parasitic eruptions must take alternate routes to reach the surface because their intrusive force and magnitude are not strong enough to overcome the barrier or blockage.

### *3.3 Topographies of the summit area in the early 20th century*

During the documented history of this volcano, extending from the 17th to the 19th centuries, all the eruptions occurred at the summit. During the following century, only the 1977 eruption occurred at the summit. Here, topographies of the summit area in the early period of the 20th century shall be referred to.

On the Military Topographical Maps published in 1896 and 1910, three small depressions were clearly expressed as ponds within the summit crater, namely "Golden Pond", "Silver Pond" and "Brown Pond" and later, only "Silver Pond" in the S part remained as a pond and the other two remained as small dry basins on the maps. It is unfortunate that no historical record exists about their formation because they may have seemed so small in diameter and insignificant. Probably the first two vents may be twin parasites (Subsection 4.1) considering that their locations are symmetric around the center of this volcano even if they formed at different periods. Furthermore, as mentioned previously, the origins of the three pyroclastic flows have not been recorded. There may be some possibility that all or any of the three ponds were their origins.

Additionally, a Japanese pioneering geologist, Katō (1909) studied geology of Usu Volcano. His geological map of the summit area is reproduced in Figure 7. It was just 1 year before the 1910 eruption. In the figure, G ("Golden Pond"), S ("Silver Pond") and B ("Brown Pond") are the three depressions at the bottoms. This may suggest that both or either of the two ponds, G and S ponds were remains of the vents of pyroclastic ejecta in the 1822 eruption.

In all the geological maps after 1909 (Figure 7), there is no indication of a central crater within the summit crater. This may be interpreted as the central crater was originally located between Ko-Usu and Oo-Usu lava domes but was hidden by the domes and their ejecta and the main conduit may exist at some depth below.

In the 1977 eruption, "Golden Pond" erupted forming Craterlet No. 4 at the early stage, and later Craterlets J, K, L and M successively formed around "Silver Pond" and finally all merged to form Craterlet Gn (J ~ M) as shown in Figure 5 (b).

## 4. Activities of the 1977 summit eruption

The 1977 eruption was the first summit eruption after the 1853 eruption, and since then no other eruptive activity occurred. We are not aware of any scientific documents regarding observations of other summit eruptions for this volcano. Geological and geophysical research reports of the 1977 eruption were published shortly after the eruption by Katsui *et al*. (1978) and Niida *et al*. (1980) and by Yokoyama *et al*. (1981) and Okada *et al*. (1981), respectively. In this paper, a brief summary of the eruption was prepared introducing some new interpretations based on geophysical considerations.

*4.1 Sequence of the 1977 eruption*

According to Katsui *et al*. (1978), the 1977 activity is divided in two stages:

**First Stage** (sub-Plinian eruptions) (Aug. 7 ~ 14, 1977)
Craterlets Nos. 1, 2 and 3: pumice eruptions and explosion columns reaching about 12 km.
Craterlet No. 4: pumice eruptions (Aug. 9).

**Second Stage** (phreatic-phreatomagmatic-magmatic eruptions)
Substage I: Typical phreatic eruptions (Nov. 16, 1977 ~ Mar. 13. 1978)
Craterlets A ~ H,
Substage II: Phreatic to phreatomagmatic eruptions (Apr. 24 ~ June 28, 1978)
Craterlet I,
Substage III: Phreatic to phreatomagmatic eruptions (July 9 ~ Oct. 27, 1978)
Craterlets J ~ N; Craterlets J. K, L and M gradually merged into Craterlet "Silver Pond".

With relation to the above external activities, related seismic activity shall be summarized as follows. Precursory earthquakes of the 1977 eruption continued for about 32 hours and their hypocenters were located at 0 ~ 5 km b.s.l. Magma took 32 hours to reach vents at the surface from depths of roughly 4 ~ 5 km making a new path or along old conduits as indicated by the relatively small magnitudes of earthquakes (max. M 3.7).

During the period of the precursory earthquakes, the expected accompanying deformations of the summit area were not initially distinguishable, but soon upheavals of "Ogari-yama" (OY in Figure 5) had become clear, and periodical theodolite surveys of point OY from the base point at about 8 km south were started on Aug. 15.

Figure 8 shows three kinds of activities: (1) daily release rates of seismic energy after Seino (1983), (2) eruption activity of each craterlet, and (3) upheaval rates of Ogari-yama (OY). In the figure, all the three have the same time axes. Coinciding rates of seismic energy and upheaval of the central part of the summit crater indicates pressure induced by rising magma below the volcano.

At 01 h 06 m on Aug. 6 (JST), 1977 precursory earthquakes started. This stage is

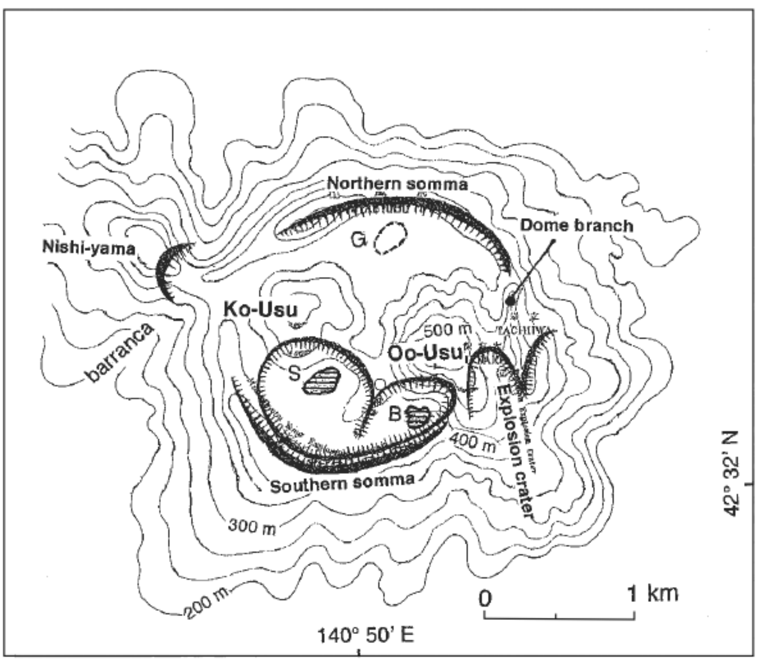



**Figure 7.** A geological sketch map of the summit crater of Usu Volcano before the 1910 eruption (Katō, 1909). Katō's "Explosion crater" indicates a depression containing two vents of S and B. In the present discussion, the two vents, G and S proved to be twin vents.

indicated by $S_0$ in Figure 8 - top. And at 09 h 12 m of Aug. 7, a violent summit eruption took place producing Craterlet No.1 at the eastern base of Ko-Usu lava dome. After the first eruptions ($S_0$), seismic energy decreased sharply for several days and recovered to Step $S_1$, and thereafter, decreased exponentially. At the end of January, 1978, both rates changed to an increase by a step ($S_2$ in the top of Figure 8) and, thereafter decreased exponentially. The next eruption activity resumed on Feb. 25, and continued to March 13, forming 6 craterlets (C, D, E, F, G and H). This development was fully discussed by Yokoyama *et al*. (1981).

According to Niida *et al*. (1980), the eruptions of Craterlets A to H were phreatic and their ejecta were almost entirely ash derived from pre-existing rocks. It is clear that magma intruded at Step $S_2$ and probably, the magma may have been more viscous than the earlier one and reached near the bottom of the volcano on Apr. 24, when Craterlet I began its activity: Before Step $S_3$ , six craterlets C ~ H were successively formed but they were not magmatic. They may have been activated by energy from the magma of Step $S_1$ . The magma of the Step $S_2$ may have been more viscous than Step $S_1$ and needed more time to reach the vent possibly because the magma had lost its $H_2O$ component. The magmas needed about 75 days to rise to an explosion depth, namely ($S_2 \rightarrow S_3$) as shown in Figure 8. Craterlet I erupted amid the three craterlets, Nos. 1, 2 and 3, and its magmatic activity lasted about 2 months from Apr. 24 to June 28, 1978, and later only fumarolic activity continued with strong emission with high temperature gases until 1982. Craterlet I may have been energized by similar mechanisms as those of Craterlets Nos. 1, 2 and 3.

