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# Holocene landscape evolution of a nivation hollow on Gassan volcano, northern Japan

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## Abstract

This paper describes the results of detailed surveys for the landscape systems (landforms, vegetation, topsoils and snow cover duration) of a nivation hollow in northern Japan and discusses their evolution in the Holocene epoch. The nivation hollow studied consists of three concentric zones whose landscapes and historical development are different. The outermost zone where snow disappears early is covered with dwarfed trees, *Sasa kurilensis* (subalpine bamboo) thicket and snowbed grasses. Fossil solifluction lobes and drainage channels are common. In this zone, slope stabilization and vegetation establishment (penetration and settlement of vegetation on slopes) followed by pedogenesis occurred after 12,350 cal BP. In the middle zone, slopes are mostly covered with snowbed plants, and turf-banked terraces and minor slumps are observed. This zone experienced slope stabilization and vegetation establishment followed by pedogenesis after 4870 cal BP. The innermost zone overlaps with the snow-induced bare ground in the centre of the nivation hollow basin. Active geomorphic processes operate here and traces of surficial wash and rills are abundant. Humic soils are not present in this zone. These differences in landscape development of the nivation hollow may reflect the temporal changes in the timings of snow disappearance associated with the Holocene climatic variabilities.


*Keywords:* Nivation hollow; Snowpatch; Alpine landscape; Holocene epoch; Snowy mountain

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## 1. Introduction

A nivation hollow (Matthes, 1900) is a typical climatic landform in a snowy region. Late-lying snowpatches strongly affect microclimates and ecosystems on the slopes (Walker et al., 2001) and thus vegetation boundaries analogous to annual changes in snow line positions occur in a nivation hollow, particularly in a circular hollow (Lewis, 1939). Snow-induced bare ground is occasionally formed in a hollow basin because the growth conditions for plant communities, such as temperatures and growing periods, are not fulfilled in a hollow occupied by a perennial to semi-perennial snowbed (Burns and Tonkin, 1982; Iwata, 1983; Kariya, 2002).

In general, the timing of snow disappearance on mountain slopes is constrained by the snow depths and climatic conditions in the ablation season (Watanabe, 1989; Daimaru et al., 2002). If major topographic changes do not occur within a nivation hollow or on the slopes surrounding the hollow, the spatio-temporal fluctuations of snow line positions in a drift-filled nivation hollow will be affected by long-term climatic variabilities. From this point of view, we are able to discuss the palaeoenvironments and palaeolandscapes in nivation hollows (Dohrenwend, 1984; Takada et al., 1990; Christiansen, 1998a,b; Daimaru et al., 2002). Although a nivation hollow is generally under an erosive situation, slope materials including soils are sometimes present in and around hollows (Henderson, 1956; Watson, 1966; Kariya et al., 1996; Christiansen, 1998a). These materials may contain various sources of information useful for reconstructing palaeoenvironments.

The high mountains in northern Japan are extremely snowy although their altitudes do not exceed 3800 m, because the northwest Siberia monsoon crosses over the Sea of Japan

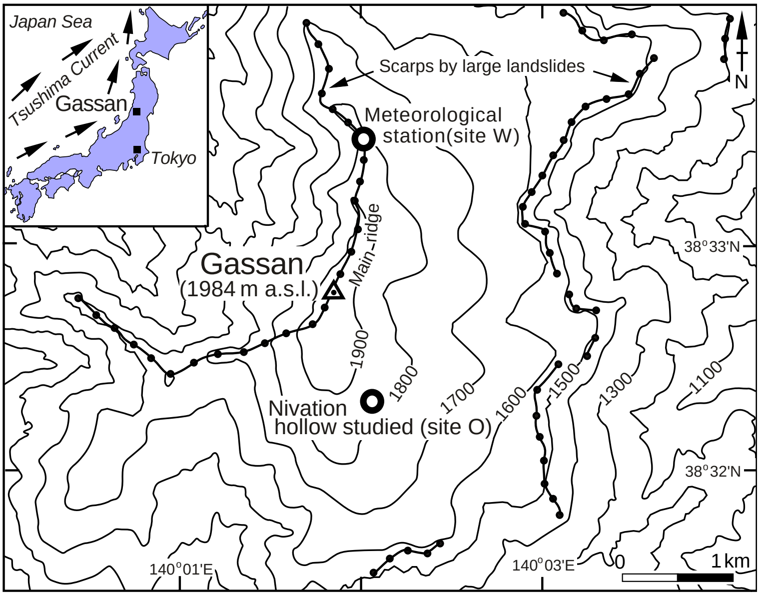



Fig. 1. Location of the nivation hollow studied (site O) on Gassan volcano, northern Japan. Contour interval is 100 m.

and receives a large amount of vapor from the warm Tsushima current flowing from the south. The monsoonal humid air mass climbs the western side of the Japanese mountains and induces substantial snow fall (Fig. 1). Warm summer weather in the Japanese mountains promotes rapid snowmelt and discharge of meltwater, permitting the growth of plant communities even in and around the snowpatch hollows. Therefore, humic soils are formed in the peripheral parts of nivation hollows although snow-induced bare slopes occur at the bottom of a hollow. In Japan, mountain glaciers existed in the last glacial period but their extent was rather small (Aoki and Hasegawa, 2003). Thus, geomorphological and pedological research on nivation hollows is important to discuss the Quaternary developments of climate landforms and landscapes in the snowy mountains of Japan.

In this paper, topography, vegetation and topsoils as landscape components of a large circular type nivation hollow (site O; Fig. 1) in the Gassan volcano, northern Japan, are surveyed to reconstruct landscape evolution of the hollow since the late glacial period. Attention is directed toward the timings of slope stabilization and vegetation establishment in the hollow.

## 2. Study site

### 2.1. Outline: geology, geomorphology, vegetation and climate

Gassan volcano is composed of mid-Pleistocene andesite (Fig. 1). The main ridge running from the north to the south shows an asymmetrical shape; the western slope has been destroyed by large failures whereas gentle lava slopes on the eastern side have been preserved. Active and fossil (i.e., vegetation-covered) nivation hollows and bogs are abundant on the eastern slopes. Although Ono et al. (2003) inferred the existence of glaciers during the last glacial period in Gassan on the basis of the estimated equilibrium line altitudes in East Asia, no evidence of glacial landforms or sediments has been discovered.

The modern climate of Gassan is characterized by abundant snow and rain as well as a strong prevalence of winds. The mean annual air temperature at the weather station on the main ridge (site W in Fig. 1; 1805 m a.s.l.) is below $-0.9$ °C (Kariya et al., 1997; Kariya, 2002). Snowfall begins in early October and continuous snow cover around the summit is established by mid-November every year. A strong wind with an average velocity over 15 m $s^{-1}$ blows from the west or the northwest along the main ridge throughout the year (Kariya et al., 1997). Consequently, in many places on the main ridge, snow thickness is below 0.5 m even in midwinter and early spring, while nivation hollows are filled with thick drifted snow. In large nivation hollows like site O, snow is as deep as 20 m or more. Similar to many other mountains in Japan, the summer season witnesses intensive rainfall due to typhoon or front activities. Total summer precipitation (June–September 1995) at site W exceeds 900 mm.

