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1 **Transition Periods in the Diurnally-Varying Atmospheric Boundary**
2 **Layer Over Land**

3

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20 **Abstract**

21 The atmospheric boundary layer undergoes transitions between stable and convective states. Over
22 land, in undisturbed conditions, these transitions occur daily in the morning and late afternoon or
23 early evening. Though less well studied and presenting more challenges than the fully stable and

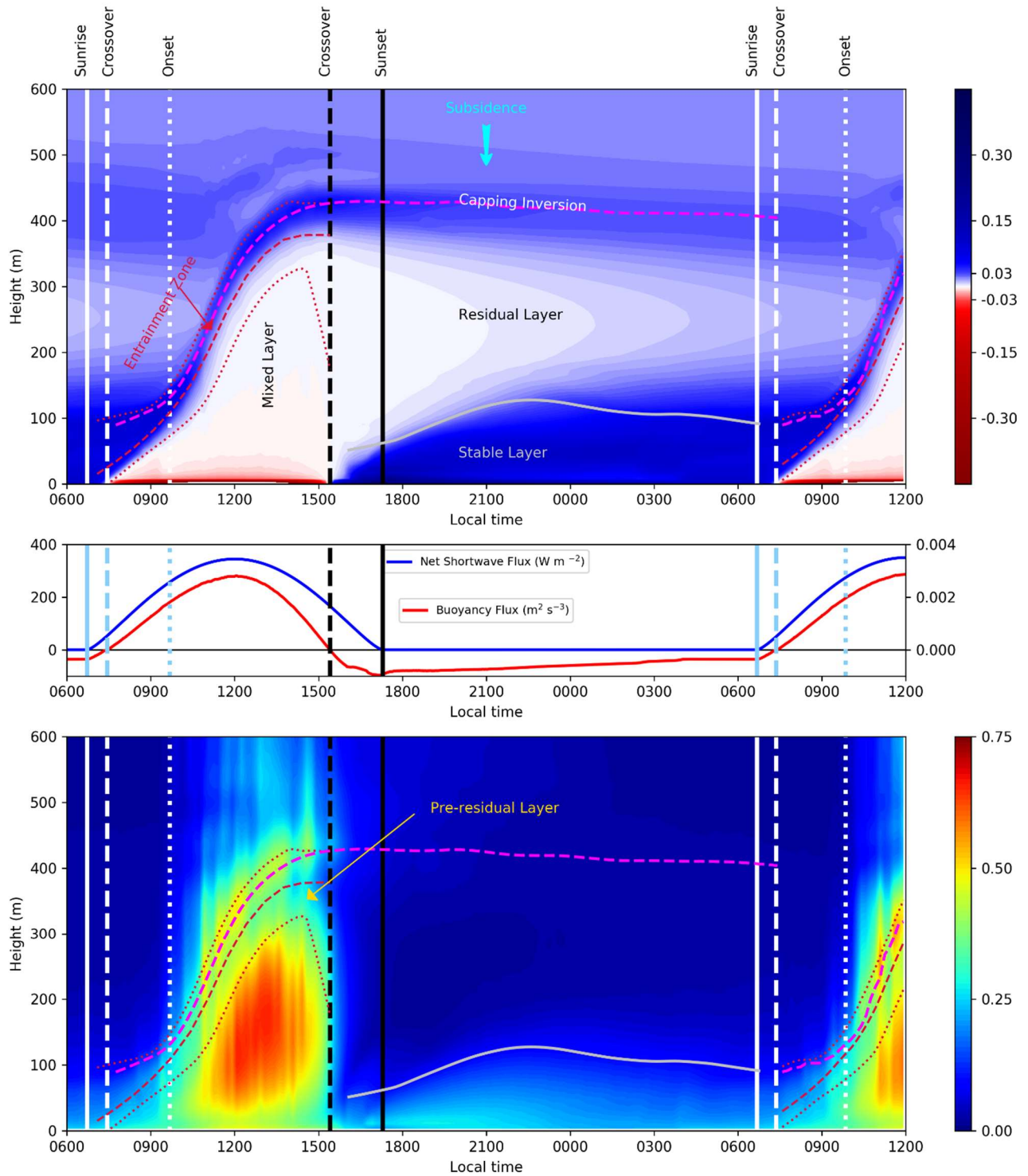
24 fully convective states, such transitions have been the subject of growing interest over the last few
25 decades. During transitions, all forcings are weak, and few simplifications are possible. Factors
26 such as terrain, radiation, advection, and subsidence can seldom be safely neglected. In this paper,
27 we review research on transitions over recent decades, with an emphasis on work published in
28 *Boundary-Layer Meteorology*. The review is brief and inevitably reflects the interests and views of
29 the authors.

30 **Keywords** Afternoon transition · Evening transition · Morning transition · Boundary-layer field
31 studies · Boundary-layer simulation

32 **1 Introduction**

33 Over land under clear skies and in the absence of any synoptic disturbance, the atmospheric
34 boundary layer (ABL) is often deep and strongly turbulent during the day, shallow and weakly
35 turbulent at night. In between, the ABL undergoes morning and afternoon/evening transitions.
36 Those transitions have been the subject of considerable research in the last 25 years, much of which
37 has been published in *Boundary-Layer Meteorology*. Figure 1 is a schematic showing a large-eddy
38 simulation (LES) of the diurnal cycle of an idealized mid-latitude ABL, see the caption for more
39 details. Shown are the times of sunrise, crossover (change of sign of the surface buoyancy flux),
40 onset of convective turbulence at a height of 200 m, rising of the entrainment zone and growth of
41 the mixed layer during the day, cessation of growth and eventual shrinkage during the afternoon,
42 and the times of crossover and sunset during the evening transition, followed by the growth of the
43 stable boundary layer (SBL), and the subsidence of the capping inversion overnight. These features
44 are discussed in more detail below. We do not provide precise definitions of the transitions
45 because they themselves evolve through the literature reviewed. Each section addresses definitions
46 as appropriate.

47 Motivations for the study of transitions are various. For air-quality forecasts and analyses, the
48 timing of changes in the depth of mixing with respect to time-varying emissions is important. The
49 evening transition is key to the evolution of the low-level jet (Smith et al. 2019), which often
50 influences severe weather. Fog and frost forecasting depend on understanding the evening
51 transition (Boutle et al. 2018; Price 2019).



