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Synoptic-scale precursors of landslides in the western Himalaya and Karakoram

Kieran M. R. Hunt¹ and A. P. $Dimri^2$

¹Department of Meteorology, University of Reading, UK ²School of Environmental Sciences, Jawaharlal Nehru University, Delhi, India

Abstract

In the Upper Indus Basin (UIB), precipitation associated with synoptic-scale circulations impinges on the complex and steep orography of the western Himalaya and Karakoram. Heavy rainfall often falls over the foothills, frequently triggering landslides there. This study explores the role of these synoptic-scale circulations – extratropical western disturbances (WDs) and tropical depressions (TDs) – in producing the conducive conditions necessary to trigger landslides, using data from the NASA Global Landslide Catalog and WD and TD track databases.

During the winter (October to April), UIB landslides peak in February and occur at a rate of 0.05 day^{-1} , 61% of which are associated with the passage of a WD. They are most common when a WD is located within a few hundred kilometres of 30° N, and significantly rarer if the WD is north of 40° N. WDs provide moist southwesterly flow from the Arabian Sea (AS) and Mediterranean Sea to the UIB, resulting in large-scale precipitation, but landslide probability is not related to WD intensity. Non-WD winter landslides are associated with small-scale orographic precipitation that we hypothesise is due to cloudbursts.

During the summer (May to September), UIB landslides peak in August and occur at a rate of 0.11 day⁻¹, 60% of which are associated with TD activity. Many of these TDs are found over central India, slightly south of the climatological monsoon trough, where they provide strong monsoonal southeasterlies to the UIB flowing along the Himalayas. Increased landslide frequency is also associated with TD activity over the southern Bay of Bengal (BoB), and it is hypothesised that this is related to monsoon break conditions. Landslide frequency is significantly correlated with TD intensity. Non-TD landslides are associated with a northwestward extension of the monsoon trough, providing southeasterly barrier flow to the UIB.

Implications for forecasting and climate change are discussed.

Keywords: landslides; Indus Basin; western disturbances; depressions; precipitation; moisture flux

1 Introduction

About 75% of all global landslides occur in Asia; of 2 these, the majority happen along the mountain ranges 3 of south Asia: the Himalaya, the Hindu Kush, and the 4 Karakoram (Froude and Petley, 2018). Along most 5 of the Himalaya, the seasonality of landslides is determined by the arrival and withdrawal of the sum-7 mer monsoon (Kirschbaum et al., 2010; Petley, 2012). 8 However, further west, over the western Himalaya and 9 Karakoram, the influence of the monsoon is less impor-10 tant, and landslides become increasingly frequent in the 11 winter and spring months (Atta-ur Rehman et al., 2011; 12 Saleem et al., 2020). This results in a complex seasonal-13 ity of landslides in the Upper Indus Basin (UIB) (Fig. 1) 14 which makes adequate preparation and mitigation a 15 challenge. Over the UIB, landslides have been responsi-16 ble for changing the course of Indus tributaries (Hewitt, 17 1998), significant loss of life (Kirschbaum et al., 2012; 18 Froude and Petley, 2018) and significant socioeconomic 19 damage (Atta-ur Rehman et al., 2011). 20

Weather patterns over the Indus Basin are highly sea-21 sonal. In the winter (October to April, i.e. outside of 22 the monsoon season), mean low-level winds are weakly 23 westerly with a strong upper-level westerly jet. Within 24 this jet, extratropical cyclones, known as western dis-25 turbances (WDs), are embedded, and are responsible for most winter precipitation in this region. In summer 27 (May to September), the westerly jet migrates poleward 28 and the frequency of WDs reduces dramatically. In-29 stead, precipitation usually occurs as a result of north-30 westward extensions of monsoon activity. Tropical de-31 pressions (TDs) are responsible for about half of sum-32 mer monsoon precipitation, though the fraction is gen-33 erally slightly lower over the Indus Basin (Hunt and 34 Fletcher, 2019). Climatological precipitation, which is 35 greatest along the foothills, is shown for both winter 36

and summer in the middle panels of Fig. 2. Many pre-37 vious studies agree that heavy rainfall is the most im-38 portant environmental precursor to landslides globally 39 (Kirschbaum et al., 2012, 2015), especially over south 40 Asia (Dahal et al., 2008; Kirschbaum et al., 2010; Zhang 41 et al., 2019), where intensity-duration thresholds are 42 usually used as predictors with some success (Sengupta 43 et al., 2010; Kirschbaum et al., 2011). It is also impor-44 tant that there has been significant antecedent rainfall 45 so that the soil is either nearly or totally saturated (Ga-46 bet et al., 2004; Ahmed et al., 2014; Kumar et al., 2014), 47 although the relative importance is highly sensitive to 48 surficial geology. A few case studies of particularly sig-49 nificant flooding and landslide events across the west-50 ern Himalaya have looked for meteorological precursors. 51 For example, the heavy rainfall that caused extensive 52 flash flooding and landslides over Jammu and Kashmir 53 in 2010 were ascribed to multiple mesoscale convective 54 storms that were steered into the region by favourable 55 mid-tropospheric jet. These coincided with southerlies 56 from the Arabian Sea (AS), which provided the mois-57 ture required for unusually sustained, heavy rainfall 58 (Kumar et al., 2014). Analysis of the 2013 Uttarak-59 hand floods and associated landslides found the an-60 tecedent rainfall came about as a strong WD interacted 61 with a TD, providing very large southeasterly moisture 62 flux to the orography (Mishra, 2015; Chevuturi and 63 Dimri, 2016; Hunt et al., 2020a). More generally, moist 64 southwesterlies embedded in WDs (Dimri et al., 2015) 65 provide heavy rainfall and sometimes cloudbursts due 66 to conjugate factors of convective triggering and oro-67 graphic locking (Dimri et al., 2017). 68