The highest forest consisting of montane broad-leaved deciduous trees (mainly *Fagus crenata* and *Quercus mongolica* var. *grosseserrata*) ends at 1400–1500 m a.s.l. Above this forest, *Sasa kurilensis* (subalpine bamboo), and dwarfed *Alnus*, *Acer* and *Pinus* dominate,

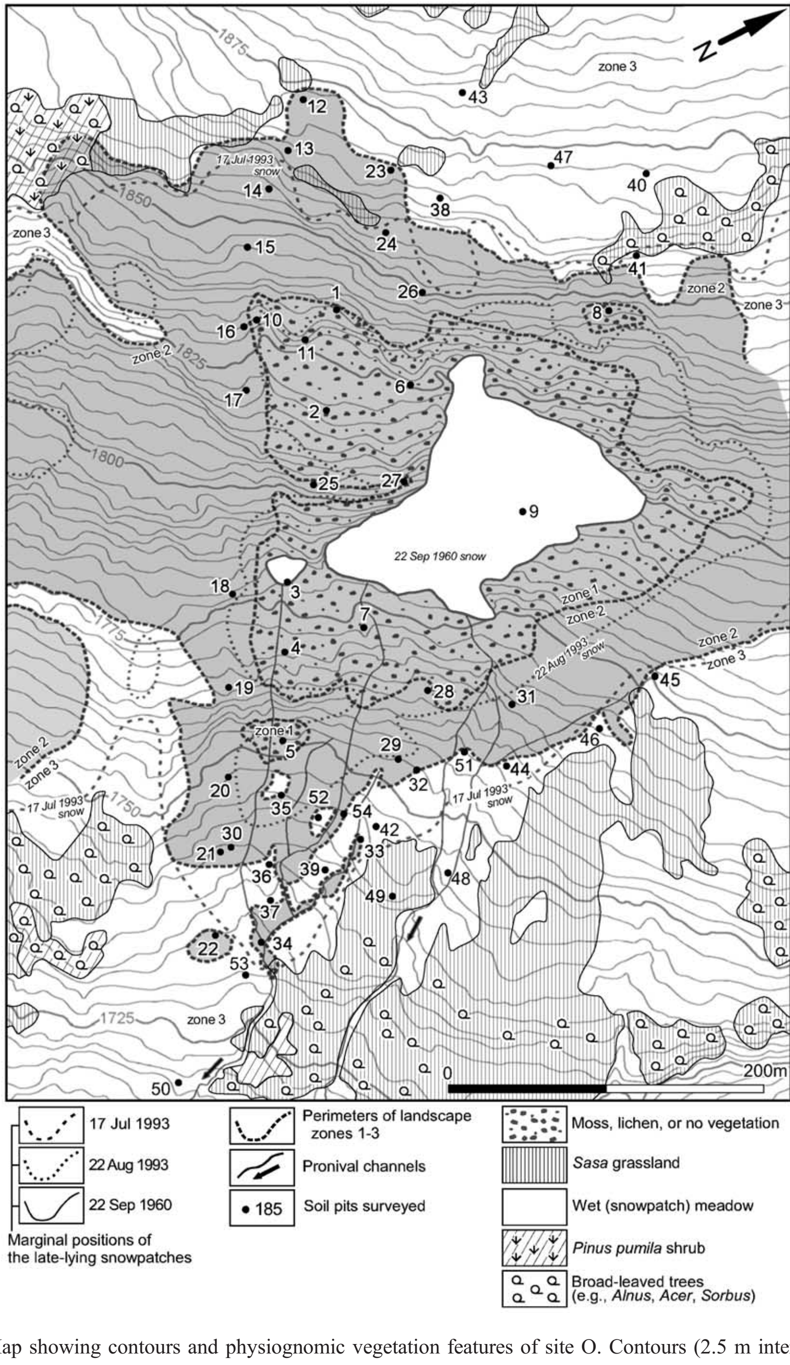



Fig. 2. Map showing contours and physiognomic vegetation features of site O. Contours (2.5 m interval) were drawn from the airphotos taken on 22 September 1960 by the use of an analytical stereoplotter. The late-lying snowpatch margins in 1993 were investigated by field measurements.

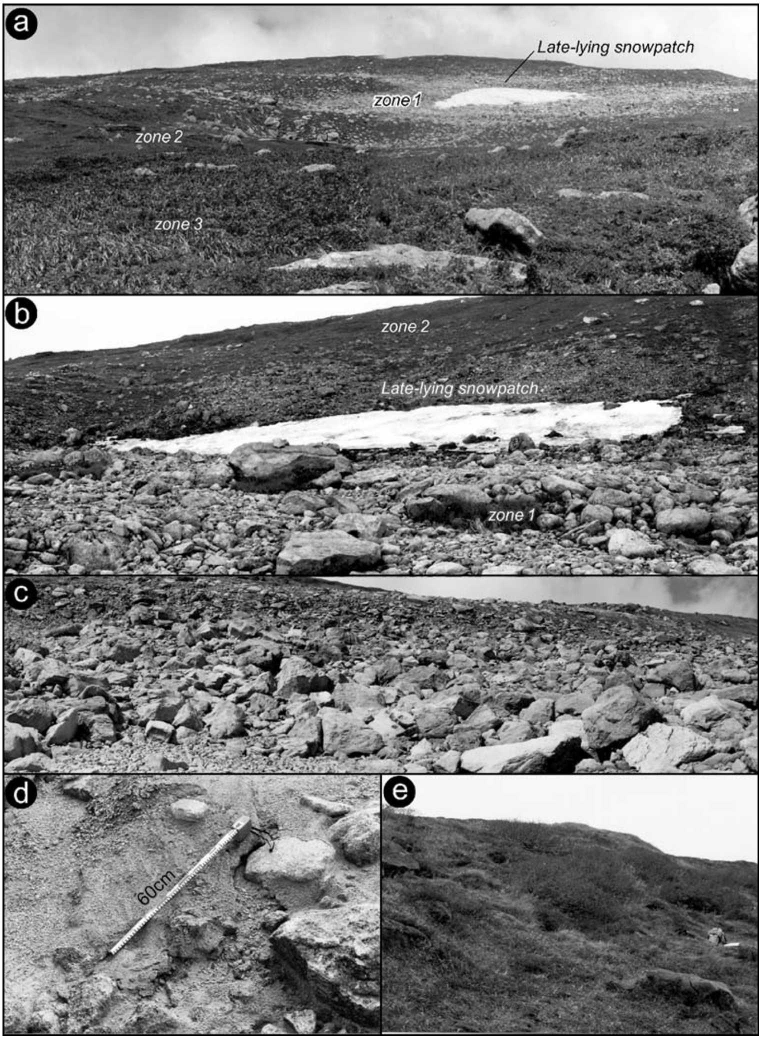



Fig. 3. Landscapes of site O. (a) Whole view of site O from southeast on 25 August 1991. The horizontal span of the snowpatch was about 80 m. (b) Snow-induced bare slopes (zone 1) near Locs. 2 and 6 on 23 August 1991. The horizontal span of the snowpatch was about 50 m. Slope process evaluation was carried out here (Kariya, 2002). (c) Snow-induced bare slopes (zone 1) near Loc. 6 on 23 August 1991. (d) Traces of surface wash at the snow-induced bare slopes near Loc. 2 on 28 August 1991. (e) Densely-vegetated slopes (zone 3) near Loc. 42 on 29 September 1991.

rather than tall trees. Wet meadows consisting mainly of Gramineae and Cyperaceae have also developed around nivation hollows and bogs. Although subalpine conifer forests normally appear above montane broad-leaved deciduous forests, the former has been replaced with dwarfed trees and grasses in some snowy mountains in Japan like Gassan. This type of system is termed a pseudo-alpine zone as it resembles a genuine alpine zone (Shidei, 1952; Sugita, 1992).

