Received: 14 May 2020 | Accepted: 19 August 2020
DOI: 10.1002/hyp.13895

RESEARCH ARTICLE

WILEY

Assessing the significance of wet-canopy evaporation from forests during extreme rainfall events for flood mitigation in mountainous regions of the United Kingdom

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Funding information
Natural Environment Research Council, Grant/Award Number: NE/R004722/1

Abstract

There is increased interest in the potential of tree planting to help mitigate flooding using *nature-based solutions* or *natural flood management*. However, many publications based upon catchment studies conclude that, as flood magnitude increases, benefit from forest cover declines and is insignificant for extreme flood events. These conclusions conflict with estimates of evaporation loss from forest plot observations of gross rainfall, through fall and stem flow. This study explores data from existing studies to assess the magnitudes of evaporation and attempts to identify the meteorological conditions under which they would be supported. This is achieved using rainfall event data collated from publications and data archives from studies undertaken in temperate environments around the world. The meteorological conditions required to drive the observed evaporation losses are explored theoretically using the Penman–Monteith equation. The results of this theoretical analysis are compared with the prevailing meteorological conditions during large and extreme rainfall events in mountainous regions of the United Kingdom to assess the likely significance of wet canopy evaporation loss. The collated dataset showed that event *Ewc* losses between approximately 2 and 38% of gross rainfall (1.5 to 39.4 mm day^{-1}) have been observed during large rainfall events (up to 118 mm day^{-1}) and that there are few data for extreme events (>150 mm day^{-1}). Event data greater than 150 mm (reported separately) included similarly high percentage evaporation losses. Theoretical estimates of wet-canopy evaporation indicated that, to reproduce the losses towards the high end of these observations, relative humidity and the aerodynamic resistance for vapour transport needed to be lower than approximately 97.5% and 0.5 to 2 s m^{-1} respectively. Surface meteorological data during large and extreme rainfall events in the United Kingdom suggest that conditions favourable for high wet-canopy evaporation are not uncommon and indicate that significant evaporation losses during large and extreme events are possible but not for all events and not at all locations. Thus the disparity with the results from catchment studies remains.

KEYWORDS
complex terrain, extreme events, interception loss, meteorological controls, natural flood management, upland United Kingdom, wet-canopy evaporation

1 | INTRODUCTION

Recently in the United Kingdom, and elsewhere, there is increased interest in *natural flood management* (NFM) or *nature-based solutions* for flood peak mitigation (Dadson et al., 2017; Hankin, Arnott, Whiteman, Burgess-Gamble, & Rose, 2017; Jongman, Winsemius, Fraser, Muis, & Ward, 2018; Lane, 2017; Wingfield, Macdonald, Peters, Spees, & Potter, 2019; World Bank, 2017). Tree planting may be one intervention that has the potential for flood peak reduction through: (a) increased soil infiltration capacity; (b) enhanced soil drying resulting from transpiration; (c) increased ground-surface roughness and (d) enhanced wet-canopy evaporation (*Ewc*). However, some studies suggest that the positive effects of tree cover on flood peaks declines as event magnitude increases, such that it is likely to be insignificant for large and extreme flood events (e.g., Bathurst et al., 2018; Bathurst, Fahey, Iroumé, & Jones, 2020; Dadson et al., 2017; Robinson & Newson, 1986; Stratford et al., 2017). These results suggest, implicitly, that *Ewc* is insignificant during large and extreme events.

For paired grassland and forest catchments on the Plynlimon massif (United Kingdom), Kirby, Newson, and Gillman (1991), p. 60) observed, using flood frequency analysis, that mature conifer cover had little or no effect on the magnitude of peak flows. They showed, using *chronological pairing* of flood peaks, that very small hydrograph peaks were consistently greater from the grassland catchment compared to the forested catchment and that moderately sized event hydrographs showed no significant difference. At another paired forest and grassland study at Coalburn, Northern England, Bathurst et al. (2018) reported that forests can reduce flood peaks for small to moderate events but that hydrograph responses tend to converge at extreme events. Bathurst et al. (2011) explicitly tested the hypothesis: *as the size of the hydrological event increases*, *the effect of forest cover becomes less important*; they concluded, for a number of study sites across Latin America, that forests do not eliminate floods and are unlikely to reduce significantly peak flows generated by extreme rainfall. Bathurst et al. (2011) however acknowledged that their analyses were based on relatively short periods with few extreme events such that conclusive support for the test hypothesis is still lacking.

Recent NFM-related literature reviews of forest effects on flood peaks support the idea of a diminishing effect with event magnitude. Stratford et al. (2017) carried out a systematic review of studies to answer the question: *Do trees in UK-relevant river catchments influence fluvial flood peaks?* Their review focussed directly on the magnitudes of flood peaks rather than on individual hydrological processes and they concluded that the evidence is uncertain for the impact of increasing tree cover on large floods but it is consistent in showing increasing tree cover reduces small floods. Dadson et al. (2017) also reviewed evidence of the effects of forest cover and reported the findings of a number of studies; they recognized that forest management practices complicate determination of forest effects but that under sustained winter rainfall, soil saturation will occur and little mitigation of high flood flows would be expected.

From a process point of view, the benefits of increased infiltration rates and drier antecedent soil moisture conditions are likely to diminish with increasing event magnitude (Calder & Aylward, 2006; Lull & Reinhart, 1972; Pereira, 1989); it is also likely that boundary layer vapour pressure deficits, which exert a strong control on *Ewc*, are likely to decrease during large and extreme rainfall events but the extent to which they decrease across large areas is not well known. The studies cited above did not explicitly included evidence from forest plot studies which estimate *Ewc* in a more direct way using a *canopy water balance* (described below), perhaps because only very few studies report *Ewc* for large or extreme events; they primarily considered the detection of hydrograph change from catchment studies globally.

Worldwide, catchment studies taken as a whole provide conflicting results regarding effects on large flood peaks; compare for example Jones and Grant (1996), Thomas and Megahan (1998) and Beschta, Pyles, Skaugset, & Surfleet, 2000. Many studies have found that the magnitudes or frequencies of large flood peaks are changed significantly by afforestation or forest harvesting (e.g., Alila, Kuraś, Schnorbus, & Hudson, 2009; Belmar, Barquín, Álvarez-Martínez, Peñas, & Del Jesus, 2018; Fahey & Payne, 2017; Guillemette, Plamondon, Prévost, & Lévesque, 2005; Jones & Grant, 1996; López-Moreno, Beguería, & García-Ruiz, 2006); many do, however, show decreasing effects on flood peak as event magnitude increases or no significant change (Beschta et al., 2000; Birkinshaw, Bathurst, & Robinson, 2014; Newson & Calder, 1989; Robinson & Newson, 1986; Thomas & Megahan, 1998; Whitehead & Robinson, 1993). Uncertainties associated with catchment studies of hydrograph change, particularly for extreme events, can be very large (Bathurst et al., 2018, 2020; Beschta et al., 2000; Carrick et al., 2018; Dadson et al., 2017). Underlying signals of change associated with specific processes (e.g., evaporation or infiltration) can also be obscured by effects resulting from forestry practices, such as road construction, drainage or harvesting method (Bathurst et al., 2018; Beschta et al., 2000; Guillemette et al., 2005; Jones & Grant, 1996; Robinson & Newson, 1986; Thomas & Megahan, 1998). These factors combined with limited observations for extreme events (Lewis, Reid, & Thomas, 2010) mean that simple conclusions regarding forest effects on large or extreme flood peaks cannot be made (Andréassian, 2004; Carrick et al., 2018).

