Observations and numerical weather forecasts of land-surface and boundary-layer evolution during an unusually dry spring at a site in central England

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Abstract
Analysis of screen-level and land-surface temperature forecasts are known to have long-standing warm nighttime, and cold daytime, temperature biases in regional models. During a record-breaking spring and subsequent summer in 2020, over 100 radiosondes were launched at the Met Office Cardington site under clear skies on ten morning and one evening transitions. We compare observations with operational Met Office UKV forecasts and a standalone land-surface model (JULES). Wind profiles show the UKV nocturnal jet was too high, suggesting too much mixing in the modelled boundary layer. The simulated nighttime surface inversion was too weak and the profile too cold immediately above the inversion. The radiosondes were in addition to comprehensive long-term observations. The evapotranspiration was too large on a seasonal timescale for both the UKV and JULES. For spring and summer, UKV mean screen temperature errors were $-0.2 \pm 1.3^\circ C$ during daytime and $1.3 \pm 1.9^\circ C$ at night. The soil temperature diurnal range was too large in both the UKV (by $3.9^\circ C$) and JULES ($2.9^\circ C$), suggesting the surface is too highly coupled to the soil in the simulations. For the spring experimental days, UKV mean maximum screen temperature error was $-0.8 \pm 0.4^\circ C$. The buoyancy-flux crossover times in the morning were slightly too early in the UKV (18 min on average) yet much earlier in JULES (52 min). Observations show the diagnosed boundary layer lifts too early in the UKV with onset of convection occurring on average 67 min too early. The UKV develops summertime boundary layers that are too deep by early afternoon. There was a time lag between the observed screen-skin temperature and buoyancy-flux crossovers in the morning that was not captured by the simulations. The evening buoyancy-flux crossovers in JULES agreed well with the observations, but the UKV crossovers were on average 55 min too early.

KEYWORDS
boundary-layer transition, buoyancy flux, evapotranspiration, JULES, Screen temperatures, soil moisture, Unified Model
1 | INTRODUCTION

Global circulation models are known to have persistent biases in near-surface variables that have been verified in both short-term numerical weather forecasts (Haiden et al., 2018; Bush et al., 2020) and long-term climate simulations (Ma et al., 2014). Errors, based on screen-level air temperature, have been demonstrated varying from diurnal to annual cycles and varying spatially from large biases in semi-arid regions (Klein et al., 2006; Van Weverberg et al., 2015) through to modest errors in temperate and arctic regions (Dutra et al., 2015). ECMWF’s Integrated Forecasting System, for example, has a general underestimation of the amplitude of the diurnal cycle of temperature, in the order of a 1–2K bias in the summertime, coinciding with a daytime low-humidity bias (Haiden et al., 2018). Model errors tend to increase with forecast lead time, as was shown for a regional version of the Unified Model (Bush et al., 2020). Although some success has been achieved in regional models to reduce biases, for example, from changes to vegetation cover which affects the momentum roughness length and therefore slows down over-speeding winds (Bush et al., 2020), errors nonetheless persist.

Errors in key forecast diagnostics that are used for model verification such as screen-level temperature, land-surface (skin) temperature, humidity and wind speed suggest problems with energy partitioning at the surface (Trigo et al., 2015). Studies have shown that, during meteorological dry-downs, land-surface models can have significant errors in evapotranspiration (Osborne and Weedon, 2021) which can have larger implications in the simulation of evaporative drought duration, magnitude and intensity (Ukkola et al., 2016). Biases related to cloudiness can be eliminated, and radiation biases can be minimised from the problem by focusing on case studies of clear-sky conditions — yet aerosols will always be present to some degree and so affect radiative transfer. Despite many studies on models biases, such analyses have tended to focus on times of nighttime minima and daytime maxima rather than on periods of transition from stable to convective regimes (or vice versa), such as the morning and evening transitions of the boundary layer (Angevine et al., 2020). It is during these periods that biases can be established through incorrect modelling of surface processes and transport of energy upwards through the surface-mixed layer and downwards into the soil.

Methods of defining the morning (and evening) transitions, when the surface layer changes from a stable to convective regime (and vice versa) vary in the literature. We define the morning transition to cover a period of time starting with sunrise and ending when the surface temperature inversion has been eroded (Angevine et al., 2001). The latter event is called onset and requires observation of turbulence substantially above the surface. The time of crossover in the sign of the near-surface heat flux occurs between sunrise and onset and is another important reference point. The development of the boundary layer through the transition is not controlled solely by solar radiation. Down-gradient diffusion of heat from above when the surface temperature reaches its minimum is known to often dominate early in the transition before surface heating dominates (Lapworth, 2006). Eddy-diffusion downward through a temperature inversion is often a poorly represented process in forecast models (Angevine et al., 2020; Edwards et al., 2020).

The growth of the convective boundary layer is influenced by the generation of vertical mixing caused by shear (mechanical turbulence) or buoyancy (convective turbulence) (Lapworth, 2006; Beare, 2008), local processes such as surface flux partitioning (Gevaert et al., 2018) and entrainment rate at the boundary-layer top (van Heerwaarden et al., 2009), in addition to large-scale processes such as advection and subsidence. Advection is a problem when analysing individual case studies as it is nearly always present but is not observed. Averaging over many cases may however largely cancel out effects of advection (Betts and Barr, 1996).

Unusual spring conditions in 2020 initiated a radiosonde-based field experiment with opportunistic intensive observational periods (IOPs), each lasting 10–12 hr, that extended into the summer. After the wettest February on record for the UK, the spring of 2020 in the UK saw further record-breaking conditions with the driest May on record since 1896, with less than 10 mm rain falling across England on average (NCIC, 2021). Another UK record was broken with 626 hrs of sunshine in spring as a whole, which exceeded the previous high of 555 hrs, set in 1948. Although exceptionally dry and sunny, spring was not notable for the air temperatures recorded. However, the whole of 2020 was the third warmest year on record for the UK, with July 31, 2020 (one of the IOPs) being the third warmest day on record for the UK. Although June and July had greater than average rainfall, we captured 11 IOPs covering an appreciable range in soil temperature, soil moisture and Bowen ratio.