### 2.2. Physical conditions of site O

Site O, one of the largest nivation hollows in Gassan, is located 1.5 km southeast of the summit and lies at 1725–1870 m a.s.l. (Figs. 1, 2 and 3). Site O is situated just below the cliffs that have originated from surficial wrinkles or terminal snouts of lava flows. This is the leeward position considering the direction of the winds that prevail in winter. The deepest basin of site O is around 1800 m a.s.l. and its relative depth from the surrounding slopes is 30–40 m. The late-lying snowpatches do not melt completely in the ablation season and persist even in late summer or early autumn, and finally become perennial in the hollow basin. The temporal changes in the snow line in 1993 are illustrated in Fig. 2 and the late-lying snowpatches as of 22 September 1960 are also mapped from the airphotos. These clearly show that snowmelt starts from the peripheral areas of site O and finally settles at the central part. The accumulation of inorganic fines on the snow surface is rare because bare slopes are scarce in the windward of site O.

The vegetation distribution at site O is closely related to the temporal changes in snow line positions (Fig. 2). In areas outside the snow line as of 17 July 1993, *Sasa kurilensis*, *Pinus pumila*, *Alnus maximowiczii* and *Acer tschonoskii* shorter than 1.5 m, are widely seen (Figs. 2 and 3). The annular area that is nestled between the positions of the snow line as of 17 July and 22 August 1993 is mostly covered with snowbed communities typical of Japanese mountains including grasses such as *Carex blepharicarpa*, *Calamagrostis fauriei* and *Fauria crista-galli* as well as dwarfed trees such as *Geum pentapetalum* and *Phyllodoce aleutica* (Figs. 2 and 3). In the areas inside the snow line as of 22 August 1993, rubble slopes and exposures of bedrock without vegetation cover are common.

## 3. Methods

The present state of the surface morphology and vegetation of site O were first analyzed on the basis of airphotos taken by the Forest Agency of Japan in September 22, 1960, which was followed by: 1) preparation of a large-scale contour map using an analytical stereoplotter, 2) preparation of a physiognomic vegetation chart, 3) surveying the geological sections of topsoils and 4) sampling topsoils and tephras for dating. For $^{14}C$ dating, samples of loamy soil layers with humus or muck layers with a vertical thickness of 1–2 cm were used for all cases. Most samples were chemically treated and humic-acid along with humin was extracted. Soil organic matters were dated using either a beta-lay counting or an AMS method. The $\delta^{13}$ C-corrected conventional radiocarbon ages were calibrated by OxCAL 3.8 (Bronk Ramsey, 1995, 2001) with INTCAL98 (Stuiver et al., 1998).

The four tephras that erupted from the remote volcanoes are significant from the chronological point of view. Their sources, ages and lithological properties were described elsewhere in detail (Kariya et al., 1996). In this paper, a brief summary of the analysis procedures is provided. The tephra samples were washed in an ultrasonic bath and dried at 60 °C for 24 h. Subsequently, lithological description of minerals and refractive index measurements of volcanic glasses were carried out. The data were compared with the previous literatures for identification. The tephras identified were the Towada a (To-a, 915AD = 1035 cal BP), the Towada Chuseri (To-Cu, 6300 cal BP), the Kikai Akahoya (K-Ah, 7300 cal BP) and the Asama Itahana Yellow / Kusatsu (As-YP / K, 16,000 cal BP).

## 4. Landscape classification of site O

Site O can be divided into three landscape zones (Figs. 2, 3 and 4) on the basis of landforms, vegetation, topsoils, snow cover duration, geomorphic processes and age. Each landscape zone has a concentric annular shape whose centre is situated around the central basin of site O. In this chapter, the features of each landscape zone are described.

Topsoils observed in the zones are classified as follows: 1) fully or partially decomposed litter layers (equivalent to A0 horizon in the pedological classification), 2) loamy soil layers with humus, 3) muck layers (fully decomposed peat), 4) gravel (granule to cobble size), sand and silt layers, 5) weathered bedrock and 6) exotic tephra layers or dense zones of exotic tephra grains in the topsoils or sediments. The loamy soil layers often contain large amounts of gravels and sands.

### *4.1. Zone 1 (innermost snow-induced bare slopes)*

As described above, the centre of the nivation hollow basin is occupied by late-lying snowpatches for an extended duration in one year and the slopes in and around such basins lack vegetation covers (Fig. 2). On the slopes where snow usually disappears by mid-August, *Phyllodoce aleutica*, *Carex blepharicarpa*, *Carex pyrenaica* and *Pencedanum multivittatum* grow sporadically. Slopes where snow persists after late August are merely covered by moss and lichen and form an almost entirely bare ground. In this study, slopes having such landscapes and topsoils as described below are denoted as zone 1.

In zone 1, traces of sheet wash and rills as well as stone-banked terraces, stone nets and nival boulder pavements are abundant (Figs. 3 and 4). These microforms are being developed by nivation such as niveofluvial action, solifluction, frost creep and cryogenic weathering (Kariya, 2002). Process measurements in the field (Kariya, 2002) demonstrate that slope lowering by wash attains 0.3 cm $a^{-1}$ near Loc. 2 (Figs. 2 and 3). In zone 1, seasonal frost occurs after early November and its maximum thickness reaches 20 cm. However, seasonal frost completely melts immediately after the disappearance of snow in mid-August. Although thawing of seasonal frost leads to slow mass-movement of topsoils from the ground surface down to a depth of 15 cm, 20 cycles of freeze–thaw occur on the ground surface and only one cycle of freeze–thaw occurs at a depth of 10 cm (Kariya, 2002). Furthermore, debris (medium sand to pebble) produced from rock surfaces due to frost weathering and wetting–drying are observed. In short, active nivation has been taking place in zone 1.

The topsoils were observed at 11 pits in zone 1 (Figs. 2 and 5). Two types of stratigraphy were identified. One (type 1a) is where the topsoils consist of poorly-sorted angular to subangular gravel layers with silty to sandy matrix from the ground surface to depths of 5–60 cm or more (e.g., Locs. 1, 5 and 9). The other (type 1b) is weathered bedrock (Locs. 10 and 11). The gravel layers of type 1a contain less humus and are brown to grayish black. In many cases, those gravel layers are loose from the ground surface to a