Heavy rainfall is not the only important ingredient ⁶⁹ for landslides. As mentioned earlier, the soil and underlying rock must usually be very close to field capacity before the shear strength within a slope is reduced enough for landslides to be initiated. In some ⁷³



Figure 1: Locations of landslides in the NASA GLC, covering 2007-2015. Blue dots mark landslides occurring between October and April; brown dots mark those occurring between May and September. The thick black line marks the boundary of the Indus Basin. Left: over the Indian peninsula and surrounding area; right: over the Indus Basin (marked in black).

locations may only be possible near the end of the wet 74 season. For example, Gabet et al. (2004) found that a 75 mean seasonal rainfall of 860 mm was required to get 76 the regolith up to field capacity over Nepal and as a 77 result, landslides peak there during the latter part of 78 the summer monsoon. Even then, several environmen-79 tal and geological factors must be taken into consider-80 ation. A review of global landslide research by Zhang 81 et al. (2019) noted that slope gradient and underlying 82 lithology were important in determining landslide size, 83 whereas climatological rainfall was important in deter-84 mining their location. Looking specifically at the UIB, 85 a detailed study by Ahmed et al. (2014) explored factors controlling landslide susceptibility. They separated 87 their analysis into environmental and causative risk fac-88 tors. For environmental factors, they found that UIB 89 landslides were most common on slopes with an angle 90 between 30° and 45° , at elevations between 2 and 4 km. 91 They also found a significant, though less important, 92 relationship with slope aspect, finding that landslides 93 were most common on slopes facing a heading of 226°-94 270° – this is the orientation at which moist southwest-95 erlies, often induced by passing WDs, would strike the orography head on. Ghosh et al. (2011) found similar 97 results for the eastern Himalaya near Darjeeling, also

highlighting the role of lithology, geomorphology, and aa land use. These environmental factors are useful in un-100 derstanding the spatial patterns of landslide frequency, 101 and thus are important in reducing false alarms. How-102 ever, being quasi-static, they are not useful for predict-103 ing when landslides might occur. Causative risk fac-104 tors, on the other hand, provide temporal information 105 and are thus potentially useful for predictions through 106 an understanding of the underlying meteorology. 107

Despite the global and regional linkages with heavy 108 rainfall and the case studies discussed earlier, very lit-109 tle work has been done to assess meteorological precur-110 sors to landslides in south Asia over climatological sam-111 ple sizes. Kamae et al. (2017) found that atmospheric 112 rivers were important in providing the heavy precipita-113 tion needed to trigger landslides over East Asia. Ma-114 madjanova et al. (2018) found that moist westerlies 115 were an important precursor for landslides in Uzbek-116 istan and assessed the relative role of cyclonic circula-117 tions. Saleem et al. (2020) noted that WDs play an 118 "equal role" to monsoonal rainfall in instigating land-119 slides over northern Pakistan. 120

A full understanding of the meteorological precursors of landslides is a vital component in constraining 122 their sensitivity to climate change, in particular be124 average precipitation and extreme precipitation events 125 to the Himalaya (Hunt et al., 2018b, 2019b; Hunt and 126 Fletcher, 2019). For example, monsoon depressions 127 are projected to fall in frequency, but move poleward 128 (Sandeep et al., 2018); and WDs are projected to bring 129 more intense rainfall (Hunt et al., 2020b), though the 130 fate of their frequency in a changing climate remains an 131 open question (Ridley et al., 2013; Krishnan et al., 2018; 132 Hunt et al., 2019a). Correspondingly, Kirschbaum et al. 133 (2020) found a projected increase in landslides over the 134 Himalaya when they applied a landslide hazard model 135 to global climate model output. Increases in the ob-136 servational record have been found for the central Hi-137 malaya (Petlev et al., 2007) and the Hindu-Kush Hi-138 malaya (You et al., 2017); though Petley et al. (2007) 139 attributed the increase in their record to be due to in-140 creased construction and infrastructure work on slopes. 141 The key research questions to be addressed, therefore, 142 are: 143

• What role does the presence of synoptic-scale sys-144 tems, such as TDs and WDs, play in initiating 145 landslides over the western Himalaya and Karako-146 ram, in particular over the UIB? 147

• Are there other weather patterns that act as im-148 portant precursors? 149

• Can this information be used to improve landslide 150 predictability? 151

$\mathbf{2}$ Data and methodology 152

$\mathbf{2.1}$ NASA Global Landslide Catalogue 153

In this study, we use landslide data from the NASA 154 Global Landslide Catalogue (GLC, Kirschbaum et al., 155 2010, 2015). The GLC is a catalogue of precipitation-156 triggered landslides identified from sources including into the forecasting system from a variety of sources, 157

cause synoptic-scale systems contribute a majority of news reports, disaster databases, and research papers. 158 Triggers are determined from the source material itself. 159 The catalogue contains over 11000 entries from 2007 160 to 2015, 327 of which occur in the UIB. Uncertainty 161 in recorded location is also included within the GLC. 162 Of the 327 UIB landslides, 88% have a location uncer-163 tainty of less than 50 km, 70% within 25 km, and 45%164 within 10 km. 11% lack spatial uncertainty data. Each 165 landslide is assigned a UTC day-of-occurrence, but un-166 certainty in temporal information is not provided. We 167 do not expect spatiotemporal uncertainties to signifi-168 cantly affect our results. However, it is worth noting 169 here that the nature of the sources used in the GLC 170 impart a bias towards populated regions - where land-171 slides are more likely to be reported – and hence towards 172 more anthropogenic triggers. The GLC also includes a 173 measure of 'size' for each landslide, ranging from small 174 (shallow; affecting one hillslope; minimal or no infras-175 tructure damage) to very large (multiple events affect-176 ing an entire region; catastrophic damage to infrastruc-177 ture; often encompassing whole villages). Of the 327 178 UIB landslides in the catalogue, 14 are rated small, 284 179 medium, 26 large, and 3 very large. For more infor-180 mation on uncertainty and size measures, the reader 181 is encouraged to read Secs. $2.1~{\rm and}~2.2$ of Kirschbaum 182 et al. (2015). 183