In this article, we investigate long-standing errors in the regional version of the Unified Model, namely skin and screen temperatures, evapotranspiration and soil processes. We make particular use of the IOP radiosonde soundings, alongside longer-term surface and subsoil observations, to allow us to compare boundary-layer thermodynamic profiles as well as near-surface conditions with operational forecasts. A standalone land-surface model has also been run and forced by local meteorology.
to investigate land-surface processes independent of atmospheric coupling.

This article is arranged as follows: Section 2 provides a description of the instrumentation deployed and the observation strategy. The regional forecast model and the land-surface model used in this evaluation are summarised. Results are presented in Section 3, including both an assessment of the seasonal evolution and analyses of IOP case studies. Section 4 presents the discussion and conclusions.

2 | METHODOLOGY

2.1 | Observations

All observations were made on the 18 Ha semi-rural Cardington field site in central England (52.105°N, 0.424°E, 29 m above sea level). The site is laid mainly to grass kept at a constant 5–10 cm height throughout the year. Further details on the site and instrumentation can be found in Osborne and Weedon (2021) and Horlacher et al. (2012).

Rainfall is measured with a Met Office Mk5 tipping-bucket gauge with a 0.2 mm accuracy. Shortwave radiative fluxes are measured with a conditioned Kipp and Zonen CM21 pyranometer. Sensible heat flux masts are located at 2, 10, 25 and 50 m using Gill HS50 3-D horizontally symmetric ultrasonic anemometers, with latent heat flux additionally determined at 10 m with a Licor Li-7500 open-path hygrometer. Eddy covariances over 10- and 30-min intervals were used to calculate the turbulent heat fluxes. Although statistical errors will increase for shorter time intervals, we are confident that 10-min intervals provide robust data during the daytime when turbulence is relatively high. The sonic data have cross-wind speed correction, coordinate rotation, detrending and despiking applied. Air temperature and relative humidity (RH) at all heights are measured with Vector Instruments T302 and Vaisala HMP155 sensors, respectively. The grass canopy air temperature and RH is determined in situ with a screened and aspirated Rotronics Hydroclip2, whilst the grass skin temperature is observed radiometrically with a Heitronics KT15D thermometer (8–14 μm) from a height of 2 m. Delta-T ML3 theta probes measure volumetric soil moisture at depths of 10, 22, 57 and 160 cm, and Delta-T ST2-396 thermistor probes measure soil temperature at 1, 4, 7, 10, 17, 35, 65 and 100 cm.

A 1.5-μm Halo Photonics Streamline lidar (Pearson et al., 2009) measures radial backscatter and Doppler velocity from aerosol (it is fully attenuated by 100–200 m penetration into cloud). It was operated here viewing in the zenith, so updraught velocity and the vertical component of turbulence, as well as the boundary-layer structure, were continuously monitored. As the lidar beam scatters off aerosol particles, the top of the scan usually represents an inversion below which the aerosol is trapped. Combination of vertical velocity and backscatter provides a continuous visualisation of the boundary-layer structure.

In addition to the Halo lidar, a Vaisala broadband differential absorption lidar (DIAL) was operating during a temporary trial in June 2020. The DIAL uses two closely spaced laser spectral lines in the near infrared that have different absorption properties for water vapour (Newsom et al., 2020). The DIAL retrieves water vapour profiles at 1-mi intervals and so provides an alternative to the Halo for visualising the boundary layer. This was important on June 24–25, 2020 when the Halo reached its maximum operating temperature in the hot weather and so went into forced shutdown for a few hours each afternoon. Part of the DIAL data processing requires filtering out untrustworthy data based on signal-to-noise. This results in a seemingly abrupt top to the boundary-layer data with no information above that in the free troposphere. We show below that the upper limit of the trustworthy DIAL data during the IOPs coincides well with the boundary-layer top.

In addition to continuous surface and near-surface measurements, we use radiosonde profiles to see the vertical extent of the model errors which may be related to the degree of mixing in the model. Vaisala RS41 radiosondes were launched during the IOPs on an hourly schedule; for the morning IOPs, these started before sunrise and continued until the boundary layer was fully developed usually by late morning, but sometimes into the afternoon. One evening IOP was captured on 24 June, so with a morning IOP the following day, this allowed nearly 24 hr of monitoring in clear-sky conditions. Table 1 presents the dates, timings and total numbers of the radiosondes for each IOP.

Individual vertical profiles of virtual potential temperature were analysed to derive the boundary-layer depth by identifying inversions of virtual potential temperature. An automated method using the bulk Richardson number which exceeds a critical value of 0.25 was tested (Vogelezang and Holtslag, 1996; Seibert et al., 2000). Using this automated method, it proved difficult on occasion to identify the top of the growing convective boundary layer within the complex temperature structure in the residual layer. At times in the afternoon where radiosonde profiles were not available, the boundary-layer depth was derived from the vertical velocity variance and attenuated back-scatter coefficient from the Halo lidar as a measure of the vertical component of turbulence.
### Table 1
Summary of intensive observation periods over spring and summer 2020

| Date  | Launch period | Number sondes | $T_{\text{scrn}}^\text{min}$ error °C | $T_{\text{scrn}}^\text{max}$ error °C | Conditions                                      |
|-------|---------------|---------------|----------------------------------|----------------------------------|-----------------------------------------------|
| 7 April | 05–09         | 5             | 2.1                              | $-0.5$                           | Clear-sky start, cumulus from 1000 UTC       |
| 15 April | 04–11       | 8             | 1.0                              | $-1.7$                           | Thin cirrus                                   |
| 20 April | 04–13        | 10            | $-0.5$                           | $-1.0$                           | Clear-sky, windy                             |
| 21 April | 05–14        | 10            | 1.2                              | $-0.3$                           | Clear-sky, windy                             |
| 22 April | 04–15        | 12            | $-0.4$                           | $-0.7$                           | Clear-sky, calmer                             |
| 23 April | 04–15        | 12            | 3.7                              | $-0.4$                           | Thin cirrus                                   |
| 25 May  | 02–11        | 10            | 5.2                              | $-0.8$                           | Clear-sky                                    |
| 1 June  | 03–12        | 10            | 1.5                              | 0.0                              | Clear-sky start, cumulus from 1000 UTC       |
| 24 June | 09, 12, 16–23| 10            | $-0.2$                           | 1.2                              | Clear-sky evening                            |
| 25 June | 03–12, 15    | 11            | 3.0                              | 1.5                              | Clear-sky morning                            |
| 31 July | 04–14        | 11            | $-0.9$                           | 0.4                              | Clear-sky morning                            |

### 2.2 Modelling

#### 2.2.1 UKV

The variable resolution UK model (UKV) for kilometre-scale forecasting is a 1.5 km deterministic configuration of the Met Office Unified Model (UM). The UKV solves non-hydrostatic, deep-atmosphere dynamics using a numerical scheme, which is semi-implicit and semi-Lagrangian (Davies et al., 2005). The UKV is one-way nested and forced by lateral boundary conditions from the Met Office global model (Walters et al., 2019). The global model is based on the Met Office Operational Suite 43 (OS43, operational between December 2019 and December 2020), which uses the UM with Global Atmosphere 7.2 configuration coupled to Joint UK Land Environment Simulator (JULES) land-surface model with Global Land 8.1 configuration.

