Comparison of the return period for landslide-triggering rainfall events in Japan based on standardization of the rainfall period

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Abstract
The intensity of rainfall events with potential to cause landslides has varying temporal characteristics. In this study, the time at which the 72-h accumulated rainfall reached its maximum was used to standardize the period of rainfall measurement. The proposed standardization of the rainfall period was used in conjunction with the return level of rainfall intensity, obtained from intensity–duration–frequency curves, to investigate rainfall intensity anomalies associated with 10 hazardous rainfall events that triggered numerous landslides at the regional scale in Japan. These landslides included shallow landslides in volcanic and non-volcanic areas, as well as deep-seated landslides. The rainfall events that triggered the shallow landslides were divided into two types: downpours that repeatedly reached close to the 100-year return level within approximately 3–4 h, and accumulated rainfall that reached close to 200–400 mm over longer time intervals but within 72 h. Lithological differences seemed unrelated to the differences between the two types of shallow-landslide-triggering rainfall; however, precipitation >1000 mm was necessary to trigger deep-seated landslides. Although the characteristics of the hyetographs differed markedly among the landslide-triggering rainfall events, all the landslides could have been triggered when the mean rainfall intensity reached the 100-year rainfall level during the standardized period. Thus, the landslide trigger can be evaluated indirectly based on the increase in the return level of the mean rainfall intensity, which could provide a means for estimating the time of landslide occurrence.

KEYWORDS
deep-seated landslide, intensity–duration–frequency curve, Japan, rainfall threshold, shallow landslide

1 | INTRODUCTION

In mountainous areas, landslides that produce voluminous sediment represent a principal driver of landscape evolution (e.g. Korup, 2005; Oguchi et al., 2001), with potential to trigger disasters that cause human casualties (e.g. Petley, 2012). Therefore, assessing the likelihood of landslide occurrence is crucial both for investigation of sediment release episodes and landscape changes and for mitigation of landslide-related disasters. Although landslides can occur in relation to earthquakes (e.g. Fan et al., 2019; Tanyaş et al., 2018) and volcanic activity (e.g. Hürlimann et al., 2000), rainfall is generally the primary trigger (e.g. Guzzetti et al., 2008).

Hence, rainfall data analysis is a valid approach for assessing the likelihood of landslide occurrence.

Current understanding remains limited regarding how rainfall data should be processed to characterize rainfall events likely to cause landslides. Landslide-triggering rainfall includes both the aspect of the ‘trigger’ that eventually initiates sliding of the sediment layer owing to the driving force exceeding the resistive force and the ‘cause’ that prepares the underlying hydrological conditions (dynamic predisposition) (Bogaard & Greco, 2018). Physical deterministic modelling, which computes the infiltration processes and/or changes in groundwater level in response to rainfall input (e.g. Borga et al., 1998; Crosta & Frattini, 2003; Iverson, 2000; Montgomery & Dietrich, 1994;
Salvucci & Entekhabi, 1994; Wu & Sidle, 1995), has been applied to
time series investigations of how rainfall input facilitates an increase
in pore-water pressure around the slip surface that acts as the trigger
and dynamic predisposition of landslides. However, in most physical
process-based models, hydrological validity (e.g. initial state of the
groundwater level and spatial variation in moisture content) is not cali-
brated rigorously at the watershed scale, primarily because of spatially
sparse or non-existent groundwater data (Bogaard & Greco, 2018).
Additionally, the strict constraints regarding the input parameters
used in physical modelling (e.g. hydraulic conductivity and soil layer
depth) highlight the difficulties of exporting models to other water-
sheds without parameter calibration or testing.

Owing to such operational problems regarding physical process-
based models, empirical rainfall thresholds have been used widely as
an alternative approach. Landslide-triggering rainfall can be
distinguished using an intensity–duration (ID) threshold applicable to
specific regions, ranging from local to national (e.g. Brunetti
et al., 2010; Saito et al., 2010) and global scales (e.g. Caine, 1980;
Guzzetti et al., 2008). These thresholds, which are derived from the
lower limit of the mean rainfall intensity and the duration of landslide-
triggering rainfall events, provide a means of forecasting the
possibility of landslide occurrence resulting from rainfall of any dura-
tion. Furthermore, normalization of the ID threshold using mean
annual precipitation or the rainy-day normal allows comparison of
landslide-triggering rainfall events between regions with different
climatic conditions (Guzzetti et al., 2007). Thus, the relationship
between mean rainfall intensity and duration is a meaningful proxy for
the trigger and dynamic predisposition of landsliding in most regions.

In the ID threshold approach, the rainfall period before the land-
slide occurrence time can be considered a proxy for the dynamic pre-
disposition for landsliding, as associated with an increase in soil water
content and/or groundwater level (e.g. Peres & Cancelliere, 2014).
Importantly, rainfall duration obviously depends on how a specific rainfall event is delimited in a sequential rainfall record. When describ-
ing rainfall events, different definitions of dry periods and of the ces-
sation of an event (e.g. period from rainfall onset to cessation of
rainfall or to time of maximum intensity) are used, which can lead to
derivation of different ID thresholds from the same data (e.g. Abancó
et al., 2016; Bel et al., 2017; Hong et al., 2018; Leonarduzzi
et al., 2017). Different criteria for rainfall event discrimination have
been proposed to improve the determination of thresholds for assess-
ment of landslide-triggering rainfall events (e.g. Mellilo et al., 2015;
Vessia et al., 2014), suggesting variations in the effective rainfall dura-
tion that prepares the trigger and dynamic predisposition for landslid-
ing. However, no rational criteria exist for determining the duration of
landslide-triggering rainfall events in terms of the characterization of
rainfall anomalies.