Judging from the sequence of the three activities shown in Figure 8, the largest amount of magmatic material and energy was afforded from the depths by the first outbreaks.

Singularity of the eruptions of Craterlet No. 4: As mentioned above, Craterlet No. 4 was formed at "Golden Pond", about 1 km NE from Ko-Usu lava dome, on the 3rd day following the outbreak. A major pumice eruption lasted 3 hours forming a crater of about 100 m across. The pumice and ash were scattered to the E of the volcano. The site of the "Golden Pond" was dry in 1909 and 1975 (Figures 7 and 5 (a), respectively). Its origin is not documented, and probably may be pre-historic. Photo. 1 shows Craterlet No.4 immediately after its formation: the first opening of this craterlet was cylindrical, not funnel shape. It was about 100 m across and about 50 m deep. The uppermost white

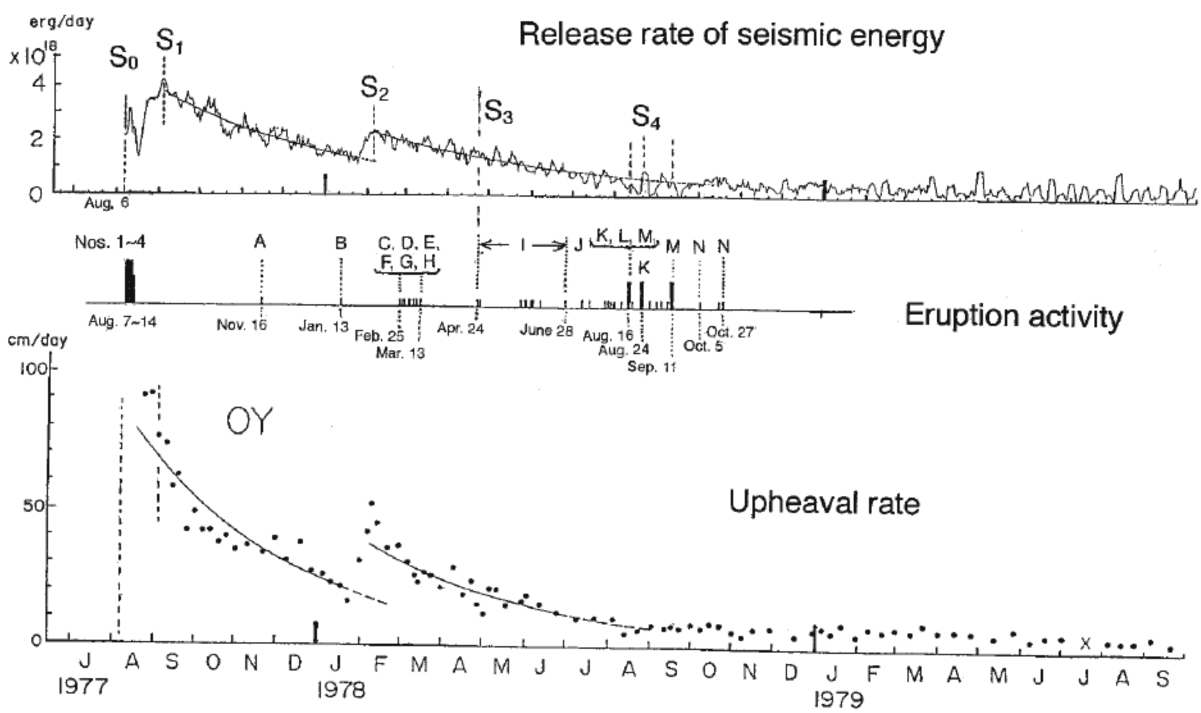



**Figure 8.** (top) Release rate of seismic energy (after Seino, 1983), (middle) Eruption activity (mainly after Niida *et al*., 1980), Nos. 1 - 4, and A ~ N are signs of the craterlets, (bottom) Upheaval rate of Ogari-yama (OY in Figure 5).

layer was the new ejecta of pumice and ashes. Leaves of the trees dropped but trunks were not damaged; explosivity was not so high and was directed vertically. In short, the eruption of Craterlet No. 4 was rather simple and readily ejected pumice and ash without expanding its crater. This may due to access through a previously formed conduit. A few months later, the surface circumference collapsed to form a funnel-shaped craterlet.

Twin parasitic vents in the summit crater: As referred above, "Golden Pond" erupted in August, 1977, and "Silver Pond" erupted in September, 1978. Here it is assumed that these two vents were twin parasitic vents formed by a dilatational source acting from the middle of the two vents. This upward pressure source serves an important role in the tilt movement of the summit region as shall be discussed in Subsection 4.3 with Figure 13.

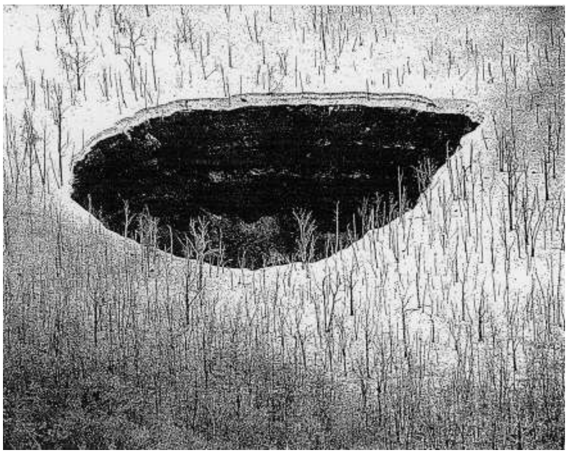


**Photo 1.** Craterlet No. 4 immediately after the formation on Aug. 9, 1977. (Hokkaido Shimbun Press).

Shear fracture model for formation of twin vents by a dilatational source: This was already discussed by Yokoyama (2015) and shall be briefly summarized in the following:

As a volcano becomes active, magmatic pressure is exerted towards the pre-existing main crater through the main conduit, and if the pressure overcomes the yielding strength of cap-rocks at its crater, an explosion should be triggered; this is usual in polygenetic volcanoes and interpretable by the maximum stress theory. Many volcanoes may have a complex of sills and dikes formed in past activities. New magma invading the volcano complex may utilize these sills and dikes for their new conduits leading to vent. In the following section, formation of parasitic vents in homogeneous volcanic bodies is discussed by applying the criteria of fracture mechanics, specifically the maximum shear stress theory. At a certain point under the flat surface of a volcano, both pressure and shear stresses exert on the structure. If either of them reach the yielding strength, the corresponding part of the volcano should fracture. Usually shearing strength of rocks is much less than the compressive one. In Figure 9 (a), the plane polar coordinates ($r$, $\theta$) at the surface are adopted and the effects of shear stress due to a dilatational source $P_0$ are considered : the maximum shear stress is equal to half of the horizontal differential stress and is expressed as:

$$1/2(\sigma_{rr} - \sigma_{\theta\theta}) \quad (1)$$

where $\sigma_{rr}$ and $\sigma_{\theta\theta}$ denote the principal stresses. The maximum shear stress occurs across a plane whose normal bisects the angle between the greatest and least principal stresses.

After some calculations, the value of term (1), positive or negative maximum, was obtained at:

$$r = \pm\, 0.82\, D \quad (2)$$

In other words, the medium undergoes the maximum horizontal differential stress at a radial distance $r = \pm\, 0.82\, D$, or where the dip angle of the pressure source from the fracture point at the surface is 51°. Consequently shear fracture develops there, in the radial direction on the surface or along the slope. Theoretically the maximum horizontal differential stress is expected to be present at two points, $r = \pm\, 0.82\, D$, or on symmetrical sides of volcanoes. In the actual volcanic fields, the media are not always uniform, and so any weaker point ruptures first eliminating any further rupturing to occur.

Anderson (1936) discussed the dynamics of the formation of cone-sheets caused by point dilatation and showed opening fractures (solid lines) and isostatic surfaces (broken lines) in Figure 9 (b). The twin parasitic fractures $S_1$ and $S_2$ are added to the figure by the present authors.