1.1 | Forest plot studies of wet-canopy evaporation: Losses during large and extreme rainfall events

Forest plot studies estimate *Ewc*, using a canopy water balance (CWB), as the difference between the gross rainfall (*Pg*) incident upon a vegetation canopy and the fraction of *Pg* that reaches the ground as *net rainfall* (*Pn*). Net rainfall comprises rainfall that bypasses or drips from the canopy (*throughfall*: *TF*) and that which flows via stems and trunks (*stem flow*: *SF*). As noted above, very few studies have focused on CWB estimated *Ewc* during large (>50 mm day^{-1} of *Pg*) or extreme (>150 mm day^{-1} of *Pg*; Collier, Fox, & Hand, 2002) rainfall events. A notable exception is the work of Keim, Skaugset, and Link, and Iroumé (2004) who report Ewc losses above 30% of *Pg* at temperate sites in Chile and Northwest USA. Equally high *Ewc* losses during large magnitude rainfall events at other locations with a temperate climate have been reported (e.g., see Deguchi, Hattori, & Park, 2006; Hashino, Yao, & Yoshida, 2002). Taken at 'face value', and recognizing that these data are also subject to uncertainties, these *Ewc* losses appear to be potentially significant in the context of flooding: removal of such large fractions of event rainfall from a catchment system are likely to have a significant effect on a flood hydrograph where tree planting covers a large proportion of a catchment (Hankin et al., 2017). Consequently, there is an apparent disparity between the publications which conclude that forest effects on flood peaks are likely to be small or insignificant for large and extreme events and the CWB observations from forest plot studies.

The significance of forest *Ewc* for flood mitigation depends upon the difference in *Ewc* between a given forest canopy and another land cover. However, as most comparative studies derive estimates from catchment or lysimeter water balances over relatively long periods (e.g., Calder, 1976, Calder & Newson, 1979; Calder, 1981), and do not separate *Ewc* and transpiration losses, these estimates are of limited use when considering individual events. Additionally, evaporation from a forest understorey or soil litter layer can be significant (Bulcock & Jewitt, 2012; Carlisle, Brown, & White, 1967; Gerrits, Pfister, & Savenije, 2010; Gerrits, Savenije, Hoffmann, & Pfister, 2007) which offsets (to an unknown degree) the difference between forest and a comparative land cover. We therefore assume that *Ewc* losses from tall canopies are likely to be significantly higher than for short vegetation under meteorological conditions favourable for wet-canopy evaporation and, consequently, that the absolute magnitude of *Ewc* from tall forest canopies is of primary relevance here. Thus it is important to determine the full extent of evidence for significant *Ewc* from forest canopies during large and extreme rainfall events as well as an understanding of the meteorological conditions under which significant losses might be supported.

1.2 | Aim and objectives of this study

This study focuses on *Ewc* from forest canopies during large and extreme rainfall events. We use the Cumbrian Mountains, UK, as a focus for some of our analyses as large catchment-scale hydrological simulations of broad-scale tree planting are required. These simulations are designed to inform UK policy on the most effective methods of NFM. To inform these simulations, pertinent event-scale *Ewc* data from temperate sites around the world are collated and are contextualized using UK meteorological conditions via theoretical analyses. In the United Kingdom, large and extreme rainfall events primarily occur as long-duration autumn and winter storms when forest canopies can be continually wet, solar radiation is low and Ewc will dominate evaporative losses (Calder, 1990); these events are the focus of this study rather than extreme summer convective storms.

Our specific objectives are:

> **Objective 1**. To collate all available *Ewc* data from UK-relevant event-scale forest plot studies to quantify the magnitude and range of observed losses, particularly during large (>50 mm day^{-1}) and extreme (>150 mm day^{-1}) rainfall events.
>
> **Objective 2**. To explore the meteorological conditions consistent with the magnitudes of *Ewc* losses from Objective 1 using the Penman–Monteith equation.
>
> **Objective 3**. To examine meteorological data from Cumbria and other mountainous regions of the United Kingdom during large and extreme rainfall events and to compare with the findings of Objective 2.
>
> **Objective 4**. To discuss the implications of the findings from objectives 1 to 3 for estimating *Ewc* across large catchments in complex terrain.

2 | METHODS

2.1 | Study area

The primary area of interest for this study, Cumbria, Northwest England (Figure 1) where the Q-NFM project (http://www.lancaster.ac.uk/lec/sites/qnfm) is tasked with simulating the effects of broad-scale tree planting scenarios on flood hydrographs using a catchment hydrological model. The study catchments are within a mountainous area where there have been four major flood events in the last 15 years. Cumbria, as for many mountainous regions of the United Kingdom, is situated towards the west coast and hence strongly influenced by temperate maritime airflows from the Atlantic Ocean. Mountainous regions of the United Kingdom are areas of extremely complex topography which, in combination with the predominant airflow, gives rise to orographically influenced and spatially heterogeneous meteorological patterns (Blackie & Simpson, 1993; Ferranti, Whyatt, & Timmis, 2009; Mayes, 2013). Annual rainfall is generally high; for example, across Cumbria the long-term annual average rainfall ranges from below 1,000 mm year^{-1} on the coast and in areas of rain shadow to greater than 3,500 mm year^{-1} across the highest mountains. Given there are no CWB *Ewc* data for this study area we consider other

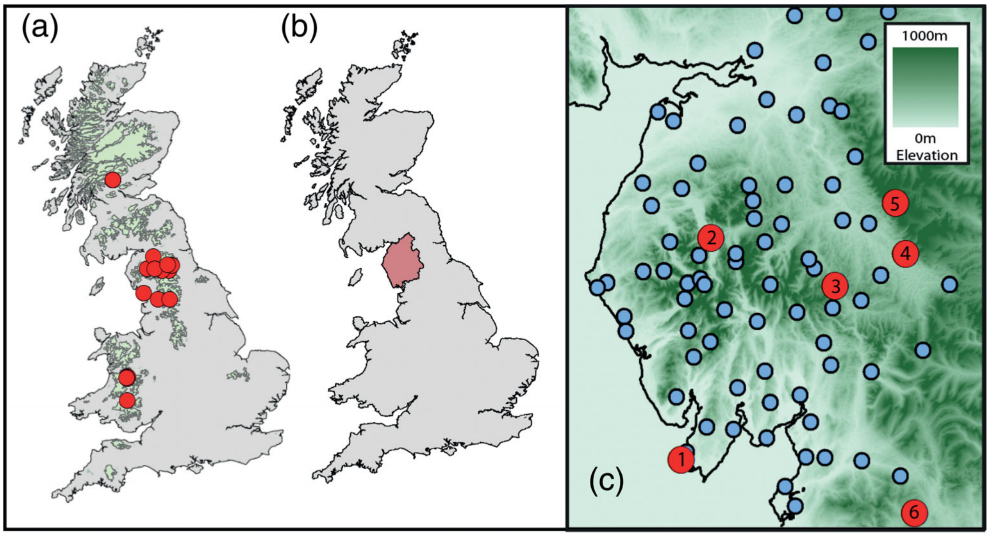

FIGURE 1 (a) UK map showing elevations of over 300 m as green shaded areas and meteorological data sites as red filled-circles. (b) UK map showing location of Cumbria. (c) Elevation map of Cumbria and Northern Lancashire showing Environment Agency rain gauges as blue filled-circles and primary meteorological sites as red filled-circles; numbers refer to the sites discussed in detail below: (1) Walney, (2) Keswick, (3) Shap, (4) Warcop, (5) Gt. Dun Fell and (6) Gisburn. Elevation data © Crown Copyright OS Panorama 2011

mountainous sites in the United Kingdom (Figure 1) and other temperate locations around the world.