52 **Fig. 1** A schematic illustration of the diurnal cycle of the ABL derived from a large-eddy simulation. **a** The evolution
 53 of the vertical gradient of the horizontally-averaged potential temperature, θ , with height above the surface. Vertical
 54 white lines indicate the times of sunrise, crossover and onset during the morning transition, while vertical black lines
 55 show the times of crossover and sunset during the evening transition. The light grey line marks the top of the stable

56 boundary layer, defined as the height at which the linearly-extrapolated horizontally-averaged buoyancy flux becomes
57 zero. The dashed magenta line marks the height of the elevated maximum of the gradient of θ . The crimson dashed line
58 indicates the height of the minimum of the buoyancy flux during the period when the boundary layer is unstable.
59 Crimson dotted lines mark the upper and lower boundaries of the entrainment zone during the same period, the former
60 defined as the lowest height above that of the minimum of the buoyancy flux at which the absolute value
61 of the buoyancy flux first falls below 5% of the surface value, and the latter as the lowest height at which the buoyancy
62 flux changes sign. The figure also shows the rising of the entrainment zone through the morning, the development of
63 the well-mixed layer, and, after the evening transition, the development of the nocturnal stable boundary layer, the
64 radiative stabilization of the residual layer, and the subsidence of the capping inversion. **b** The evolution of the net
65 downward shortwave flux at the surface (left axis) and the surface buoyancy flux (right axis). **c** The evolution of the
66 square root of the variance of the vertical velocity with height. Lines as in **a** and also showing the pre-residual layer.
67

68 Transitions are complex because many processes are roughly equally important, and the
69 simplifications and approximations that work well for the fully convective or the fully stable
70 boundary layer do not apply. In budget analyses, more terms must be retained, often including
71 direct radiative heating of the air, mixing due to shear, and advection. Stationarity may not apply.
72 Full descriptions of the surface energy balance and of the energy balance at the top of the ABL are
73 not included in this review, reflecting the contents of the literature we are considering.

74 Most of our increased understanding of the transition periods has been enabled by ground-based
75 remote sensing measurements (Wilczak et al. 1996), since measurements at the surface provide
76 only a limited perspective. The ability to view continuously the full depth of the boundary layer
77 and especially its upper region is critical to understanding transitions. Traditional tall-tower
78 measurements and observations taken by aircraft have also contributed.

79 While this should not be mistaken for a full description of the diverse possibilities, considering the
80 diurnal cycle is an improvement over looking only at the fully convective and fully stable states.
81 Throughout this review, we remain mindful of ways in which the real atmosphere frequently
82 departs from the idealization. Horizontal inhomogeneity of the surface, land use, soil and
83 vegetation properties has strong impacts. Coasts create land and sea breezes. Terrain induces
84 mountain and valley breeze systems. The daytime ABL is not always strongly turbulent, and the
85 nocturnal boundary layer is not always shallow.

86 The depth of the boundary layer is an ambiguous but important variable in discussions of the ABL
87 in general and transition periods in particular. When the convective boundary layer (CBL) is
88 topped by a well-defined increase in virtual potential temperature, its depth is fairly precisely
89 measured by any of several definitions (e.g. turbulence intensity, virtual-temperature profile,
90 virtual-temperature-flux profiles, or Richardson number) (LeMone et al. 2013). The depth of the
91 stable boundary layer is fundamentally more difficult to define, since the turbulence-intensity
92 profile may have a variety of shapes. Clouds add another level of complexity. Measurements of
93 the depth of an aerosol layer, for example by lidar, can be useful but must be carefully examined to
94 avoid mistaking the residual layer for the current boundary layer (Hogan et al. 2009; Tucker et al.
95 2009). The preferred technique is to measure the profile of vertical velocity variance, but this
96 approach can produce ambiguous results because of gravity waves near the boundary-layer top
97 (Lothon et al. 2009). Whether the ABL is stable or convective, some definitions can be readily
98 derived from NWP or LES model output, but rarely or with difficulty in observations. Throughout
99 this review, we will use the terms “boundary-layer depth” and “boundary-layer top” as if they were
100 well-defined, without tediously acknowledging all of this complexity every time the term comes
101 up.

102 The review proceeds as follows: First, the morning and afternoon/evening transitions as seen in
103 observations are described in Sects. 2 and 3. In Sect. 4, key processes involved are discussed,
104 including entrainment, radiation, advection and subsidence, length and time scales, and clouds.
105 Sections 5 and 6 describe the current understanding provided by LES and mesoscale numerical
106 weather prediction (NWP) models respectively, and the challenges that continue to arise in the
107 modelling of the diurnal cycle. Finally, we provide an outlook and thoughts on needs for ongoing
108 research.

109 **2 The Morning Transition in Observations**

110 The role of mixing from above the near-surface layer relative to heating from the surface was an
111 important question in early research on the morning transition. Data from the Cabauw tower in the
112 Netherlands were used by Driedonks (1982) to test analytical models of the early-morning
113 boundary-layer growth. He recognized that entrainment was important in many cases, and his
114 preferred model included a term proportional to the surface stress, based on theory provided by
115 Tennekes (1973). Aircraft data in Australia were used by Coulman (1978) to test predictions of
116 morning boundary-layer growth, with mixed results. He found that entrainment was probably
117 important, and that estimating some parameters with sufficient precision was not always possible,
118 particularly the inversion temperature gradient. Several expressions were provided by Garratt
119 (1992) along with a useful figure showing the change in potential-temperature profile for a
120 transition driven by surface heating vs. shear. A large dataset was examined by Angevine et al.
121 (2001) using data from the Cabauw tower (van Ulden et al. 1996) and radar wind-profiler data from
122 the Cabauw site and from the Flatland (Bondville, Illinois) site (Angevine et al. 1998). They
123 defined two milestones in course of the transition, “crossover” (when the surface buoyancy flux
124 first becomes positive) and “onset” (when convective turbulence is first detected at 200 m above

125 the ground). Sunrise was used as a reference time (Fig. 1). The most interesting discovery from
126 this study was that the time required to erode the stable layer from below, using the measured heat
127 flux and ignoring entrainment, was much longer than the time over which the layer was actually
128 eroded. In other words, the surface heat flux was insufficient to warm the stable layer in the
129 observed time. The conclusion was, and remains, that much of the warming arises from
130 entrainment at the top of the eroding stable layer. The data showed a situation more like Garratt's
131 Fig. 6.7b for a shear-driven transition. This was foreshadowed by Wyngaard (1983) in a brief
132 remark indicating that only a small heat flux was required to switch from a stable to a convective
133 layer.