On the actual volcanoes, the shear fracture model may be useful when the volcano is still young and remains uniform in structure.

Yokoyama (2015) mentions several examples of twin parasitic vents on volcanoes of the world. Their origins were not always documented and one or both of them sometimes would be omitted from the maps after substantial time because they are usually not so large in diameter.

Considering the past volcanic activities of Usu Volcano, the shear fracture model shall not be strictly applicable to future formation of twin parasites on this volcano because the volcano is not homogeneous any longer. Magma may proceed to preexisting sills and dikes to form a new conduit and vents.

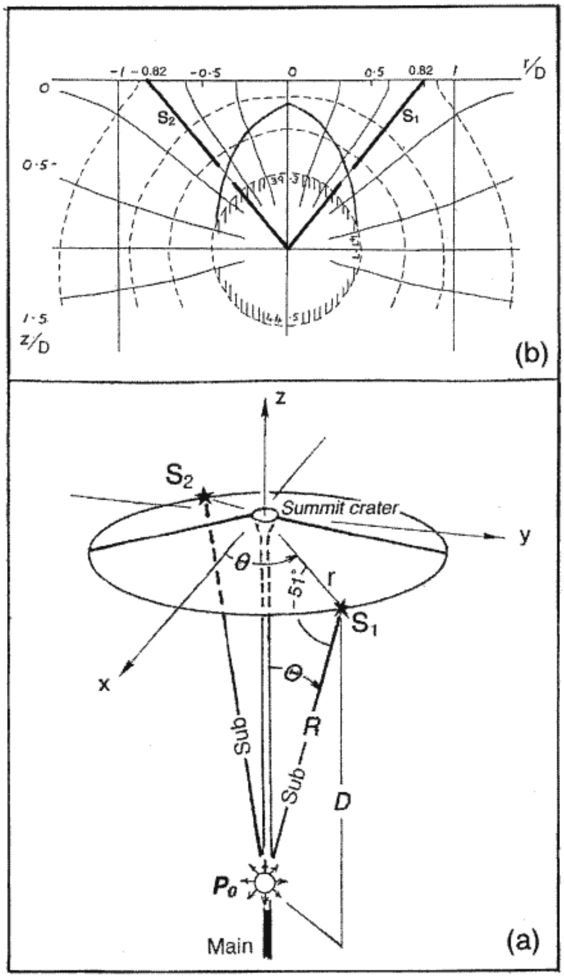



**Figure 9.** Shear fracture model: (a) Point dilatational model, (b) Isostatic surfaces (broken lines) produced by point dilatation after Anderson (1936).

*4.2 Deformations in and around summit crater*

a) Ko-Usu lava dome: It was formed in 1769 and stood 608.8 m a.s.l. before the 1977 eruption. The dome began to subside as a whole, simultaneously with the pumice eruptions of Craterlets Nos. 1, 2 and 3, accumulating at the E base of this lava dome. Finally the height reached 557 m, about 50 m drop in about 3 days from Aug. 7 to 9, and it was clearly visible by naked eye. No one expected such subsidence of the dome, and so no instrumental surveys were conducted. Pumice explosions at Craterlets Nos. 1, 2 and 3 probably may have dug the base beneath the lava dome, or the conduit of the lava dome was, in part, common with those of the new craterlets. The 1977 eruption started after 32 hours of precursory earthquakes, and the maximum magnitude of the earthquakes was M 3.7, both quantities were minimal relative to other historical eruptions. The present authors suspect that the magma passed through some preexisting cracks of the conduit of Ko-Usu lava dome and reached the surface in a short time with minimal resistance.

b) Oo-Usu lava dome: It was formed in the 1853 eruption. Before the 1977 eruption, the main peak was 727 m a.s.l. and a dome branch ("Tate-iwa" or obelisk) separated at the northern side was about 650 m a.s.l. (Figure 4 and Figure 5a). In the early period of the 1977 eruption, a branch of the dome collapsed by earthquake movements or tilting of the summit area. Frequent earthquakes occurring directly beneath the dome partly destroyed it (cf. Figure 5), and its height decreased while the area of Oo-Usu dome was upheaved by tilting of the summit block. Finally, the height decreased to about 670 m a.s.l. and the highest peak apparently shifted northward about 400m and attained a height of 733 m (a.s.l.). As shall be discussed later (cf. Subsection 4.3), a summit block bounded by a line from Oo-Usu toward N 30° W tilted gradually since the outbreak of the eruption. Such deformation caused upheavals of Oo-Usu dome as a whole. In short, Oo-Usu lava dome suffered from upheavals and collapses. This suggests that the upward magmatic force acted under [NM-OY] line (Figures 5b and 13) accompanied by relatively large earthquakes (M 4). This shall be discussed later (Subsection 4.4).

c) Horizontal displacements: The summit crater of the volcano deformed according to the progress of volcanic activity. The Geographic Survey Institute of Japan had carried out photogrammetric surveys over the summit crater three times after the starting of the 1977 eruption: The results are shown in Figure 11 where the S half of the crater rim has not deformed. OY denotes a mound formed in the 1769 eruption and its relative height was about 10 m as for 1976. A curved line connecting OY and the base of Ko-Usu lava dome had displaced about 250 m and NR-point of the northern rim about 130 m, both NE-ward during 2 years and 3 months after the beginning of the 1977 eruption. This pattern of the horizontal deformation indicates that the summit block displaced toward the NE, and the SW rim (SR) of the crater had not deformed probably because crater deposits are explosion ejecta, and are soft and not elastic.

d) Upheaval or tilt movement of the summit block: The 1977 eruption began on August 7 and after a few days, Ogari-yama (ex-OY in Figure 5a) was noticed to have been rising daily. Routine theodolite-observations of the target OY commenced in the middle of August 1977 and its upheaval rates are shown in Figure 8. Later, point NM (new mountain) was installed as another target.

Harada *et al*. (1979) installed an EDM measuring line between the N rim (NR) and the base station (HK) near the lake as shown in Figure 5. According to them, the line had shortened about 100 m during 1 year after September 1977. Complementing the above measurements, Maekawa and Watanabe (1981, Figure 3) monitored the N-ward displacement of the caldera wall, along the line NR – HK. The results are shown in Figure 12 where there is a stationary point (X in Figure 12) in the displacement. For 11 months from July 1980 to May 1981, NR point on the northern rim had displaced about 1.3 m towards the NE in reference to the base station HK while point X and the lower point (UVO) displaced very little and discontinuously.

These results mean that point X is located at the boundary line between the movable summit part and the fixed base. The boundary line is the tectonic structural line (T.S.L.). It is remarkable that the 1910 eruption formed many craterlets roughly along T.S.L. The ground at the lake side of T.S.L. had remained stationary while the Usu summit side had tilted. The block movement of the summit part shall be discussed in Subsection 5.3.

A summit block containing NM – OY line in Figure 5 started to rise simultaneously with the eruptive activity. The driving force was believed to be magmatic pressure that caused the summit block to tilt. The block sheared along NM – OY line. A hypothetical profile of the summit block is shown in Figure 13.

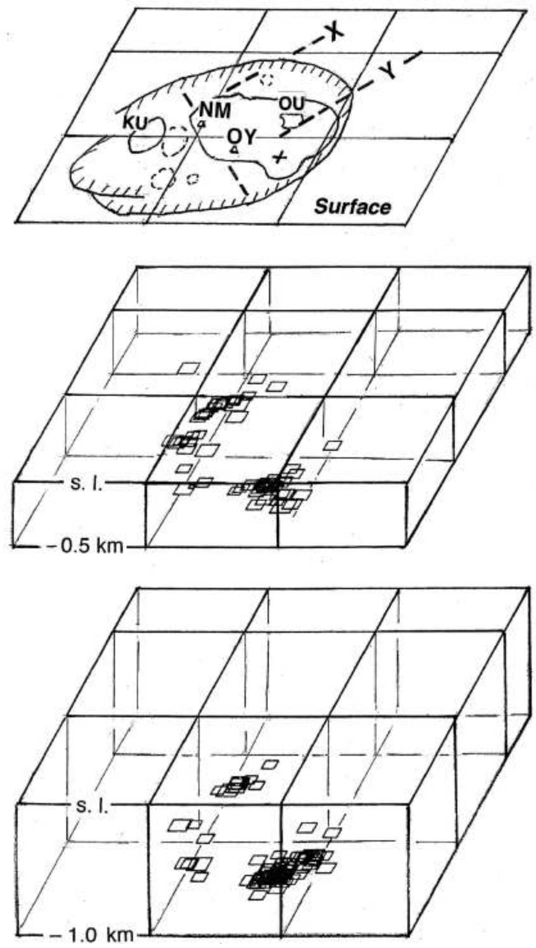



**Figure 10.** Hypocenter distribution of magnitude M 3.8 ~ 4.3 during October 1977 ~ July 1979 classified according to depths. At the deeper depths, earthquakes are of smaller magnitude. The surface shows the topographies as of 1982.