The first Regional Atmosphere and Land (RAL1) science configuration used in the UKV is documented in Bush et al. (2020). The second regional science configuration, RAL2, became operational in December 2019 and is the configuration of the UKV used in this study. The differences between RAL1 and RAL2 are expected to have a limited impact in the cases considered here. The UKV uses hourly cycling whereby the forecasts initialised at 0000 UTC, 0600 UTC, 1200 UTC, and 1800 UTC are run for 36 hrs and all other forecasts are run for 3 hrs. In this study, the UKV is initialised from its own analysis at 0000 UTC on each IOP date. Hourly cycling regional land-surface data assimilation is used in the UKV, in which pseudo-observations (analysis increments from the atmospheric assimilation) of screen temperature and humidity and satellite-derived soil wetness (ASCAT scatterometer) observations, when available, are assimilated (Gómez et al., 2020). Recent improvements to forecasts of the near-surface air temperature and humidity have been demonstrated through regional land-surface data assimilation (Gómez et al., 2020).

The UKV has 70 levels in the vertical, of which six levels are below 150 m. The lowest vertical level is at 2.5 m for horizontal winds and 5 m for temperature and humidity (Boutle et al., 2016). Diagnostics at 1.5 m or screen level are a vertical interpolation between the surface and the lowest model level based on Monin–Obukhov similarity theory in the surface layer (Edwards et al., 2020). This assumption holds true where the constant flux approximation is applicable, that is, where fluxes of heat and momentum do not change significantly from their values at the surface up to the lowest model level (Edwards, 2009).

The UKV uses a blended boundary-layer parametrization as described in Boutle et al. (2014). The scheme transitions between the one-dimensional (1-D) boundary-layer scheme for vertical mixing of Lock et al. (2000) to a Smagorinsky sub-grid turbulence scheme based on Smagorinsky (1963). Radiative processes are represented using the SOCRATES radiative transfer scheme (Edwards and Slingo, 1996). Land-surface processes and the exchange with the atmosphere are represented using the JULES scheme of Best et al. (2011) that we describe next.

#### 2.2.2 JULES

While the UKV represents an atmosphere coupled to the JULES land-surface scheme, JULES can also be run cheaply as a standalone research tool when sufficient
initialisation and driving variables are available, for example from observations or model analyses. We compare operational UKV forecast data with JULES run as a standalone scheme, set up using the UKV RAL2 configuration, for a single grid point. JULES is initialised from March 1, 2020 with observed radiometric skin temperature, and soil temperature and moisture in all subsoil layers. These parameters thereafter run freely in the scheme whilst the required drive parameters of air temperature (10 m), specific humidity (10 m), wind speed (10 m), precipitation, barometric pressure and short- and longwave downwelling irradiances are prescribed at every time step from the site observations. Such drive data as well as all observations presented below, whether for 30- or 10-min time steps, represent means over the intervals.

JULES uses a tiled scheme to represent sub-grid scale heterogeneity of the land surface. It has the option of four non-vegetated surface types and five plant functional types. As the research site is laid to grass, we have assumed 100 % C3 grass canopy for our configured simulation. JULES is run here with the standard UKV configuration of the C3 canopy and the four soil layers: 0–0.10 m, 0.10–0.35 m, 0.35–1 m and 1–3 m. The root fractional distribution decreases exponentially with an e-folding depth of 0.5 m (operational UKV value). The soil parameters, such as hydraulic and thermal conductivities, and the saturation (44.9%), wilting (34.1%) and critical (22.6%) soil moisture contents have been set to the standard values for the relevant gridbox in the UKV (see Osborne and Weedon (2021) for more details on the subsoil set-up). Saturation point is the maximum moisture content that the soil can hold when all pores are filled with water; critical point is when the ability of the plant to transpire water starts to reduce; wilting point is reached when plants are no longer able to draw water up and transpiration and photosynthesis ceases. These soil moisture points are equal for all four soil layers in order to mimic the UKV, although we appreciate that soil compacts with depth and so in reality these values should decrease through the layers.

3 | RESULTS

3.1 | Seasonal evolution

3.1.1 | Soil moisture

This section and 3.1.2 that follows show 30-min data covering meteorological spring and summer, that is, March 1 to August 31, 2020. Figure 1a–c shows the evolution of soil moisture as simulated in soil layers 1–3 for standalone JULES and the coupled UKV as compared with the observations. The times of the IOPs are shown by the vertical pink lines. The first IOP on 7 April took place with the layer 1 soil moisture above the critical point, the next five IOPs (still within April) had soil moistures between the critical and wilting points, and the remaining five IOPs took place after the main dry-down with the soil moisture below the wilting point. Despite large data outages in March for the observed soil moistures, it is assumed that the soil at all three layers was saturated or near saturated for all of March. The assumption in the model of the same saturation, critical and wilting points for all soil layers is unrealistic. This explains why JULES is too wet in layer 3 throughout the whole period because the JULES saturation point is too high. Observations from the soil layer equivalent to layer 4 of JULES (1.6 m sensor depth, not shown) were saturated at all times with a constant value similar to the layer 3 saturation; JULES layer 4 shows a slow decrease over the period, a little above the observed value, due to deep roots drawing up water in the simulation.