134 We must caution that the definition of entrainment used in the above papers is imprecise.
135 Formally, entrainment is "The process by which turbulent fluid within a mixed layer incorporates
136 adjacent fluid that is nonturbulent, or much less turbulent." (American Meteorological Society
137 *Glossary of Meteorology*). However, in the context of the morning transition, within a statically
138 stable layer, the turbulent state of the air at each level is often not known. Some of the air that is
139 mixed to produce a near-neutrally-stratified layer may already have been turbulent and therefore
140 formally part of the boundary layer. The unknown turbulent state makes the formal definition
141 difficult to apply rigorously.

142 It should be noted that the measurements used by Angevine et al. (2001) were influenced by a
143 general effort in previous years to understand and exploit the capabilities of ground-based remote
144 sensing instruments (e.g. (Angevine et al. 1994; Beyrich et al. 1995; Carter et al. 1995).

145 A six-year dataset from the rural research site at Cardington in central England was used by
146 Lapworth (2006) to examine the morning and evening transitions, with an emphasis on wind speed
147 and forecasting rules. He confirmed the finding that most of the warming during the morning

148 transition was due to mixing from above. Before crossover, the warming is better described as
149 eddy diffusion rather than entrainment, since the boundary-layer top may not be rising. The near-
150 surface air temperature is controlled by the balance between radiative cooling and turbulent mixing,
151 which, under weak to moderate stability, is continuous. As the radiative cooling decreases, the
152 turbulent heat flux from above warms the near-surface layer, before the surface heat flux becomes
153 positive (upward).

154 Figure 2 shows observed soundings during a morning transition at Cardington. Entrainment is
155 evidenced by cooling in the upper part of the growing boundary layer in the progression of
156 observed soundings, especially between 0400 and 0700 UTC for example. Profiles from an
157 operational NWP model are also shown. The simulated boundary layer is consistently too cool and
158 too shallow, probably because of insufficient entrainment. Advection is important in most
159 individual cases, but is not observed, so attribution of the changes in profiles due to entrainment is
160 suggestive rather than conclusive; see Angevine et al. (2001) for more cases and discussion of their
161 variations.

162 An interesting feature of the morning transition is a peak in the time series of near-surface specific
163 humidity that is sometimes observed, e.g. (LeMone et al. 2002). When transpiration by vegetation
164 becomes significant while the mixed layer is still shallow, the humidity may exhibit a local
165 maximum, and then decline when the mixed layer grows. The morning peak is not seen in the
166 Flatland or Cabauw data of Angevine et al. (2001), but was seen at Cabauw on the day used by
167 Holtslag et al. (1995), and seems therefore to depend on the season and vegetation type. This has
168 been further studied for the afternoon/evening transition by Blumberg et al. (2019).

169 It is notable that multi-day averages are necessary to make understandable behaviour emerge from
170 data. Individual days exhibit a wide range of behaviour, dominated by advection, which is hardly

171 ever negligible (Baas et al. 2010), but sometimes averages out over multiple days (Betts et al.
172 1996). Even mild terrain can produce phenomena that overwhelm column-wise analyses, as was
173 found by Lenschow et al. (1979). Some influence of heterogeneity cannot be ruled out in the
174 Flatland and Cabauw cases, although the Flatland site in particular is extremely uniform for a land
175 site.

176 Further development of observational equipment and techniques continues to add to our knowledge
177 of the morning transition. Wildmann et al. (2015) conducted a study with a small remotely-piloted
178 aircraft, emphasizing the scaling behaviour of variances compared with theory, and they include a
179 thorough discussion of uncertainties and challenges.

180 **3 Afternoon and Evening Transition in Observations**

181 From a surface perspective, the boundary layer changes from unstable to stable in the “evening,”
182 some time near sunset, but looking at the boundary-layer top, changes happen much earlier in the
183 day. This has led to a lack of clarity and consensus in terminology. In this review, we refer to the
184 afternoon or evening transition according to the terms used by the cited papers. Many of these
185 terms are illustrated in Fig. 1. In general, “afternoon transition” denotes the period when the surface
186 buoyancy flux decreases (from midday to sunset), and “evening transition” refers to the period near
187 sunset including the change of sign of the surface buoyancy flux. A sunset-relative frame of
188 reference was used by Wingo et al. (2015). In complex terrain, local sunset is a key time,
189 especially when slopes are steep (Nadeau et al. 2011). In Fig. 1, we have indicated sunset and
190 crossover, newly defined here for the evening transition as the time when the surface buoyancy flux
191 changes sign from upward to downward, the opposite of morning crossover.

192 Mahrt (1981) made an early observational study of the “early evening” transition based on data
193 from the Wangara (Australia) study, emphasizing the behavior of the wind and the low-level jet.
194 Mahrt observed that the wind changes began well before sunset. Many studies of the evening
195 transition consider slightly or moderately complex terrain. Acevedo et al. (2001) and Mahrt (2017)
196 emphasized the evolution of temperature and wind through three phases. The cooling is small at the
197 start (before zero flux), greater in the second phase (around zero net radiation), and decreases again
198 after sunset. The wind decelerates, changes to calm or local drainage, and evolves toward regional
199 drainage or synoptic flow, as also shown by LeMone et al. (2003), Busse et al. (2012), Román-
200 Cascón et al. (2015), and Nilsson et al. (2016a). Those phases all depend on the characteristics of
201 the studied sites. A peak in the time series of specific humidity may also be present, and is
202 dependent on the season, location, and crop behavior (Blumberg et al. 2019).

203 Recent works have highlighted new features like the counter-gradient flux (Blay-Carreras et al.
204 2014a), caused by a delay between the times when the surface flux changes sign and when the
205 temperature gradient changes sign.. It has been shown to be partly related to the eddy turnover
206 time, and dependent on the vegetation cover and height (Jensen et al. 2016). Also the lifted
207 temperature minimum is worth mentioning (Blay-Carreras et al. 2015), that is, a minimum of
208 temperature at 0.1-1 m above the surface.

209 A detailed study by Grant (1997) used data from tethered balloon equipment at the Cardington site
210 to describe the latter part of the transition period, after the surface buoyancy flux had changed to
211 negative (downward). Grant showed profiles and spectra to explore the applicability of proposed
212 scaling frameworks during the evening transition. Further data from the same site and similar
213 equipment were used by Edwards et al. (2006). Observations of turbulent kinetic energy near the
214 surface were used by Nadeau et al. (2011), who fitted a simple analytical model which

215 approximately showed a t^2 power-law decay of turbulent kinetic energy (TKE) early in the
216 transition, and t^6 later (t being time).