In forecasting rainfall, the return period of a rainfall event of given
intensity and duration can be evaluated using an intensity–duration–
frequency (IDF) curve based on rainfall records (e.g. Chow et al., 1988;
D’Odorico et al., 2005; Frattini et al., 2009; lida, 2004; Koutsouyannis
et al., 1998). As a high rainfall return level can be considered a proxy
of rainfall with unprecedented intensity, its return period could be
used to evaluate whether forecast rainfall might be likely to cause
landslides. However, similar to the ID threshold approach, the overall
time span used in IDF analysis in relation to the trigger and dynamic
predisposition for landsliding remains obscure, which hampers

2 | LANDSLIDE EVENTS

In analysis using rainfall data recorded by rain gauges, the degree to
which the recorded rainfall represents the actual rainfall that triggered
a landslide is important (e.g. Leonarduzzi et al., 2017). Nikolopoulos
et al. (2014) highlighted that increase in the distance between a rain
gauge station and the point of debris flow occurrence can cause
underestimation of triggering rainfall. For a case in which a rainfall
event triggers multiple landslides at the regional scale, because the
weather system responsible probably covers a wide region, the spatial
variations in rainfall characteristics are possibly unremarkable
compared with those of rainfall events that locally trigger only a few
landslides. Moreover, information regarding the timing of a landslide
can help inform rainfall analysis. However, the timing of landslide
occurrence is not necessarily recorded in landslide databases
(e.g. Leonarduzzi et al., 2017; Segoni et al., 2018; Staley et al., 2013).

To remedy these uncertainties, this study considered 10 landslide
events that induced numerous landslides in the vicinity of a rain gauge
(Figures 1 and 2) and for which the occurrence time of at least one
landslide could be estimated reliably. The selected landslides included
two shallow landslide events near volcanoes, seven shallow landslide
events in non-volcanic areas, and one deep-seated landslide event
(Tables 1–3). Strictly, almost all regions in Japan have been affected
by earlier volcanic activity. In this study, after considering differences
in lithological settings, the shallow landslides in volcanic areas were
defined as ‘shallow landslides that occurred at the boundary between
tephra and other layers’ (see the online Supporting Information for
further details of each landslide event).

The dataset focused only on rainfall events that triggered
numerous landslides at the regional scale; therefore, it was expected
that their return period would indicate a rainfall level that rarely arises
in the examined regions. Additionally, it was assumed that the rainfall
data acquired by the closest Japan Meteorological Agency rain
gauge could produce reasonable hyetographs of the rainfall events that induced the landslides. This assumption, which is inevitable regarding the use of rainfall data recorded by rain gauges, is responsible for unverifiable uncertainties without resort to either a dense rain gauge network or rainfall data with reasonable spatial distribution (e.g. Destro et al., 2017; Marra et al., 2017). The analysis in this study was based on consideration of these assumptions and limitations.

3 | METHODS

3.1 | Rainfall IDF curves

Multiple IDF curves corresponding to different return periods can be plotted as a function of rainfall duration (e.g. Koutsoyiannis et al., 1998; Sane et al., 2018). To derive IDF curves, a statistical model that conforms to the distribution of extreme values of rainfall intensity is necessary. The Gumbel distribution was used in this study because its fewer shape parameters might reduce the uncertainty in parameter estimation (e.g. Frattini et al., 2009). Based on extreme value theory (Fischer & Tippett, 1928; Gnedenko, 1943), application of the Gumbel distribution assumes asymptotic behaviour of the dataset (i.e. an infinite number of rainfall events should occur annually). The asymptotic assumption has not been validated rigorously, but current alternative methods can relax this assumption and reduce the estimation uncertainty of the return period (e.g. Zorzetto et al., 2016). However, to simplify the rainfall data processing, this study followed the traditional extreme value theory.

In this study, extremes of rainfall intensity were defined as annual rainfall maxima (e.g. Sane et al., 2018). The annual maxima of rainfall intensity over 1-, 2-, 3-, 6-, 12-, 24-, 48-, and 72-h periods were extracted from rainfall records and used to estimate the parameters describing the Gumbel distributions based on the L-moments method (Hosking, 1990). To verify whether the rainfall intensity distributions corresponding to the analysed observational data and the calibrated Gumbel distribution matched, goodness-of-fit tests (i.e. the Kolmogorov–Smirnov and Anderson–Darling tests) were conducted.

![Figure 1](wileyonlinelibrary.com) Location of landslide events. Red circles indicate the locations of the rain gauge stations used [Color figure can be viewed at wileyonlinelibrary.com]

![Figure 2](wileyonlinelibrary.com) Locations of landslides and rain gauge stations (red circles). Black points in (a) and (c–j) indicate the locations of the landslide scars. Brown polygons in (b) indicate landslides and inundation areas. Note that the black points do not indicate all the landslide scars because the data cover only limited areas, especially in the cases of (a) Aso, (d) Okaya, and (e) Tokachi [Color figure can be viewed at wileyonlinelibrary.com]
The null hypothesis was that the sample had been fitted to the calibrated model (i.e. the Gumbel distribution); therefore, the calculated $p$-value should have been $>0.1$ (e.g. Sane et al., 2018).

Even if the goodness-of-fit tests suggest statistical reliability of the IDF curves, this statistical analysis approach includes inevitable uncertainties. The extrema of hourly rainfall acquired by the calibrated model (i.e. the Gumbel distribution); therefore, the calculated $p$-value should have been $>0.1$ (e.g. Sane et al., 2018).