2.2**ERA-Interim**

To analyse the structure of dynamical fields and mois-185 ture in the atmosphere, we use the European Centre 186 for Medium-Range Weather Forecasts Interim reanaly-187 sis (ERA-I: Dee et al., 2011). All fields are available at 188 six-hourly intervals with a horizontal resolution of T255 189 (~ 78 km at the equator), with the three-dimensional 190 fields further distributed over 37 vertical levels span-191 ning from the surface to 1 hPa. Data are assimilated 192 193

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including satellites, ships, buoys, radiosondes, aircraft,
and scatterometers. In this study, we use wind and humidity data at all pressure levels to compute verticallyintegrated moisture flux.

198 2.3 GPM-IMERG

For our precipitation dataset, we use the gridded sur-199 face product Integrated Multi-Satellite Retrievals for 200 GPM (IMERG; Huffman et al., 2015). This has global 201 coverage at a half-hourly, 0.1° resolution, starting June 202 2000 and continuing to the present day. Over the trop-203 ics, IMERG primarily ingests retrievals from (for 2000-204 2014) the now-defunct Tropical Rainfall Measuring Mis-205 sion (TRMM; Kummerow et al., 1998, 2000) 13.8 GHz 206 precipitation radar and microwave imager (Kozu et al., 207 2001) and (for 2014-) the Global Precipitation Mea-208 surement (GPM; Hou et al., 2014) Ka/Ku-band dual-209 frequency precipitation radar. Where an overpass is 210 not available, precipitation is estimated by calibrat-211 ing infrared measurements from geostationary satel-212 lites. While GPM-IMERG performs well when com-213 pared against gauge-based products, performance falls 214 at higher elevations or when quantifying extreme rain-215 fall events (anj; Prakash et al., 2018). Given the nature 216 of our study, this introduces some uncertainty into our 217 results. As such, all key results are additionally verified 218 using precipitation output from the Indian Monsoon 219 Data Assimilation and Analysis (IMDAA) reanalysis 220 project (Ashrit et al., 2020), as high resolution reanaly-221 ses have been shown to perform well in the Indus Basin 222 (Baudouin et al., 2020). Full half-hourly resolution is 223 used in the antecedent rainfall analysis in Fig. 8, but 224 daily accumulations (00UTC-00UTC) are used for the 225 composite analysis in Figs. 2 and 6. 226

2.4 Track databases

We use the database of WD tracks from Hunt et al. 228 (2018a) in this study. Using six-hourly ERA-Interim 229 data, they tracked WDs by computing the mean relative 230 vorticity in the 450–300 hPa layer, then performed a 231 spectral truncation at T63 to filter out short-wavelength 232 noise. They then identified positive-definite vorticity 233 regions within this field and determined the centroid 234 location for each one. These centroids were then linked 235 in time, subject to constraints in distance and steering 236 winds, to form candidate WD tracks. Finally, those 237 candidate tracks that did not pass through South Asia 238 $[20^{\circ}N-36.5^{\circ}N, 60^{\circ}E-80^{\circ}E]$, have a lysis to the east of 239 their genesis, or last at least 48 hours were rejected. 240 This catalogue is publicly available at http://dx.doi. 241 org/10.5285/233cf64c54e946e0bb691a07970ec245. 242

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We use the database of TD tracks from Hunt 243 and Fletcher (2019) in this study. The core of the 244 algorithm used for this is identical to that used to 245 develop the WD catalogue above, except the input 246 is the truncated 900-800 hPa relative vorticity field. 247 There is no domain filtering at the end but tracks 248 shorter than 48 hours are still rejected. This catalogue 249 is publicly available at http://gws-access.jasmin. 250 ac.uk/public/incompass/kieran/track_data/lps-251 tracks_v1_1979-2014.csv. Note that in this study, 252 tropical depression is used as a collective term to 253 refer to monsoon low-pressure systems, monsoon 254 depressions, tropical lows, and tropical storms. 255

3 Results

As discussed in the introduction, the overwhelming majority of landslides in the Himalaya occur after heavy 258 precipitation. To show what form it takes, mean precipitation is plotted for days on which a landslide occurs in the Upper Indus Basin (UIB; an area that con-261



Figure 2: Mean precipitation for days in (a) October–April and (b) May–September in which a landslide occurs in the UIB (left panels) compared with the mean for all days in the same period (2007–2015, middle panels). Difference plots shown on the right. Data from GPM-IMERG. The solid black line marks the edge of the Indus River Basin. The same figure, computed using IMDAA reanalysis, is shown in supplementary Fig. S1.

tains both the western Himalaya and the Karakoram) 262 in Fig. 2. For comparison, the precipitation climatol-263 ogy is shown, and the results are partitioned by season 264 into the summer monsoon (May through September) 265 and winter (October through April). Before proceed-266 ing, we remind the reader of the inherent uncertainties 267 and biases present in the landslide dataset, as discussed 268 in Sec. 2.1. Landslides are approximately twice as fre-269 quent in the UIB during the summer monsoon (occur-270 ring on 11% of days) as during the winter (occurring 271 on 5% of days); this contrasts the central and east-272 ern Himalaya, for which landslides almost exclusively 273 occur during the summer monsoon (e.g. Kirschbaum 274 et al., 2010). For a complete understanding of pre-275 cursor weather, we must, therefore, consider the syn-276