At the Cardington grid-point, the UKV is universally too wet in soil layer 1 and rarely falls below the wilting point, with deeper layers also a little too wet. The behaviour of the UKV soil water is impacted by hourly cycling land-surface data assimilation which includes remotely sensed ASCAT soil wetness observations (Gómez et al., 2020). JULES achieves drier soils in the summer, substantially below the wilting point via bare-soil evaporation, yet is still at all times moister than observed. Periods of dry-down starting below the critical point are too slow in JULES, whereas dry-down starting from above the critical point is more realistic. JULES does not remain saturated, however, and readily loses soil water early on in a simulation if initialised at or very near saturated. This is because the water must balance in JULES for the single point in the simulation, whereas in reality the site can remain saturated with the water table at or near the surface because of the sub-surface advection of water.

Table 2 summarises the monthly mean daytime maximum Bowen ratios from the observations, JULES and UKV. Figure 1d shows the seasonal evolution of the Bowen ratio. Whilst JULES captures some of the variability in the monthly mean Bowen ratio, the UKV shows no seasonality. For example, the UKV monthly mean Bowen ratios range from 0.9 to 1.1. JULES varies between 1.3 and 2.2, whilst the observations span 1.4–3.3. JULES Bowen ratios are much closer to the observations, particularly in spring. For example, the May monthly mean Bowen ratio from JULES and the observations are in agreement at 2.0 ± 1.0, which corresponds to a period where the soil moisture in JULES is well simulated. In contrast, JULES does not simulate the highest Bowen ratios in July and August. The August monthly mean Bowen ratio for JULES
FIGURE 1  Seasonal evolution during spring and summer (March 1 to August 31) 2020 of the observed ('obs') and modelled JULES and UKV (a–c) volumetric soil moisture content for soil layers 1, 2 and 3, (d) Bowen ratio (observations at 10 m height) and (e) cumulative rain and evapotranspiration (observations at 10 m height). The moisture contents at the saturation, wilting and critical points are taken from the relevant gridbox in the UKV. Days of the IOPs are marked [Colour figure can be viewed at wileyonlinelibrary.com]

TABLE 2  Monthly means (± one standard deviation) of the daytime Bowen ratio maxima

| Date | Obs. | UKV | JULES |
|------|------|-----|-------|
| March | 1.4 ± 0.6 | 1.1 ± 0.4 | 1.8 ± 0.8 |
| April | 1.4 ± 0.3 | 0.9 ± 0.3 | 1.3 ± 0.6 |
| May | 2.0 ± 1.0 | 1.0 ± 0.4 | 2.0 ± 1.0 |
| June | 2.7 ± 1.7 | 0.9 ± 0.4 | 2.2 ± 0.8 |
| July | 2.2 ± 0.9 | 0.9 ± 0.3 | 1.7 ± 0.7 |
| August | 3.3 ± 1.9 | 1.0 ± 0.5 | 1.7 ± 1.1 |

The excessive evapotranspiration in the models can be summarised by cumulative totals shown in Figure 1e, with the UKV showing 429 mm equivalent evapotranspiration by 31 August compared with the observed value of 304 mm. Although JULES is closer to the observations (accumulation of 386 mm), the excessive evapotranspiration in the surface scheme is of concern given it is forced with the observed precipitation.

3.1.2  |  Temperatures

Figure 2 shows screen-level, skin and soil temperatures across spring and summer for the UKV and JULES compared with the observations. The long-standing issue of UKV screen and skin temperatures having too small a diurnal range, that is, a warm night bias and a cold day bias, is seen here in Figure 2a,b almost universally (Bush et al., 2020). The UKV skin temperature biases (mean
Seasonal evolution during spring and summer (March 1 to August 31) 2020 of the observed ('obs') and modelled JULES and UKV (a) screen air temperature, (b) skin temperature and (c–e) soil temperature within layers 1, 2 and 3. Days of the IOPs are marked [Colour figure can be viewed at wileyonlinelibrary.com]

The daytime error of $-2.2 \pm 3.4^\circ C$ and mean nighttime error of $2.3 \pm 2.0^\circ C$ across the period) are larger than the equivalent screen biases (daytime error of $-0.2 \pm 1.3^\circ C$, nighttime error of $1.3 \pm 1.9^\circ C$) as might be expected. The cold daytime bias in the model is a long-standing issue that can be partially explained by the soil moisture being too high, resulting in too much evapotranspiration, leading to suppressed daytime temperatures.

The UKV temperatures differ from JULES for three main reasons: the UKV is impacted by land-surface data assimilation, different meteorological forcing and coupling feedbacks between the land and atmosphere. In addition, JULES is forced throughout the period with observed 10-m temperature, with JULES therefore showing closer agreement to the screen and skin temperature observations. The JULES daytime screen temperature error is $0.7 \pm 0.8^\circ C$, and the nighttime error is $0.4 \pm 2.0^\circ C$. For skin temperature, the daytime error is $0.8 \pm 1.9^\circ C$, with a nighttime error of $1.3 \pm 2.1^\circ C$. Whilst a warm nighttime bias is to be expected, the warm daytime bias in JULES is at odds with the cold UKV bias.

The UKV and JULES nighttime soil temperature minima in layer 1 (Figure 2c) are fairly close to the observations (mean errors of $-0.73 \pm 1.3^\circ C$ for UKV, and $-0.4 \pm 1.0^\circ C$ for JULES) but with the daytime maxima showing significant errors ($3.2 \pm 2.1^\circ C$ for UKV, and $2.5 \pm 1.7^\circ C$ for JULES) in both simulations, especially during cloud-free, settled conditions such as the IOPs. The warm daytime bias in soil temperatures within JULES and UKV, for a soil layer 0–10 cm deep relative to the mid-layer measurement at 4 cm, suggests that the soil is too closely coupled to the canopy. Although the diurnal range is naturally suppressed at layer 2 (Figure 2d), the simulations again show too much energy reaching this layer, leading to slightly too warm daytime soil temperatures (mean errors of $0.2 \pm 1.4^\circ C$ for UKV and $0.6 \pm 0.7^\circ C$, respectively), noticeable for extended periods of up to a few weeks. In layer 3 shown in Figure 2e, the simulated soil
temperatures tend to be a little too cold except at the warmest time in early August which also coincides with the largest biases in layers 1 and 2.