217 Observations through the depth of the boundary layer in the afternoon and evening were rare until
218 ground-based remote sensing instruments became available. The first full analysis was undertaken
219 by Grimsdell et al. (2002) using wind-profiling radar observations in conjunction with surface-flux
220 measurements. They found that the height of the boundary layer decreases in the afternoon,
221 starting some hours before sunset. The explanation is this: as solar heating of the surface begins to
222 decline, and the air continues to warm, thermals become less vigorous. Since the bulk of the
223 boundary layer is roughly adiabatic, the weaker thermals can still reach the boundary-layer top at
224 first. Meanwhile, several other processes are at work, and all of these tend to reduce the height of
225 thermals. Subsidence directly lowers the boundary-layer top. Subsidence, shear-driven
226 entrainment, and direct radiative heating all act to stabilize the upper part of the boundary layer. In
227 the case of subsidence, regime selection is at work: If large-scale upward motion is present, clouds
228 and precipitation often form, and these days are seldom considered for further case study analysis.
229 Direct measures of turbulence show the decreasing top height, while indirect measures, particularly
230 aerosol layer height, may not. Since the boundary layer is defined as being in recent contact with
231 the surface (Stull 1988; LeMone et al. 2019), the direct turbulence measurement is preferred
232 (Tucker et al. 2009; Barlow et al. 2011; Harvey et al. 2013; Manninen et al. 2018).

233 Lidar observations of the boundary layer over London, UK covering several full diurnal cycles
234 were published by Barlow et al. (2011). The urban boundary layer has additional complexity not
235 seen in simpler situations, but some of its characteristics are similar. Long lidar datasets from two
236 sites were analyzed by Manninen et al. (2018).

237 The Boundary Layer Late Afternoon and Sunset Turbulence (BLLAST) field study was focused on
238 these transitions. It provided a comprehensive set of measurements by all types of instruments and
239 platforms. Unlike most earlier studies, the BLLAST study embraced the challenge of complex
240 terrain and heterogeneous land use. Lothon et al. (2014) provides an overview, early results, and a
241 rich bibliography. The project started with careful definitions, and in addition to measurements,
242 incorporated theory and modelling. Frequent and continuous observations through the whole depth
243 of the boundary layer allowed more complete analysis than had previously been possible. The
244 decay of TKE in a vertically resolved sense, and the ways in which turbulent length scales change,
245 were particular foci.

246 In terms of the scaling of turbulence within the boundary layer, the BLLAST study results from
247 aircraft (Lothon et al. 2014; Darbieu et al. 2015) and near-surface measurements (Nilsson et al.
248 2016a; Nilsson et al. 2016b) indicated that statistics of turbulence on days with weak shear are
249 quasi-stationary until late afternoon. Shear delays the decay, or even prevents decay if strong
250 enough (Pino et al. 2006). While the decay remains slow enough for statistics to be quasi-
251 stationary, the turbulence structure does not evolve substantially, and remains scaled with the
252 boundary-layer depth. Later in the afternoon, the decay accelerates and the turbulence structure
253 departs from CBL scaling. The decay was found to start from the top (Darbieu et al. 2015),
254 although this strongly depends on the shear at surface and at boundary layer top. Consistent with
255 this, Nilsson et al. (2016b) used a simple model to identify a region of reduced turbulence that
256 appears earlier in the upper part of the boundary layer, before the stable layer develops, and
257 expands downward with time. They described this as a “pre-residual” layer. Scaling by the
258 buoyancy flux clearly becomes questionable as it approaches zero.

259 **4 Processes of Boundary-Layer Transitions**

260 The afternoon transition is nearly unforced; all contributing processes are weak. The morning
261 transition is forced by the rising sun, but this forcing is surprisingly weak as well. Many processes
262 influence the boundary layer through the diurnal cycle, and the key challenge to understanding is
263 the need to include all relevant processes. Turbulent eddy sizes and time scales change.
264 Entrainment at the boundary-layer top and mixing of entrained air play important roles. Advection
265 is almost always important when examining individual days and hours, and is difficult to estimate.
266 Direct radiative heating or cooling of the air (as opposed to heating or cooling of the surface) is
267 often small, but cannot be safely neglected during transitions. When fluxes or turbulent energy are
268 changing rapidly, methods of analysis that neglect time rates of change may fail. Clouds may
269 affect the profiles of mean and turbulent quantities, especially near the top of the boundary layer.
270 This section describes these processes in more detail.

271 4.1 Eddy Length and Time Scales

272 Time scales are fundamental to transitions, in fact to the idea of transition itself. Some relevant
273 time scales include the eddy turnover time, the time required for the boundary-layer height to
274 change by a given fraction, and the time required for the surface buoyancy flux to change by a
275 given fraction (or to reverse sign). Around midday, the eddy turnover time (10–20 min) is much
276 shorter than the time scale of the evolution of the boundary-layer height, which changes only
277 slowly. In the morning growth phase, however, the boundary layer may grow as rapidly as a
278 thermal rises (Sorbjan 1997).

279 The time scale for turbulence to decay is important in the afternoon and evening, and depends on
280 the eddy size (Pino et al. 2006; Lothon et al. 2014). Larger eddies decay more slowly, leading to
281 an effective increase in the length scale above the surface layer.

282 The impact of variation in surface fluxes on time scales in the CBL was explored by Van Driel et
283 al. (2011) in an idealized framework. They made several interesting findings: the time scale for
284 changes in the boundary-layer depth is very long, of order 10 h, because it is controlled by the
285 entrainment velocity. A phase lag between the changing surface flux and the mixed-layer height is
286 due to these differing time scales. For rapidly varying fluxes, the production and dissipation of
287 TKE become important. Their simulations included subsidence and radiative cooling specifically
288 tuned to produce a steady boundary-layer depth and temperature. The application of these results
289 to transitions is challenging, because, for example, the entrainment velocity during the morning
290 transition can be very large (shortening the time scale). . Subsidence and other processes that are
291 roughly in balance at midday, allowing the boundary-layer depth to remain approximately constant,
292 become unbalanced when the surface flux permanently weakens in the afternoon, resulting in rapid
293 changes of the boundary-layer depth. Nonetheless, the concept of competing time scales and
294 transfer functions introduced by Van Driel et al. (2011) is a promising framework for future
295 exploration.