2.2 | Observed event *Ewc* estimates from forest plot studies in temperate locations

Studies relevant to UK conditions with forest plot CWB *Ewc* observations were identified. Relevant studies were defined climatologically using the revised Köppen climate classification (Chen & Chen, 2013) with the recognition that the classification provides general climatic classes within which there is a large degree of variability. Although this variability exists, Köppen climate classifications were used with the rationale that hydrometerological conditions will have similarities to those of UK mountainous regions during large and extreme rainfall events. The specific classifications deemed acceptable for the purposes of this study were: Cfa (mild temperate; fully humid; hot summer), Cfb (mild temperate; fully humid; warm summer), Cfc (mild temperate; fully humid; cool summer), Csb (mild temperate; dry summer; warm summer), Dfb (snow; fully humid; warm summer), Dfc (snow; fully humid; cool summer), Dfd (snow; fully humid; cold summer): the classifications Cfb and Cfc encompass all mountainous regions of the United Kingdom. For the climate classes where snow can form a significant part of winter precipitation (i.e., Dfb, Dfc and Dfd), care was taken not to include data affected by snow falls: in most cases this was already carried out in the original study.

From the studies identified as relevant (Table S1), only CWB *Ewc* observations reported on a rainfall event basis or as a daily total were collated for analyses. For consistency, events reported to be over 24 hr (but no more than 48 hr) duration were standardized by calculating a *normalized 24-hr rainfall* (i.e., $Pg = 24.\left(\frac{Pe}{De}\right)$) where *Pe* is the total event precipitation and *De* is event duration in hours): note that, in the text below, *Pg* relates to a daily or normalized daily total rainfall unless otherwise stated. It is accepted that using daily observations is somewhat artificial in as much as the duration of rainfall events may be truncated where they span multiple days; it is also recognized that both daily and event data may include periods without rain. Event-based data where event duration was not reported were also collated and are presented separately.

Data were gathered using values provided in tables or by digitizing data presented as figures in published material and by abstracting data from field log sheets; in the case of Aussenac (1968) and Reynolds and Henderson (1967) *Pg-Ewc* relationships were digitized. Where observations were obtained from digitized figures, and where there were many data points, obscured or overlapping data will have resulted in some values not being included; these data were, however, invariably for relatively low *Pg* magnitudes. In some cases, the dates associated with individual rainfall events were not provided in the published material, which does not allow separation by season (or by leafed or leafless period for deciduous species). Furthermore, as event data were limited and data for deciduous forest plots were very few, all events irrespective whether they were evergreen or deciduous were combined for the comparison with theoretical estimates, but are identified separately in figures. The following additional published studies provided data for the analysis: Andre, Jonard, Jonard, & Ponette (2011), Calder (1985), Calder, Smyle, & Aylward (2007), Chappell (2018a), Chappell (2018b), Cisneros Vaca, van der Tol, & Ghimire (2018), Crabtree, & Trudgill (1985), Crockford, & Richardson (1990), Crockford, & Richardson (2000), Dunin, Oloughlin, & Reyenga (1988), Eden, & Burt (2010), Gavazzi, et al. (2016), Giacomin, & Trucchi, (1992), Hankin, et al. (2017), Kelliher, Whitehead, & Pollock (1992), Klaassen, Lankreijer, & Veen (1996), Lankreijer, et al. (1999), Law, (1956), Link, Unsworth, & Marks (2004), Loustau, Berbigier, & Granier, (1992), Lu, Sun, McNulty, & Amatya, (2005), Massman, (1983), Pook, Moore, & Hall, (1991), Price, & Carlyle-Moses (2003), Robins, (1969), Saito, et al. (2013), Staelens, De Schrijver, Verheyen, & Verhoest (2008), and Toba, & Ohta (2005).

2.3 | Theoretical *Ewc* estimates

2.3.1 | Exploring the drivers of observed *Ewc* losses using the Penman–Monteith equation

To explore the meteorological conditions required for consistency with observed *Ewc* losses, a broad range of Ewc estimates were made using the Penman–Monteith equation (Monteith, 1965;Equation (1)); estimates were made using stratified samples from ranges of relative humidity (85 to 100%; expressed via vapour pressure in Equation (1) where the temperature at the observation height is assumed to be 10°C) and aerodynamic resistances (0.5 to 12 s m^{-1}). This analysis creates a response surface where different combinations of meteorological variables that lead to similar *Ewc* losses can easily be visualized. The Penman–Monteith equation takes the form:

$$\lambda E_{PM} = \frac{\Delta_e H + \rho_a c_p \left(e_s(Tz) - e_z\right)/r_{a\_}s}{\Delta_e + \gamma\left(1 + r_g/r_{a\_}s\right)}, \tag{1}$$

where λ is the latent heat of vaporization (J kg^{-1}), E_{PM} is the evapotranspiration rate (kg m^{-2} s^{-1}), $\Delta_e$is the slope of the saturation vapour pressure curve versus temperature relationship (Pa K^{-1}) at temperature, *Tz*, where *z* is the observation height in metres, *H* is the total energy available for evaporation (J m^{-2}), ρ_a is the density of air at *Tz* (kg m^{-3}), c_p is the specific heat capacity of air (J kg^{-1} K^{-1}), $e_s(Tz)$ is the saturation vapour pressure at *Tz*, e_z is the actual vapour pressure at *z*, γ is the psychrometric constant (≈66 x 10^{-3} Pa K^{-1}) and $r_{a\_}s$ and r_g are the resistance to the aerodynamic exchange for scalars (sensible heat and vapour) and surface resistances respectively (s m^{-1}). Note that in all calculations made here, the canopy is assumed to be wet and r_g is assumed to be zero such that the term (1 + $r_g/r_{a\_}s$) disappears (Stewart, 1977; van Dijk et al., 2015). The total energy available (*H*) was assumed to be the approximate net radiation (R_n) for a cloudy day during winter in Northern England: nominally 2.5 MJ m^{-2} d^{-1}.

