

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

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## 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, throughfall 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 UK to assess the likely significance of wet canopy evaporation loss. The collated dataset showed that *Ewc* losses between approximately 2 and 38% of gross rainfall (1.5 to 39.4 mm d<sup>-1</sup>) have been observed during large rainfall events (up to 118 mm d<sup>-1</sup>) and limited data for extreme events (> 150 mm d<sup>-1</sup>). Event data greater than 150 mm, where duration was not reported, showed similarly high percentage evaporation losses. Theoretical estimates of wet-canopy evaporation indicated that, to reproduce these high losses, relative humidity and the aerodynamic resistance for vapour transport needed to be within an envelope of approximately 90 to 97.5% and 0.5 to 2 s m<sup>-1</sup> respectively. Surface meteorological data during large and extreme rainfall events in the UK 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

Wet-canopy evaporation; interception loss; extreme events; Natural Flood Management; meteorological controls; complex terrain; upland UK

## 1. Introduction

Recently in the UK, and elsewhere, there is increased interest in ‘natural flood management’ (NFM) or ‘nature-based solutions’ for flood peak mitigation (Dadson *et al.*, 2017; Environment Agency, 2018; Jongman, Winsemius, Fraser, Muis, & Ward, 2018; Lane, 2017; Wingfield, *et al.*, 2019; World Bank, 2017). Tree planting may be one intervention that has the potential for flood peak reduction through: 1. increased soil infiltration capacity; 2. enhanced soil drying resulting from transpiration; 3. increased ground-surface roughness and 4. 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; 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 (UK), Kirby, Newson, & Gillman (1991, p60) 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.* . (2018) also reviewed evidence of the effects of forest cover and reported the findings of a number of studies; they recognised 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 and Reinhart 1972; Pereiea, 1989); it is also likely that 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 & Grant (1996), Thomas and Megahan (1998) and Beschta *et al.* , 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 *et al.* , 2009; Belmar *et al.* , 2018; Fahey and Payne, 2017; Guillemette, Plamondon, Prévost & Lévesque, 2005; Jones & Grant, 1996; López-Moreno *et al.* , 2006); many do, however, show decreasing effects on flood peak as event magnitude increases or no significant change (Beschta *et al.* , 2000, Birkinshaw *et al.* , 2014; Robinson & Newson, 1986; Thomas and Megahan, 1998; Whitehead and Robinson, 1993; Newson & Calder, 1989). Uncertainties associated with catchment studies of hydrograph change, particularly for extreme events, can be very large (Bathurst *et al.* , 2018; 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; Jones & Grant, 1996; Guillemette *et al.* , 2005; Robinson & Newson, 1986; Thomas and 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 (Andreassian, 2004; Carrick *et al.* . 2017).

## 1.1 Forest plot studies of wet-canopy evaporation: losses during large and extreme rainfall

## events

Forest plot studies estimate  $E_{wc}$ , using a canopy water balance (CWB), as the difference between the gross rainfall (??) incident upon a vegetation canopy and the fraction of ?? that reaches the ground as *net rainfall* ( $P_n$ ). 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  $E_{wc}$  during large ( $> 50 \text{ mm d}^{-1}$  of ??) or extreme ( $> 150 \text{ mm d}^{-1}$  of ??; Collier, Fox & Hand, 2002) rainfall events. A notable exception is the work of Keim, Skaugset, Link and Iroume (2004) who report  $E_{wc}$  losses above 30% of ?? at temperate sites in Chile and Northwest USA. Equally high  $E_{wc}$  losses during large magnitude rainfall events at other locations with a temperate climate have been reported (e.g. see Deguchi *et al.* , 2006 and Hashino *et al.* , 2002). Taken at “face value”, these  $E_{wc}$  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  $E_{wc}$  for flood mitigation depends upon the difference in  $E_{wc}$  between a given forest canopy and another land cover. For example, moorland vegetation species such as Heather (*Calluna vulgaris*) tend to have *long-term evapotranspiration* losses approximately 30% to 60% of that for tall forests and short semi-natural grasses lose around 10 to 40% of that for forests (e.g. Calder, 1976, Calder & Newson, 1979; Calder *et al.* , 1981). However, as most comparative studies derive estimates from catchment or lysimeter water balances over relatively long periods, and do not separate  $E_{wc}$  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 *et al.* , 1967; Gerrits *et al.* , 2007, 2010) and effectively increase the difference in  $E_{wc}$  loss (e.g. between forest and grass). We therefore assume that  $E_{wc}$  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  $E_{wc}$  from tall forest canopies is of primary relevance here. Thus it is important to determine the full extent of evidence for significant  $E_{wc}$  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  $E_{wc}$  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  $E_{wc}$  data from temperate sites around the world are collated and are contextualised using UK meteorological conditions via theoretical analyses. In the UK, 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  $E_{wc}$  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  $E_{wc}$  data from UK-relevant event-scale forest plot studies to quantify the magnitude and range of observed losses, particularly during large ( $> 50 \text{ mm d}^{-1}$ ) and extreme ( $> 150 \text{ mm d}^{-1}$ ) rainfall events; **Objective 2** - to explore the meteorological conditions consistent with the magnitudes of  $E_{wc}$  losses from Objective 1 using the Penman-Monteith equation; **Objective 3** - to examine meteorological data from Cumbria and other mountainous regions of the UK 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  $E_{wc}$  across large catchments in complex terrain.