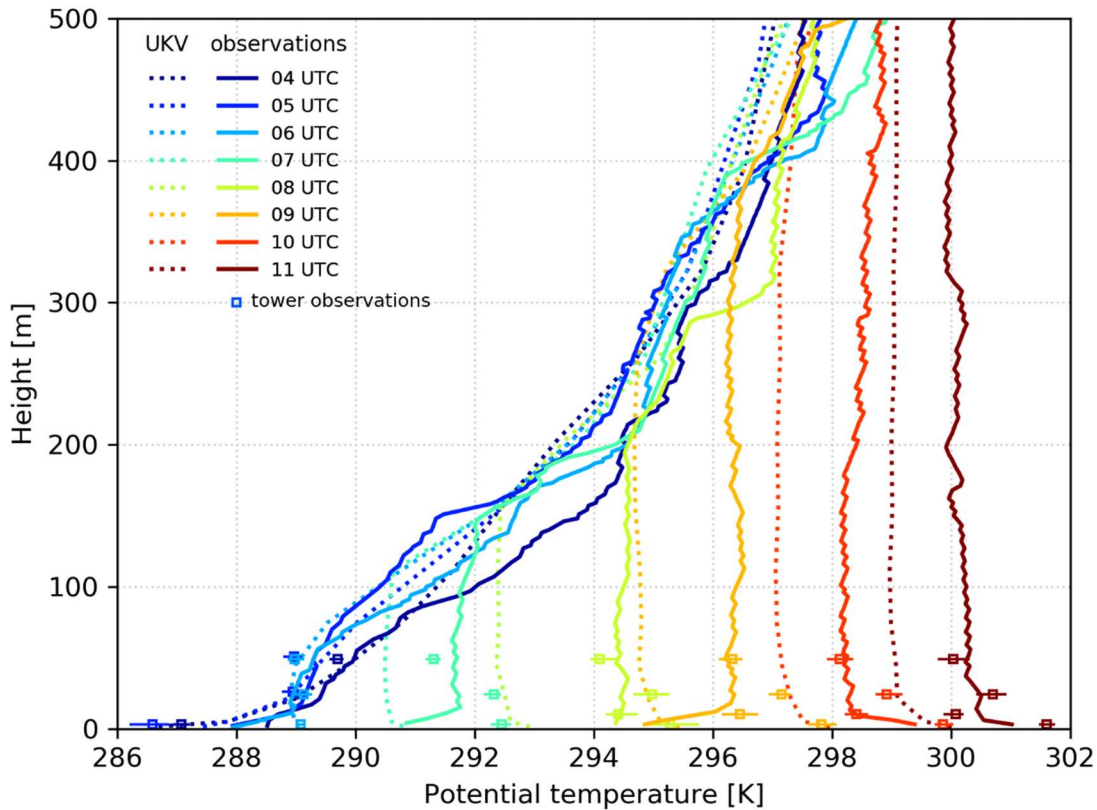
296 Another important time scale from a practical point of view is the averaging time required in
297 processing turbulence measurements, whether from a sonic anemometer or a remote sensor. If
298 important changes happen during the averaging time, the measurements are hard to interpret (Mahrt
299 1981; Grant 1997). On the other hand, averaging over too short a time gives noisy and excessively
300 uncertain results, and can lead to systematic underestimation of fluxes.

301 Length scales and their variation are also vital to understanding boundary-layer behaviour. With
302 the dataset from the BLLAST study, Darbieu et al. (2015) found that the length scales of turbulence
303 remain scaled with the boundary-layer depth until late afternoon (more precisely until about 0.75
304 times the decay-forcing time scale). Later, they observed a sharp decrease of the length scale in the

305 surface layer, and an increase of the length scale above. The changes occurred first at the top of the
306 boundary layer, although this depends on shear.

307 4.2 Entrainment

308 As described in sections 2 and 3 above and sections 5 and 6 below, entrainment at the boundary-
309 layer top is critical to the transitions. In the morning, most of the warming during the transition is
310 due to mixing from above, either from within the upper part of the boundary layer or from
311 entrainment. In the afternoon, entrainment is one of the processes that causes warming in the upper
312 part of the boundary layer and thereby reduces the depth of mixing. Entrainment is a complex
313 topic that deserves a whole review by itself; see Angevine (2008) for further discussion and
314 references. It is particularly worth noting that entrainment relationships depend strongly on the
315 interpretive framework, e.g. Liu et al. (2017). Figure 2 shows the importance of entrainment in the
316 morning transition, and insufficient entrainment appears to be the cause of the cool and shallow
317 bias in the morning boundary layer in the UKV NWP model.



318 **Fig. 2** Profiles of potential temperature at the Cardington field site during the clear-sky morning of 19 July 2016.

319 Radiosonde ascents launched on each hour are shown as solid lines, together with tower sensors (square points) at 2 m,
 320 10 m, 25 m and 50 m; horizontal bars on the mast data show variations of +/- 5 mins centred on each hour. Model data
 321 (dotted lines) are the instantaneous hourly forecast profiles from the Met Office UKV operational 1.5 km grid spacing
 322 model, initialized at 0300 UTC.

323

324 A nearly adiabatic residual layer often exists above the nocturnal boundary layer. In the morning,
 325 when the growing boundary layer reaches the base of the residual layer, its growth rate increases
 326 until the residual-layer top is reached (Deardorff 1974; Sorbjan 1996; Blay-Carreras et al. 2014b).
 327 It is useful to note, however, that the residual layer is not simply the remains of the previous day's
 328 mixed layer at the same location; because of differential advection, the mixed layer found in the
 329 morning above a certain site was likely formed somewhere else.

330 4.3 Advection and Subsidence

331 Experience with setting up cases for single-column model (SCM) and LES comparisons has shown
332 that advection is rarely negligible, even at nearly ideal sites. Measurements of advection are
333 difficult or impossible. The best that can be done is to estimate advection terms from observations
334 at levels that are thought not to be affected by turbulent mixing, or to use output from mesoscale
335 models (Barr et al. 1997; Angevine 1999). Estimates of subsidence are at least as problematic.
336 Efforts involving many dropsondes have succeeded in producing subsidence measurements for the
337 simplest possible situation over water (Bony et al. 2018). It should be noted, however, that
338 comparisons between LES and other types of models using the same forcings are not subject to
339 these uncertainties.

340 Baas et al. (2010) discuss the issue of specifying forcing (advection) for SCM studies. They
341 demonstrate strong impacts on the formation of the low-level jet (an evening-transition issue) from
342 different plausible specifications of advection. These differences make it difficult to determine
343 which vertical mixing formulation is best. They also show that a composite case, constructed by
344 averaging several similar cases, removes some of the “noise” and allows for clearer conclusions.
345 Different choices of forcing, all from plausible sources, had a strong impact on shallow cumulus
346 simulations shown by Angevine et al. (2018). Averaging over many cases was done by Angevine et
347 al. (2001) to understand the morning transition. Systematic advection can mislead the analyst
348 despite averaging. For example, in the central United States, the prevailing wind direction is
349 southerly (warm and moist) in summer and northerly in winter, so advection cannot be expected to
350 average out within a season.