2.3.2 | Estimates of aerodynamic exchange and E_{PM} using meteorological observations

To explore the potential for *Ewc* across mountainous regions of the United Kingdom during large and extreme rainfall events, E_{PM} was calculated using meteorological data for 17 sites (Figure 1a; Table S2; Met Office, 2006) using Equation (1). The aerodynamic resistance for momentum ($r_{a\_}m$) was estimated using Equation (2):

$$r_{a\_}m = \frac{\ln\left(\frac{z-d}{z0\_m}\right)^2}{\kappa^2 U_z}, \tag{2}$$

where *z* is the wind speed observation height, *d* is the zero-plane displacement, *z0_m* is the roughness length for momentum (all in metres), U_z is the wind speed (m s^{-1}) at *z* and κ is the dimensionless von Karman constant (≈0.41). The canopy height (*Zc*) was arbitrarily assumed to be 20 m, *d* to be 0.75(*Zc*) and *z0_m* as 0.1(*Zc*) in accordance with Szeicz, Endrödi, and Tajchman (1969) and Rutter, Robins, Morton, and Kershaw (1972). However, studies have shown an enhancement of exchange compared to estimates assuming these approximations for *z0_m* (e.g., Holwerda, Bruijnzeel, Scatena, Vugts, & Meesters, 2012). Enhancement of momentum exchange has been observed both for tall canopies and in complex terrain owing to breakdown of theoretical vertical logarithmic wind profiles (Cellier & Brunet, 1992; Raupach, 1979; Simpson, Thurtell, Neumann, Den Hartog, & Edwards, 1998). For the indicative calculations made here, where z < *Zc*, wind speed was extrapolated to *Zc* using a logarithmic wind profile relationship. Wind speed observations used here are taken over short grass surfaces and extrapolated to hypothetical canopy height as if the logarithmic profile assumption is valid. It is recognized that this may not be the case in complex terrain but, as the degree of enhancement of momentum exchange is not easily estimated and because the calculations made here are purely indicative no enhancements have been made for $r_{a\_}m$.

It is often assumed that $r_{a\_}s$ is equal to $r_{a\_}m$, but this assumption can to lead to considerable error (Brutsaert, 1982, p. 62) owing to so-called *excess resistance* for scalars. Excess resistance occurs because pressure forces associated with *form drag* increase momentum exchange, but not scalar exchange and because of differences in source and sink distributions for these entities (Brutsaert, 1982; Moors, 2012; Simpson et al., 1998; Stewart & Thom, 1973). Although there can be differences between the magnitude of exchange for different scalars, we assume that the exchange of heat and vapour are equal for the purposes of this study and hence only explore differences between the magnitude of scalar exchange compared to the exchange of momentum. Aerodynamic exchange estimated using Equation (2) is more sensitive to the value of *z*0 than it is to the value of *d* (Gash, Wright, & Lloyd, 1980). The value of *z*0 has been shown to vary significantly with wind speed for forest canopies, whilst *d* tends to remain relatively constant (Bosveld, 1999; Szeicz et al., 1969). Consequently, *d* is fixed as specified above for all calculations made here and it is assumed that the primary differences between $r_{a\_}s$ and $r_{a\_}m$ are driven by differences in *z0_m* and *z0_s*. It is worth noting that *d* may vary significantly for very sparse canopies or for deciduous canopies during the leafless period (Brutsaert, 1982, p. 116; Dolman, 1986).

The ratio of *z0_s*/*z0_m* used in previous studies varies over approximately an order of magnitude as it is influenced by canopy roughness, canopy density, atmospheric stability and wind speed (Bosveld, 1999; Brutsaert, 1982, p. 114; Lalic, Mihailovic, Rajkovic, Arsenic, & Radlovic, 2003; Raupach, 1979; Thom, Stewart, Oliver, & Gash, 1975). The sensitivity of $r_{a\_}s$ and E_{PM} to the ratio *z0_s*/*z0_m* is explored here using three scenarios:

Scenario 1 - $\frac{z0_s}{z0_m} = 1.0$; that is, $z0_s = 0.1(Zc)$;

Scenario 2 - $\frac{z0_s}{z0_m} = 0.5$; that is, $z0_s = 0.05(Zc)$;

Scenario 3 - $\frac{z0_s}{z0_m} = 0.1$; that is, $z0_s = 0.01(Zc)$.

It is likely that vapour pressure deficit (also expressed as relative humidity, *RH*) observations over grassland meteorological observation sites are likely to be lower than those over an adjacent forested area (e.g., see Pearce, Gash, & Stewart, 1980). No attempt has been made to correct *RH* observations for this study owing to the complexities associated with such a correction and the indicative nature of our calculations; this is also the case for *Tz* which is likely to be lower above a forest canopy (Rutter, 1967). Additionally, Equation (2) is strictly only valid for neutral atmospheric conditions (Szeicz et al., 1969) but corrections for non-neutral conditions are often assumed to be insignificant during rainfall (e.g., Morton, 1984; van Dijk et al., 2015) and we assume they are negligible here.

3 | RESULTS AND DISCUSSION

3.1 | Objective 1: Canopy water balance observations of wet-canopy evaporation during large and extreme events

The data search provided observations from 18 study sites that have CWB observations of *Ewc* for large storms associated with either daily or normalized event *Pg* data (see Table S3): none of the events were extreme events based upon the classification employed (>150 mm day^{-1}). From these sites 1,387 *Pg*-*Ewc* pairs were obtained with a maximum *Pg* of approximately 118 mm. Only 35 of these pairs were associated with *Pg* observations over 50 mm day^{-1} (Table S3). In absolute terms, the *Ewc* data include some high magnitudes; they include a maximum *Ewc* loss of 39.4 mm day^{-1} with 40 events where *Ewc* loss was over 10 mm day^{-1}. When *Ewc* is expressed as a percentage of *Pg* (%*Ewc*), there is a decreasing trend in %*Ewc* as *Pg* increases (Figure 2). This pattern of declining relative loss has been shown previously in many studies (e.g., Bulcock & Jewitt, 2012; Iroumé & Huber, 2002), including the seminal review by Horton and E. (1919). From these 1,387 data pairs, 58 gave a negative %*Ewc* and 25 were over 100%*Ewc*; these data lie predominantly at low *Pg* magnitudes and are not presented in Figure 2 and can be caused by error, sampling truncation of events and fog-drip and will be incorporated into uncertainty analysis in future work. The subset of the data from UK catchments are overlain by green filled-circles in Figure 2; the individual numbered events where *Pg* > 50 mm are specified in Table S3. These UK data tend to span the higher rates of %*Ewc* for higher *Pg* magnitudes and include absolute losses up to 26.3 mm day^{-1}. The few data which are associated with plots under entirely deciduous species are highlighted as orange-filled circles in Figure 2: only one data pair was associated with *Pg* greater than 50 mm day^{-1} of which approximately 11% was lost to *Ewc*. The degree to which %*Ewc* continues to decrease with increasing *Pg*, or whether it has reached a stable range is unclear from Figure 2 given the few data available at higher *Pg* magnitudes. This is important as the highest *Pg* from these data is significantly lower than extreme daily rainfall totals recorded in the Cumbrian Mountains which has been observed to be as high as 341 mm in a 24-hr period at Honister Pass, Cumbria (Met. Office, 2018) which led to widespread severe flooding.

Observations associated with studies where no event duration data were available provided 1,144 *Pg*-*Ewc* pairs including some high %*Ewc* losses (ranging from approximately 7 to 35%) for very large *Pg* values (up to 435 mm: see Table S2). These data are plotted as black filled-circles in Figure 3 with plots under entirely deciduous species highlighted as orange-filled circles. Although these data are difficult to compare to the daily or normalized daily data, the high losses observed are significant given that these are potentially extreme events, likely to be of maximum 3 or 4 days in duration and hence serve as a useful reference. An exception to this rule are the data from Deguchi et al. (2006) where it is possible that observations were made over a period of up to 2 weeks; however, the largest event from their study (which was identified as taking place on September 11-12th, 2000) deposited 347 mm *Pg*, with 14% of this being lost to *Ewc*. The magnitude of the losses from these non-normalized events are qualitatively consistent with the normalized events presented in Figure 2.