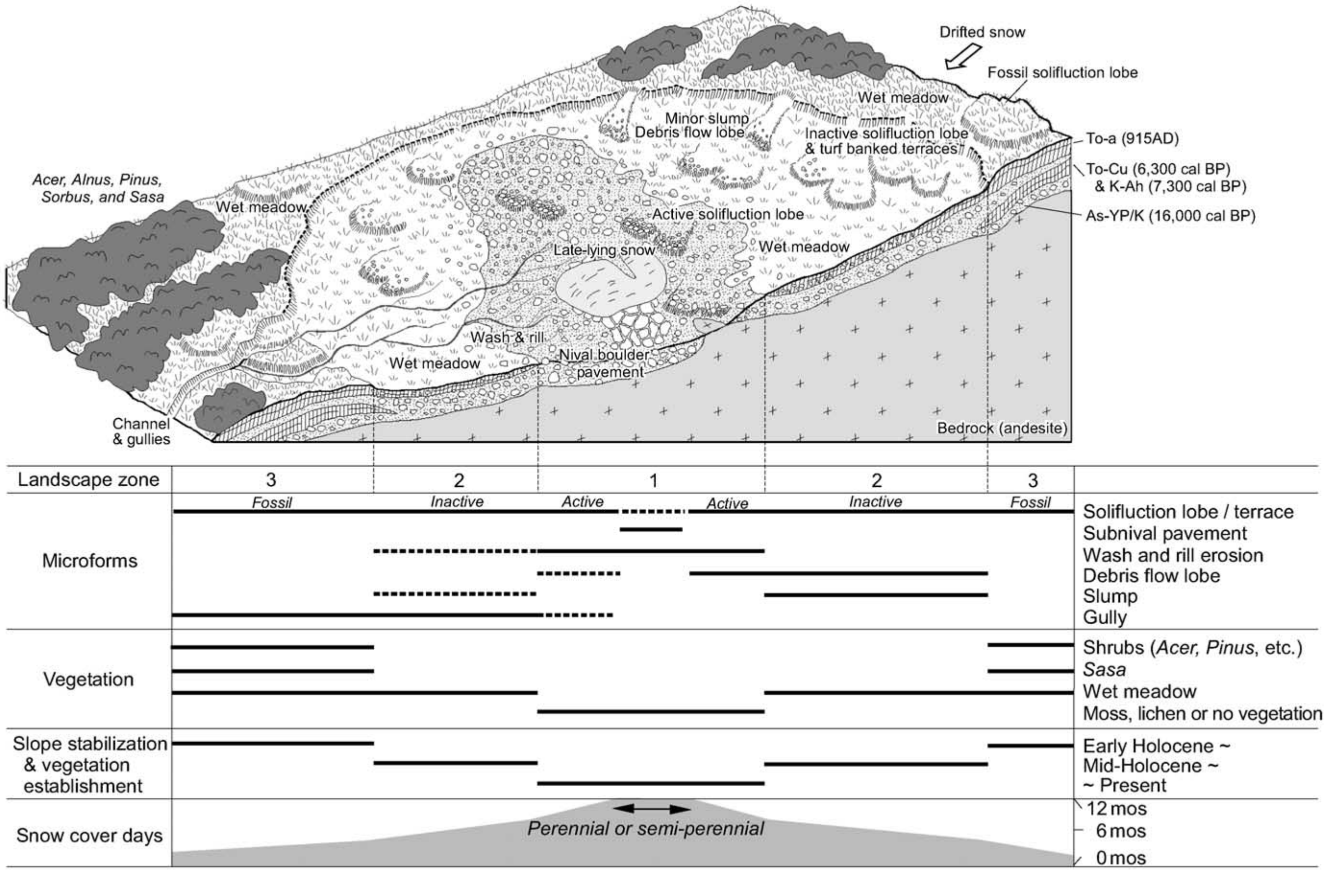
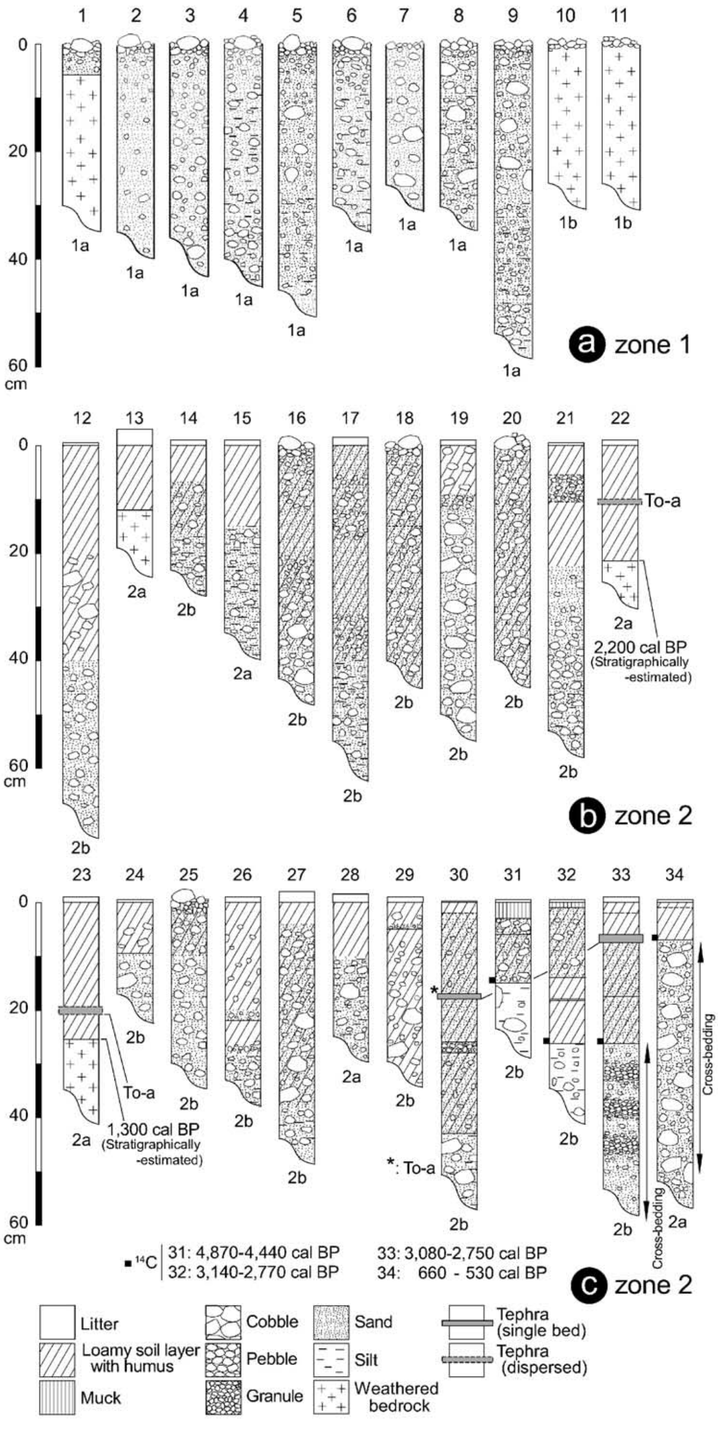
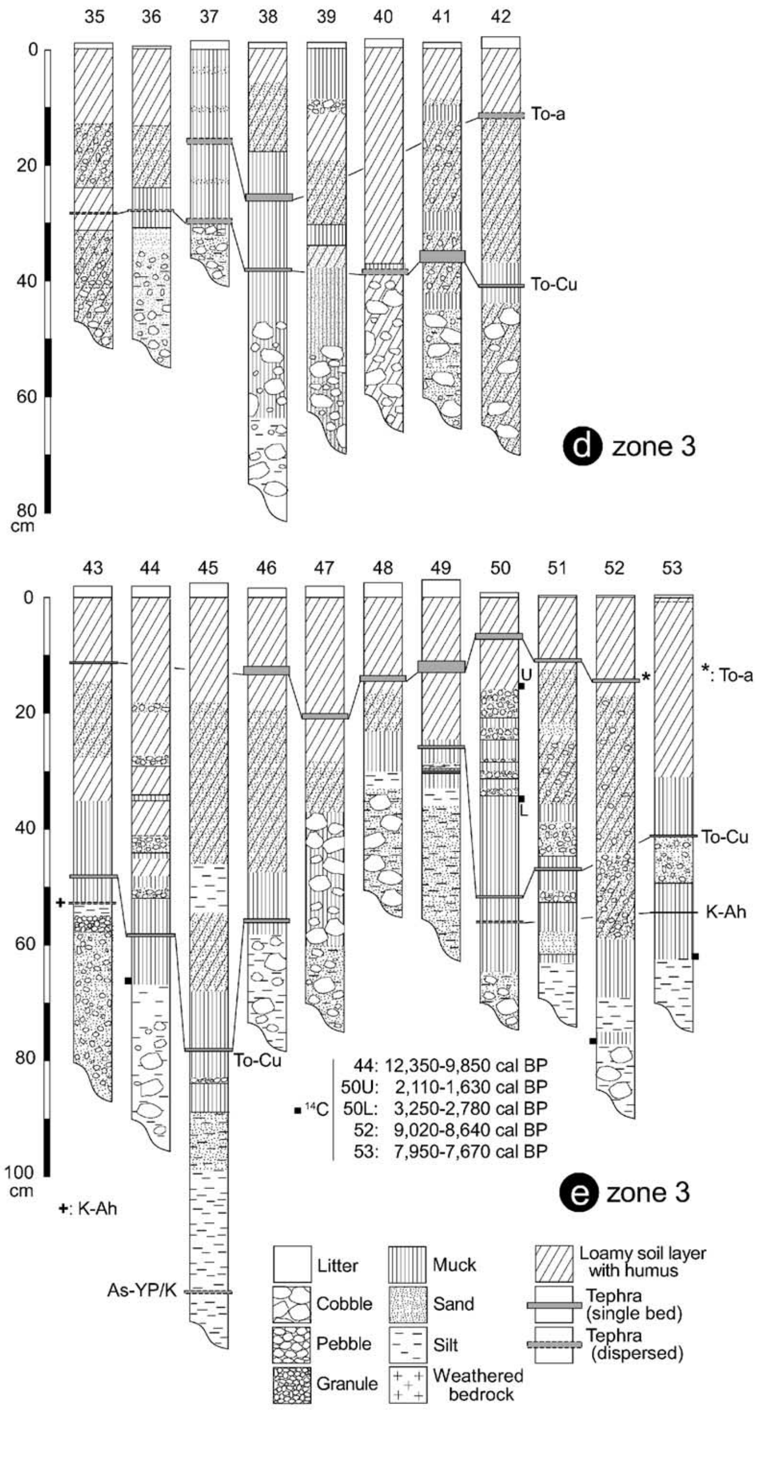