*4.3 Block movements of the summit part*

Deformations of the summit part of Usu Volcano observed during the 1977~1982 activity suggest block movements. A working hypothesis to interpret the block movements is proposed. To prove validity of the model, the block movements will be correlared with fault movements at the block boundary. Yokoyama and Seino (2000) presented a hypothesis that the summit part extending from Craterlet No. 4 to the tectonic structural line (ab. T.S.L.) at the N base and to the bottom of 500 m (b.s.l.) tilted accordingly as volcanic activity developed. In the present discussion, the previous model is improved. On Figure 5 (b), NM-OY line apparently moved toward the NE, but actually the summit block tilted as shown in Figure 13 where the tilting axes of the three points, NM, Craterlet No.4 and NR, converge approximately at P point at a depth of about 1.5 km b.s.l. Thus, the block is determined to be larger than in the previous model. In Figure 11, the N part of Ko-Usu and OY line had moved NE-ward while the SW part did not move. Their tilt angles amounted to 8, 8 and 5 degrees, respectively. The last one may have resulted from resistance, because the N ground is relatively immobile at T.S.L Thus it is presumed that the ground remained fixed at a depth of P point (1.5 km b.s.l.). The NR-SR profile of the summit part of the volcano is shown in Figure 13 where NM-OY line in Figure 5 (b) and NR (the N rim) in Figure 11, shifted NE-ward or actually tilted about 8 degrees around the common pivotal point P which is assumed to be located at 1.5 km b.s.l. This trial model is very approximate.

According to the shear fracture model, the branch of the conduit leading to Craterlet No. 4 should be located at about 0.5 km b.s.l. in Figure 13. A substantially strong force N acted at point Q to tilt the block Q-Cr-X-P around pivotal point P. In Figure 13, the 1910 eruption started along the tectonic structural line (T.S.L.), and its magma conduit may have been derived from a deeper source below point Q (cf. Figure 15).

The upper surface of the block can be defined, but deep structure is uncertain and may be shallower or deeper around the pivotal

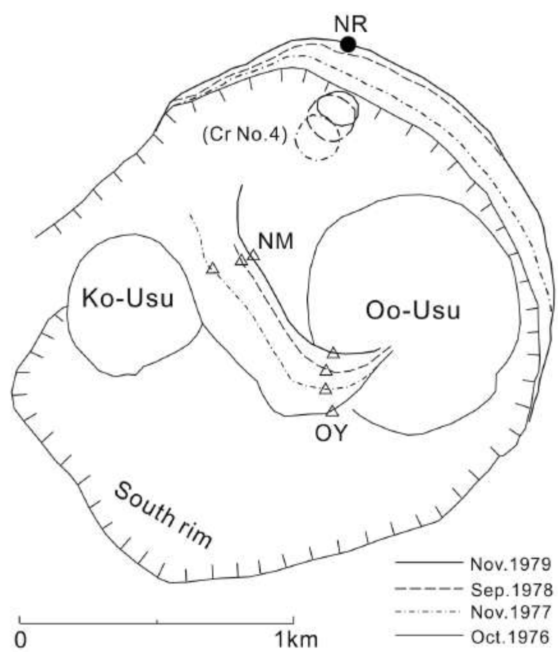



**Figure 11.** Horizontal displacements in the summit crater during Oct. 1976 to Nov. 1978 (after Geographical Survey Institute).

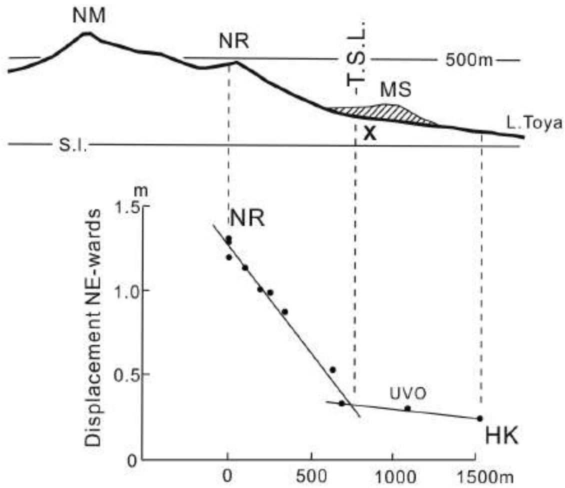



**Figure 12.** Distance measurements between the northern slope of the volcano and the base point (NR – X – HK in Fig. 5(b)) for the period July 1980 to May 1981. MS: the 1910 mound shown in Figure 4. Point X is the hinge-line of both parts (after Maekawa and Watanable, 1981).

point. The rotational movement is defined in Figure 13, which shows deformation of the NE-ward profiles observed during the period from Oct. 1976 (before the eruption) to Nov. 1980. At some particular points such as the main craterlets (Cr), Craterlet No. 4 and NR (the N rim), their rotation angles around pivotal point P were estimated. As shown in Figure 13, the first two rotated about 8 degrees and the last one about 5 degrees: The last (NR) was at the edge of the summit block and could not deform further. Additionally, the summit block may have tilted about 8 degrees around pivotal point P.

The crater deposits between the S crater rim (SR) and Cr – Q line (in Figure 13) may be soft and inelastic as previously referred in the horizontal deformation of the summit part (Figure 11) and thus did not present any detectable deformation. The depth or configuration of the crater deposits is uncertain but assumed to be at the level of Q point at a depth of 1.5 km b.s.l. Thus Cr – Q line may have deformed by upward pressure N acting at point Q.

*Rotational kinetic energy and earthquake occurrence:*

The dynamics of the summit block shall be interpreted from mechanical movements. It is assumed that the summit block repeats rotational and upheaval movement stepwise around the pivotal point. Here angular velocity, moment of inertia and force are denoted as $\omega$, $I$ and $N$, respectively. The following relationships are then used:

$$N = I\frac{d\omega}{dt} \tag{3}$$

and

$$I = M \bullet r^2 \tag{4}$$

where $M$ and $r$ denote mass of the block and lever length between pivot P and the center of gravity G of the block, respectively. In the following, configuration of the block and the above factors cannot definitely be determined and shall be estimated within an order of magnitude on Figures 5 and 13: the side length is roughly equal to NM – OY line, about 1.5 km, and the depth probably range from the surface of the summit (about 0.5 km a.s.l,) to the level of 1.5 km b.s.l. as indicated by volcanic earthquakes that occurred above this depth (cf. Figure 10).