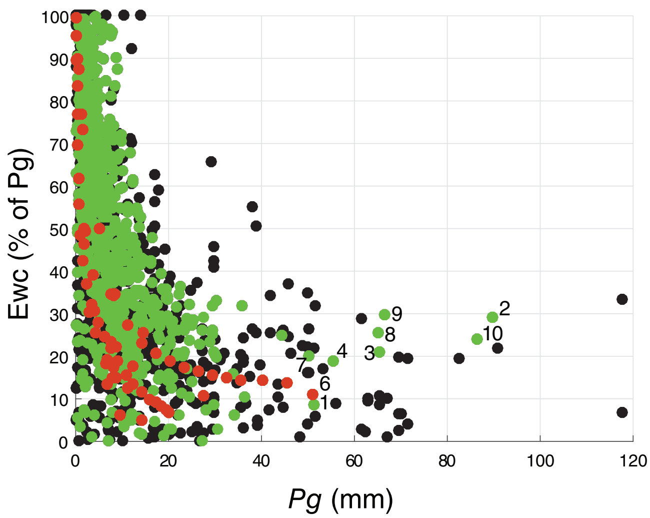

FIGURE 2 Observed *Ewc* versus *Pg* from studies with daily or normalized daily event data with the subset of UK studies shown as green filled-circles: numbers next to circles relate to the numbered events in Table S3. The orange filled-circles show observations within deciduous forests

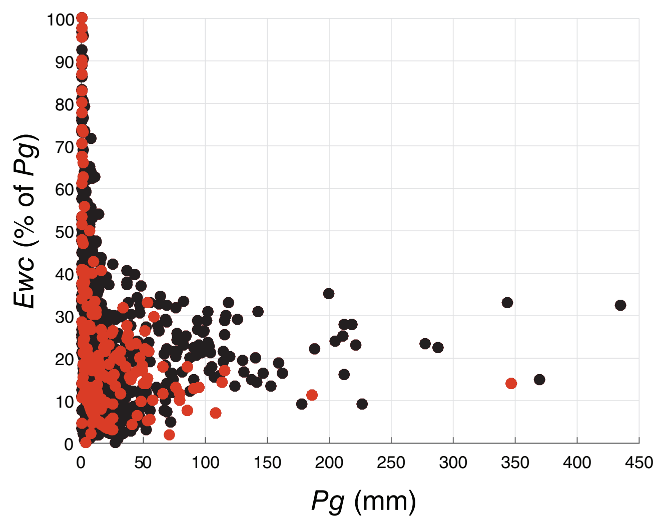

FIGURE 3 Observed *Ewc* versus *Pg* from studies data pairs for 'non-normalized' event studies in Table S1; the orange filled-circles show observations within deciduous forests

The data for large and extreme events presented in Figures 2 and 3 show a large range of *Ewc* loss: approximately 2–38% of *Pg*. These losses are apparently significant in the context of flood mitigation with absolute losses of up to approximately 40 mm day^{-1}. Unfortunately, concurrent meteorological observations were generally not reported for events greater than 50 mm day^{-1}: concurrent observations were only available for 4 events at one site (Dolydd, mid-Wales; events 7–10; Table S3). This lack of meteorological data means that it is, in general, not possible to link the observed *Ewc* losses with the magnitude of important meteorological variables which would allow some form of model calibration. Model calibration of this kind is problematic and also needs to include canopy storage limitation of *Ewc* (e.g., see Calder, 1977).

3.2 | Objective 2: Estimation of meteorological conditions consistent with observed *Ewc* losses

3.2.1 | The likely magnitude of meteorological controls driving observed *Ewc* losses

Penman–Monteith potential evaporation estimates, E_{PM}, were made across the ranges described in section 2.3 above. A representation of how E_{PM} varies with $r_{a\_}s$ and saturation vapour pressure deficit (expressed as *RH*) is shown in Figure 4. With respect to Objective 2, given the Penman–Monteith equation and the assumptions of the analysis, to achieve the higher end of absolute *Ewc* losses observed (≈ 20 to 40 mm day^{-1}) either fairly low *RH* or very low $r_{a\_}s$ values are required, or an equivalent combination of *RH* and $r_{a\_}s$. Relative humidity needs to be below approximately 90% where $r_{a\_}s$ is around 2 s m^{-1} or around 97.5% as $r_{a\_}s$ approaches 0.5 s m^{-1}. Figure 4 also highlights how E_{PM}becomes increasingly sensitive to small changes in $r_{a\_}s$ at lower values: i.e., for higher wind speeds and rougher canopies (as previously shown by Beven, 1979 & Dolman, 1986). At these low $r_{a\_}s$ values, E_{PM} is also considerably more sensitive to changes in *RH*, and even at relatively high *RH*, the potential for significant evaporation loss exists. Owing to this extreme sensitivity at low $r_{a\_}s$ values, uncertainties associated with estimating *effective* $r_{a\_}s$ values and *RH* become critical in the interpretation of the results presented here and are discussed in more detail below.

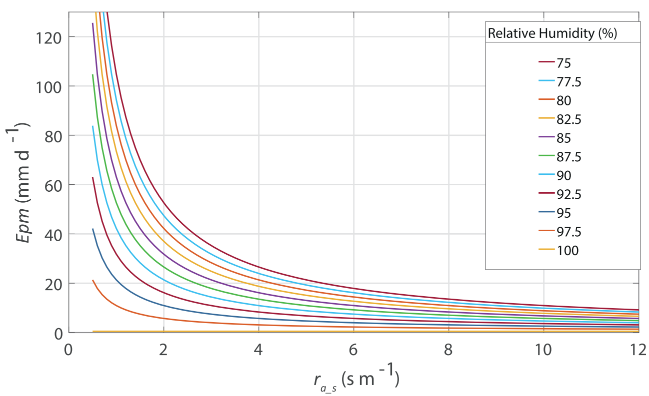

FIGURE 4 A psuedo-three-dimensional response surface of E_{PM} estimates for stratified samples of aerodynamic resistance and relative humidity

3.3 | Objective 3: Meteorological conditions and wet-canopy evaporation estimates for mountainous regions of the United Kingdom

3.3.1 | Penman–Monteith wet-canopy evaporation estimates for mountainous regions of the United Kingdom

Estimates of E_{PM} made using the meteorological data for the 17 sites specified in section 2.3.2 show that, given the assumptions of our analysis, within-storm conditions for potentially high *Ewc* loss are possible in mountainous regions of the United Kingdom. High wind speeds and relatively low *RH* can prevail during days with significant rainfall. This is illustrated in Figure 5a where hourly average $r_{a\_}s$ versus *RH* data are plotted for the sites identified in Table S2. The points plotted in Figure 5a relate to hourly periods within a 24-hr period with over 50 mm of rainfall ***and*** where the hourly rainfall total was above