Fig. 4. Schematic diagram showing the landscape system of site O. Slope gradient and dimensions of topography are not to scale. Symbols for the subsurface geology are the same as those used in Fig. 5.



a zone 1
b zone 2
c zone 2
0
20
40
60
cm
1a
1b
2a
2b
To-a
2,200 cal BP
(Stratigraphically
-estimated)
1,300 cal BP
(Stratigraphically
-estimated)
*: To-a
Cross-bedding
■ 14C
31: 4,870-4,440 cal BP
32: 3,140-2,770 cal BP
33: 3,080-2,750 cal BP
34: 660 - 530 cal BP
Litter
Loamy soil layer with humus
Muck
Cobble
Pebble
Granule
Sand
Silt
Weathered bedrock
Tephra (single bed)
Tephra (dispersed)


35 36 37 38 39 40 41 42
0
20
40
60
80
cm
To-a
To-Cu
d zone 3
43 44 45 46 47 48 49 50 51 52 53
U
L
*: To-a
To-Cu
K-Ah
To-Cu
100
cm
+: K-Ah
As-YP/K
■ ¹⁴C
44: 12,350-9,850 cal BP
50U: 2,110-1,630 cal BP
50L: 3,250-2,780 cal BP
52: 9,020-8,640 cal BP
53: 7,950-7,670 cal BP
e zone 3
Litter
Cobble
Pebble
Granule
Muck
Sand
Silt
Weathered bedrock
Loamy soil layer with humus
Tephra (single bed)
Tephra (dispersed)

depth of 4–12 cm. The degree of consolidation increases for the deeper parts of the section. The thickness of those gravel layers approximately coincides with the maximum depth of slow mass-movement induced by the thawing of seasonal frost, suggesting that the topsoil is disturbed down to this depth. Unlike other zones, the topsoils in zone 1 contain neither buried humic soil layers nor tephras.

In addition to active slope processes, the lack of higher plant growth and plant litter deposited on the snow surface suggests that the addition of humus to topsoils has seldom occurred in zone 1. In summary, zone 1 is an assemblage of slopes that are being formed and where pedogenesis of humic soil layers does not occur at present.

### 4.2. Zone 2 (intermediate vegetated slopes)

The slopes where the time of snow disappearance is relatively early exist around zone 1 (Figs. 2, 3 and 4). These slopes are normally free of snow cover after mid-July to early August. *Fauria crista-galli* and *Veratrum stamineum* grow on particularly wet slopes, while *Phyllodoce aleutica*, *Calamagrostis fauriei*, *Carex blepharicarpa*, *Peucedanum multivittatum*, *Geum pentapetalum* and *Anemone narcissiflora* are dominant on other drier slopes. Although turf-banked terraces with sparsely-vegetated treads occur sporadically, most slopes are covered with the snowbed plant communities described above (Figs. 3 and 4). Slopes having these landscapes and topsoils (described below) are denoted as zone 2.

In zone 2, minor slumps with scars 0.2–2 m high as well as turf-banked terraces are distributed. Small-scale debris flow tracks and lobes are also seen. Every slump permits vegetation to encroach and thus it seems inactive. Gullies shallower than a depth of 1 m are also developed. Sheet wash sometimes appears during the snow ablation period; nevertheless, specific microforms are not produced by this slope process. Zone 2 is considered to be comprised of fossil geomorphic surfaces formed in the past because vegetation and topsoils have developed and few microforms have been formed except for gullies.

A total of 23 soil sections were observed in zone 2 (Figs. 2 and 5). Overall soil stratigraphy can be divided into two subtypes, 2a and 2b. The difference between the two subtypes is whether the loamy soil layers in the upper part of the soil sections contain coarser inorganic materials at least partly. The basic stratigraphy and the basal ages of loamy soil layers are fundamentally the same between these two subtypes. The features of both the types are described below.

For subtype 2a (e.g., Locs. 13, 15, 23 and 34), silt–loamy soil layers with humus 10–25 cm thick appear in the upper section, and poorly-sorted sand and gravel layers thicker than 20 cm emerge in the lower portion. A thin bed or a dense zone of tephric grains of To-a is occasionally found in the upper loamy soil layers, whereas none of the mid-Holocene tephras, To-Cu or K-Ah, are detected. Although the lower sand and gravel layers do not

Fig. 5. Stratigraphy of topsoils in site O (zones 1 and 2). The calibrated $^{14}C$ ages are indicated with a $2\sigma$ confidence level except for the stratigraphically-estimated ages. Characters under columnar sections indicate subtypes of soil stratigraphy. Stratigraphy of topsoils in site O (zone 3). The calibrated $^{14}C$ ages are indicated with a $2\sigma$ confidence level.

have specific sedimentary structures in most cases, those at Loc. 34 exhibit small-scale cross bedding.

The basal age of the upper silt–loamy soil layer of type 2a was determined only at Loc. 34: 660–530 cal BP (2σ, and so forth; Beta-85101). The basal ages of the upper loamy soil layer can also be estimated by using the stratigraphic relations with To-a at both Locs. 22 and 23; the pedofacies of the upper loamy soil layers at these two localities do not vary through their overall sections and therefore the accumulation rate of soil layers might be almost constant. The stratigraphically-estimated ages are 2200 (Loc. 22) and 1300 cal BP (Loc. 23).