351 The effects of subsidence (Blay-Carreras et al. 2014b) and advection (Pietersen et al. 2015) were
352 studied for some BLLAST cases. At the BLLAST site, there was a persistent diurnal pattern of

353 subsidence, induced by the nearby mountain range and accentuated by moist convection over the
354 mountains.

355 4.4 Radiation

356 Vertical turbulent heat fluxes driven by net radiation at the surface play a substantial role in the
357 diurnal evolution of the boundary layer, but direct atmospheric radiative heating is also relevant.
358 Radiative cooling at cloud top can generate turbulence and affect the entrainment rate (Stevens
359 2002; de Lozar et al. 2015). Clear-sky radiative processes have received less attention, but can play
360 a significant role in transitions. Longwave radiative cooling erodes the capping inversion and
361 accelerates the growth of the boundary layer through the morning transition (Edwards et al. 2014).
362 Radiative cooling near the surface during the evening transition stabilizes the developing stable
363 boundary layer and suppresses turbulence (Garratt et al. 1981; Ha et al. 2003). In near-calm
364 conditions radiation can be the dominant process determining the temperature profile (Savijarvi
365 2006; Edwards 2009). Longwave radiative cooling rates vary on short length scales near the
366 surface, so poor vertical resolution can produce incorrect results in either observations or models
367 (Ha et al. 2003; Edwards 2009; Steeneveld et al. 2010).

368 Aerosols in the boundary layer may have large effects on radiative heating or cooling when the
369 aerosol loading is significant or the aerosols are strongly absorbing. Aerosol layers within or above
370 the boundary layer reduce the net radiation reaching the surface (Barbaro et al. 2014; Liu et al.
371 2019), thereby affecting transition timing. Aerosols within the boundary layer or the residual layer
372 can directly heat those layers enough to affect timing as well. The effects are complex and subtle
373 because the sun is low in the sky during transitions. Barbaro et al. (2014) give a detailed discussion
374 of these effects, supported by model studies of two observed cases.

375 4.5 Clouds

376 Clouds are very common at the top of the boundary layer. The University of Washington Cloud
377 Atlas (<https://atmos.uw.edu/CloudMap/index.html>; accessed 15 March 2019) shows that 26% of
378 land and 55% of the oceans is covered with low cloud on an annual average basis. Most of these
379 clouds are in the boundary layer. Nonetheless, strongly cloud-influenced conditions are commonly
380 avoided or selected against in boundary-layer field research. During the morning transition,
381 Angevine et al. (2001) noted that the most extreme day-to-day variations in timing were found on
382 days with cloud. In their data from the Flatland site, cloud-free days showed more rapid increases
383 in temperature than cloudy days, but no systematic differences were found between cloud-free and
384 cloud-influenced days at Cabauw. In the afternoon, Grimsdell et al. (2002) found little systematic
385 influence of clouds. Over water, where the diurnal cycle is damped, Pennell et al. (1974) and
386 LeMone (1978) found little influence of clouds on subcloud-layer TKE budgets. Brown et al.
387 (2002) found that subcloud-layer buoyancy-flux profiles were insensitive to cloud cover. In all of
388 these analyses, regime selection probably prevented strong cloud influences.

389 An interesting and complex but not uncommon case is presented by Hogan et al. (2009). A thin
390 stratocumulus layer in the early morning is replaced by a convective boundary layer with shallow
391 cumulus. Radiative cooling in the stratocumulus layer produces mixing, which must affect the
392 timing of the morning transition. Similar cases probably arise commonly, and deserve further
393 exploration. Clouds and their effects are also seen in the data of Harvey et al. (2013) and
394 Manninen et al. (2018).

395 **5 Transitions in large-eddy simulations and direct numerical simulations**

396 Large-eddy simulations of transitions are challenging because they require both fine resolution and
397 large domains, and are thus computationally expensive (Beare 2008). Several LES studies of
398 transitions have been done, however. Beare (2008) produced simulations of the morning transition
399 by starting with a small, finely resolved domain and expanding to a larger, more coarsely resolved
400 domain partway through the transition. These simulations started at the time of heat-flux
401 crossover, so they describe the late phase of the transition. Beare found a shallow convective
402 boundary layer with penetrating, entraining updrafts under the remnant stable layer, which he
403 named the “mixed CBL-SBL state”. The morning transition in the third GEWEX Atmospheric
404 Boundary Layer Studies (GABLS3) case was simulated by Edwards et al. (2014), who
405 demonstrated the effect of radiation in weakening the capping inversion, thus accelerating the
406 growth of the CBL.

407 Evening-transition LESs have a longer history. It seems to have been assumed that the resolution
408 suitable for convective boundary layers was also sufficient to represent the transition. Nieuwstadt
409 et al. (1986) used an abrupt cut-off of the surface heat flux. Sorbjan (1997) introduced and showed
410 the importance of a smoother reduction, which introduces a second timescale. Pino et al. (2006)
411 used LES to show that horizontal length scales become larger during the afternoon transition. Beare
412 et al. (2006) may have been the first to simulate the evening transition at fine resolution, with more
413 realistic fluxes. Early simulations had simplified representations of non-turbulent processes such
414 as radiation, which was shown to be important in light winds by Edwards (2009). Few LES models
415 have an interactive land surface, and the choice of surface boundary condition (flux or temperature)
416 has been controversial. Details of initialization and forcing were found to be important, e.g. Kumar
417 et al. (2010). Observed surface temperature was used by Rizza et al. (2013) to study the evening
418 transition, and they found TKE-decay power-law exponents of -2 early and -6 late in the

419 transition, similar to those observed by Nadeau et al. (2011). El Guernaoui et al. (2019) explored
420 the TKE decay in the context of interacting time scales.