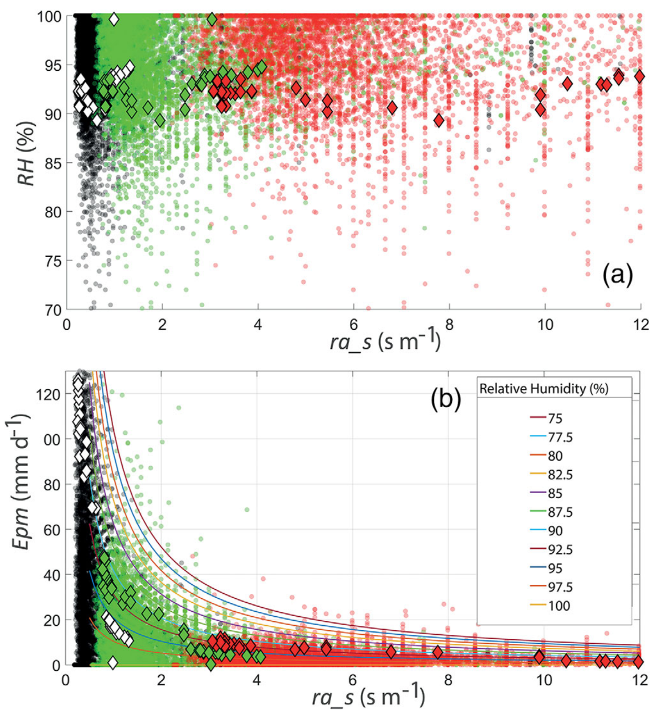

FIGURE 5 (a) Hourly average values of relative humidity versus aerodynamic resistance calculated from the datasets listed in Table S2. Calculations are for timesteps associated with a 24 hr period with over 50 mm of rainfall and where hourly rainfall was above zero. The different colours represent *ra_s* calculated for three different values of *z0_s*: 0.1*z* black filled-circles, 0.05*z* green filled-circles and 0.01*z* red filled-circles; (b) The hourly timesteps presented in (a) with Penman–Monteith estimated *Ewc* versus *ra_s* superimposed upon the response surface of Figure 4; diamond symbols identify timesteps associated with 24 hr periods with over 150 mm of rainfall

zero. The estimates of r_a_s, using the three scenarios for $z0\_s$ as described above, are represented by: black filled-circles for $z0\_s$= 0.1 (Zc), green filled-circles for $z0\_s$= 0.05(Zc) and red filled-circles for $z0\_s$= 0.01(Zc). Figure 5a demonstrates that very low r_a_s values can occur within 24-hour periods where Pg is greater than 50 mm and that the majority of these periods were associated with RH values predominantly in the range 85 to 98% which shows significant overlap with the conditions required for significant E_{PM} estimated for Objective 2. Meteorological conditions during more extreme events (>150 mm in 24-hr and where the hourly rainfall >0), also shown in Figure 5a as diamonds; this figure suggests that, particularly for RH, conditions can be even more favourable for high E_{PM} but are associated with the caveat that there are relatively few observations during very few events of this magnitude. The potential for high E_{PM} is shown more explicitly in Figure 5b (which uses the same data as Figure 5a). The difference between the estimates made using the 3 $z0\_s$ highlights again how sensitive E_{PM} magnitude is to $z0\_s$. However, fairly high rates of E_{PM} are estimated for all $z0\_s$ scenarios

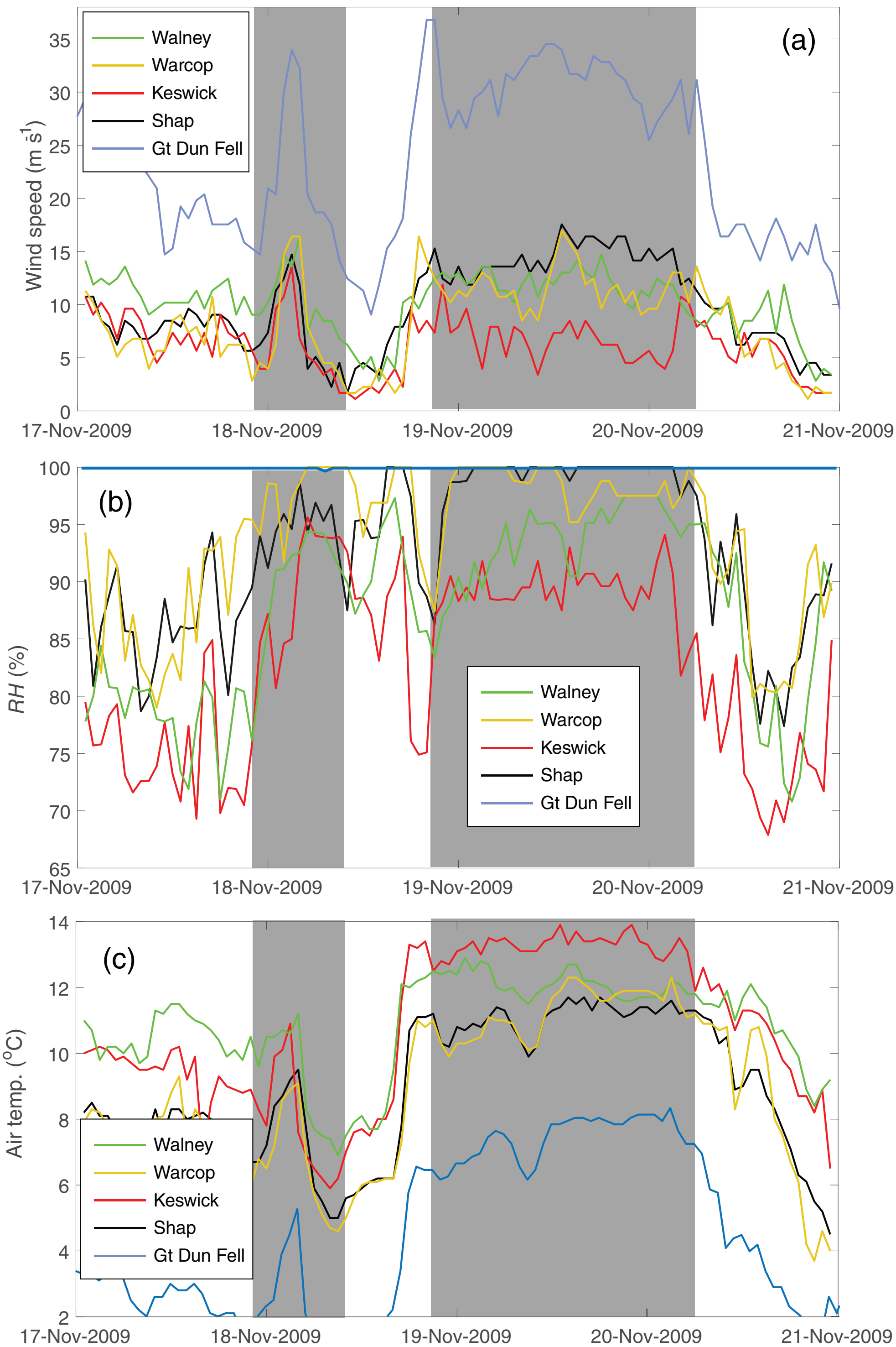

FIGURE 6 Meteorological time series (hourly averages of 1-min frequency observations) during the extreme event that led to the Cockermouth floods of 2009 for (a) wind speed, (b) relative humidity and (c) air temperature. The grey shaded areas indicates the main periods of rainfall

although the $z0\_s$ = 0.01(Zc) scenario is mainly limited to losses of below 12 mm day^{-1}.