optic dynamics at work in both seasons. Computing 277 the dispersion statistic, σ , (not shown), we find that 278 UIB landslides are highly temporally clustered in the 279 summer season ($\sigma \sim 4$), and quite temporally clus-280 tered in the winter season ($\sigma \sim 1.5$). It is evident from 281 Fig. 2 that the precipitation is much heavier on land-282 slide days in both seasons (in some places averaging 283 20 mm day⁻¹ more than the climatology). For events 284 that occur during the summer monsoon (Fig. 2(b)), the 285 anomalous precipitation has a footprint that extends 286 southeast, far beyond the UIB, over the Indian subcon-287 tinent. This suggests that the anomalous rainfall over 288 the UIB during the monsoon probably comes about as 289 a result of enhanced monsoon activity, either through 290 an active phase or through the passage of a monsoonal 291

tropical depression (TD). Disentangling active phases 292 of the monsoon from TD activity is difficult, and some 293 authors (e.g. Rajeevan et al., 2010) consider them to be 294 synonymous; however, TDs have been shown to provide 295 considerable monsoon rainfall to northwest India and 296 Pakistan if they sufficiently penetrate the subcontinent 297 or start in the Arabian Sea (Hunt and Fletcher, 2019). 298 It is reasonable, therefore, to assume that the source 299 of anomalous precipitation over the UIB during sum-300 mer monsoon landslide is increased TD activity. For 301 winter landslides (Fig. 2(a)), the footprint of anoma-302 lous precipitation is, for the most part, confined to the 303 UIB, suggesting either a local source, or a more distant 304 source capable of providing significant moisture flux. 305 WDs are the predominant source of both climatolog-306 ical precipitation and heavy precipitation events over 307 the winter UIB (e.g. Hunt et al., 2018b), but there are 308 other important sources such as non-WD cloudbursts 309 and orographic precipitation (Dimri, 2006). 310

311 **3.1** Synoptic controls on landslide fre-312 quency

We will now test the hypothesis that TDs and WDs 313 are responsible for the majority of monsoon and winter 314 landslides respectively in the UIB. Using the track cat-315 alogues described in Sec. 2.4, we find the nearest TD 316 or WD to each UIB landslide event. To identify the 317 nearest TD and WD, all track points for the respective 318 systems logged on the UTC day on which the landslide 319 occurred are considered. Among those, the point (one 320 for each type of system) with the shortest great-circle 321 distance to the landslide is used. The results that follow 322 in this study are not qualitatively sensitive to the choice 323 of temporal window. The distances between landslide 324 and system are tallied by month in Fig. 3. WDs have a 325 strong annual cycle and are most common in the win-326 ter, and this projects strongly onto their relationship 327

with UIB landslide frequency (Fig. 3(a)). From Novem-328 ber through May (narrowly excepting April), a WD is 329 present within 5000 km of a majority of UIB landslides. 330 Integrated over the non-monsoon months of October 331 to May, this value is 61%, of which 45% occur within 332 2000 km. TDs also have a strong annual cycle, al-333 though they do continue in significant numbers outside 334 the monsoon. Again, this projects strongly onto their 335 relationship with UIB landslide frequency (Fig. 3(b)), a 336 TD is present with 5000 km of 60% of all UIB landslides 337 occurring between May and September, of which it is 338 within 2000 km 35% of the time. There is a secondary 339 peak of TD contribution in February and March, due 340 to pre-monsoon tropical cyclones in the Arabian Sea. 341

So, we have seen that landslide hazard in the UIB 342 has a relatively strong dependence on both the presence 343 and proximity of synoptic circulations such as WDs and 344 TDs. It is therefore reasonable to assume that land-345 slides are also sensitive to system location. For ex-346 ample, if a WD is too far east, then the associated 347 southerly moisture flux will impinge on the central or 348 eastern Himalaya instead of the western Himalaya, and 349 the chance of a landslide in the UIB will not be sig-350 nificantly increased. To test this hypothesis, we use 351 the extract from the TD and WD track catalogues all 352 systems that occur during the GLC period (2007–2015). 353 Individual track points are then binned into $2^{\circ} \times 2^{\circ}$ grid-354 boxes, and the mean frequency of UIB landslide occur-355 rence given a system present in that gridbox is com-356 puted. These maps, filtered by their respective seasons 357 (Mav–September for TDs; October–April for WDs) are 358 shown in Fig. 4. The map for WDs (Fig. 4(a)) has sev-359 eral key features. Firstly, we see that the latitude of 360 the WD, tightly controlled by the latitude of the sub-361 tropical westerly jet in which it is embedded (Dimri 362 and Chevuturi, 2016), is very important in determining 363 whether the likelihood of a UIB landslide is increased or 364



Figure 3: Monthly frequencies of landslides in the UIB. Bars coloured by proximity of (a) western disturbance and (b) tropical depression. Black indicates no system present within 5000 km of the landslide.



Figure 4: Mean likelihood of a landslide in the UIB, given presence of (a) a western disturbance in October–April or (b) a tropical depression in May–September in a given $2^{\circ} \times 2^{\circ}$ gridbox. The climatological values of 5% day⁻¹ and 11% day⁻¹ for October–April and May–September respectively are drawn with a solid blue line. Grid boxes with fewer than five systems in are not shown.

decreased. Systems within $3-4^{\circ}$ of $30^{\circ}N$ result in a sig-365 nificantly increased chance of a UIB landslide, whereas 366 those much further north, particularly beyond about 367 40°N, result in a significantly decreased chance. Sec-368 ondly, there are three zonal maxima in frequency at 369 about 30°E, 50°E, and 70°E respectively. This approx-370 imate wavelength of 2000 km corresponds to the spatial 371 scale of WDs in the subtropical westerly jet (Rao and 372 Srinivasan, 1969) and highlight the fact that when a 373 WD is over, or very close to, the UIB, a younger one 374 is often ~ 2000 km upstream. It is quite possible that 375 these upstream WDs play an important role in trigger-376 ing landslides over the UIB, and we will briefly discuss 377 this in the next section; however a full treatment of the 378

role of coupled WD dynamics in bringing heavy precipitation to the Indus Basin is left for future work. 380