In the present case, pivot P is assumed to be located roughly at the same level as the bottom. Thus the summit block is determined to be cubic, having the side [Cr – X – P – Q] in Figure 13 and its volume is roughly estimated at 8 km$^3$. Its mass (M) is estimated at 2 × 10$^{13}$ kg, assuming a density for the summit part of 2.5 × 10$^3$ kg / m$^3$. In this case, the lever length r is equal to the distance between the pivot and the center of gravity (G) of the block, which is about 1.4 km. Subsequently, the rotational kinetic energy $K$ is given as

$$K = 1/2 \bullet I\omega^2. \tag{5}$$

In this case, angular velocity ω is evaluated by streaks on the upheaval boundary indicated by earthquake faulting exposed at the surface and duration times of the upheavals. The former is indicated by a photo of the streaks (Photo. 2), and the latter is estimated by seismograms registered by upheaval movements. Photo. 2 was taken by T. Maekawa

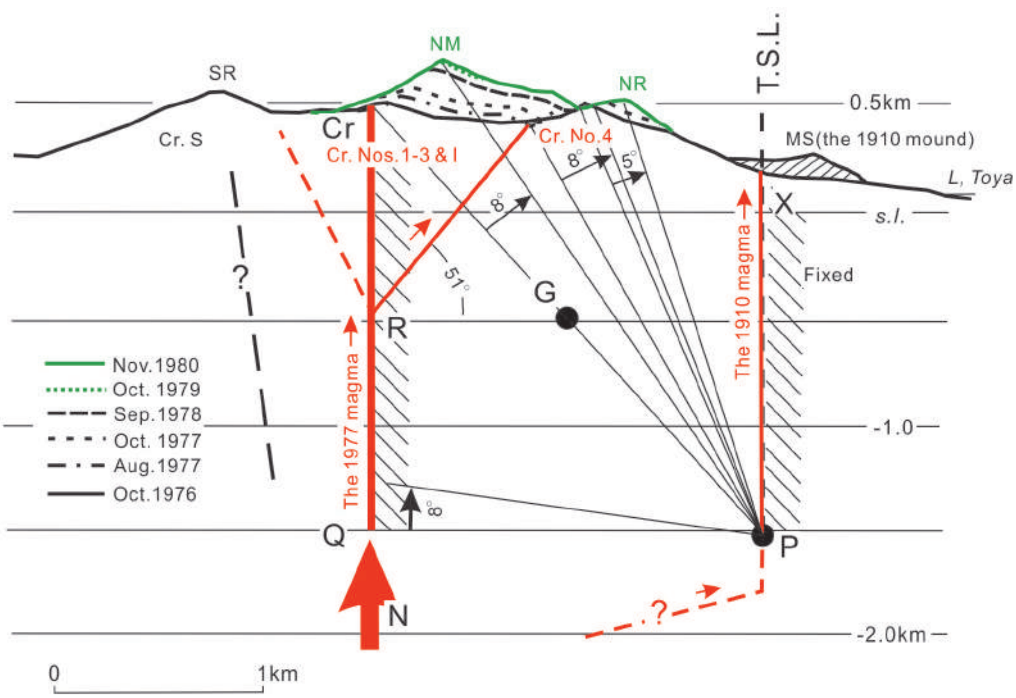



**Figure 13.** Schematic figure of tilting of the summit block. A vertical profile of Usu Volcano in the NE direction, in Figure 5.
G: Center of gravity of the summit block (Cr – X – P – Q),
NM: New mountain, NR: North rim, MS: Meiji-Shinzan (a new mound formed in 1910),
T.S.L. or X – P line: Tectonic structural line, P: Pivotal center of the rotation,
Q : Assumed magma front pressing the summit block in 1976 ~ 1982,
R : Branch point of magma conduits for Craterlet No. 4 ("Golden Pond") and Craterlet S ("Silver Pond") determined by the shear-fracture model,
**Cr**: Site of Craterlets group Nos. 1 ~ 3 and I. Point Cr moved to point NM during Aug. 1977 to Nov. 1980.
**Cr. S**: Craterlet "Silver" composed of craterlets. J, K, L and M, not on the same plane as Cr. No. 4.

and H. Watanabe (pers. com.) on Sep. 12, 1978 at a scarp marked by an asterisk at the S of OY in Figure 5b. During this period, the block uplifted stepwise 2.5, 3.5 and 5 cm, all being accompanied by relatively large earthquakes of M 3 ~ 4, located at levels of – 0.5 and – 1.0 km below mark (+) on the surface in Figure 10. Though their occurrence times are not exactly known, reference to the seismograms observed at approximately similar periods are made. Mizukoshi and Moriya (1979) observed volcanic earthquakes from December, 1977, to October, 1978, by accelerographs and displacement seismographs of low magnification at the Usu Volcano Observatory (UVO in Figure 5b). The three displacement seismograms indicated the block movement occurring from the S base of Oo-Usu lava dome (+ mark in Figure 10) are as shown in Figure 14.

The authors interpret that NM – OY line in Figure 5b formed by an upheaval of a summit part and the upheaving boundary is at a surface earthquake fault, and a slip plane (a kind of slickensides) produced by upheavals of one-sided land block. Probably in the case of formation of Showa-Shinzan (SS) lava dome in 1943 may have produced similar slip planes under the ground surface, even though they were not found. In case of Showa-Shinzan, solidified magmas continuously upheaved causing slickensides while in case of NM – OY line, magmas upheaved stepwise causing fault-like traces.

In Figure 14, the rise-time is defined as the time needed for a certain place on the fault plane to displace from the beginning to the end. The rise-times derived from the seismograms shown in Figure 14 and are about 1 sec. in average. Then, angular velocity $\omega$ is estimated as 0.03 m / 3 km / 1 sec.

Thus rotational kinetic energy $K$ is calculated as:

$$K = 1/2 \bullet Mr^2 \omega^2 = 2 \text{ x } 10^9 J \qquad (6)$$

A part of the kinetic energy can be attributed to the occurrence of earthquakes of magnitude 3 which discharge kinetic energy of $2 \times 10^9$ *Joule*. In the above calculation, the energy necessary to overcome frictional resistance is not considered during the movements. If this effect is considered, the related earthquakes of magnitude M 3 ~ 4 can be explained. By the above calculations, it is proved that kinetic energy necessary to tilt the summit part is just enough to cause the seismic movements observed at an edge of the block.

*4.4 Eruption mechanism of Usu Volcano determined from the observations in the 20th century*

Eruption activity of Usu Volcano discussed in Sections 3 and 4 should afford an inductive model of its eruption system even though it may be revised with further analyses.

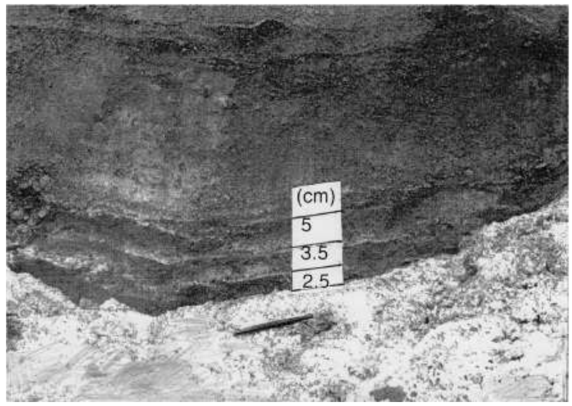



**Photo 2.** A fault scarp as of Sep. 12, 1982. The location is shown by an asterisk in Figure 5b. This side is covered with sublimates and a ball pen is a scale.

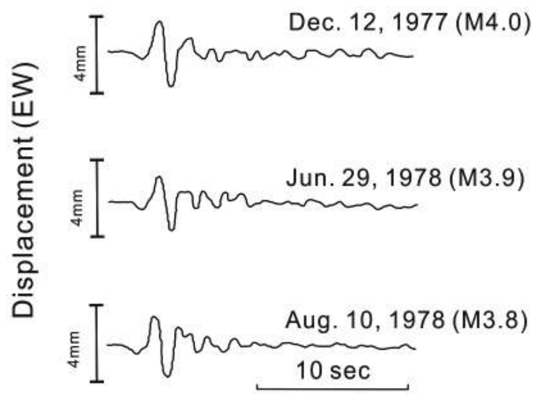



**Figure 14.** Seismogram-samples of earthquake family having large amplitude, mostly located at the southern base of Oo-Usu lava dome (+ mark in Figure 10 top) (The traces are taken from Mizukoshi and Moriya, 1979).