Table 2: Characteristics of representative landslides and their lithological settings

| Region       | Number | Area (m$^2$) | Depth (m) | Slope (°) | Lithological setting                                      | Elevation (m)$^a$ |
|--------------|--------|--------------|-----------|-----------|----------------------------------------------------------|------------------|
| Aso          | >850   | ~50–5000     | ~1.2–1.6  | ~35–40    | Tephra, scoria, and loam                                 | Min: 626.6, Max: 940.3, Mean: 741.7 |
| Izu-Oshima   | ~100   | ~136–466     | <1        | ~32–41    | Tephra, scoria, and loam                                 | 141.3, 531.7, 372.9 |
| Hidaka       | >4000  | ~400         | ~0.5      | ~30–40    | Shale, sandstone, conglomerate, and tuff                 | 157.7, 444.6, 273.0 |
| Okaya        | >121   | ~187–800     | ~1.5–5.0  | ~20–30    | Tuff, andesite, and pyroclastic rocks                    | 821.9, 1052.7, 940.9 |
| Tokachi      | >30    | ~21–33       | ~10–20    | ~33–40    | Granodiorite and pelitic schist                          | 39.6, 844.7, 308.3 |
| Asakura      | >2000  | ~800–1000    | ~0.8      | ~35       | Accretionary complex of sandstone and shale              | 33.0, 251.0, 166.9 |
| Igashihiroshima | >8000 | ~10–10 000   | ~1.0      | ~30–40    | Granodiorite and pelitic schist                          | 6.7, 643.1, 134.0 |
| Uwajima      | >1200  | ~50–1000     | ~1.0      | ~15–35    | Granite and granodiorite                                | 22.9, 621.9, 254.9 |
| Marumori     | >150   | ~3000        | ~1.0      | ~15–35    | Granite and granodiorite                                | 211.2, 1417.1, 746.1 |

$^a$As elevations were investigated based on the locations of landslides shown in Figure 2 that did not completely cover all triggered landslides, the values of elevation include unverifiable uncertainty.

3.2 | Comparison of landslide-triggering rainfall events

As explained in the Introduction, the calculated duration of each rainfall event could differ depending on the definition of the dry period, which might lead to discrepancies in the time spans investigated among the defined rainfall events. As the rainfall duration that affects the dynamic predisposition (e.g. increase in groundwater level) for landsliding can vary depending on the characteristics of individual landslides, the use of different time spans in rainfall analysis is probably unreasonable for inter-regional comparison of the characteristics of landslide-triggering rainfall events, especially when their hyetographs differ substantially.

Landslide-triggering rainfall can generally be characterized as significant rainfall of short duration (i.e. a downpour), significant accumulation resulting from continuous rainfall with high average intensity, and the coexistence of both features (e.g. Leonarduzzi & Molnar, 2020). In this study, for most landslide-triggering rainfall events, it was expected that the hyetographs would show that the maximum longer-period intensity would be reached a certain time after the shorter-period intensity peak. For example, the maximum hourly rainfall intensity within a 72-h period would probably occur before the peak 72-h intensity was reached (Figure 3).

Based on this assumption, the time at which the maximum 72-h rainfall intensity occurred (hereafter, referred to as $t_{72}$) was tentatively defined as the hypothetical end of each landslide-triggering rainfall event (Figure 3). The use of 72-h rainfall was motivated by the...
TABLE 3 Characteristics of the rain gauges used

| Label of region | Name of station | Period  | Elevation (m) | Min (km) | Max (km) | Mean (km) |
|-----------------|-----------------|---------|--------------|----------|----------|-----------|
| Aso             | Aso-Otohime     | 1978–2019 | 487          | 6.2      | 7.8      | 7.1       |
| Izu-Oshima      | Ohshima         | 1976–2019 | 74           | 0.7      | 2.6      | 1.7       |
| Hidaka          | Asahi           | 1986–2019 | 280          | 4.8      | 8.6      | 6.5       |
| Okaya           | Tatsuno         | 1979–2019 | 729          | 7.9      | 9.2      | 8.8       |
| Tokachi         | Shintoku       | 1976–2019 | 178          | 9.0      | 16.5     | 12.6      |
| Asakura         | Asakura         | 1976–2019 | 38           | 2.1      | 26.4     | 12.4      |
| Higashihiroshima| Higashihiroshima| 1976–2019 | 224          | 1.6      | 35.3     | 15.4      |
| Uwajima         | Uwa             | 1976–2019 | 200          | 3.6      | 23.4     | 8.1       |
| Marumori        | Hippo           | 1979–2019 | 305          | 0.4      | 18.6     | 7.4       |
| Kii-Peninsula   | Kamikitayama    | 1976–2019 | 334          | 9.5      | 73.1     | 27.3      |

*As distances from landslides were estimated based on the locations of landslides shown in Figure 2 that did not completely cover all triggered landslides, the values of the distances include unverifiable uncertainty.

ASUNETAKA

RESULTS

4.1 Annual maxima of rainfall intensity over different periods

By comparing the annual maxima of rainfall intensity over different periods, the characteristics of the landslide-triggering rainfall events in each region could be investigated. When the plots of the maximum rainfall intensities during $P_{\text{std}}$ (red points in Figure 4) are not overlapped with the line denoting the annual maxima in the year of landslide occurrence (red line in Figure 4), the annual maximum intensities did not occur within $P_{\text{std}}$ (Figure 4). These discrepancies were observed only in relation to rainfall intensity over periods of 3 h or less in the Okaya, Tokachi, and Kii Peninsula events (Figures 4d, e, and j), indicating that the annual maxima of rainfall intensities for durations of <3 h occurred during rainfall in periods that excluded $P_{\text{std}}$. In other words, for the years in which landslides occurred in all the examined regions, the annual maximum rainfall intensities for longer durations (6–72 h) did occur within $P_{\text{std}}$ (Figure 4).