## 2. Methods

## 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 4 major flood events in the last 15 years. Cumbria, as for many mountainous regions of the UK, is situated towards the west coast and hence strongly influenced by temperate maritime airflows from the Atlantic Ocean. Mountainous regions of the UK are areas of extremely complex topography which, in combination with the predominant airflow, gives rise to orographically-influenced and spatially heterogeneous meteorological patterns (Blackie and 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 1000 mm yr<sup>-1</sup> on the coast and in areas of rain shadow to greater than 3500 mm yr<sup>-1</sup> across the highest mountains. Given there are no CWB *Ewc* data for this study area we consider other mountainous sites in the UK (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 Koppen 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, Koppen climate classifications were used with the rationale that hydrometeorological 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 UK. 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 Supp. 1), 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 hours (but no more than 48 hours) duration were standardised by calculating a *normalised 24-hour rainfall* (i.e.  $P_g = 24 \left( \frac{P_e}{D_e} \right)$  where  $P_e$  is the total event precipitation and  $D_e$  is event duration in hours): note that, in the text below,  $P_g$  relates to a daily or normalised 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 recognised 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 digitising 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)  $P_g$ -*Ewc* relationships were digitised. Where observations were obtained from digitised 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  $P_g$  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.

## 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  $E_{wc}$  losses, a broad range of  $E_{wc}$  estimates were made using the Penman-Monteith equation (Monteith, 1965 ;Eqn. 1); estimates were made using stratified samples from ranges of relative humidity (85% to 100%; expressed via  $e_z$  where  $T_z$  is assumed to be 10°C) and aerodynamic resistances (0.5 s m<sup>-1</sup> to 12 s m<sup>-1</sup>). This analysis creates a response surface where different combinations of meteorological variables that lead to similar  $E_{wc}$  losses can easily be visualised. The Penman-Monteith equation takes the form:

$$\lambda E_{PM} = \frac{\Delta_e H + \rho_a c_p (e_s(T_z) - e_z)/r_{a\_s}}{\Delta_e + \gamma(1 + r_g/r_{a\_s})} (1)$$

where  $\lambda$  is the latent heat of vaporisation (J kg<sup>-1</sup>),  $E_{PM}$  is the evapotranspiration rate (kg m<sup>-2</sup> s<sup>-1</sup>),  $\Delta_e$  is the slope of the saturation vapour pressure curve versus temperature relationship (Pa K<sup>-1</sup>) at temperature,  $T_z$ , where  $z$  is the observation height in metres,  $H$  is the total energy available for evaporation (J m<sup>-2</sup>),  $\rho_a$  is the density of air at  $T_z$  (kg m<sup>-3</sup>),  $c_p$  is the specific heat capacity of air (J kg<sup>-1</sup> K<sup>-1</sup>),  $e_s(T_z)$  is the saturation vapour pressure at  $T_z$ ,  $e_z$  is the actual vapour pressure at  $z$ ,  $\gamma$  is the psychrometric constant ( $\approx 66 \times 10^{-3}$  Pa K<sup>-1</sup>) 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<sup>-1</sup>). 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; VanDijk 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<sup>-2</sup> d<sup>-1</sup>.