Originally, the main vent was thought to be roughly at the center of the summit crater, measuring about 2 km in diameter, and the vent was suspected to be blocked with Oo-Usu and Ko-Usu lava domes. At the beginning of the 1977 eruption, magmas may have ascended through the main conduit and erupted beside Ko-Usu dome, forming the three craterlets. Also, magma channeled through the main conduit caused eruptions of the twin parasitic vents, "Golden Pond" and "Silver Pond" in pre-historic times. The magma pressure from the conduit system presumably caused tilting of the summit block. Magma intrusion at the upper part of the main conduit was verified by the electrical resistivity surveys (cf. 5.3). The first three eruptions may have partly deviated from the main conduit, and passed beneath the old conduit to Ko-Usu lava dome because the dome subsided about 50 m simultaneously with the early eruptions. The central conduit may have been active in supplying magmatic material to parasitic eruptions.

Magma supply systems of Usu Volcano that functioned in historical times are schematically shown in Figure 15 in the SW – NE cross-section. In a previous study, De la Cruz-Reyna and Yokoyama (2011, Figure 11) brought up the parasitic vents of Usu using a fracture criterion for brittle materials. Here, the present authors include the block movements and local structure of the NE sector of the volcano in Figure 15. We assume that Ko-Usu, Oo-Usu and SS-lava domes formed in 1769, 1853 and 1944, respectively, may have developed directly from the magma reservoir because supply of the juvenile dacitic magmas may need special conditions. On the other hand, parasitic eruptions, such as the 1910 and the 2000 eruptions, may have been fed through branches from the main central conduit. In Figure 15, only the path of the 1910 eruption along a lateral side of the tilting block and T.S.L is indicated; the 2000 eruption took other paths that branched from the central conduit. Such idea may develop to an assumption that magma-supply routes are related to local melting condition at starting point from the reservoir or that a magma reservoir is not always uniform in melting condition.

In the future, the above suppositions should be revised with more advanced studies of geophysics and petrography.

## 5. Subsurface structure of Usu Volcano

Knowledge of the subsurface structure of an active volcano is usually the result of multiple observations, from monitoring data during periods of activity and from exploration methods during quiescent periods.

To further understand volcanic eruptions, it is essential to explore the detailed subsurface structure of the volcano. The research directions and methods should vary with magma types, history of the eruptions and the present knowledge of the structure. One of the actual ultimate purposes on studying the subsurface structure of Usu Volcano is defining the parameters of its magma reservoir, such as its location, structure and action mechanisms. Data bases covering extended periods of time are essential for understanding the long-term changes in volcanic structure and behavior.

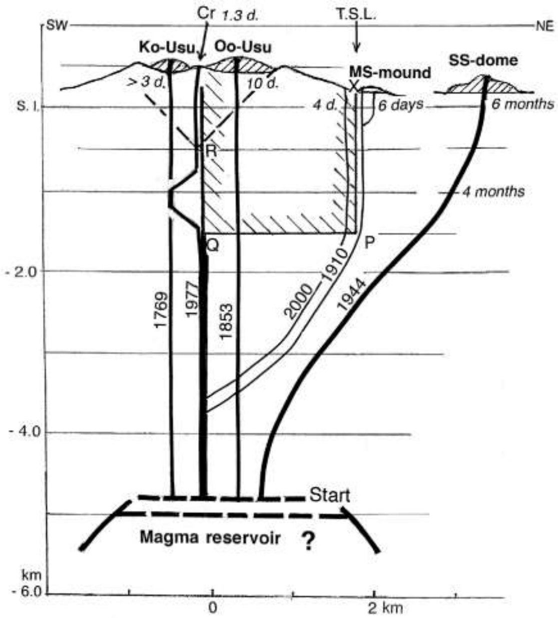



**Figure 15.** An assumptive and schematic model of eruption system of Usu Volcano, projected on the SW-NE direction. Days and months denote precursory times before the outbursts. Cr denotes the positions of Craterlets Nos. 1, 2 and 3 of the 1977 eruption. T. S. L. denotes a tectonic structural line appeared in the 1910 eruption at the eastern foot of the volcano. This figure overlaps partly with Figure 13.

### *5.1 Seismic structure*

Real-time acquisition and precise analyses of seismic data are most important processes to interpret the dynamic state of the volcano. At the time of the 1977 eruption of Usu Volcano, a new volcano observatory was just under construction. Okada *et al*. (1981) located the hypocenters of the volcanic earthquakes as mentioned in Subsection 4.2 and proved that some relatively large earthquakes were related with tilt movements of the summit block.

During the 2000 eruption of Usu Volcano, Onizawa *et al*. (2002) studied three-dimensional P- and S-wave velocity structure around the volcano using traveltime data of the volcanic earthquakes. After the eruption, Onizawa *et al*. (2007) studied the three-dimensional P-wave velocity structure beneath the volcano using 288 temporary seismic stations and seven programmed dynamite-explosions. The velocity structure down to the Pre-Neogene basement below the volcano or to a depth of about 3 km was revealed and low-velocity (1.5 ~ 2 km /

s) area at depths 3 ~ 4 km was found. The most prominent feature of the velocity model was the deepening of the basement toward the S-SW. This feature agrees with decrease of the gravity anomaly toward the S, at the S of the volcano as shown in Figure 2.

Accurately determined velocity structure is necessary to precisely locate hypocenters of volcanic earthquakes, and permitted the relocation of the precursory earthquakes of the 2000 eruption by Onizawa *et al*. (2007), as shown in Figure 16. The seismic activity was then divided into three patterns: (1) a quasi-vertical distribution indicating magma ascent toward the summit, (2) a N-ward distribution indicating the subsequent eruptions at the NW base, and (3) S-ward distribution indicating S-ward intrusion of sills. The present authors notice that the last pattern (3) indicates possibility of parasitic eruptions at the S base. Additionally there is Minami-yama (MY) mound of unknown age on the topographic maps (Figure 4).

*5.2 Aquifer structure beneath Usu Volcano*

Aquifer structure beneath and around any volcano fundamentally affects magmatic activity. Usu Volcano is located between Lake Tōya (83 m a.s.l.) and Eruption Bay (Figure 4). In general, underground water flows from the lake towards the sea smoothly even if there are temporal rises of the water level due to volcanic activity. In fact, coinciding with the 1977 eruption of Usu Volcano, the water level at GS-R1 well (Figure 4) at the E base rose 37 m and then recovered after about 4 years. Watanabe (1983) interpreted these changes as a result of increased pore pressure diffusion beneath the summit crater. Such changes in water level are relatively small in comparison with the volcano height, of about 600 m, and similar changes were not reported in the case of the 2000 eruption. Also, the flow rates may not be high because the gradient of the flows is roughly 83 m / 7 km. Thus, the aquifer beneath Usu Volcano may be almost flat, flowing from 83 m (a.s.l.) to the sea. Ascending magmas should first contact the Quaternary aquifers which are located at the upper boundary of Neogene period layers, as shall be discussed in the next subsection.

*5.3 Electrical-resistivity structure of Usu Volcano*

Underground structure of Usu Volcano was investigated using magnetotelluric soundings (Ogawa *et al*., 1998; Matsushima *et al*., 2001). A characteristic structure of the edifice is a

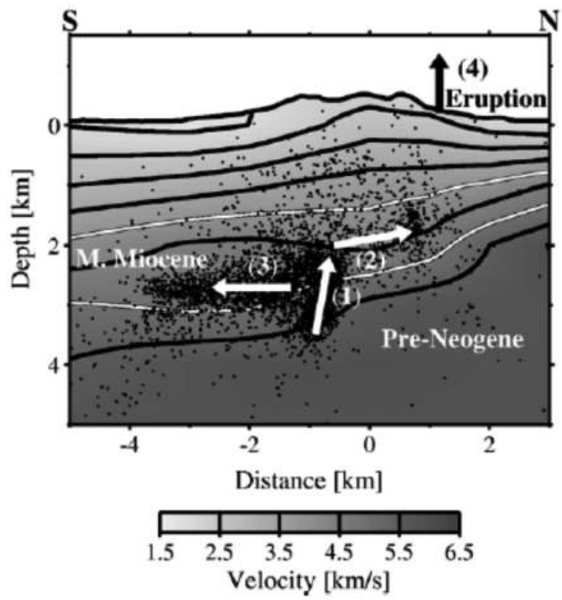



**Figure 16.** Earthquake hypocenters during the precursory stage of the 2000 eruption and the sequence of the possible magma movements projected on the N–S velocity cross-section after Onizawa *et al*. (2007, Figure 14). The dots represent the hypocenters during the precursory stage, and two white lines denote the boundaries of geological strata. The white arrows indicate direction of the magma movements.