In the case of the Aso and Izu-Oshima events (i.e. shallow landslides in volcanic areas), in the years in which the landslides occurred, all the rainfall intensity maxima for periods in the range of 1–72 h within $P_{\text{std}}$ were higher than the corresponding maxima in other years (Figures 4a and b). Particularly, rainfall intensities for durations of <3 h exceeded 90 mm h$^{-1}$, indicating that a downpour producing at least 270 mm of rain occurred within a 3-h period. Although the Hidaka and Asakura events occurred in non-volcanic areas, all the rainfall intensity maxima for the years in which the landslides occurred were the highest in the record, regardless of duration, similar to the situation regarding shallow landslide events in volcanic areas (Figures 4c and f). The maximum hourly rainfall intensities of the Hidaka and Asakura events were ~75 and ~100 mm h$^{-1}$, respectively, and all the

![Graphical illustration of the definitions of the rainfall parameters $t_{27}$ and $P_{\text{std}}$: $t_{27}$ indicates the time that the 3-day (72-h) rainfall reached its maximum value; $P_{\text{std}}$ indicates the 72-h period before $t_{27}$](image)

Following reasons: (1) landslide-triggering rainfall occurred within ~72 h (3 days) at longest among the examined landslide events (during the Higashihiroshima, Uwajima, and Kii Peninsula events, see the online Supporting Information); and (2) although analysis to generate IDF curves is viable without the restriction of rainfall duration, an excessively long duration could increase the total precipitation, potentially being responsible for uncertainty in the estimated return periods that exceeded accepted levels of tolerance. Thus, the hypothetical rainfall end was intended to be set as the necessary minimum duration over which rainfall patterns among the examined landslide-triggering rainfall events.

The hyetographs for the 72-h periods before $t_{27}$ (taken as the standardized rainfall period and hereafter referred to as $P_{\text{std}}$) of the different landslide-triggering rainfall events were compared. The use of this arbitrary rainfall period, $P_{\text{std}}$, meant that comparison between rainfall periods corresponding to different landslide events could be standardized. Thus, the characteristics of landslide-triggering rainfall events were analysed in terms of the overarching time span from short-term (hourly) to long-term (3-day) intensities.

To trace the fluctuation in rainfall intensity from the hyetographs, the change in the mean rainfall intensity with time was plotted on the log–log $D$–$I$ plane and compared with the IDF curves. The intention was to characterize the time at which the return level of the mean rainfall intensity increased significantly in the log–log $D$–$I$ plane, thereby revealing the duration of the significant rainfall that triggered the landslides.
maximum rainfall intensities for periods of <3 h exceeded 50 mm h\(^{-1}\) (Figures 4c and f). In the case of the Marumori event, the rainfall intensities for most durations were the highest in the rainfall record during 1979–2019 (Figure 4i). However, the maximum hourly rainfall intensity was the second highest and not the highest intensity in the record of this region. In the case of the rainfall associated with the Okaya, Higashihiroshima, and Uwajima events, the maximum rainfall intensities for durations of <4 h during \(P_{std}\) were lower than or similar to the highest intensities in the records; however, the maximum rainfall intensities for periods >24 h were clearly higher than the maxima recorded in other years (Figures 4d, g, and h). Thus, for the examined landslide events, the characteristics of the landslide-triggering rainfall events were manifested in the rainfall intensities over periods of different lengths.

Unlike the above landslide events, in the case of the Tokachi event, the maximum rainfall intensities in 2–72-h periods during 1981 (i.e. rainfall that did not trigger any landslide events) were clearly greater than those during \(P_{std}\) (i.e. the landslide-triggering rainfall event in 2016; Figure 4e). The maximum hourly rainfall intensity during 1976–2019 was observed in 2012 (34 mm h\(^{-1}\)), and the corresponding values for 1981 and 2016 were close to the highest intensity in the record at 32 and 33 mm h\(^{-1}\), respectively. Thus, comparison between maximum rainfall intensities cannot completely distinguish landslide-triggering rainfall events from non-landslide-triggering rainfall events, because even rainfall events that do not exceed the record maximum rainfall intensity can trigger landslides.

4.2 | Return periods for rainfall intensities measured over different lengths of time

All the p-values that resulted from the Kolmogorov–Smirnov and Anderson–Darling goodness-of-fit tests were >0.1, suggesting that the frequency distributions for the respective rainfall intensities could be well represented by the calibrated Gumbel distributions (Tables 4 and 5). For the Aso, Izu-Oshima, Asakura, and Kii Peninsula regions, the IDF curves indicate that the 100-year return-level intensity for hourly rainfall exceeds 90 mm h\(^{-1}\) (Figures 5a, b, f, and i), which is clearly high in comparison with the corresponding intensity in other regions (Figure 5). The 100-year return-level intensity of 72-h rainfall is >20 mm h\(^{-1}\) in the Kii Peninsula region (Figure 5i) but <5 mm h\(^{-1}\) in the Hidaka, Okaya, and Tokachi regions (Figures 5c, d, and e). Thus, owing to differences in regional rainfall characteristics, the IDF curves derived for the various examined regions are substantially different.

Hereafter, the earliest time within the estimated time range for the occurrence of a landslide is defined as the time at which that landslide occurred (see Table 1; for the Kii Peninsula event, this time was assumed to be 19:00 JST). The distance between the location of a rain gauge and the landslide area (Figure 2) can account for the time lag between the rainfall recorded by the rain gauge at the estimated time of landsliding and the actual landslide-triggering rainfall around hillslopes. Thus, some rainfall intensities at the estimated time of landsliding did not necessarily reach high return levels (Figure 5). Hence, it was necessary to evaluate the rainfall anomalies over periods of time that included the time of landsliding.
Importantly, the rainfall intensity maxima for different rainfall durations within $P_{\text{std}}$ reached exceptional return levels. For the Izu-Oshima, Hidaka, and Asakura events, all the maxima of rainfall intensities within $P_{\text{std}}$ were above the IDF curve for the 100-year return level (Figures 5b, c, and f). In the case of the Aso event, the return levels of the intensities corresponding to durations of <12 h exceeded the 100-year return levels (Figure 5a). For the Marumori event, although the hourly intensity within $P_{\text{std}}$ was below the IDF curve for the 100-year return level, all other intensities within $P_{\text{std}}$ exceeded the 100-year return level (Figure 5i).