3.2 | Intensive observing periods (IOPs)

3.2.1 | Buoyancy fluxes

Figure 3 presents the observed and modelled diurnal cycles of kinematic buoyancy flux, that is, the turbulent heat flux calculated with the virtual potential temperature, together with the observed net shortwave irradiance at the surface. The contribution of water vapour, that is, via the latent heat flux, to the buoyancy flux is typically only 5–10% in the middle of the day, so buoyancy flux tends to be dominated by the sensible heat flux. Although the water vapour contribution can be large (>50%) around the times of the transitions, the total buoyancy flux is small at such times compared with midday. Although we have used 15-min intervals from the UKV and 10-min intervals from JULES and observations to determine the timings below, the observations are presented at both 10 and 30 min, the latter demonstrating that the 10-min intervals are sufficient to calculate fluxes with the eddy-covariance technique.

Convective onset is defined where the variance in vertical velocity at a height of 200 m, as measured with the Halo lidar, is greater than 0.12 m²·s⁻² when combined with a positive skewness (Hogan et al., 2009). Angevine et al. (2001) used 200 m because it was the height of the Cabauw tower. Given we are using the Doppler lidar to measure turbulence, we could have chosen any height above 100 m, but a depth of 200 m captures most surface inversions found at Cardington based on radiosonde launches. We define crossover as the change in sign of the near-surface buoyancy flux, i.e. when the flux becomes positive. The evening transition is dynamically different from the morning counterpart, with a relatively rapid collapse of the
Table 3 summarises the morning and evening buoyancy-flux crossover times with model timing errors, combined with the observed onset times and the local sunrise and sunset as points of reference. The observed mean difference between sunrise and morning crossover is 106 min, with the UKV time to crossover being on average 18 min early and JULES 52 min early. During the evening, however, the JULES mean difference is negligible (5 min) compared with the UKV being 55 min early. JULES is highlighting a model issue in the morning where the turbulent fluxes respond too quickly and the developing boundary layer lifts too early. The early reversal of the buoyancy flux in the UKV during the afternoon is more prominent in the summer cases, leading to evening crossover occurring up to 80 min early.

On IOP days with higher wind speeds, such as 20–22 April, the timing of the morning crossover is well represented by the UKV and JULES (e.g., errors of 10 min for the UKV), as is the magnitude of the buoyancy flux at convective onset. Conversely, on IOPs with lower wind speeds, such as 7 and 15 April, the UKV morning crossover is early and has a larger buoyancy flux at convective onset compared with the observations. Angevine et al. (2001) demonstrated a correlation of higher 10-m wind speeds leading to a time delay between sunrise and the surface sensible heat flux going positive, leading to a shorter time for convective onset to occur. This correlation is also evident in this study, particularly during the consecutive four-day period between 20 and 23 April. The highest wind speeds were observed on 21 April, with an average 10-m wind speed of 4.4 m s\(^{-1}\) during the morning transition, and a low-level jet of 17 m s\(^{-1}\) at 400 m (Section 3.2.2). The corresponding observed time between crossover to convective onset was 60 min. On 23 April, the low-level jet at crossover had slowed to 7 m s\(^{-1}\) with a slight slowing of the the 10-m wind speeds to 3.2 m s\(^{-1}\). The time between crossover and convective onset increased considerably to 140 min.

We analyse crossover timings further in Figure 4, where the skin minus screen temperatures and the skin minus soil temperatures are shown for two IOPs of low (<1 m s\(^{-1}\) on 15 April) and appreciable (≈4 m s\(^{-1}\) on 21 April) near-surface wind speeds at sunrise. The crossover times shown by vertical lines are compared with the crossover in sensible heat flux shown in the top panels. The larger negative heat fluxes in both the observations and simulations can be seen on 21 April at nighttime during more turbulent conditions. The parameterization of the sensible heat flux is proportional to the skin to air temperature difference, and as such the modelled heat flux crossover times are similar to the skin minus screen temperature crossovers (although they are not always identical because there are differences in the UKV diagnostics in that some are instantaneous values while others are means over the time interval). The model skin minus screen temperature crossovers occurred at similar times to the observations both morning and evening. However, there tends to be a lag in the observed morning sensible heat flux crossover compared with the skin minus screen temperature crossover on calm days that is not consistently modelled. A similar lag is also seen in the latent heat flux (not shown). On the more stable day of 15 April,
the relatively large model errors in skin temperature are shown at night when the simulations for both JULES and UKV do not get cold enough. The skin minus soil temperatures demonstrate that the simulated crossovers are too early both morning and evening relative to the observations, indicating there is too much thermal coupling in the model between the surface and soil, as was indicated in Figure 2.

To estimate onset for the UKV, the time at which the turbulent mixing length, that is, the boundary layer depth, reaches 200 m is applied. The UKV boundary depth is a diagnostic parameter which applies a local Richardson number-based formulation to determine the length scale of turbulence (Lock et al., 2000). As the boundary-layer transitions from a stable to unstable boundary layer, a non-local component to the turbulent flux is computed based on a non-local K-profile for surface-driven turbulence. Therefore the UKV boundary depth can be considered the same as the length scale of turbulence. On average across all IOP days, the UKV convective onset is $67 \pm 39$ min too early compared with the estimated time of onset from the Halo. This emphasises that the model morning transition events, whether it is flux crossover or onset, are consistently too early.

3.2.2 Boundary-layer structure: vertical wind profiles

Figure 5 presents the vertical wind profiles in the lowest 600 m for each IOP and compares the radiosondes and the UKV during the morning transition. The comparisons are shown for the times closest to sunrise, buoyancy-flux crossover and convective onset. There is significant day-to-day variability in the structure of the vertical wind profiles, although a distinctive feature is a peak
in the vertical profile of wind velocity, indicative of the nocturnal low-level jet. At sunrise and morning crossover, the maximum in the UKV winds tended to be situated too high compared with the radiosonde. The mean height of the low-level jet is 270 ± 96 m, compared with 327 ± 120 m for the UKV. This may be suggestive of too much mixing in the stable boundary layer in the UKV.

The magnitude of the maximum wind speeds of the low-level jet are reasonably well simulated. The wind speed maxima in the low-level jet are seen at sunrise, except on 7 April, and range from 4 m s⁻¹ (25 May) to 17 m s⁻¹ (21 April, 31 July). The radiosonde profiles show a slowing of the peak wind velocity at convective onset, as turbulent mixing increases through surface heating, and momentum in the low-level jet is mixed downwards. At convective onset, the radiosonde profiles show a characteristic shape with uniform decreasing winds between 100 and 300 m, particularly on IOPs where wind shear dominates, for example 20–21 April.