For monsoonal TDs (Fig. 4(b)), we see that prox-381 imity to the UIB is the most important parameter in 382 increased landslide frequency. TDs aligned along the 383 southern edge of the climatological monsoon trough re-384 sult in UIB landslide frequencies of up to 20% day⁻¹. 385 Since TDs very rarely penetrate the subcontinent as 386 far as the Indus Basin itself, they cannot provide the 387 rain directly; instead, they must enhance the south-388 easterly monsoonal moisture flux that impinges on the 389 UIB orography. This also explains why it is preferable 390 for them to be near the south of the typical monsoon 391 trough location; too far north and they would direct 392



Figure 5: Change in likelihood of a landslide occurring in the UIB on a given day as a function of the intensity percentile of western disturbances (blue) and tropical depressions (orange) within 2000 km. For each season (October–April, left; May–September, right), the rarer type of system is plotted with a dotted line. A dashed grey line marks the climatological landslide frequency for each season. The grey areas indicate where the populations of the more common system are too small for the results to be significant. Note that the *y*-axes have different scales in each subplot.

those moist southwesterlies into the central or eastern 303 Himalaya instead. We will explore this moisture flux 394 framework more in the next section. Of additional note 395 are the pronounced minimum in the northeastern AS – systems here would result in anomalous dry northerlies 397 or northeasterlies passing over the Indus Basin - and 398 the secondary maximum located towards the southern 399 Bay of Bengal. We propose that this maximum comes 400 about indirectly through monsoon breaks; during such 401 periods, TD activity is highly favoured over Sri Lanka 402 and the southern Bay of Bengal (Deoras et al., 2020), 403 meanwhile the monsoon westerly jet is deflected north, 404 where it impinges upon the western Himalaya. Mon-405 soonal TDs are climatologically far less frequent in this 406 region than over the head of the Bay of Bengal or within 407 the monsoon trough, so this rainfall signal may not be 408 immediately obvious. We will explore this more in the 409 next section. 410

Along with system location, system intensity may
also be an important control on landslide frequency.
For example, a strong low-pressure system over central
India would be capable of providing much larger mois-

ture flux to the Indus Basin than a very weak one. To 415 determine this relationship, we consider only systems 416 within 2000 km of the landslide, which excludes the up-417 stream WDs and monsoon break TDs discussed previ-418 ously. The probability of a landslide occurrence is then 419 computed at each percentile of system intensity (using 420 850 hPa relative vorticity for TDs and 350 hPa relative 421 vorticity for WDs), such that the zeroth percentile in-422 cludes all systems, the fiftieth percentile includes those 423 stronger than the median, and so on. These intensity-424 probability charts are shown in Fig. 5. For October-425 April (left panel), we see that WD intensity has very 426 little bearing on UIB landslide probability. A maxi-427 mum at about the sixtieth percentile corresponds to a 428 frequency of 5.5% day⁻¹, only marginally higher than 429 the climatological value of 5% day⁻¹. The presence of 430 strong TDs during the winter is, on average, detrimen-431 tal to landslide probability, which falls below $4\% \text{ day}^{-1}$ 432 at the fiftieth percentile of TD intensity. This is corrob-433 orated by Fig. 4(a), since TDs are much more common 434 in the Arabian Sea during winter than summer. During 435 May–September (right panel), we see that TD intensity 436

is strongly correlated with landslide occurrence, which 437 is 50% higher during the passage of the most intense 438 TDs (i.e. eightieth percentile and up) than for the aver-439 age. The same is true of WDs, but they are rare during 440 the summer months. 441

In summary, there is a significantly increased chance 442 of a landslide in the UIB during winter if a WD is po-443 sitioned at or around 30° N either over or ~ 2000 km to 444 the west of the Indus Basin. The intensity of the WD is 445 not important. There is a significantly increased chance 446 of a landslide in the UIB during the summer monsoon if 447 a TD is present along the southern boundary of the cli-448 matological location of the monsoon trough, and within 449 2000 km of the basin. Stronger TDs result in a further 450 increase in landslide probability. 451

3.2Synoptic controls on precursor pre-452 cipitation 453

We now know that system location (and in some cases 454 intensity) as well as precipitation is important in fore-455 casting landslide occurrence. Here, we will now explore 456 the relationship between the two. We start by com-457 paring mean precipitation on days in which a landslide 458 occurs in the UIB for instances where a TD or WD is 459 within 2000 km with instances where they are not, as 460 shown in Fig. 6. This is a relatively strict criterion, 461 as we will see, but still gives a respectable sample size: 462 32 out of 116 winter UIB landslides occur with a WD 463 within 2000 km and 47 out of 211 summer UIB land-464 slides occur with a TD within 2000 km. During winter 465 UIB landslides (Fig. 6(a)), the presence of a local WD 466 causes heavier precipitation over the basin as a whole. 467 but also results in the region of heaviest precipitation 468 moving northeastwards, penetrating deeper into the Hi-469 malayan range (see right panel). We might expect this 470 to result in landslides also occurring deeper into the 471 472

is present, but the difference in mean landslide loca-473 tion between the two populations (WD present and no 474 WD present) is not statistically significant. What are 475 the reasons for these differences in precipitation? As 476 discussed earlier, WDs are the major source of winter 477 precipitation in the UIB, but there is also a consider-478 able contribution from smaller-scale cloudbursts and lo-479 calised orographic precipitation. These smaller storms 480 do not usually result in significant moisture flux into 481 the UIB, unlike WDs, and thus provide more isolated 482 rainfall and snowfall, which is what we see in the right 483 panel of Fig. 6(a). Cloudbursts have been previously 484 been associated with cases of severe landslides in the 485 western Himalaya (Mishra, 2015). The broad mois-486 ture flux commonly associated with WDs can penetrate 487 further inland than the weaker, smaller-scale flux as-488 sociated with cloudbursts and orographic storms, and 489 so this may also be the reason for the enhanced up-490 slope precipitation. WDs also bring some additional 491 moisture along the subtropical westerly jet (e.g. Singh 492 et al., 1981), which impacts orography at higher lati-493 tudes than southwesterly moisture flux from the AS. 494