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

To explore the potential for  $E_{wc}$  across mountainous regions of the UK during large and extreme rainfall events,  $E_{PM}$  was calculated using meteorological data for 17 sites (Figure 1a and Table Supp. 2) using Eqn. 1. The aerodynamic resistance for momentum ( $r_{a\_m}$ ) was estimated using Eqn. 2:

$$r_{a\_m} = \frac{\ln\left(\frac{z-d}{z_0\_m}\right)^2}{\kappa^2 U_z} (2)$$

where  $z$  is the wind speed observation height,  $d$  is the zero-plane displacement,  $z_0\_m$  is the roughness length for momentum (all in metres),  $U_z$  is the wind speed (m s<sup>-1</sup>) at  $z$  and  $\kappa$  is the dimensionless von Karman constant ( $\approx 0.41$ ). The canopy height ( $Z_c$ ) was arbitrarily assumed to be 20 m,  $d$  to be 0.75( $Z_c$ ) and  $z_0\_m$  as 0.1( $Z_c$ ) in accordance with Szeicz, Endrodi and Tajhman (1969) and Rutter, Robins, Morton and Kershaw (1972). However, studies have shown an enhancement of exchange compared to estimates assuming these approximations for  $z_0\_m$  (e.g. Holwerda *et al.*, 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 and Brunet, 1992; Raupach, 1979; Simpson, Thurtell, Neumann, Den Hartog, & Edwards, 1998). For the indicative calculations made here, where  $z < Z_c$ , wind speed was extrapolated to  $Z_c$  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 recognised 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 lead to considerable error (Brutstaert, 1982, p62) 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 (Brutstaert, 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 Eqn. 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 (Bosvelt, 1999; Szeicz, & Endrodi, 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 (Brutstaert, 1982, p116; 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 (Bosvelt, 1999; Brutstaert, 1982, p114; 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 -  $z0\_s/z0\_m = 1.0$  ; i.e.  $z0\_s = 0.1(Zc)$ ;

Scenario 2 -  $z0\_s/z0\_m = 0.5$ ; i.e.  $z0\_s = 0.05(Zc)$ ;

Scenario 3 -  $z0\_s/z0\_m = 0.1$ ; i.e.  $z0\_s = 0.01(Zc)$ .

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, Eqn. 2 is strictly only valid for neutral atmospheric conditions (Szeicz, and Endrodi, 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  $E_{wc}$  for large storms associated with either daily or normalised event  $??$  (see Table Supp. 3): none of the events were extreme events based upon the classification employed ( $>150 \text{ mm d}^{-1}$ ). From these sites 1387  $??-E_{wc}$  pairs were obtained with a maximum  $??$  of approximately 118 mm. Only 35 of these pairs were associated with  $??$  observations over  $50 \text{ mm d}^{-1}$  (Table Supp. 3). In absolute terms, the  $E_{wc}$  data include some high magnitudes; they include a maximum  $E_{wc}$  loss of  $39.4 \text{ mm d}^{-1}$  with 40 events where  $E_{wc}$  loss was over  $10 \text{ mm d}^{-1}$ . When  $E_{wc}$  is expressed as a percentage of  $??$  ( $\%E_{wc}$ ), there is a decreasing trend in  $\%E_{wc}$  as  $??$  increases (Figure 2). This pattern of declining relative loss has been shown previously in many studies (e.g. Iroume & Huber, 2002; Bulcock & Jewitt, 2012), including the seminal review by Horton (1919). From these 1387 data pairs, 58 gave a negative  $\%E_{wc}$  and 25 were over  $100 \%$   $\%E_{wc}$ ; these data lie predominantly at low  $??$  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  $?? > 50 \text{ mm}$  are specified in Table Supp. 3. These UK data tend to span the higher rates of  $\%E_{wc}$  for higher  $??$  magnitudes and include absolute losses up to  $26.3 \text{ mm d}^{-1}$ . 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  $??$  greater than  $50 \text{ mm}$  per day of which approximately 11% was lost to  $E_{wc}$ . The degree to which  $\%E_{wc}$  continues to decrease with increasing  $??$ , or whether it has reached a stable range is unclear from Figure 2 given the few data available at higher  $??$  magnitudes. This is important as the highest  $??$  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-hour 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 1144  $??-E_{wc}$  pairs including some high  $\%E_{wc}$  losses (ranging from approximately 7 to 35 %) for very large  $??$  values (up to 435 mm: see Table Supp. 2). 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 normalised 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, Hattori, & Park (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 11-12<sup>th</sup>, 2000) deposited 347 mm ??, with 14% of this being lost to *Ewc*. The magnitude of the losses from these non-normalised events are qualitatively consistent with the normalised events presented in Figure 2.