421 Both morning and afternoon transitions were simulated by Conzemius et al. (2008) for one day in
422 Oklahoma. In their LES model, they simulated very large relative contributions of shear-driven
423 entrainment to the total warming during the morning and afternoon. These are caused by the
424 realistic nature of the case. Profiles and TKE budgets are shown. Basu et al. (2008) simulated a
425 diurnal cycle from Wangara, emphasizing the low-level jet that appeared after the evening
426 transition. Their evening transition was abrupt, probably due to the lack of radiation and/or
427 subsidence in their setup. Full diurnal cycles were also simulated by Sharma et al. (2017), again
428 finding an abrupt evening transition without a representation of radiation and with no subsidence.
429 Both morning and evening transitions were included in the LES of Blay-Carreras et al. (2014b),
430 who emphasized the importance of the profiles above the growing boundary layer in the morning
431 and the forcing by subsidence in the afternoon. Darbieu et al. (2015) used a BLLAST case to set
432 the surface flux and initial conditions, and NWP model outputs to account for total advection
433 (including subsidence). Although the amount of TKE was different than in observations, and the
434 real conditions were much more complex than in the LES, the way that the TKE and the turbulence
435 structure evolved were very similar to the observations, showing a certain universality of the decay,
436 well reproduced by LES.

437 Figure 1 illustrates the evolving sophistication of coupled LES models. It was performed using the
438 large-eddy Met Office-NERC Cloud model (MONC) (Brown et al. 2015) coupled to the Joint UK
439 Land Environment Simulator (JULES) land-surface model (Best et al. 2011) and the Suite of
440 Community Radiative Transfer codes based on Edwards and Slingo (SOCRATES) radiation code
441 (Manners et al. 2015). The specifications of the simulation were chosen for visual clarity in the

442 figure. The latitude was set as 57° N and a geostrophic wind speed of 8 m s^{-1} was prescribed. The
443 land surface was specified as short grass and the initial atmospheric profiles of temperature and
444 specific humidity were chosen to avoid the formation of cloud. The period shown begins on 2
445 March.

446 Direct numerical simulation (DNS) has recently become practical for application to atmospheric
447 problems. Mellado et al. (2016) simulated a free-encroachment case relevant to the late morning
448 transition. Fedorovich et al. (2017) performed DNS simulations relevant to the formation of the
449 low-level jet at the end of the evening transition.

450 **6 Transitions in Mesoscale Numerical Models**

451 Operational NWP models simulate transitions and the diurnal cycle regularly over large areas.
452 These models inherently simulate subsidence and advection. Radiation, boundary-layer physics,
453 clouds, and interaction of the atmosphere with the surface are parametrized. Climate-model
454 simulations likewise include many diurnal cycles under all kinds of conditions. However, research
455 specifically targeting transitions in mesoscale models is relatively rare in the literature. The
456 GEWEX Atmospheric Boundary Layer Study (GABLS) is a long-running project to improve the
457 simulation of the boundary layer in models and has proceeded through increasingly realistic
458 intercomparisons. Two of the three GABLS studies looked at diurnal cycles over land, and provide
459 the most complete view of mesoscale model performance during the transitions. The results show
460 considerable room for improvement. An overview of GABLS is given by Holtslag et al. (2013).

461 For GABLS2, Svensson et al. (2011) simulated a diurnal cycle from the CASES-99 experiment in
462 Kansas, USA. Many single-column models participated. Their behaviour during the transitions
463 was diverse. One conclusion was that choosing the most realistic or most useful boundary

464 conditions and forcings is both difficult and critical. Specified surface temperatures were used as
465 the bottom boundary condition, and it became clear that the fluxes produced by the different
466 models for the same atmospheric and surface temperatures diverged. Heat flux during the morning
467 transition was too small and the boundary layer grew too slowly. This may have been due to too
468 little entrainment (see above), which also led to wind speeds that were too low. Different kinds of
469 boundary-layer scheme in the models had different kinds of errors, but no type of scheme was
470 clearly better for the whole diurnal cycle.

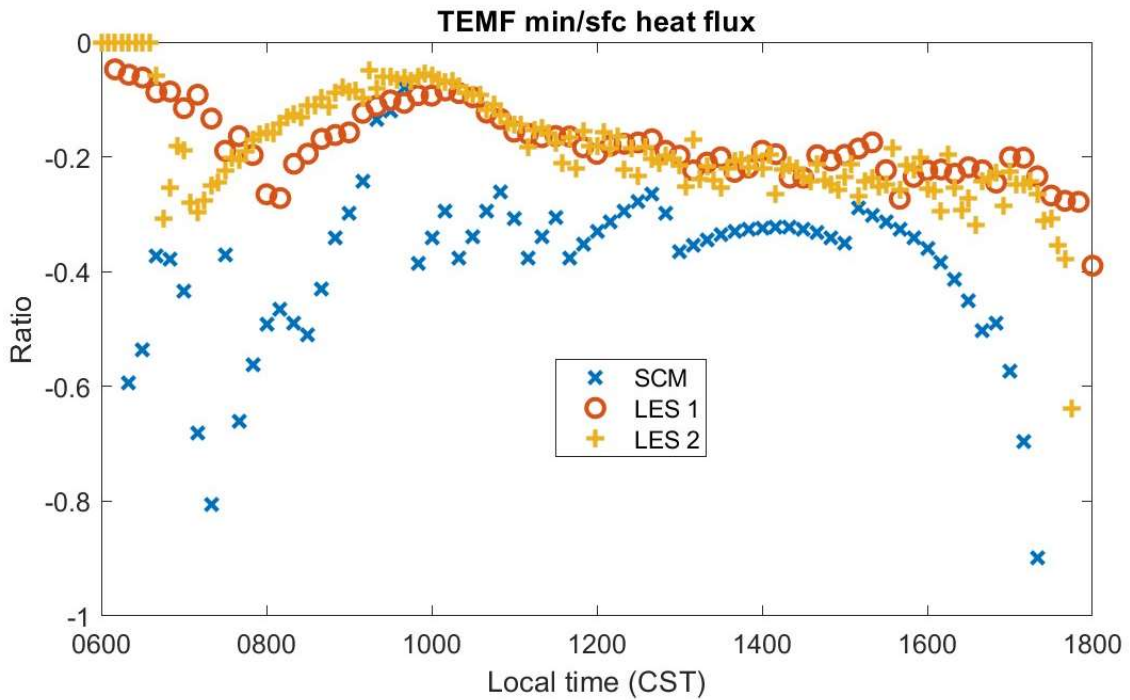
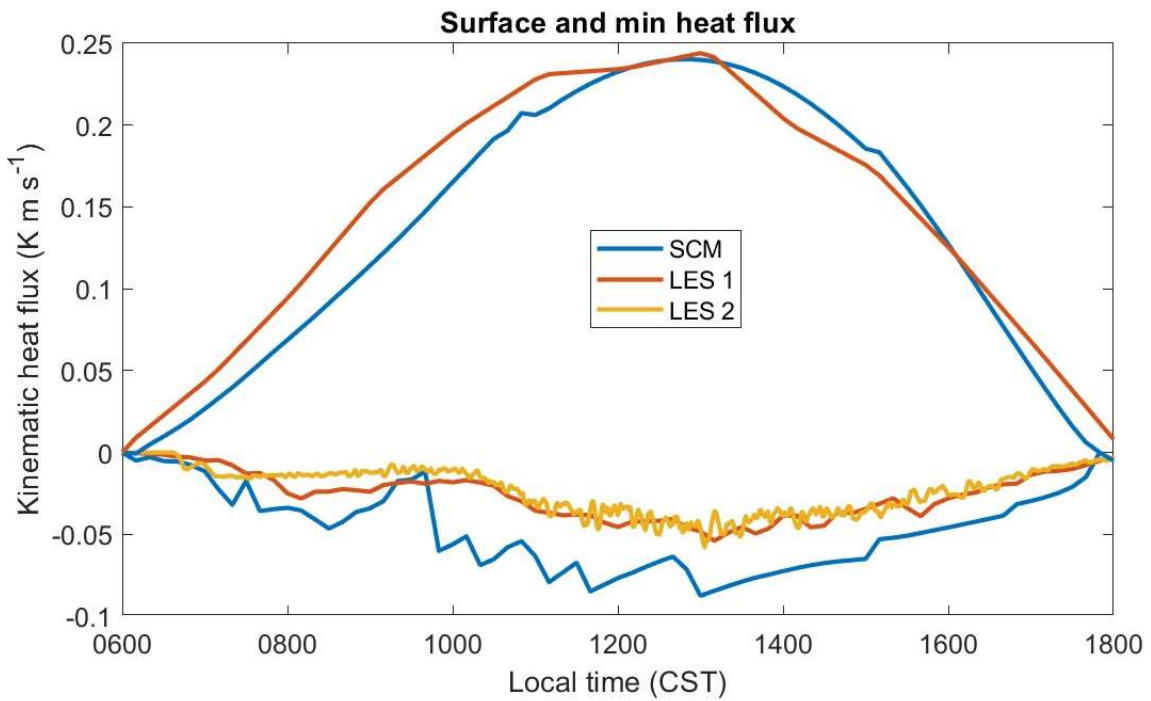
471 The GABLS3 study (Bosveld et al. 2014) benefited from the lessons learned in GABLS2. The case
472 was derived from measurements at Cabauw, The Netherlands. The single-column models included
473 land-surface and radiation modules, removing the need for specified lower boundary conditions.
474 Several of the conclusions of GABLS3 involved the land-surface formulations, which dominated
475 the transitions and stable boundary-layer behaviour. Another important conclusion was the
476 importance of longwave radiation from within the stable boundary layer (see above). Evening-
477 transition timing varied by 1.5 h among the models. In the morning, the simulated time of
478 crossover from downward to upward surface flux varied by 2 hours between models, and the
479 interval from crossover to onset of turbulence at 200 m was about half an hour too long.