3.3.2 | Meteorological conditions during extreme events across Cumbria

Figures 6 and 7 (Figures S1 and S2) show time series of meteorological variables during the 4 extreme Cumbrian rainfall events that occurred since 2005. Each event was associated with a frontal system, the 2005, 2009 and 2015 events being classified as atmospheric rivers, where enhanced horizontal water vapour transport from the Atlantic Ocean occurs (Lavers et al., 2011, 2013; Matthews, Murphy, McCarthy, Broderick, & Wilby, 2018). The meteorological time series show that, at many of the locations, during the main periods of rainfall (indicated by the grey shaded area in Figures 6b and 7b; Supporting Information Figures S1b and S2b) relatively low RH (in the context of the results presented above) occurs along with high wind speeds

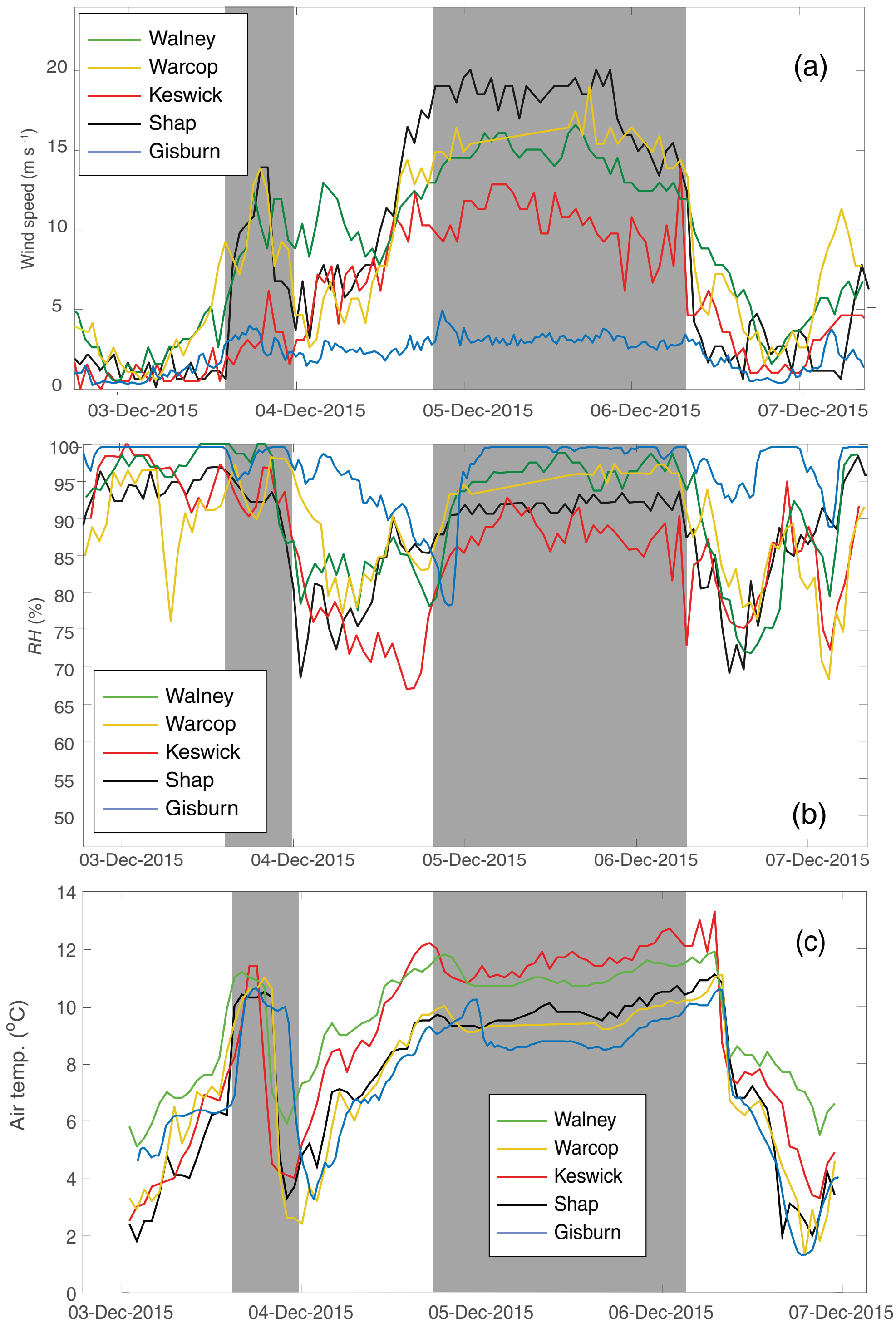

FIGURE 7 Meteorological time series (hourly averages of 1-min frequency observations) during the extreme event Storm Desmond that led to the severe flooding of 2,105 for (a) wind speed, (b) relative humidity and (c) air temperature. The grey shaded areas indicates the main periods of rainfall

(Figures 6a and 7a; Supporting Information Figures S1a and S2a) and an increase in air temperature (Figures 6 and 7c; Supporting Information Figures S1a and S2a). A notable exception to these general patterns are the differences in *RH* between sites for the 2009 event where Keswick and Walney were at approximately 90–95% *RH* during the main rainfall period but the Shap and Walney sites were at, or close to, saturation. This was also the case at the higher elevation site of Great Dun Fell where the wind speeds were very high but the *RH* remained at 100% throughout the entire period of rainfall (Figure 7a, b, respectively). Similarly, but for the 2015 flood event, in a clearing in Gisburn Forest, Lancashire the *RH* remained very high or at saturation for a large part of the storm (Figure 7b). There is also consistency in that Keswick, which is a less exposed site at relatively low elevation on the lee side of one of the highest regions of mountains, tends to have lower wind speeds, higher temperatures and lower *RH* than the other locations during all extreme events considered (Figures 6 and 7; Supporting Information Figures S1 and S2). It is worth highlighting the caveat that the Keswick, Shap and Warcop sites are located on the leeward side (relative to the dominant south westerly flows) of mountain ridges such that the observed favourable conditions for *Ewc* loss shown are not necessarily representative across the region as a whole: as shown by the less favourable conditions for *Ewc* loss at Gisburn Forest and Great Dun Fell. These results show that even during the 4 most extreme events in Cumbria over the last 15 years, high windspeeds and surprisingly low *RH* provide favourable conditions for significant *Ewc* loss at some locations. Consequently the need to estimate *Ewc* losses across large catchments, particularly in complex mountainous terrain, requires a representation of the spatial variability of meteorological controls which will be challenging where only sparse meteorological observations are available.

3.4 | Objective 4: Implications for estimating the magnitude of *Ewc* across large catchments in complex terrain

The results from objectives 1 to 3 show that *Ewc* losses up to approximately 40 mm day^{-1} have been observed at temperate sites around the world and that meteorological conditions that have the potential to give rise to such large losses can exist in mountainous regions of the United Kingdom. However, these findings must be treated with caution because concurrent meteorological observations are rarely reported with CWB *Ewc* data, particularly during extreme events, and Penman–Monteith estimates are extremely sensitive to *estimated* aerodynamic exchange and small changes in *RH* at the higher windspeeds that often prevail during the large rainfall events considered here. The analysis has also shown that both wind speed and *RH* varies significantly with spatial location. Given that Ewc estimates are required for hydrological simulation of large catchments, a representation of this spatially variable control of *Ewc* magnitude is required; it is not appropriate to sample a statistical distribution of Ewc loss generated from the worldwide observations of *Ewc* data (for a given gross rainfall total) as the autocorrelation of *Ewc*, controlled by autocorrelated meteorological variables, through sequences of <u>real</u> events is needed. With respect to this requirement, even a spatially sparse time series of meteorological observations contains important information describing temporal patterns of some of the primary controls on *Ewc* and this information must be retained. Spatial interpolation and extrapolation from these sparse meteorological observations will inevitably be inherently uncertain but is an important prerequisite for appropriate estimation of *Ewc* losses.