By contrast, for the Okaya, Higashihiroshima, Uwajima, and Kii Peninsula events, although the return levels of rainfall intensities within $P_{\text{std}}$ for durations of over $\sim 24$ h were above or comparable to the 100-year return level, those for durations of <6 h were substantially below the IDF curve for the 100-year return level (Figures 5d, g, h, and i). The Tokachi event produced rather different results in terms of rainfall intensities within $P_{\text{std}}$ (i.e. all maxima were below the IDF curve for the 100-year return level; Figure 5e). Most rainfall maxima within $P_{\text{std}}$ coincided well with the rainfall intensities at the time of landsliding, and both the hourly rainfall intensity and the rainfall intensities for periods of >48 h reached relatively high return levels.

### 4.3 Hyetographs of landslide-triggering rainfall events

The hyetographs during $P_{\text{std}}$ corresponding to the different landslide-triggering rainfall events varied (Figure 6). Particularly, the time of appearance and duration of the non-rainfall period before landslide occurrence varied widely, indicating differences in the duration of valid rainfall between examined events in terms of landslide initiation. In all cases apart from the Tokachi event, the total 72-h rainfall at $t_{72}$ (i.e. the right-hand end of the lower panel of Figure 6) was the highest in the rainfall record, reaching or exceeding the 95% confidence interval of the 100-year return level (Figure 6). Although the 72-h rainfall recorded at $t_{72}$ for the Tokachi event was not the highest in the record, it still reached the 95% confidence interval of the 100-year return level (Figure 6e). Additionally, in most cases, at the start of the hyetograph within $P_{\text{std}}$ (i.e. the right-hand end of the lower panel of Figure 6) was the highest in the rainfall record, reaching or exceeding the 95% confidence interval of the 100-year return level (Figure 6).

### 4.3.1 Kolmogorov–Smirnov goodness-of-fit test $p$-values

| Region       | Rainfall duration (h) | 1 | 2 | 3 | 6 | 12 | 24 | 48 | 72 |
|--------------|-----------------------|---|---|---|---|----|----|----|----|
| Aso          | 0.74 0.54 0.47 0.64 0.89 0.96 0.96 0.99 |     |     |     |     |     |     |     |     |
| Izu-Oshima   | 0.88 0.82 0.84 0.55 0.61 0.89 0.94 0.83 |     |     |     |     |     |     |     |     |
| Hidaka       | 0.51 0.43 0.53 0.70 0.54 0.31 0.15 0.43 |     |     |     |     |     |     |     |     |
| Okaya        | 0.69 0.40 0.74 0.38 0.90 0.99 0.99 0.82 |     |     |     |     |     |     |     |     |
| Tokachi      | 0.77 0.34 0.52 0.51 0.92 0.75 0.89 0.82 |     |     |     |     |     |     |     |     |
| Asakura      | 0.91 0.77 0.45 0.49 0.21 0.60 0.92 1.00 |     |     |     |     |     |     |     |     |
| Higashihiroshima | 0.88 0.98 0.94 0.96 0.98 0.81 0.95 0.64 |     |     |     |     |     |     |     |     |
| Uwajima      | 0.89 0.45 0.31 0.30 0.89 0.96 0.99 1.00 |     |     |     |     |     |     |     |     |
| Marumori     | 0.92 0.50 0.41 0.47 0.41 0.31 0.93 0.93 |     |     |     |     |     |     |     |     |
| Kii Peninsula| 0.80 0.72 0.95 0.99 0.93 0.66 0.97 0.91 |     |     |     |     |     |     |     |     |

### 4.3.2 Anderson–Darling goodness-of-fit test $p$-values

| Region       | Rainfall duration (h) | 1 | 2 | 3 | 6 | 12 | 24 | 48 | 72 |
|--------------|-----------------------|---|---|---|---|----|----|----|----|
| Aso          | 0.77 0.64 0.47 0.56 0.86 0.99 0.95 0.99 |     |     |     |     |     |     |     |     |
| Izu-Oshima   | 0.91 0.80 0.74 0.45 0.62 0.59 0.67 0.63 |     |     |     |     |     |     |     |     |
| Hidaka       | 0.53 0.43 0.32 0.40 0.45 0.23 0.22 0.24 |     |     |     |     |     |     |     |     |
| Okaya        | 0.53 0.70 0.82 0.75 0.96 0.96 0.99 0.84 |     |     |     |     |     |     |     |     |
| Tokachi      | 0.74 0.34 0.58 0.80 0.80 0.72 0.83 0.91 |     |     |     |     |     |     |     |     |
| Asakura      | 0.66 0.50 0.62 0.32 0.14 0.73 0.99 0.99 |     |     |     |     |     |     |     |     |
| Higashihiroshima | 0.92 0.93 0.96 0.99 0.99 0.94 0.97 0.93 |     |     |     |     |     |     |     |     |
| Uwajima      | 0.90 0.75 0.58 0.56 0.96 0.95 1.00 1.00 |     |     |     |     |     |     |     |     |
| Marumori     | 0.79 0.43 0.37 0.42 0.40 0.57 0.85 0.95 |     |     |     |     |     |     |     |     |
| Kii Peninsula| 0.45 0.77 1.00 0.99 0.94 0.73 0.92 0.89 |     |     |     |     |     |     |     |     |
For the Aso, Izu-Ohshima, Hidaka, and Asakura events, around the time that the landslides occurred, the hourly rainfall intensity reached the 95% confidence interval of the 100-year return level between two and four times in each case (Figures 6a, b, c, and f). These downpours corresponded to notable maxima in rainfall intensity for periods of <3 h, indicating that these short downpours were responsible for the landsliding. In the case of the Marumori event, the hourly rainfall intensity reached the 95% confidence interval of the 100-year return level 4 h before the landslides occurred, although it was also high both before and after this peak (Figure 6i). Although the maximum hourly rainfall intensity was relatively low (i.e. approximately 75 mm h$^{-1}$), this downpour continued for several hours (Figure 6i). Thus, this consecutive downpour could have triggered the Marumori event, similar to the situation of the Aso, Izu-Ohshima, and Asakura events. In the case of the Aso, Izu-Ohshima, and Asakura events, the 72-h rainfall at the time of landsliding was <100-year return level (Figures 6a, b, and f). Therefore, rainfall that did not relate to the trigger and dynamic predisposition for landsliding in these regions. The length of time between landslide initiation and $t_{72}$ (i.e. the length of time ‘non-effective’ rainfall fell after the landslides had occurred) varied widely up to 12 h among the examined events.