thick low resistive layer (<10 Ωm) located from sea level to 2 km b. s. l. (Figure 17). This low resistive layer is thought to contain highly altered rocks produced in the Neogene period. The base line (point Q and point P in Figure 13) of the block tilt model proposed in the present paper corresponds to the lower boundary of the low resistive layer. At this boundary, the physical property changes among the different formations. The earthquakes associated with the 2000 eruption of Usu Volcano showed a unique distribution which spreads horizontally with time (Onizawa *et al*., 2007). The precisely relocated hypocenters, based on a three-dimensional P-wave velocity structure, are located just below the lower boundary of the low resistive layer (Figure 17). This distribution also suggests the existence of a sharp boundary of the physical property. A sufficiently large amount of magma was located at near point Q to induce tilt movement of the volcanic edifice. Figure 16 indicates that the number of volcanic earthquakes is few at the location of the presumed magma body. The limited number of earthquakes may indicate that the magma body is still at high temperature and molten or partially molten. However, the magma body could not be identified from resistivity data. This is due to the insufficient sensitivity of the anomaly below the thick low resistivity layer

(Matsushima *et al*., 2001), or lack of resistivity contrast between the cooling magma body and surrounding formations.

A relatively high resistive part (>500 Ωm) is located below Craterlet I at a depth from 200 m (a.s.l.) to 400 m (b.s.l.). This resistive part is interpreted as the cooling magma which may have intruded to a shallower part of the 1977 eruption site because the corresponding resistive part was not observed by the surveys just after the eruption (Ogawa *et al*., 1998). This intrusion is probably derived from the deep-seated magma body near point Q (Figure 13) which induced the tilting of the volcanic edifice. The inhomogeneous structures such as the original vents associated with Craterlet "Silver Pond", and Ko-Usu lava dome acted as a buffer between the NE inclined block and SW stationary block. The magma intruded into the shallow part at the boundary between the two blocks to compensate for the horizontal extension produced by the inclined movement. The upheaval of "New Mountain" (NM) continued to 1982, suggesting the deep-seated magma body had sufficient magmatic force causing further intrusion into shallow part. The phreatomagmatic explosions occurred repeatedly at Craterlet "Silver Pond" in 1978. The scale of the explosion and the content of fresh magma output increased gradually after the onset of the explosions (Niida *et al*., 1980). At the 1945 eruption of Usu Volcano, phreatomagmatic explosion started when the rising magma contacted the Quaternary aquifer (Yokoyama and Seino, 2000). The increase of pore pressure with vaporization of the water which leads to the rock failure, depends on the permeability of the formation: the lower the permeability, the larger the pore pressure (Delaney, 1982). On the other hand, an explosion, which is induced by vaporization, needs sufficient water in a permeable formation (Wohletz, 1986). A phreatomagmatic explosion occurs by satisfying these contradicting conditions. Although the hydrological condition at the summit of Usu Volcano is unknown, phreatomagmatic explosions at Craterlet "Silver Pond" are supposed to start when the intrusive magma passed through the Neogene layer constituted by less permeable altered rocks. The highly resistant part observed by magnetotelluric soundings probably indicates cooling of the intrusion at a shallow depth. The accompanying heat discharge from the surface after the 1977 eruption suggests that the pore of the high resistive part was filled with superheated vapor (Matsushima, 2003). Geochemical analysis of fumarole at Craterlet I indicates that the gas contains foreign water and the water vapor must be heated sufficiently before the mixture with the volcanic gas (Tomiya *et al*., 2012). These observations indicate that the highly resistive part also contains the high temperature rocks heated by magma and volcanic gas.

*5.4 Structure of the summit part surveyed by muography*

Installation sites of muographic equipment for volcanoes are limited mainly by topographic conditions: On and around Usu Volcano, some of the extruded structures were studied by muography, such as, the 1943 lava dome (SS) by Tanaka and Yokoyama (2008), and the 1910 mound (MS) by the same authors (2013).

Kusagaya and Tanaka (2015) developed a multi-layered telescope using seven detectors for selecting linear trajectories and preliminarily carried out a muographic survey for the summit part of this volcano with this muographer installed at the Usu Volcano

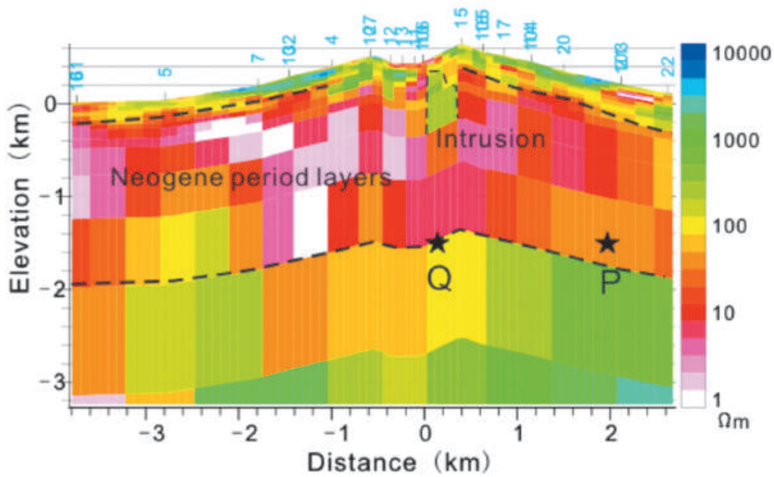



**Figure 17.** Resistivity cross-section through the center of Usu Volcano from SW to NE, obtained by the two-dimensional inversion of AMT and MT data, and its geological interpretations after Matsushima *et al.* (2001, Figure 5). "Intrusion" in the central part corresponds to the central magma conduit in Figure 13. Underground structure is divided to three layers by two broken lines; Quaternary, Neogene and Pre-Neogene period layers from surface to depth. The low resistivity area bounded by broken lines corresponds to the altered Neogene-period layers. Q and P are the same as those in Figure 13, and they are located on the lower boundary of Neogene period layers.

Observatory (UVO in Figure 5). A muograph obtained after 38 days is shown in Figure 18b which shows a two-dimensional density map of the vertical cross-section along line [A–B] in Figure 18a.

A muograph of the summit part is shown in Figure 18b where the present muographic observation can detect the volcano structure roughly above the elevation angle of 150 mrad from the muographic telescope, or roughly, above 300 m (a.s.l.) in average. We may assume this anomaly below [OY-NM] line or [A-B] line, and here the former case is examined:

Along the elevation angle 150 mrad, anomalous material with density $\rho$, and range of muon path $\Delta l$ are expected. The total path length of muon is about 2000 m in Figure 18c, the width of the intrusion is $\Delta l$, and the average density $\rho$ along the path is assumed to be $2.1 \times 10^3$ kg/m$^3$. Then, we obtain the following equation:

$$\Delta l \times \rho + (2000 - \Delta l) \times 2.0 = 2000 \times 2.1$$

where $\rho$ and $\Delta l$ are unknown. (7)

Here, a probable $\Delta l$ as 200 m was adopted considering the results of electric-resistivity surveys in Subsection 5.3, then the following relation is obtained:

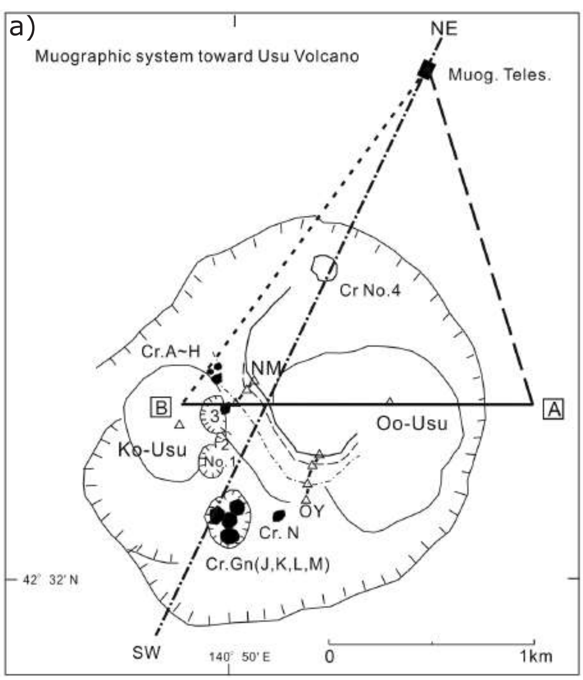
$$200 \times \rho + (2000\text{-}200) \times 2.0 = 2000 \times 2.1 \qquad (8)$$

The density of the anomalous material should be $3.0 \times 10^3$ kg/m$^3$. The horizontal position of the material of high density on the 150 mrad line cannot be precisely determined. Such material may be the remnant of intruded magmas in the past, not in the 1977 eruption. Another observation from the EW direction is needed. In the present case, the time for the anomalous material intrusion could not be determined.