The data for large and extreme events presented in Figures 2 and 3 show a large range of *Ewc* loss: approximately 2-38 % of ?. These losses are apparently significant in the context of flood mitigation with absolute losses of up to approximately 40 mm d<sup>-1</sup>. Unfortunately, concurrent meteorological observations were generally not reported for events greater than 50 mm d<sup>-1</sup>: concurrent observations were only available for 4 events at one site (Dolydd, mid-Wales; events 7-10, Table Supp. 3). 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* (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 ( $\approx 20$  to 40 mm d<sup>-1</sup>) 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<sup>-1</sup> or around 97.5% as  $r_{a\_s}$  approaches 0.5 s m<sup>-1</sup>. 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.

### 3.3 Objective 3: Meteorological conditions and wet-canopy evaporation estimates for mountainous regions of the UK

#### 3.3.1 Penman-Monteith wet-canopy evaporation estimates for mountainous regions of the UK

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 UK. 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 Supp. 2. The points plotted in Figure 5a relate to hourly periods within a 24-hour period with over 50 mm of rainfall and where the hourly rainfall total was above zero. The estimates of  $r_{a\_s}$ , using the 3 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 ?? 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-hours 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 although the  $z0\_s = 0.01(\text{Zc})$  scenario is mainly limited to losses of below 12 mm d<sup>-1</sup>.

### 3.3.2 Meteorological conditions during extreme events across Cumbria

Figures 6 and 7 (and Figures Supp. 1 & 2) 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, 7b, Supp. 1b & 2b) relatively low  $RH$  (in the context of the results presented above) occurs along with high wind speeds (Figures 6a, 7a, Supp. 1a & 2a) and an increase in air temperature (Figures 6, 7c, Supp. 1a & 2a). 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 (Figures 7a and 7b 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, 7 and Supp. 1 & 2). 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  $E_{wc}$  loss shown are not necessarily representative across the region as a whole: as shown by the less favourable conditions for  $E_{wc}$  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  $E_{wc}$  loss at some locations. Consequently the need to estimate  $E_{wc}$  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 $E_{wc}$ across large catchments in complex terrain

The results from objectives 1 to 3 show that  $E_{wc}$  losses up to approximately 40 mm d<sup>-1</sup> 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 UK. However, these findings must be treated with caution because concurrent meteorological observations are rarely reported with CWB  $E_{wc}$  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  $E_{wc}$  estimates are required for hydrological simulation of large catchments, a representation of this spatially variable control of  $E_{wc}$  magnitude is required; it is not appropriate to sample a statistical distribution of  $E_{wc}$  loss generated from the worldwide observations of  $E_{wc}$  data (for a given gross rainfall total) as the autocorrelation of  $E_{wc}$ , controlled by autocorrelated meteorological variables, through sequences of *real* 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  $E_{wc}$  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  $E_{wc}$  losses.



Although simple empirical models can be used to estimate  $E_{wc}$  where there is a scarcity of adequate meteorological data and knowledge of appropriate parameter values for more complex models (e.g. see Lu, McNulty & Amatya, 1995), 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  $E_{wc}$  models (Muzylo *et al.*, 2009). Thus, Penman-Monteith equation remains a useful method to determine the potential for  $E_{wc}$  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  $E_{wc}$  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  $E_{wc}$  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 (Brutstaert, 1982, p114; Stewart & Thom, 1973) and around 1 in some cases (Bosvelt, 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 (Brutstaert, 1982, Fig. 5.1; Cellier & Brunet, 1992; Holwerda *et al.*, 2012) as well as atmospheric stability and wind speed (Bosvelt, 1999; Cellier & Brunet, 1992; Szeicz, & Endrodi, 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  $E_{wc}$  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  $E_{wc}$  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 d<sup>-1</sup> have been recorded for large rainfall events (>50 mm d<sup>-1</sup>) 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 UK 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  $E_{wc}$  over forest to occur during large flood events

in mountainous regions of the UK 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 with the large uncertainties associated with the estimation of aerodynamic exchange with forest canopies such that wet-canopy evaporation estimates may be represented best as scenarios. Simulations must also be able to represent the limitation on evaporation imposed by storage on the surfaces of different vegetation canopies (e.g. between foliated and unfoliated deciduous trees) meaning that scenarios of parameterisations must be incorporated into the simulations. 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  $E_{wc}$  across large areas such as 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.

### Data Availability Statement

For part of the data which are digitized from this work, data sharing is not applicable new data were created or analyzed in this study. Other data which has been newly gathered from historical archives requires the permission of the data holder.

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