In the Tokachi region, both the hourly rainfall and the 72-h rainfall reached the 100-year return level at the same time (Figure 6e). For the non-landslide-triggering rainfall that produced most of the highest rainfall maxima (i.e. the rainfall of 5 August 1981), the hourly rainfall intensity reached its maximum before the peak of the 72-h rainfall (Figure 7). This difference in the hyetographs within $P_{std}$ could explain why the 1981 rainfall did not trigger landslides. Therefore, there is the possibility that if the times at which the hourly rainfall intensity and the 72-h rainfall reach their maxima coincide, the accumulated rainfall and the subsequent intense rainfall of short duration could be considered as the trigger and dynamic predisposition for landsliding, respectively.

4.4 | Changes in mean rainfall intensity on a log–log D–I plot

Depending on the differences in the hyetographs, the tendency of fluctuations in the mean rainfall intensity during $P_{std}$ differed among the examined events (Figure 8). For all the landslide-triggering rainfall events, the mean rainfall intensity reached the global ID threshold derived by Caine (1980) before the estimated time of landslide occurrence. In the case of the Kii Peninsula event, the relatively high mean rainfall intensity of >10 mm h$^{-1}$ was maintained throughout
Therefore, the relationships between changes in the mean rainfall intensity during $P_{\text{std}}$ and the ID threshold highlight the difficulty in forecasting the time of landslide occurrence using rainfall thresholds.

Except for the Kii Peninsula event (i.e. the deep-seated landslide event), the landslides occurred at around the times that the graphs of the mean rainfall intensity and the lines depicting total precipitation of 200–400 mm intersected (Figure 8). The range of $P_{\text{std}}$. Therefore, the relationships between changes in the mean rainfall intensity during $P_{\text{std}}$ and the ID threshold highlight the difficulty in forecasting the time of landslide occurrence using rainfall thresholds.
total precipitation of approximately 200–400 mm corresponds to the lower limit of the 95% confidence interval of the 100-year return level for 72-h rainfall (Figure 6). In other words, for the examined rainfall events that triggered shallow landslides, the landsliding was triggered when the cumulative rainfall reached approximately 200–400 mm within <72 h, irrespective of differences in landslide characteristics and the relevant hyetographs. Cumulative precipitation of >1000 mm could have been responsible for the deep-seated landslides that occurred in the Kii Peninsula region.

In most cases, the times at which the landslides occurred corresponded well with the intersection between the graph of the mean rainfall intensity during $P_{\text{std}}$ and the IDF curve for the 100-year return level (Figure 9). As differences in the locations of the IDF curves in the log-log $D$–$I$ plane can be slight for high return levels, other IDF curves over a 100-year return level might result in similar correspondence. The key point in this context is that the use of the IDF curve for the 100-year return level could be an alternative to rainfall thresholds, and it might offer the possibility of investigating whether and when a rainfall event could potentially trigger landslides.

As most of the rainfall intensity maxima of the landslide-triggering rainfall events were substantially higher than other annual maxima (Figure 4), the inclusion/absence of the landslide-triggering rainfall in the rainfall record changes the IDF curves (i.e. the return level of rainfall intensity). Excluding those rainfall events known to have triggered landslides systematically reduces the 100-year return period owing to the decrease in the frequency of high-intensity rainfall (Figure 9). However, in the examined cases, the changes in the IDF curves for the 100-year return level (i.e. the differences between the thick and narrow dotted black lines in Figure 9) were slight. Therefore, when estimating the potential for landslide occurrence using IDF curves and $P_{\text{std}}$, differences in the availability and processing of data related to previous landslide-triggering rainfall events might produce false positives but are unlikely to produce false negatives.

5 | DISCUSSION

5.1 | Differences in characteristics of time spans of rainfall events that triggered numerous landslides at the regional scale

The characteristics of the examined shallow-landslide-triggering rainfall events could be divided broadly into two types with different time spans: downpours whose rainfall intensity repeatedly reached the ~100-year return level over a period of a few hours (Figures 6a, b, c, f, and i); and rainfall that produced totals of approximately 200–400 mm over longer periods but within 72 h (Figures 6d, g, and h). The former type of rainfall triggered the landslide events in the
volcanic areas (i.e. the Aso and Izu-Oshima events), as well as the landslide events in the Hidaka, Asakura, and Marumori regions, where the lithological setting is dominated mainly by granite and/or granodiorite. In contrast, despite a similar lithological setting comprising granite, the latter type of rainfall triggered the landslides in the Higashihiroshima region. Therefore, the time span of rainfall events in relation to landslide occurrence cannot be constrained only by lithological setting. As emphasized by Bogaard and Greco (2018), this suggests that the total volume of rainfall is the principal influencing factor in the process that initiates shallow landslides.