On the days where wind shear dominates, the UKV over-speeds at convective onset, whereas on days where wind shear is less dominant — for example on 23 April and 1 June — the UKV captures the structure of the wind profile and the slowing of the peak wind velocity at convective onset. With regards to the near-surface winds, no mean difference in the 10-m wind speed during the morning transition is seen between the observations (2.4 ± 1.4 m s⁻¹) and the UKV (2.2 ± 1.2 m s⁻¹).
3.2.3 Boundary-layer structure: potential temperature profiles

The UKV minimum \( T_{\text{scrn}}^{\text{min}} \) and maximum \( T_{\text{scrn}}^{\text{max}} \) screen-level temperature errors for the IOP days are summarised in Table 1. The mean screen-level \( T_{\text{scrn}}^{\text{min}} \) error is 1.4°C across all IOP days, although on stable nights such as 23 April and 25 May, the \( T_{\text{scrn}}^{\text{min}} \) error is 3.7 and 5.2°C respectively. The UKV mean \( T_{\text{scrn}}^{\text{max}} \) error in the screen level is \(-0.8 \pm 0.4°C\) for the spring clear-sky IOP days, and \(0.8 \pm 0.6°C\) for summer IOP days. Although this suggests a seasonal dependence on screen-level temperature errors, the seasonal discussion in Section 3.1.2 demonstrated the long-standing cold daytime bias in the UKV. Large-scale warm advection partly contributed to the development of the very unusual warm daytime bias on June 24 and June 25, and these two IOP days dominate the summer mean \( T_{\text{scrn}}^{\text{max}} \) error.

Figure 6 presents potential temperature profiles for each IOP and compares the radiosondes and UKV during the morning transition. The comparisons are shown for the times closest to sunrise, buoyancy-flux crossover and convective onset. The observations from the 50 m tower are presented. Generally, good agreement is shown between the radiosonde and the tower observations within the variability in potential temperature over a 30-min period at the time of the radiosonde. At the time of sunrise and crossover, the UKV has a warm nighttime temperature bias.
which persists throughout the lowest 100 m and is not confined only to the near-surface layer. The bias dominates on nights with stable nocturnal boundary layer with a strong gradient in temperature confined to a shallow near-surface layer, suggesting model errors representing near-surface processes including radiative surface cooling.

The UKV potential temperature profile is generally too cold above the surface inversion. For example, on 7 April (15 April) the profiles indicate the surface inversion in the UKV is too weak between 100 and 200 m, and too cold immediately above the inversion, which extends to a depth of 1,500 m (500 m for 15 April). The radiosonde temperature profile nearest to the time of crossover shows that the inversion at the boundary-layer top (150–220 m) becomes sharper than at sunrise. This was noted by Lapworth (2006). It is most evident on IOPs with the most stable near-surface conditions (e.g., 7 April, 15 April, 25 May, 31 July), although it is not apparent on IOP days where shear-driven turbulence dominates (e.g., 20–22 April).

On 7 April, the largest cold air temperature bias develops in the UKV during the morning transition. A bias of $-2.4^\circ C$ is evident throughout the depth of the mixed layer at convective onset (0900 UTC). The boundary-layer depth on 7 April is the only IOP day where at convective onset the UKV is too shallow (180 m) compared with the radiosonde profile (315 m). By mid-afternoon on 7 April, the near-surface maximum temperature error has reduced ($T_{\text{max}}$ error of $-0.5^\circ C$), yet unfortunately radiosonde profiles are only available during the morning transition on this day. On April 15, in contrast, a cold bias of $-1.1^\circ C$ is evident throughout the depth of the mixed layer at convective onset, and the magnitude of the bias continues to develop leading to $T_{\text{max}}$ error of $-1.7^\circ C$. The inset panel in Figure 8 extends the potential temperature profiles beyond the morning transition, and demonstrates the evolving cold bias in the mixed layer. The buoyancy flux in the UKV for both of these days (Figure 3) was seen to be slightly too large and therefore not suggestive of errors in surface heating. The role of the entrainment from aloft is examined in Section 3.2.4.

During the consecutive four-day April IOP, the deepest boundary layers at convective onset (0800 UTC) were observed on 20 April and 21 April, with a mean depth of 518 m. The convective boundary layer has grown through the top of the former stable boundary at the time of onset. On 23 April, however, the boundary layer is shallower at convective onset (0900 UTC), still capped by strong temperature inversion. The potential temperature profiles at onset are suggestive of a warm bias, although by mid-afternoon a cold maximum temperature error has developed.

At convective onset (0900 UTC), the mean boundary-layer depth is $383 \pm 94$ m compared with $372 \pm 120$ m for the UKV, which indicates there is no systematic error on the IOP days. The capping inversion at onset tends to be diffuse over several model levels, and does not capture the sharp temperature gradient. On 20 April, for example, the observed temperature inversion at convective onset (0800 UTC) is approximately $4.5^\circ C$ and less than 50 m in depth. The UKV in contrast represents the temperature inversion across four model levels and over a depth of 250 m. Although the vertical resolution in the UKV at these heights is sub-100 m, the representation of sharp gradients is a challenge at the current resolution.

### 3.2.4 Entrainment

Angerveine et al. (2020) compared soundings during a morning transition with operational forecast profiles and hypothesised insufficient entrainment contributed to a simulated boundary layer which was too cold and shallow. In Figure 7, the surface and entrainment contributions of heat and moisture energy to the developing boundary layer are separated. Measurements of entrainment are sparse in literature and difficult to carry out. Entrained heat and moisture through the top of the surface mixed layer can be estimated using the LoCo mixing diagram approach (Santanello et al., 2018). The entrained heat and moisture can be inferred from the daytime co-evolution of 2-m potential temperature and humidity by accounting for the flux contributions of surface heat (sensible) and moisture (latent). Using this approach the inferred processes are the combined contribution from entrainment and large-scale advection of heat and moisture, but for brevity will be referred to here as entrainment.