During monsoonal UIB landslides (Fig. 6(b)), the 495 presence of a nearby TD causes a significant intensifi-496 cation of the rainfall band along the western Himalaya 497 compared to when one is not present. As discussed ear-498 lier, this is likely due to an intensification of the mon-499 soon trough combined with an anomalously southern 500 TD, resulting in anomalously large southeasterly mois-501 ture flux penetrating deep into the Indus Basin. The 502 region of north India to the east of the Indus Basin, 503 south of the central Himalava, experiences simultane-504 ous anomalous drying. It is unlikely that this is due 505 to monsoon breaks, the secondary cause proposed ear-506 lier, as we have filtered out contributions from the more 507 distant TDs that would be associated with this – this 508 Himalaya (i.e. further to the northeast) when a WD is in fact confirmed by extending the map southwards 509



Figure 6: Mean precipitation $[mm day^{-1}]$ on days in which a landslide occurred in the UIB in (a) October to April and (b) May to September. Separated into days where (a) a WD or (b) a TD (b) is within 2000 km of the landslide (left panels) and days where the respective system is either absent or farther than 2000 km away (middle panels). The right panels show the difference between the two. Individual landslide locations are marked with black dots. Precipitation data from GPM-IMERG.

The same figure, computed using IMDAA reanalysis, is shown in supplementary Fig. S2.

(not shown), which shows excess rainfall over much of
the monsoon core zone. Instead, we propose that TDs
over central India, which as we previously showed, were
likely to be further south than their climatology, no
longer cause moist barrier flow southeasterlies along the
central Himalayan foothills; rather, they advect moisture towards the UIB from further south.

We will now test and explore these claims, and those 517 made regarding WDs, in a moisture flux framework. 518 Fig. 7 is constructed in the same way as Fig. 2, but 519 instead uses vertically integrated moisture flux. The 520 composites in Fig. 7(a) confirm our earlier hypothesis 521 that the presence of a local WD during or immedi-522 ately before a UIB landslide results in a much wider 523 and stronger branch of westerly/southwesterly mois-524 ture flux over the Indus Basin than during its absence. 525

This causes more widespread precipitation and deeper 526 penetration of moisture into the continent, and hence 527 greater precipitation deeper into the mountain ranges, 528 as we saw in Fig. 2(a). The scale and westward (i.e. up-529 stream) extent of this anomalous moisture flux suggests 530 that upstream WDs may also play an important role, 531 though further work is needed to confirm this. We also 532 see that during landslide days when a WD is not present 533 (or within 2000 km), there is still a significant easterly 534 atmospheric river passing over the Indus Basin. Though 535 it is located too far south to provide significant moisture 536 to the foothills or mountains where landslides are most 537 frequent, it does show evidence of northward excursions 538 that would provide a sufficient source of moisture for 539 cloudbursts and orographic storms, as proposed earlier. 540

For UIB landslides occurring during the monsoon sea-



Figure 7: Mean vertically integrated moisture flux $[\text{kg m}^{-1} \text{ s}^{-1}]$ on days in which a landslide occurred in the UIB in (a) October to April and (b) May to September. Arrows indicate the vector field, and coloured contours show its magnitude. As in Fig. 6, these are separated into days where (a) a WD or (b) a TD (b) is within 2000 km of the landslide (left panels) and days where the respective system is either absent or farther than 2000 km away (middle panels). The right panels show the difference between the two.

son (Fig. 7(b)), the structure of the monsoon domi-542 nates, regardless of whether a TD is present or not. 543 Strong monsoonal westerlies bring a significant quantity 544 of moisture over the Indian peninsula, some of which 545 reaches the BoB and is subsequently directed north-546 westward by the monsoon trough. This results in a 547 thin but intense stream of moisture - the so-called bar-548 rier flow - reaching the edge of the Indus Basin, enough 549 to support thunderstorms over the western Himalaya. 550 When a TD is present, the monsoon trough is zonally 551 extended and deepened (right panel). Perhaps most 552 importantly, however, the whole circulation is further 553 south than in the non-TD composite. As we hypothe-554 sised, this reduces the barrier flow (which tends to rain 555 out over the foothills of the central Himalaya) but en-556

hances southeasterly moisture flux into the Indus Basin and onto the western Himalaya and Karakoram. 558

We have spoken at length about precipitation oc-559 curring on the day of the landslide, which previous 560 authors have also given the greatest importance (e.g. 561 Dahal et al., 2008; Kirschbaum et al., 2010). How-562 ever, other authors have pointed out that in addition to 563 this, sustained antecedent rainfall, snowmelt, or runoff 564 is required to raise the local soil moisture and reduce 565 slope stability (Gabet et al., 2004; Kumar et al., 2014). 566 We will now explore the relative importance of these 567 two contributing timescales. To do this, we composite 568 precipitation over two scales: maximum precipitation 569 within 25 km of the landslide, and mean precipitation 570 within 100 km of the landslide. The latter captures 571



Figure 8: Antecedent rainfall, computed using GPM-IMERG, for UIB landslides occurring in (a) October–April and (b) May–September. Both mean rainfall within 100 km of the landslide (cyan, magenta) and maximum rainfall within 25 km of the landslide (blue, red) are partitioned according to whether a synoptic system was (cyan, blue) or was not (magenta, red) present and within 2000 km of the event. The grey area indicates the 24-hour period in which the landslide occurred, aligned so that 0 hr on the x-axis always corresponds to 0000 UTC. An 8-hour low-pass smoothing is applied to all data to reduce noise. The same figure, computed using IMDAA reanalysis, is shown in supplementary Fig. S3.