In defining a national-scale (country-wide) rainfall threshold for landslide occurrence in Italy, the influence of differences in regional climatic setting (Peruccacci et al., 2017). In contrast, if rainfall thresholds for landslide occurrence were defined for subregions with a similar lithological setting, the performance of the thresholds was found to increase in terms of distinguishing the likelihood of landslide occurrence (Segoni et al., 2014). Thus, depending on the focal scale of landslide events, the effect of lithology can vary. The relationships obtained in the current study between the time span that triggered landslides and the lithological setting could have been biased, because only those rainfall events that triggered numerous landslides at the regional scale were analysed. For the local shallow-landslide event that occurred in Hiroshima Prefecture (Japan), rainfall with similar hyetographs triggered landslides through different mechanisms depending on the lithological setting.

**FIGURE 9** Comparison between IDF curves and changes in mean rainfall intensity within $P_{10}$. Thick and narrow dotted black lines indicate the IDF curves for the 100-year return level derived from the complete rainfall record and the record excluding the year of landslide occurrence, respectively. Circles denote times of landslide occurrence [Color figure can be viewed at wileyonlinelibrary.com]
(Watake & Matsushi, 2019). Therefore, even if rainfall characteristics were seemingly similar, the landslide-triggering processes could potentially have differed, indicating that retrospective analysis of landsliding mechanisms would go beyond the scope of this study.

The rainfall anomalies related to all the analysed landslide events could be characterized based on analysis using $P_{\text{std}}$. It has been considered that a substantially larger amount of total rainfall is essential to trigger deep-seated landslides than to trigger shallow landslides (Uchida et al., 2013), which accords with the observation in this study that precipitation of >1000 mm was required to trigger deep-seated landslides (Figure 8). Deep-seated landslides sometimes occur more than 12 h after the peak hourly rainfall intensity (e.g. Tsou et al., 2011). In fact, at least seven deep-seated landslides occurred after $t_{72}$ (i.e. end of $P_{\text{std}}$) during the Kii Peninsula event (Yamada et al., 2012). Given this time lag, although the rainfall that triggered the deep-seated landslides could be characterized using $P_{\text{std}}$, it probably corresponded to the landslides that occurred on the more vulnerable hilltops in the case of the Kii Peninsula event.

5.2 Implications and limitations of standardizing the rainfall time span using $P_{\text{std}}$

For the examined landslide events, standardization of the rainfall time span using $P_{\text{std}}$ revealed that the mean rainfall intensity reached the 100-year return level at around the time of landslide occurrence (Figure 9). Therefore, the time at which the 72-h rainfall reached its maximum (i.e. $t_{72}$) and the changes in the mean rainfall intensity in the preceding 72 h (shown by the hyetograph for $P_{\text{std}}$) are operationally meaningful pieces of information. If this information could be obtained from forecast rainfall, the potential for and timing of numerous landsliding events at the regional scale could be estimated based on analysis of when or whether the mean rainfall intensity was expected to reach the 100-year return level (i.e. the IDF curve).

Even if $t_{72}$ for the forecast rainfall was unknown, evaluation of the return level of the mean rainfall intensity during arbitrary 72-h time periods would contribute to the decision regarding whether impending rainfall could potentially trigger numerous landslides at the regional scale.

These findings might allow the potential for landslides to occur as a result of forecast rainfall to be evaluated without need for records related to previous landslides in the target region (e.g. local rainfall thresholds). Analysis of rainfall based on $P_{\text{std}}$ requires only rainfall data, and it overcomes problems regarding the lack of other types of data (e.g. geological and topographical data). Most IDF curves are derived from annual rainfall maxima that are extracted by applying a moving interval to the rainfall record (e.g. Koutsoyiannis et al., 1998; Sane et al., 2018); thus, rainfall analysis using $P_{\text{std}}$ can be conducted without need to define rainfall events based on an arbitrary dry period. This approach would be valid, particularly when rainfall is likely to cause landslide-related hazards at the regional scale, where recordable landslides have not occurred previously, and/or where there is no available rainfall threshold.

Analysis of IDF curves based on rainfall extreme frequency usually demands a multidecadal rainfall record (e.g. Sane et al., 2018). The general sparsity of rain gauge stations has hampered the unravelling of local rainfall characteristics (Kidd et al., 2017). Even in Switzerland, where shallow landslides have resulted in economic losses of at least EUR 520 million during 1972–2007 (Hilker et al., 2009), the rain gauge network could not produce adequate multidecadal records of hourly rainfall covering the mountainous areas (Leonarduzzi & Molnar, 2020). Thus, the availability of rainfall records is the primary limiting factor regarding whether landslide likelihood could be evaluated using $P_{\text{std}}$ and IDF curves.

For the Tokachi region, the non-landslide-triggering rainfall of 1981 and the landslide-triggering rainfall of 2016 reached the 100-year return level at a similar time during $P_{\text{std}}$ as plotted on the log–log $D$–$I$ plane (Figure 9e). It indicates that this standardization method cannot be used to decide whether the peaks of short- or long-term rainfall intensity (e.g. the hourly intensity or 72-h intensity; Figures 6e and 7) will coincide. This means that more false-positive errors might result for regions where these peaks must coincide for landslides to be triggered, as was the case for the Tokachi event. In fact, physical modelling of slope stability has demonstrated that the appearance of short-term rainfall intensity peaks in the latter part of the relevant hyetographs could increase the probability of landsliding, because the earlier rainfall would have already produced wet conditions on the hillslopes (D’Odorico et al., 2005). Thus, in situations where this earlier rainfall reached the 100-year return level but did not trigger landslides, making direct comparisons between hyetographs might be a more valid approach than using mean rainfall intensity for investigating the potential for landslide occurrence (e.g. Peres & Cancelliere, 2016).