Although simple empirical models can be used to estimate *Ewc* where there is a scarcity of adequate meteorological data and knowledge of appropriate parameter values for more complex models (e.g., see Lu, Sun, McNulty & Amatya, 2005), their use is limited as they may not explicitly include important meteorological controls. Consequently, the Penman–Monteith equation is still used to simulate evaporation from wetted surfaces in the majority of *Ewc* models (Muzylo et al., 2009). Thus, Penman–Monteith equation remains a useful method to determine the potential for *Ewc* loss but the magnitude of any estimates made will be highly uncertain without meaningful calibration of critical and sensitive parameters such as r_a_s. However, as there are so few *Ewc* data associated with concurrent meteorological observations, particularly large rainfall events, it is rarely possible to calibrate the parameters of the Penman–Monteith equation and any calibration would need to include the joint-calibration of parameters of an (e.g., Rutter-type) *effective* canopy store model (e.g., see Calder, 1977).

Our theoretical analyses show that it is possible to get a very wide range of *Ewc* estimates depending upon, in particular, the way that r_a_s is estimated. These analyses used 3 scenarios of r_a_s which were based upon a range of published values derived both directly from micrometeorological observations and via model calibration. Ratios of $z0\_s/z0\_m$ have been reported to be: of the order 0.1–0.2 (Klingaman, Levia, & Frost, 2007; Lankreijer, Hendriks, & Klaassen, 1993); approximately 0.3–0.5 (Brutsaert, 1982, p. 114; Stewart & Thom, 1973) and around 1 in some cases (Bosveld, 1999; Gash, Valente, & David, 1999; Moors, 2012). Significant uncertainties exist when estimating r_a_m and the relative magnitude of r_a_s compared to r_a_m. When only momentum is considered, representing the degree of exchange is not simple as it has been shown to vary, and to be enhanced compared to theoretical estimates, in complex terrain and over tall canopies (Cellier & Brunet, 1992; Holwerda et al., 2012); r_a_m also varies with canopy roughness and canopy density (Brutsaert, 1982, figure 5.1; Cellier & Brunet, 1992; Holwerda et al., 2012) as well as atmospheric stability and wind speed (Bosveld, 1999; Cellier & Brunet, 1992; Szeicz et al., 1969). The ratio $z0\_s/z0\_m$ also varies widely and with the same factors as r_a_m and current understanding of scalar exchange for tall canopies in complex terrain remains rudimentary (Belcher, Harman, & Finnigan, 2012). There are, however, a relatively large number of published studies which report r_a_m and r_a_s for various vegetation of differing roughness which may help elucidate the relevant range of r_a_s for use in *Ewc* estimation for a given application: a review of these studies is, however, beyond the scope of this paper.

Given the need for interpolation and extrapolation from sparse meteorological data to estimate meteorological controls on *Ewc* spatially, uncertainties will be very large such that a scenario-based approach may be most appropriate. Any defined scenario will be *conditional* on the evidence base used in its development and any additional modelling assumptions. The conditionality of each scenario must be made explicit and each scenario can be associated with a confidence-weighting which can be propagated to simulation results. This will be the subject of future publications.

4 | CONCLUSIONS

At temperate locations around the world, high wet-canopy evaporation losses have been observed from forests using canopy water balance methods during large and extreme rainfall events and are associated with significant variability. Wet-canopy evaporation of up to approximately 40 mm day^{-1} have been recorded for large rainfall events (>50 mm day^{-1}) and across all events range between approximately 2 and 38% of gross rainfall. Taken at 'face value' these evaporation losses are *qualitatively* significant in the context of flood mitigation resulting from tree planting. Theoretical wet-canopy evaporation estimates made using the Penman–Monteith model for large and extreme events in mountainous regions of the United Kingdom suggest consistency with these high observed losses but uncertainties associated with the estimation of, in particular, aerodynamic exchange are so large that this test of consistency remains weak. During 4 major flood events in the Cumbrian Mountains, UK, meteorological conditions were favourable for high rates of wet-canopy evaporation: high windspeeds prevailed and surprisingly low relative humidity was observed at some locations. Thus the disparity regarding the significance of wet-canopy evaporation for flood mitigation between conclusions based upon results from catchments studies of forest cover effects and results from forest plot studies remains.

Our results suggest that it is possible for high rates of *Ewc* over forest to occur during large flood events in mountainous regions of the United Kingdom but not in all locations and not for all events. To be able to determine the potential of tree planting scenarios on flood hydrographs using hydrological models, estimates of the spatial and temporal patterns of wet-canopy evaporation through sequences of rainfall events are needed. Appropriate estimates require simulation of the control imposed by meteorological variables on wet-canopy evaporation to be made necessitating interpolation and extrapolation from (normally) sparse meteorological observation sites. This is difficult to implement with any accuracy and the uncertainties associated with this step are compounded by the large uncertainties inherent with the estimation of the magnitude of aerodynamic exchange with forest canopies. Owing to these large uncertainties, wet-canopy evaporation estimates may be represented best as scenarios based upon explicit assumptions. Hydrological simulations must also represent the limitation on evaporation imposed by storage on the surfaces of different vegetation canopies (e.g., between foliated and unfoliated deciduous trees) such that scenarios of canopy model structure and the associated parameterisation are also required. Simulation scenarios must be associated with a confidence weighting that can be propagated to simulation results as expressions of modelling uncertainty.

If the considerable uncertainties associated with estimating meteorological conditions and *Ewc* across large areas are to be constrained, collection and analysis of a larger number of well-placed and well-distributed meteorological observations is required, combined with concurrent wet-canopy evaporation observations.

ACKNOWLEDGEMENTS

We would like to thank The Environment Agency of England for the rainfall data (licence CL77737MG), Andres Iroume for supplying the raw data for the Chilean sites and Dave Norris and Matt Fry of the UK Centre for Ecology and Hydrology (UKCEH) for the Plynlimon and Balquidder data (no data quality checks were made by UKCEH and UKCEH do not take responsibility for the data or any interpretations made from the data). Thanks are also due to the Eden Defra Demonstration Test Catchments team for their meteorological data collected under Defra projects WQ0210 & LM0304. This study was funded by Natural Environment Research Council grant NE/R004722/1: Quantifying the likely magnitude of nature-based flood mitigation effects across large catchments (Q-NFM). We also thank Richard Keim and an anonymous reviewer for their helpful comments.

CONFLICT OF INTEREST

The authors declare no potential conflict of interest.

DATA AVAILABILITY STATEMENT

For part of the data which are digitized from this work, data sharing is not applicable as no new data were created or analyzed in this study. Other data which have been newly gathered from historical archives require the permission of the data holder.

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How to cite this article: Page T, Chappell NA, Beven KJ, Hankin B, Kretzschmar A. Assessing the significance of wet-canopy evaporation from forests during extreme rainfall events for flood mitigation in mountainous regions of the United Kingdom. *Hydrological Processes*. 2020;34:4740–4754. https://doi.org/10.1002/hyp.13895