*Magma reservoirs under Usu Volcano*

Location and properties of magma reservoirs under active volcanoes are important parameters that influence their activities. After the 2000 eruption of Usu Volcano, magma reservoirs under the volcano were investigated with seismological and petrological methods.

Seismological methods: Generally the standards for detection of magma reservoirs by seismometric methods should be seismic wave velocities propagating through them, and reflections and refractions at their boundaries. First, the seismological method and results at Usu Volcano are summarized as follows: Onizawa *et al*. (2007) studied the three-dimensional P-wave velocity structure of Usu Volcano by using first arrival time data of the 2001 active seismic survey. Hypocenters of precursory earthquakes of the eruption were accurately determined (Onizawa *et al*., 2002, 2007) and most of these were confined to depths of less than about 4 km. The scarcity



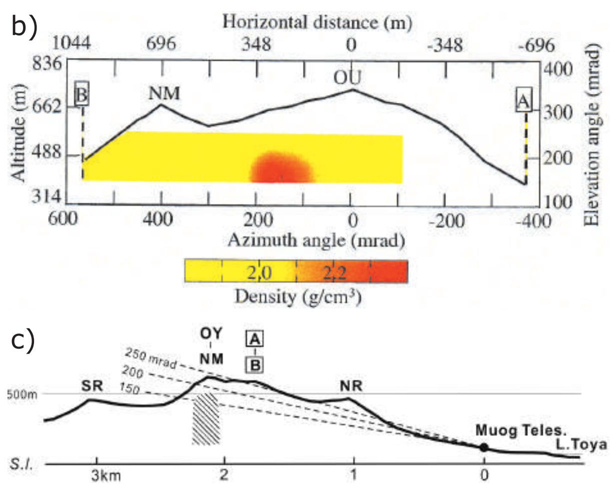




**Figure 18.** Muographic structure of Usu Volcano.
a) Muographic instrumentation after Kusagaya and Tanaka (2015)
b) A muograph along line [B-A] in Figure 18a (Kusagaya and Tanaka, 2015),
c) Interpretation: a hatched part is an assumed magma intrusion. Its muon path-length is expressed as $\Delta l$.

of deeper earthquakes is considered to be caused by the ductile behavior of the crust as a result of its elevated temperature (Onizawa *et al*., 2007), suggesting the presence of magma bodies below 4 km depth. Large magma bodies have not been detected to a depth of about 7 km by seismological observations (Onizawa *et al*., 2007) nor electrical measurements (Matsushima *et al*., 2001).

Activity related to magma reservoirs beneath Usu Volcano, if any, should be governed by characteristics of dacitic magmas, of which viscosity is relatively high, and drastically changes with temperature and pressure. Goto (1997) measured viscosity of dry melts of Showa-Shinzan (SS dome) lavas, such as $10^7$ Pa•s at 950 °C, and $10^9$ Pa•s at 850 °C. The viscosities of magmas would increase by degassing (mainly $H_2O$) at the same temperature. At the boundary zone of a magma reservoir, all factors such as the material, temperature, mechanical strength and viscosity, would change continuously toward the center. Therefore such magma reservoirs cannot be considered as a simple sphere. Furthermore configurations of reservoirs may vary, from simple spheres to complicated assemblies of sills or dykes which are related to the genesis of the reservoirs. At present, their location can only be reliably discussed by seismic wave velocities.

The 2000 eruption was observed from various standpoints of seismology. In the present paper, a hypothesis is presented considering that usual parasitic eruptions of this volcano may be driven by magma branches or derivatives of the magma reservoir not by the magma reservoir itself (cf. 4.4 and Figure 15); only one exception is the 1943 eruption of SS dome. The 2000 eruption was not directly fed by the magma reservoir. Hence, within the reservoirs, physical parameters such as temperature and viscosity, may not always be uniform.

Petrological methods: Tomiya and Miyagi (2002) considered magma movement during the 2000 eruption of Usu Volcano as follows: (1) the 2000 eruption started by the ascent of magma from the deep (high-P) chamber; (2) this magma from the deep chamber was injected into the shallow (low-P) chamber; (3) this injection triggered an eruption at the surface from the shallow chamber. This process is similar to those proposed for other historical eruptions, all based on petrographic observations (Tomiya and Takahashi, 1995).

Tomiya *et al*. (2010) studied the rhyolite pumice from mafic magma that existed in the 1663 eruption (Us-b in Table 1) by high-pressure melting experiments, and determined the depth of magma reservoir just before the 1663 eruption as around 200 – 250 MPa (high-P), corresponding to a depth of about 8~10 km based on a crustal density structure. They consider this depth to be the level of mafic magma that existed beneath the rhyolitic magmas just before the 1663 eruption. On the other hand, Tomiya and Takahashi (1995) studied the eruptive products of Usu Volcano after the 1663 eruption and proposed that a new shallow magma reservoir (low-P) formed during or just after the 1663 eruption. On the basis of experimental results on the eruptive products, the pressure for the low-P magma is estimated at 100 ~ 150 MPa, corresponding to a depth of about 4 ~ 6 km. According to Tomiya *et al*. (2010), this depth may also correspond to the level of neutral buoyancy for the dacite.

Geophysical methods, such as seismometric observations, have to be used to detect any deeper magma reservoirs beneath Usu Volcano and to confirm the results from petrological studies.

## 6. Concluding remarks

Interpretations of various eruptive events of Usu Volcano are presented in the above discussions. A hypothetical model of eruption activities mainly observed in the 20th century is tentatively developed below.

Based on P- and S-wave velocity structure and electrical resistivity structure, the shallow part of the volcano was modeled after 2000 by Onizawa *et al*. (2007) and Matsushima *et al*. (2001), respectively. Location of all of the magma reservoirs has not yet been established.

In the present discussion, the following results were reached:

a) Parasitic vents in the summit crater, “Golden Pond” and “Silver Pond” formed as twin parasites before historical time and reawakened in the 1977 eruption.

b) On Usu Volcano, during the last 350 years, its eruptions occurred almost regularly in time, and the ejected volumes have been dispersed, and generally the summit eruptions ejected larger volumes than parasitic ones.

c) The 1944 lava dome (SS dome) formed as a parasite at the base. This may be exceptional in the history of this volcano.

Juvenile magma rarely reached the basal surface. It is recognized that the conduit to this lava dome was derived from a central part of the parental volcano. At present, this is the southernmost branch from the center of the volcano. It is not clear why the magma changed the direction at YH point. Based on the magma activities during the 1944 eruption, it seems possible that magma will extrude towards the S base in the future. S-ward migration of magmas just before the 2000 eruption observed by Onizawa *et al*. (2007) may support this possibility.

d) The 1977 eruption took place very near Ko-Usu lava dome that formed in 1769 AD. The conduits of the 1977 eruption have probably been very close to the lava dome. The magma conduit of Ko-Usu lava dome played a role at the first outbreak. Its magma reached the summit crater partly via the old conduit of Ko-Usu, rapidly in 32 hours, and without strong resistance accompanied only by small earthquakes, M≦3.7 in magnitude.

e) Analyses of the block movements at the summit part in the 1977 eruption indicate strong magmatic pressure that may cause the mountain to collapse.

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