Even $P_{\text{std}}$ (72 h) might not be an adequate time span for evaluating the accumulated rainfall that prepares the dynamic predisposition for landsliding in certain cases. To remedy this, a longer duration (>72 h) could be used for $P_{\text{std}}$. However, for the Tokachi event, the maximum rainfall intensity for the period 144–168 h occurred ~8 days before the estimated time of landslide occurrence owing to the unusual antecedent precipitation produced by sequential typhoons (Figure S1). This is potentially responsible for the mismatch between $P_{\text{std}}$ and the time of landslide occurrence; indeed, the Tokachi event did not occur during the $P_{\text{std}}$ based on 144- and 168-h rainfall maxima (Figures S1 and S2). The evaluated characteristics of the rainfall occurring within $P_{\text{std}}$ comprised factors related to both the trigger and the dynamic predisposition for landsliding, and the temporal boundary between the trigger and the dynamic predisposition remains fundamentally unclear, similar to results of the different approach using the longer antecedent period (~15 days before landsliding; Leonarduzzi & Molnar, 2020).

In other cases, even if the times at which the maximum 96–, 120-, 144-, and 168-h rainfall intensities occurred were defined separately as the hypothetical end of the landslide-triggering rainfall, the intersection of the mean rainfall intensity and IDF curves for the 100-year return level might correspond to the timing of landsliding (Figures S2–S12). Here, depending on the hyetograph, the total precipitation during $P_{\text{std}}$ might increase substantially with extension of the rainfall duration. As shown in Figure 6j, the 95% confidence interval of the 100-year return level for 72-h rainfall intensity usually has a wider range (~1048–1635 mm) in accordance with the increase in total precipitation, which suggests that errors exceeding the accepted tolerance for the IDF curves are likely to occur. Considering these operational problems, $P_{\text{std}}$ should be set as the necessary minimum rainfall duration, implying that scrutiny of the appropriate time span
for $P_{\text{std}}$ using a large dataset of landslide-triggering rainfall events is worthwhile.

Most of the mean rainfall intensities had 100-year return levels that were greater than the mean rainfall intensities indicated by the global ID thresholds (Figures 8 and 9). These ID thresholds are mainly derived from the lower limits of landslide-triggering rainfall events, but might also include rainfall events that triggered debris flows owing to the entrainment of channel deposits as well as local landslides at locations far from rain gauge stations (e.g., Bogaard & Greco, 2018; Segoni et al., 2018). Additionally, for landslide-triggering rainfall events, the duration of the non-effective rainfall both before and after landslide initiation is often considered owing to the lack of information regarding the actual time of landslide occurrence (e.g., Hong et al., 2018). These uncertainties hamper direct comparison between the ID thresholds and IDF curves. However, at least the lower position of the ID thresholds in the log-log D–I plane, in comparison with the IDF curves, implies that rainfall that is less intense than the corresponding 100-year return level might also trigger landslides. Thus, standardization of the rainfall measurement period using $P_{\text{std}}$ to some extent characterizes rainfall events that trigger numerous landslides at the regional scale, but is probably less effective for rainfall that triggers fewer landslides at the hillslope scale. Therefore, the approach using $P_{\text{std}}$ and IDF curves could support the provision of alerts for landslide hazards at the regional scale (i.e., the maximum alert level for landslide hazards overarching from hillslope to regional scales), but it should be combined with the use of other thresholds for more comprehensive evaluation of local risks.

6 CONCLUSIONS

To compare the characteristics of rainfall intensities over different lengths of time, this study examined the return levels of rainfall related to 10 landslide events, which included shallow landslides in volcanic and non-volcanic areas as well as deep-seated landslides. To ensure consistent comparison, based on the time that the 72-h (3-day) rainfall reached a maximum ($t_2$), the rainfall periods used for the analysis were standardized as $P_{\text{std}}$ (i.e., the 72 h before $t_2$).

In most cases, the 72-h rainfall intensity at $t_2$ was the highest in the respective rainfall records, and reached or exceeded the 95% confidence interval of the 100-year return level. For the rainfall events that triggered shallow landslides, the rainfall patterns within $P_{\text{std}}$ could be divided broadly into two types: downpours that repeatedly reached the 100-year return level during a period of approximately 3 h; and rainfall that produced totals of approximately 200–400 mm over longer periods but within 72 h. Although all the examined shallow landslide events that occurred in volcanic areas were categorized as corresponding to the former type, some shallow landslide events in non-volcanic areas were also found to have been triggered by the former type of intense rainfall. This suggests that the rainfall patterns that trigger landslides are probably not constrained by lithological setting alone. It was also found that rainfall of >1000 mm was required to initiate the examined deep-seated landslide. Thus, there were clear differences between the hyetographs corresponding to the different examined landslide events.

Irrespective of differences in the hyetographs, most landslides occurred when the mean rainfall intensity within $P_{\text{std}}$ reached the level of the IDF curve for the 100-year return level. Therefore, changes in the rainfall return level of the mean rainfall intensity within $P_{\text{std}}$ might allow estimation of the potential for and timing of landslides. This approach to processing rainfall data could be implemented independently of rainfall thresholds, and could contribute to assessment of whether forecast rainfall has potential to trigger landslides, especially in regions where data on earlier landslides are limited but a long-term rainfall record is available.

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CONFLICT OF INTEREST

The author has no conflict of interest.

DATA AVAILABILITY STATEMENT

The polygon and point data of the landslide scars and inundated areas are available on the websites of the Association of Japanese Geographers (http://ajg-disaster.blogspot.com/2018/07/3077.html) and the Geospatial Information Authority of Japan (https://www.gsi.go.jp/bousai.html). The rainfall data are available on the website of the Japan Meteorological Agency (https://www.data.jma.go.jp/obd/stats/etrn/).

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ABSTRACT

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