For subtype 2b (Locs. 16, 19, 30 and many others), sands/gravels bearing loamy soil layers with humus 15–45 cm thick or more appear in the upper section. Gravel sizes from granule to pebble. Poorly-sorted sand and gravel layers thicker than 10 cm lie beneath the upper loamy layers. Although the lower sand and gravel layers normally do not have specific sedimentary structures, those at Loc. 33 exhibit small-scale cross bedding similar to that at Loc. 34. A thin bed or a dense zone of tephric grains of To-a is sometimes intercalated in the upper loamy soil layers of this type, while the mid-Holocene tephras are not detected. The three dates obtained for the basal part of the upper loamy soil layers are as follows: 4870–4440 cal BP (Loc. 31; Beta-85097), 3140–2770 cal BP (Loc. 32; Beta-85098) and 3080–2750 cal BP (Loc. 33; Beta-85099).

### *4.3. Zone 3 (outermost vegetated slopes)*

The slopes where the time of snow disappearance is the earliest are distributed to the outerside regions of zone 2 (Figs. 2, 3 and 4). These slopes are for the most part normally free from snow cover after mid-July. *Alnus maximowiczii*, *Sorbus matsumurana*, *Spiraea betulifolia* and *Athyrium melanolepis* are particularly common on such slopes. Although these species also grow in zone 2, the vegetation cover in the outer slopes is denser. *Sasa kurilensis*, *Acer micranthum*, *Juniperus communis* var.*nipponica* and *Rhododendron brachycarpum* grow on the slopes where snow disappears earliest, i.e., before mid-July. Lobate forms with vegetation cover whose risers are 0.5–2 m high and pronival channels 1.5–2 m deep exist on those slopes (Figs. 3 and 4). Slopes having these landscapes are denoted as zone 3. Slope mass-movement and erosion processes are fairly weak except for within pronival channels. Microforms caused by contemporary slope processes do not exist in this zone. Due to the development of vegetation and topsoils, zone 3 can be considered to comprise of fossil geomorphic surfaces.

A total of 20 soil sections were observed in zone 3 (Figs. 2, 5 and 6). The topsoils possess some of the following characteristics of lithofacies and stratigraphy: 1) the upper part of topsoils, from the ground surface down to a depth of 30–70 cm or more, composed of loamy soil layers with humus, 2) loamy soil layers with one or a few muck layer(s), 3) loamy soil layers and muck layer(s) with visible sand and gravel layer(s) having cross bedding, or with large amounts of sands and gravels, 5) occasional thin beds or a dense zone of tephric grains of To-a, To-Cu and K-Ah in loamy soil layers or muck layer(s), 6) poorly-sorted sand and gravel layer(s) below the loamy soil layers or the muck layer(s), 7) a platy soil structure in the lower sand and gravel layer(s) with abundant fine materials and 9) tephric grains of As-YP/K contained in the lower sand and gravel layer.

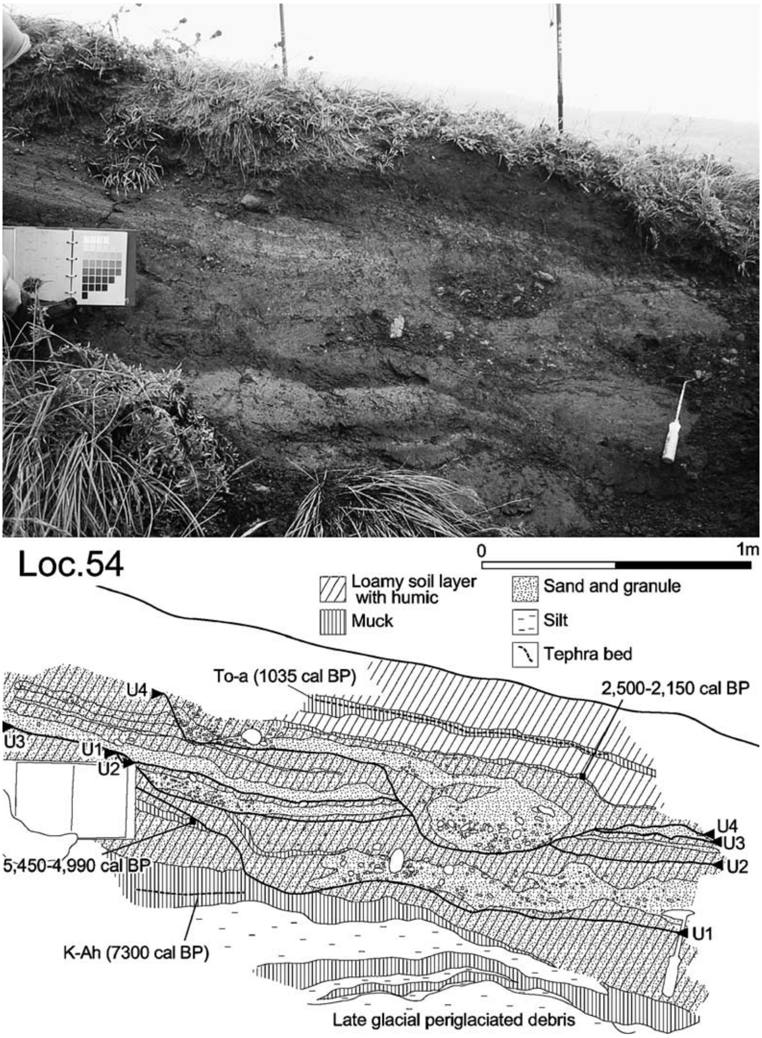



Fig. 6. Geological section at Loc. 54 showing the mid-to late Holocene channel incisions in front of the late-lying snowpatches in site O. The calibrated $^{14}C$ ages are indicated with a 2σ confidence level.

Seven $^{14}C$ ages were obtained from several horizons (Figs. 5 and 6). For example, at Loc. 44, the basal part of the lowermost muck layer gave readings of 12,350–9850 cal BP (GaK-16119). Similarly, the basal part of the lowermost muck layer was dated as 9020–8640 cal BP (Beta-85100) at Loc. 52 and as 7950–7670 cal BP (Beta-85102) at Loc. 53. Thin beds or dense horizons of tephric grains of To-Cu or K-Ah were also found in the muck layers at many localities (e.g., Locs. 35, 38, 42, 43, 46 and 53). Furthermore, several $^{14}C$ ages were obtained from the sand and gravel layers intercalated into the loamy soil layers or muck layers. For example, at Loc. 54, the muck layer was dated as 5450–4990 cal BP (Beta-183785) and another layer with the K-Ah bed are cut by sand and gravel layers with cross bedding (U1; Fig. 6). At this locality, similar cycles of erosion/deposition took place three times after the first deposition of a sand and gravel layer (U2–U4). The uppermost sand and gravel layer is covered by a thin muck layer dated as 2500–2150 cal BP (Beta-183784). These facts show that a series of deposition of sand and gravel layers had occurred from 5450 to 2150 cal BP. A similar instance was found at Loc. 50 (Fig. 5). Here, the bottom and the top of sand and gravel layers gave readings of 3250–2780 cal BP (N-6688) and 2110–1630 cal BP (N-6687), respectively. The timings of the influxes of sands and gravels from the surrounding slopes to these two sites partly overlap.