The first example on 15 April is for a very stable boundary layer with low turbulence transitioning to a convective boundary layer, whilst the second example on 21 April is for a less stable, shear-driven turbulent boundary layer. The plots are for the same IOP days as presented in Figure 4. The surface energy time series Figure 7a,c shows the surface moisture is too large and highlights the excessive evapotranspiration which is a long-standing issue in the UKV. The entrainment energy time series in Figure 7b,d demonstrates the importance of entrained heat in warming the boundary layer during the morning transition. On both days, the observed surface heat contributes an average of 500 J·kg$^{-1}$ of the warming throughout the morning period, whereas the entrained heat accounts for over three times as much to the observed warming (1,800 J·kg$^{-1}$ at convective onset, 0900 UTC) on 15 April, and twice as much to the observed warming (1,000 J·kg$^{-1}$ at convective onset, 0800 UTC) on 21 April. The entrained heat in the UKV tends to be too small through the morning transition. The bias in entrained heat energy is at its
largest on IOP days that coincide with a cold potential temperature bias throughout the depth of the mixed layer at convective onset (e.g., 7 April, 15 April, 25 June, 31 July). On less stable days, with moderate-to-high wind shear (e.g., 20–22 April), the entrainment energy is better simulated.

Error in the derived entrainment energy based on the observations results from contribution of errors in the surface heat and moisture fluxes, and also temperature and humidity evolution at screen level. The role of larger-scale advection might also be significant yet remains unknown. Therefore the budget cannot discriminate between entrainment and advection. The long-standing UKV screen temperatures having too small a diurnal range associated with a cold daytime bias (Bush et al., 2020), however, indicates the error in the advection term may be minimal.

It is worth noting the disparity between the underestimation in the entrained heat that coincides with excessive deepening of the diagnosed boundary layer in the UKV. This suggests a more detailed investigation of the heat budget, taking account of the role of large-scale advection, is needed. This is out of the scope of the present study.

### 3.2.5 Lidar boundary-layer structure

Figures 8 and 9 contrast the boundary-layer evolution in terms of vertical velocity variance from the Halo Doppler lidar, for a spring (15 April) and summer (1 June) IOP, respectively. The minimum height of the lidar-derived vertical velocity statistics is approximately 100 m, and the variance data below this are from mast observations at 10, 25 and 50 m. The mean maximum boundary-layer depth during the spring is 1,431 ± 302 m and is generally well simulated in the UKV (1,405 ± 387 m). The mean summer maximum boundary-layer depth is 1,935 ± 370 m in the UKV and was too deep compared with the observations (1,755 ± 291 m).

On 15 April, the velocity variance for the convective period ranges from 0.5 to 0.75 m²·s⁻², with the maximum velocity variance of 1.3 m²·s⁻² coincident with the time of maximum boundary-layer depth. The maximum depth of the convective boundary layer is not particularly distinct from the depth of the nighttime residual layer of the previous day. The plot highlights that the boundary layer develops more quickly in the UKV during the morning transition. The boundary-layer depth at the time of the observed convective onset is 390 m, compared with 488 m for the UKV for the same time. The convective onset was 82 min earlier compared with the estimated time of onset from the Halo.

On 1 June, the boundary-layer evolution is well simulated by the UKV during the morning transition in terms of the potential temperature profile and boundary-layer depth. Observed convective onset on 1 June (0652 UTC) is 2 hrs earlier than on 15 April (0852 UTC), and maximum velocity variances of >2.0 m²·s⁻² are observed. The height of the convective boundary layer reaches a maximum depth of 1,890 m, in contrast with the UKV which reaches a depth of 2,170 m. The nighttime residual layer has a variable depth of 1,200–1,500 m. Equivalent plots of vertical velocity skewness (not shown) for 1 June help explain the
FIGURE 8  Evolution of vertical velocity variance with height from the Halo Doppler lidar on April 15, 2020. Data just above the surface are from the 10-, 25- and 50-m masts. The grey line shows the boundary-layer depth simulated by the UKV. Vertical red lines indicate the timing of sunrise and sunset. The insets show the potential temperature profiles on April 15 for (left) morning transition below 50 m, (middle) morning transition below 600 m and (right) the convective boundary layer [Colour figure can be viewed at wileyonlinelibrary.com]

FIGURE 9  Evolution of vertical velocity variance with height from the Halo Doppler lidar on June 1, 2020. Data just above the surface are from the 10-, 25- and 50-m masts. The grey line shows the boundary-layer depth simulated by the UKV. Vertical red lines indicate the timing of sunrise and sunset. The inset panels show the potential temperature profiles on June 1, 2020 for (left) morning transition below 50 m, (middle) morning transition below 600 m and (right) the convective boundary layer [Colour figure can be viewed at wileyonlinelibrary.com]
bi-modal distribution in the vertical velocity variance, and which is evident in Figure 9. The morning skewness values are strongly positive, indicating turbulence which is driven by surface heating, and correlated with maximum velocity variance of 2.0 m$^2$s$^{-2}$. Suppressed vertical velocity variances can be seen between 1300 and 1400 UTC, which is coincident with period of negative skewness, and is possibly indicative of turbulence generated by cloud-top cooling (Hogan et al., 2009). On 1 June, a cumulus-topped boundary layer develops shortly after convective onset, with maximum coverage of 4–5 Oktas of shallow cumulus resembling broken stratocumulus between 1300 and 1500 UTC; the impact on the down-welling shortwave flux can be seen in Figure 3.

The evolution of the boundary-layer structure for the double IOP on June 24–25, 2020 that captured an evening and morning transition is shown by the specific humidity retrievals of the DIAL lidar in Figures 10 and 11. The DIAL is able to sense the nocturnal residual layer and the growing convective layer which agrees with the developing UKV depth until mid-morning on both days. We have confidence that the DIAL boundary-layer depth is accurate from its agreement with the soundings, for example on 25 June in Figure 11 from a depth of 500 m at 0900 UTC up to 1,900 m at 1500 UTC. The UKV convective layer carries on deepening into the afternoon such that it overestimates the boundary-layer depth by about 450 m for most of the afternoon on 24 June and about the same on 25 June but with a peak error of 800 m at 1600 UTC.