precipitation features on a larger scale (either convec-572 tive or stratiform), which tend to respond to both syn-573 optic dynamics and local forcing; the former captures 574 high intensity small-scale precipitation, which is typi-575 cally convective in nature and more sensitive to local 576 forcing. Another reason to include two scales is the 577 spatial error in GPM-IMERG precipitation, which can 578 suffer from anvil bias in regions of large vertical wind 579 shear (Shrestha et al., 2015), such as south Asia under 580 the presence of the winter subtropical westerly jet or 581 monsoonal easterly upper-tropospheric jet. When con-582 structing these composites, we also align the time axes 583 of each event such that the diurnal cycle is consistent; 584 that is, 0 hr always represents 0000 UTC on the day 585 in which the landslide occurs, -12 hr always represents 586 1200 UTC on the day before the landslide occurs, and 587 so on. This is done for two reasons: firstly, the diurnal 588 cycle of precipitation has considerable magnitude in the 589 orographic regions of south Asia (Ahrens et al., 2020) 590 and so it makes sense to preserve it when compositing; 591 secondly, precise timing of the landslide is not always 592 available in the NASA GLC, so we can only consistently 593 attach each landslide to the day, rather than the hour, 594

in which it happened.

Fig. 8 shows the antecedent precipitation timeline for 596 all UIB landslides in the catalogue, partitioned by sea-597 son, scale, and whether a WD/TD was present and 598 within 2000 km. During winter landslides, the 100 km 599 mean precipitation differs little from its climatology in 600 the absence of a WD. When a WD is present, how-601 ever, there is a significant increase in antecedent rain-602 fall, starting at about 0600 UTC on the day before the 603 landslide (i.e. -18 hours). This lends further support to 604 our claim that non-WD winter landslides are caused by 605 very localised events such as cloudbursts, which would 606 have little impact on precipitation over scales of 100 km 607 or so. On the smaller 25-km scale, the WD cases again 608 have higher precipitation rates than non-WD cases, and 609 spread over a longer period, too: a positive anomaly in 610 precipitation extends back three days prior to the land-611 slide when a WD is present, but only two days when it 612 is not. In summary, winter landslides associated with 613 WDs receive more precipitation for longer and on a 614 larger scale; therefore, they are likely to be more se-615 vere, given the positive correlation between antecedent 616 rain rate and landslide fatalities (Froude and Petley, 617

595

618 2018).

During summer landslides (Fig. 8(b)), the effect of a 619 local TD is less clear than it was for WDs in the winter. 620 This is because, as we saw in Fig. 7(b), the presence of 621 a TD does not modulate the synoptic-scale dynamics 622 or synoptic-scale moisture flux over the Indus Basin in 623 the way that a WD does. Even so, there are some in-624 teresting features in the antecedent rainfall timelines. 625 The mean rainfall within 100 km is never significantly 626 higher than the climatology, regardless of whether a 627 TD is present or not, highlighting the role of small-628 scale convective storms in producing the rainfall. The 629 maximum rainfall within 25 km shows a very prominent 630 diurnal cycle in both TD and non-TD cases, peaking at 631 0000 UTC (0530 LT), in agreement with satellite obser-632 vations of the region (Sahany et al., 2010). When a TD 633 is present, the highest rate occurs at 0000 UTC on the 634 preceding day, with a second, smaller peak occurring at 635 0000 UTC on the day of the landslide itself. For cases 636 without a local TD, there is a singular peak, larger than 637 for TDs, at 0000 UTC on the day of the landslide. It 638 is not clear what the source of this difference is, but we 639 hypothesise that a likely cause is that monsoonal TDs 640 provide a steadier (i.e. longer-lived) source of moisture 641 to the UIB than surges in barrier flow. High rainfall 642 the day before the landslide means that less rainfall 643 is then needed to trigger it on the following day. In 644 summary, the majority of summer monsoon landslides 645 in the UIB occur as the result of small-scale storms, 646 whether or not a TD is present. If a TD is present, 647 however, antecedent rainfall is considerably higher the 648 day before the landslide, resulting in favourable condi-649 tions for one to be triggered by lighter rainfall the next 650 day. Comparison of the results from IMDAA, shown in 651 Fig. S3, with Fig. 8 confirms our findings. However, the 652 reader will have noticed that the diurnal cycle of IM-653 DAA rainfall has a much larger amplitude. Errors in 654

this amplitude are common in models and reanalyses, such as IMDAA, that use convective parameterisation (e.g. Dirmeyer et al., 2012).

3.3 Discussion 658

One of the key research questions we asked in the in-659 troduction was whether we can improve landslide pre-660 dictability through an improved understanding of the 661 meteorology that causes episodes of heavy precipita-662 tion in the UIB. We have shown that UIB landslide fre-663 quency is highly sensitive to specific attributes of both 664 nearby WDs and nearby TDs, such as intensity (in the 665 case of TDs), location, and associated large-scale mois-666 ture flux patterns. Combined, these results could be of 667 value to local forecasters looking to assess landslide risk 668 in the UIB in short-range forecasts. This may be fur-669 ther improved when used in conjunction with the south 670 Asian weather regime analysis presented in Neal et al. 671 (2020).672

One shortcoming of this work is the uncertainty introduced by the significant relationship between underlying geology (e.g., lithology, geomorphology and land use) and the relative importance of antecedent rainfall.