## 5. Discussion: landscape evolution since the late glacial period

### 5.1. Interrelation among slope stability, vegetation encroachment and pedogenesis

Topsoils with humus are being formed in both zones 2 and 3, except for the pronival gullies and channels. On the other hand, such soils and sediments are hardly produced in zone 1, because both the existence of active nivation processes and the lack of vegetation would prevent continuing pedogenesis. Field observations suggest that attenuation of slope processes by vegetation cover (Thorn, 1976; Strömequist, 1985), per contra, permitted the formation of topsoils with humus in zones 2 and 3. Therefore, if the onset age for pedogenesis of topsoils is evaluated, the timing of slope stabilization and vegetation establishment on specific slopes can be discussed (Takada et al., 1990;Kariya et al., 2004).

### 5.2. Late glacial period to the early Holocene epoch

Despite observations at numerous localities in site O, As-K/YP tephra was not found in the upper loamy soil layers with humus and muck layer(s). Topsoils with humus formed prior to 12,350–9850 cal BP have also not yet been discovered. These facts suggest that slope stabilization and vegetation establishment were not complete during the period ranging from the late glacial period to the earliest Holocene epoch (16,000 cal BP at the earliest to 9850 cal BP at the latest).

Both the platy structures in soils and the buried lobate topographies suggest that the poorly-sorted sand and gravel layers lying in the lowermost part in soil section were deposited under periglacial conditions. The platy structures in the lower sand and gravel layers can be associated with segregation of ice lenses and it resembles platy or bladed structures reported from the deep seasonal frozen regions (Harris, 1985; Van Vliet-Lanoe, 1985; FitzPatric, 1993). Present-day seasonal freezing of topsoils reaches a depth of 20–30 cm in the snow-induced bare ground (zone 1) and deeper than 1.5 m in the periglacial slopes with less snow in Gassan (Kariya, 2002). According to Van Vliet-Lanoe (1985), these platy structures occur particularly in poorly-drained slope materials and soils suitable for ice segregations. The slope condition of site O, inundated by snowmelt water, agrees well with this requirement. Furthermore, lobate microforms in zone 3 are indicative of this. They are blanketed by loamy soil layers with humus and muck layer(s), and are therefore the buried topography formed before the onset of pedogenesis of these soil layers. Lobate microforms are often formed by periglacial processes such as solifluction and minor slumping. Although the existence of the glacial environment in Gassan during the last glacial period remains equivocal, the possibility of severe periglacial environment accompanying nivation processes exists (late glacial to early Holocene in Fig. 7).

### 5.3. Early Holocene epoch

The basal ages of the lowermost muck layers indicate that the initial slope stabilization and vegetation establishment had occurred between 12,350–7670 cal BP in zone 3 (Fig. 5). In addition, the detection of K-Ah or To-Cu in the topsoils at many localities in zone 3 suggests that the slope stabilization and vegetation establishment occurred before the mid Holocene epoch.

As described later, the slope stabilization and vegetation establishment in zone 3 occurred several millennia earlier than those in zone 2. In zone 3, the total thickness of the upper soil layers (i.e., ground surface to the lowest loamy soil layer with humus or the lowest muck layer) is mostly less than 1 m and thus such soil layers merely mantle the lower slope materials. Thus, the contours on the slopes in zone 3 (Fig. 2) approximately show the outline of palaeotopography prior to the early Holocene pedogenesis. Based on this idea, it is assumed that a broad shallow valley which drained the watershed area of the past nivation hollow of site O existed around the slopes from Loc. 36 to Loc. 45 and from Loc. 22 to Loc. 48. The prototype of the present-day nivation hollow would have already existed in the late glacial period to the early Holocene epoch and the tendency of the timing of snow disappearance around site O has remained unchanged since then. This implies that the historical snowmelt in the previous period in zone 3 should be earliest in site O in the same manner as it is today. Slope stabilization, vegetation establishment and the resulting pedogenesis would have started from such slopes (Fig. 7; late glacial to early Holocene).

However, on the basis of the spatial distribution of the onset timing of pedogenesis in zone 3, it is suggested that the slope stabilizations and vegetation establishment proceeded with recognizable spatial patterns. Basically, the slope stabilization and vegetation establishment started from the outermost regions of site O since the late glacial period; however, it should be recognized to be somewhat sporadic from a local point of view.

The early Holocene muck layers have also been found in other snowy mountains in central and northern Japan and they are considered to be formed in snowbed plant communities in a relatively warmer climate during the early to mid-Holocene epoch (Iwata, 1983; Takada et al., 1990, Kariya et al., 1996, 2004).

### 5.4. Mid Holocene epoch

The extent of the early Holocene slope stabilization and vegetation establishment is to be determined. Unfortunately, there is no correct answer to this requirement, because none of the topsoils formed in the early Holocene epoch have been found in zones 1 or 2. In the case of site O, two models of landscape development are considered. One is that the warmer climates in the climatic optimum could have permitted the formation of humic soil layers in zones 1 and 2; subsequently, such soil layers might have been eroded (Fig. 7; mid-Holocene in model B). Another is that the early Holocene slope stabilization and vegetation establishment in zone 3 did not spread to zone 2 (Fig. 7; model A). Information

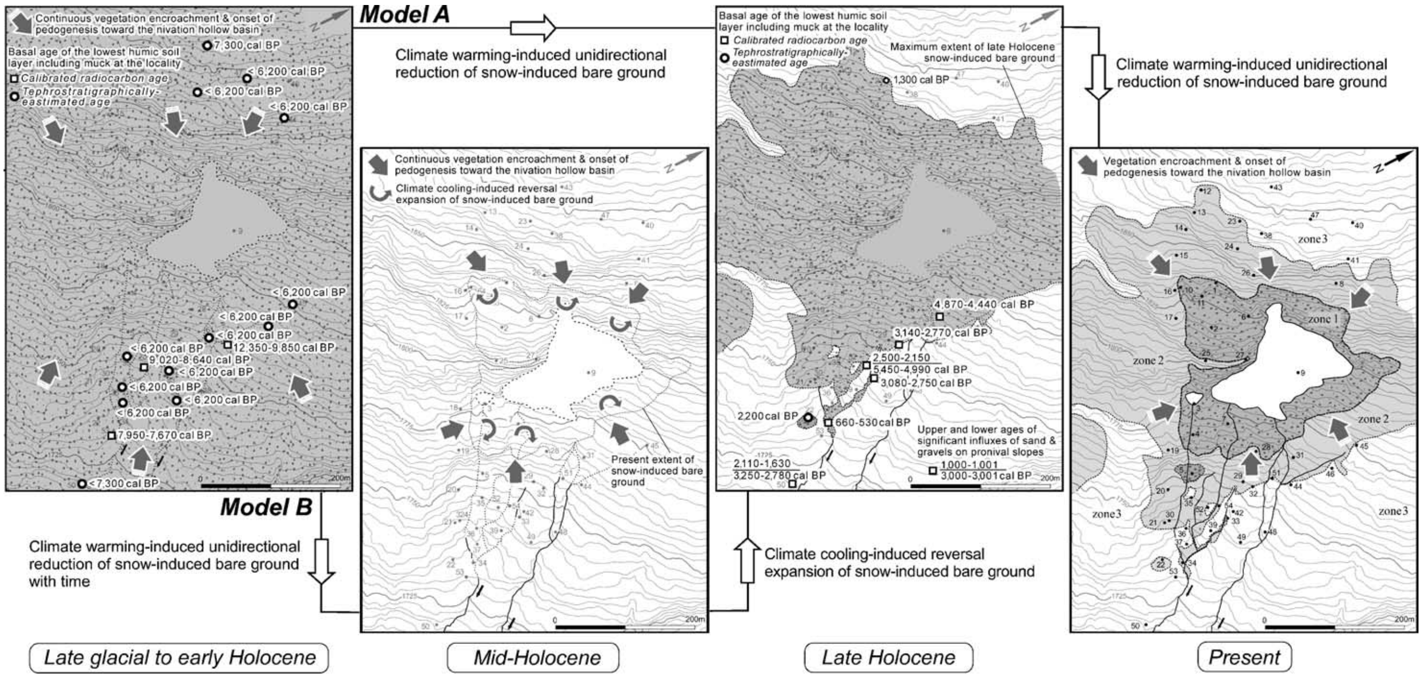



Fig. 7. Landscape evolution of the large nivation hollow (site O) on Gassan volcano since the late glacial period.

on the ages and palaeoclimates of the slopes to verify the validity of either possibility is so far insufficient.