The latter in particular seems like an unrealistic convective depth (i.e., 2,800 m) for summer in central England. The strength of the humidity gradients at the top of the boundary layer tend to be too weak in the UKV, especially so on 25 June. The UKV also tends to blur some of the structure such as the intermediate layer between 1,500 and 2,000 m at 1100 and 1200 UTC in Figure 11. Although DIAL captures the growing internal boundary layer well on these days, the evening collapse cannot be observed via the humidity structure. So we cannot criticise the timing of the UKV collapse. The marked higher humidity feature seen on 25 June at about 1630 UTC coincides with the convective collapse, but it has been identified as a sea-breeze front driven in from the coast (120 km away) on the east-southeasterly flow. This density current also had elevated aerosol loading (e.g., there was an increase in diffuse solar irradiance), a dramatic increase in the mean specific humidity between 2000 and 1000 UTC.
wind speed (from 3 to 7 m s$^{-1}$ at 10 m) and a small decrease in air temperature (1–2°C). This feature was also captured by the UKV forecast of the wind speed and specific humidity profiles.

4 | CONCLUSIONS

During a record-breaking spring and subsequent summer of 2020, over 100 radiosondes were launched at the Cardington research site on 11 IOP days. These captured ten morning transitions and one evening transition during clear-sky conditions, although on a few of the days cumulus clouds developed after the morning transitions. Even though the spring was unusually dry, the context of these IOPs was mostly typical in terms of the UKV forecast model performance. The UKV soil moisture is universally too wet, meaning there was too much evapotranspiration on a seasonal timescale, and Bowen ratios that are too small, that is dominated by latent heat rather than sensible heat. This can partially explain why UKV skin and screen-level temperatures biases are too warm at night and too cold during the day. Similar shortcomings were found in the JULES land-surface model during the drought year of 2018 in Osborne and Weedon (2021). The long-standing screen-level temperature errors in the UKV are noted by Bush et al. (2020). The surface temperature inversions are strongest, with model errors at screen level greatest under calm conditions. Two exceptions were June 24 and 25, 2020, where the UKV model predicted daytime screen maxima that were too warm. Model errors on clear days like this, where a cold bias would be expected, are rare and have been attributed to the representation of large-scale forcing in the model. Further analysis is out of the scope of this study.

Running the standalone surface scheme JULES forced by local meteorology shows improved Bowen ratios via a decrease in evapotranspiration, although a bias still exists. The skin temperature bias is also reduced relative to the operational UKV with a much better diurnal range. However, JULES soil temperatures, like the UKV, are too warm during the day with a diurnal range that is too large, suggesting the soil is coupled too tightly to the surface. The seasonal trends show that the mean late winter soil temperatures in the model are too cold, and the summertime temperatures are too warm, again suggesting there is too much coupling to the surface.
The UKV buoyancy-flux crossover times are slightly too early on average in the morning, yet very early in the evening particularly during the summer IOPs. The JULES evening buoyancy-flux crossovers agree well with observations, but the morning crossover was far too early. This is because there tended to be a lag between the observed skin minus screen temperature crossover and the turbulent heat flux crossover (both sensible and latent) in the morning that was not simulated consistently.

The developing boundary layer lifts too early in the UKV, with convective onset found to be 67 ± 39 min too early compared with the estimated time of onset from the Doppler lidar. Lidar comparisons also show that the UKV tends to develop boundary layers that are too deep by early afternoon, particular in the summer.

Although the magnitude of the maximum wind speeds of the low-level jet are reasonably well simulated, the radiosonde profiles on IOP days indicated that the height of UKV maximum wind speeds at sunrise and morning crossover of the low-level jet tends to be higher than the radiosonde. Profiles of observed potential temperature suggest the simulated stable boundary-layer surface inversion is too weak and the air is too cold immediately above the inversion. The importance of entrained heat in warming the boundary layer during the morning transition was demonstrated, and accounts for two-to-three times more than the surface heat flux to the observed warming. The entrained heat in the UKV tends to be too small through the morning transition, and on clear-sky days where the error is at its largest also coincides with IOP days where a cold potential temperature bias has developed throughout the depth of the mixed layer at convective onset.

Forecast verification metrics in general include screen temperature, wind, relative humidity and precipitation amounts, and are key parameters in evaluating numerical weather forecasts. Diagnosing model systematic errors associated with such parameters is rarely straightforward and often involves identifying existing compensating errors, such as within physical parameterisations, or arising through data assimilation. In this paper we have run both standalone JULES and the coupled UKV and compared the outputs with observations in an attempt to reconcile errors that lie within the surface scheme. For example, we have demonstrated there is too much thermal coupling between the skin and soil in both models where the daytime soil temperatures are too warm through spring and summer. Overestimation of evapotranspiration is evident in both JULES and coupled UKV, although JULES does show an improvement. It is not possible to scrutinise screen-level temperatures from JULES in the same manner as the coupled UKV as JULES is forced with 10-m meteorological observations. On a general note, optimising the land-surface model for a specific site does not necessarily directly translate into improvements when coupled to the atmosphere; therefore it is important to develop both surface schemes and coupled forecast models simultaneously.

The dataset presented in this paper has detailed observations of the land-surface and boundary-layer evolution during spring and summer, and we anticipate the dataset will be used in future assessment of science configurations in the Unified Model development. For example, the dataset is currently being used in the evaluation of the third Regional Atmosphere and Land (RAL3) science configuration. The numerical modelling activities so far have not isolated the role of land–atmosphere feedbacks in the model biases described. To this end, single-column and large-eddy simulation experiments will be utilised in the next stage of the analysis to explore the role of these processes in detail.

**OBTAINING THE UM**

The Met Office Unified Model is available for use under licence. A number of research organisations and national meteorological services use the UM in collaboration with the Met Office to undertake atmospheric process research, produce forecasts, develop the UM code and build and evaluate Earth system models. For further information on how to apply for a licence, see [https://www.metoffice.gov.uk/research/modelling-systems/unified-model](https://www.metoffice.gov.uk/research/modelling-systems/unified-model) (last access: September 30, 2021).

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**AUTHOR CONTRIBUTIONS**

**J. K. Brooke:** conceptualization; data curation; formal analysis; investigation; methodology; visualization; writing – original draft; writing – review and editing.

**S. R. Osborne:** conceptualization; data curation; formal analysis; investigation; methodology; visualization; writing – original draft; writing – review and editing.

**DATA AVAILABILITY**

Data from the simulations are archived at the Met Office and available for research use through the Centre for Environmental Data Analysis JASMIN platform (http://www.jasmin.ac.uk/); please contact the authors for details. Data from the Met Office Cardington site used in this study are archived and available for research use on request; please contact the authors for details.

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