We are also able to make some hypotheses about how 677 climate change will affect UIB landslides in the future, 678 notwithstanding additional effects from deforestation or 679 construction. Monsoonal TDs are projected to move 680 poleward and decline in frequency Sandeep et al. (2018). 681 Given that UIB landslides favour TDs further south 682 than usual, both changes would contribute to a decline 683 in summer UIB landslide risk. However, a warmer at-684 mosphere can hold more moisture, and so heavy mon-685 soonal precipitation that does occur over the UIB would 686 likely be heavier still. Thus, the individual precipitation 687 events that typically precede clusters of UIB landslides 688 would likely become rarer but more intense. 689

As discussed in the introduction, the sign of change of 690

future WD frequency remains a topic of debate. How-691 ever, a significant poleward movement of the subtrop-692 ical westerly jet would reduce winter UIB frequency, 693 based on our results. In contrast, previous work (Hunt 694 et al., 2020b) has shown that precipitation associated 695 with individual winter WDs is likely to get significantly 696 more intense in a warmer atmosphere and so, like above, 697 we may find that the storms that cause landslides in the 698 UIB get rarer but more intense. 699

The work presented here opens a number of important questions for future research, which are briefly summarised below.

Precipitation associated with landslides is typically
 heavier when a WD or TD is nearby. Does this result in landslides occurring in locations that are not
 otherwise usually susceptible to slope instability?

• We have shown that WDs far upstream in the westerly jet can be associated with increased UIB landslide activity. They are too distant for this to be a direct effect (i.e. through moisture advection), so does their presence in a cluster result in amplification of nearer, downstream WDs?

• Landslides occurring when neither a WD or TD 713 are present are associated with small-scale precip-714 itation, which we have hypothesised are due to 715 mesoscale convective systems and/or cloudbursts. 716 What conditions increase the likelihood of such sys-717 tems in the UIB? A complete inventory of convec-718 tive system tracks in this region is required to im-719 prove understanding of the conditions behind non-720 WD, non-TD landslides. 721

We speculated on the relationship of monsoon
break conditions and UIB landslides, due to the
secondary maximum of TD occurrence in the south
of the Bay of Bengal, where they are typically
found during monsoon breaks. Monsoon break

conditions often bring anomalous rainfall to northwest India, but does this result in a significant increase in landslide risk in the UIB? 729

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749

4 Concluding remarks

The objective of this study was to determine and un-731 derstand the meteorological precursors to landslides in 732 the Upper Indus Basin (UIB), a region which con-733 tains the orography of both the western Himalaya and 734 the Karakoram. An overwhelming majority of land-735 slides in the UIB are precipitation-triggered, rather 736 than seismic, in nature (e.g. Froude and Petley, 2018), 737 and so we used the NASA Global Landslide Catalogue 738 (GLC; Kirschbaum et al., 2010, 2015), an inventory of 739 precipitation-triggered landslides covering 2007–2015. 740 327 UIB landslides in this period were analysed in 741 combination with track databases of common tropical 742 (tropical depressions; TDs) and extratropical (western 743 disturbances; WDs) synoptic-scale systems over south 744 Asia (Hunt et al., 2018a; Hunt and Fletcher, 2019) to 745 explore the underlying statistical and meteorological re-746 lationships between them. A summary of the main re-747 sults follows. 748

4.1 Winter: October to April

Based on the available landslide data, and recognising 750 the biases in the GLC, UIB landslides occur on about 751 5% of all winter days, peaking in February. WDs are 752 associated, either directly or indirectly, with 61% of 753 these. UIB landslides are significantly more common 754 when a WD, and the subtropical westerly jet in which 755 they are embedded, is situated within $3-4^{\circ}$ of 30° N. 756 WDs further north than 40°N significantly reduce the 757 likelihood of a winter UIB landslide. In contrast to po-758 sition, WD intensity (measured using 350 hPa relative 759 vorticity) does not have a significant relationship with 760

UIB landslide likelihood. Analysis of composite precip-761 itation over the days preceding and during landslides 762 showed that winter landslides associated with WDs re-763 ceive significantly more precipitation and over a longer 764 time period than those not associated with WDs. In ad-765 dition, the spatial scale of precipitation preceding non-766 WD landslides is significantly smaller (< 100 km) than 767 that of WD landslides. In both WD and non-WD land-768 slide cases, moisture flux into the Indus Basin is pre-769 dominantly westerly or southwesterly. In cases where a 770 WD is near, the southwesterly moisture flux is signif-771 icantly enhanced by the associated cyclonic winds, re-772 sulting in moisture penetrating further inland and into 773 the mountain ranges than in non-WD cases. When a 774 WD is not present, westerlies supply significantly less 775 moisture to the UIB, though occasional northward ex-776 cursions are sufficient to support cloudbursts and thun-777 derstorms there. 778

779 4.2 Summer: May to September

UIB landslides occur on about 11% of all summer 780 (i.e. monsoon or late pre-monsoon) days, peaking in 781 August. TDs – used as a collective term for monsoon 782 low-pressure systems, monsoon depressions and other 783 synoptic-scale tropical storms - were associated with 784 60% of these. Landslides are significantly more com-785 mon (rising to about 20% per day) when a TD is located 786 over the centre or towards the northwest of the Indian 787 peninsula, slightly to the south of the climatological 788 position of the monsoon trough. The resulting deep-789 ening and southward adjustment of the trough causes 790 strong, moist, monsoonal southeasterlies to enter the 791 Indus Basin. For non-TD summer landslides, the re-792 quired moisture flux is also provided by southeasterlies, 793 but these take the form of an elongated barrier flow 794 passing parallel to the Himalayan foothills, caused by 795 a northwestward extension of the monsoon trough. In 796

both cases, rainfall preceding the landslide has a strong 797 diurnal cycle peaking at local dawn and is associated 798 with relatively small-scale systems (< 100 km), such as 799 thunderstorms or mesoscale convective systems. TDs 800 are associated with higher antecedent rainfall, bringing 801 the soil closer to field capacity, and meaning that less 802 rainfall is needed to trigger the landslide on the day 803 itself. Unlike WDs, landslide likelihood is sensitive to 804 TD intensity (measured using 850 hPa relative vortic-805 ity): TDs in the 80th percentile of intensity are about 806 50% more likely to be associated with a UIB landslide. 807

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