Ogasawara et al. (1956) presumed climatological requirements for the complete melt of late-lying snowpatches at site O. They investigated the differences of air temperature between summer and winter in Yamagata City (153 m a.s.l.), approximately 40 km southeast of Gassan, in 1955 when the snowpatches at site O melted completely. They also examined the cumulative snowfall from winter 1955 to spring 1956. By comparing these data with the instrumental meteorological records before 1955 in Yamagata, Ogasawara et al. (1956) pointed out the top 9 specific years indicated a high possibility of the snowpatches at site O disappearing (1894, 1899, 1901, 1914, 1916, 1922, 1933, 1946 and 1950). Mean temperature differences of such years were +4.6 °C in summer seasons and +2.0 °C in winter seasons. Moreover, mean cumulative snowfall in those periods was 1232 mm, equivalent to half of the total mean value. It is suggested that complete snowmelt of the late-lying snowpatches at site O requires an increase in mean annual air temperature of 2–4 °C and reduction of snowfall by half. Furthermore, Daimaru et al. (2002) recently pointed out that a decrease in winter westerlies at a 850 hPa surface over Akita 135 km north of Gassan might induce a decline of snowdrift in high mountains in northern Japan, resulting in an accelerated snow melt in snowpatch hollows. Although the past aerological conditions in the mid-Holocene epoch are not known, the mid-latitude westerlies may have become rather weak due to the global climatic warming. In the mid-Holocene epoch, perennial snowpatches might not have formed at site O, and bare slopes in the snowpatch hollow basin might have been extremely small or might have disappeared (Fig. 7; mid-Holocene in model B).

### *5.5. Late Holocene epoch*

In zone 2, the slope stabilization and vegetation establishment took place in the late Holocene epoch. The upper part of topsoils in zone 2 mainly consists of loamy soil layers with humus containing a lot of sands and gravels, and the onset timing of their pedogenesis, estimated from $^{14}C$ chronology, is the late Holocene epoch between 4870–530 cal BP. Approximate estimation based on soil depositional rates, 2200 cal BP and 1300 cal BP, is also possible.

In zone 2, bare slopes would have been developed before the beginning of pedogenesis of soil layers with humus, and such non-vegetated slopes finally shrank to the areas of zone 1 (Fig. 7; late Holocene, and present). In this regard, it is unknown whether or not the slope stabilizations and vegetation establishment preceded in one direction toward the snowpatch basins, because the radiocarbon dates for pedogenesis are unevenly distributed in the study slopes (Fig. 7). However, the sediment structures at Loc. 54 in zone 3 (Fig. 6), close to zone 2 at present gives some information about this problem. As described earlier, the formation of muck layers since the early Holocene epoch at Loc. 54 continued till 4990 cal BP at the latest, nevertheless channel erosion took place at least four times by 2150 cal BP. These facts show that snow-induced bare slopes that were located on the upslope side of Loc. 54 began to expand after 5450–4990 cal BP, or the magnitudes of surface wash around the site might have fluctuated, or both occurred. Such fluctuations of slope conditions are considered to have ceased after 2500–2150 cal BP. Similarly, the intercalation of sand and gravel layers into the upper loamy soil layers is often seen in the horizon between K-Ah or To-Cu and To-a in zone 3 (Fig. 5), reflecting the area pulsation of snow-induced bare slopes and changes in surface running water conditions in the late Holocene epoch.

### 5.6. Linkage between landscape evolution and global changes

The landscape development since the late glacial period in site O seems to be related to the global changes. Although no glacigenic landforms or sediments have been discovered in Gassan, ice bodies or large perennial snowpatches might have existed at site O during the last glacial period and a severe periglacial environment might have been established around them. Under such circumstances, pedogenesis of soil layers with humus might be impossible.

The pollen records from the annually laminated sediments of Lake Suigetsu in west Japan indicates that the termination of cool climate correlatable with the Younger Dryas event in the North Atlantic occurred around 11,250 cal BP (Nakagawa et al., 2003). In site O, the slope stabilizations and vegetation establishment would have occurred around this period, as suggested by the early pedogenesis in zone 3. On the other hand, the evidence for the landscape changes related to the climatic cooling around 8200 cal BP (Alley et al., 1997) has not been obtained from site O.

In the late Holocene epoch, the Neoglacial cooling arose in various parts of the world. Although no reliable evidence for either the late Holocene glacier advances or development of mountain permafrost has been discovered in the Japanese high mountains, possibilities of snowmelt delay have been pointed out in the snowy mountains of central Japan based on pollen and plant phytolith analyses (Takada et al., 1990; Kariya et al., 2004). During these environmental changes, the slopes inside zone 2 became unstable and thus the establishment of vegetation was difficult.

Many proxies have indicated that the cooling occurred in various places in the Little Ice Age (LIA) (Briffa and Osborn, 1999; Christiansen, 1998b; Luckman, 2000; Briffa et al., 2003). Although a glacier advance was not reported in Japan in the LIA, the snowmelt delay was presumed based on a nivation hollow study combined with pollen analysis in central Japan (Takada et al., 1990). Despite them, the slope instabilizations and vegetation retreat related to the LIA climate were not evident in site O. Similarly, although the early snowmelt and the resulting new deposition of soil layers around the snowpatch hollow basins in the Medieval Period were detected in another mountain in Japan (Daimaru et al., 2002), they were not recognized at site O at all. It is considered that the climatic changes in the LIA and the Medieval Period would not have been remarkable and the response of vegetation and topography to the climatic changes would have been longer than the duration of these climatic changes in the late Holocene epoch.

## 6. Concluding remarks

The circular type large nivation hollow (site O) in Gassan volcano, northern Japan, consists of three landscape zones whose micro-landforms, vegetation, snowmelt periods and topsoils are different (Figs. 2, 3 and 4). Zone 3, the outermost zone, is covered by dwarfed scrub, *Sasa* and snowbed grasses. Fossil lobate microforms and pronival channels are developed. In zone 3, since the earliest Holocene epoch, slope stabilization and vegetation establishment occurred in lieu of the preceding periglacial environments without vegetation. As a result, humic soil layers including muck layers were formed. Zone 2, the intermediate zone, is covered by snowbed grasses but their coverage ratios are lower than that in zone 3. Inactive turf-banked terraces, debris flows and minor slumping are developed. Zone 2 had experienced slope stabilization and vegetation establishment and the resulting pedogenesis of soil layers with humus in the late Holocene epoch. Finally, zone 1, the innermost zone, coincides largely with the snow-induced bare slopes in the nivation hollow basin. Active stone-banked terraces, nival boulder pavements and wash-rill erosion are developed on those slopes and soil layers with humus are not formed at present. The historical development of landscapes in the nivation hollow would have been related to the global to semi-global changes, although they do not clearly record minor climatic changes during the Holocene epoch.

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