480 Using a single-column model, Edwards et al. (2006) simulated two evening-transition cases, as
481 observed at Cardington. Observed surface temperature was prescribed to the model as a lower
482 boundary condition. They emphasized the importance of specifying realistic advection. In both
483 cases, the simulated boundary layer was too shallow during the transition. They found limited
484 sensitivity to the resolution or the parameterization of turbulence in the stable boundary layer. At
485 the moderate wind speeds in these cases, longwave radiative cooling was a significant, but not the
486 dominant process in the development of the nocturnal boundary layer. Edwards (2009) described a

487 case of very light winds (about 1 m s^{-1}) and strong surface cooling and showed that radiative
488 cooling to the surface was the dominant process determining the evolution of the near-surface air
489 temperature through the transition. It was noted that the assumptions underlying standard surface-
490 layer similarity theory failed during the transition, leading to cold biases in operational forecasts in
491 such conditions.

492 Extensive observations of temperature profiles were examined by van Hooijdonk et al. (2017) to
493 study the growth rate of the surface-based temperature inversion, and compared with bulk and
494 single-column models of varying complexity. They found that the simplest models could not
495 reproduce the observations, because history effects were important. This reinforces the idea that
496 the boundary layer is not in equilibrium during some parts of the evening transition.

497 Mesoscale modelling for the BLLAST campaign also pointed out the importance of the land
498 surface. Lothon et al. (2014) and Couvreux et al. (2016) found variations of the vegetation among
499 different measurement sites to be an important contribution to variations in afternoon transition
500 timing. Angevine et al. (2014) devised a technique for making soil moisture match a limited-area
501 model better to produce more realistic simulations. Couvreux et al. (2016) found that a finer-
502 resolution NWP model (around 2 km grid spacing) resolves the vertical structures better than the
503 coarser models (around 10 km), in particular a well-defined maximum of the water-vapour mixing
504 ratio that arises during the evening in the upper part of the residual layer due to fine-scale
505 advection. The finer- resolution model also provides a good simulation of the diurnal variability of
506 the turbulent kinetic energy, including its late afternoon decay, while the coarser model still shows
507 the right order of magnitude.



508 **Fig. 3** (Top) surface and minimum sensible heat flux from a single-column simulation with the TEMF scheme and two
 509 LES models for 6 June 2015 at the ARM SGP site in Oklahoma. Potential temperature flux is shown for SCM and
 510 LES1, virtual potential temperature flux for LES2. See text for details. (Bottom) Ratio of minimum flux to surface
 511 flux, one way of calculating the entrainment ratio.

512 Diagnosing entrainment in mesoscale or NWP models is difficult because they rarely produce flux-
513 profile output by default. One exception is the Total Energy Mass Flux (TEMF) boundary-layer
514 scheme available in the Weather Research and Forecast model (WRF) (Angevine et al. 2018).
515 Figure 3 shows the diurnal cycle of the surface kinematic heat flux and the minimum flux in the
516 profile within the boundary layer for 6 June 2015 at the Atmospheric Radiation Measurement
517 Southern Great Plains site in Oklahoma. Along with the single-column model running TEMF, two
518 LESs of that day are also shown. The ratio of entrainment flux to surface flux is commonly used in
519 parameterizations, even though it is strictly applicable only to pure free convection driven from the
520 surface, and different methods of calculation yield different ratios (Conzemius et al. 2006; Liu et al.
521 2017). With those caveats in mind, the lower portion of Fig. 3 shows the entrainment ratio through
522 the diurnal cycle. TEMF has large ratios of the magnitude of entrainment flux to that of the surface
523 flux early and late in the day, consistent with some of the observational studies. TEMF also has a
524 relatively large ratio in midday. Breaks in the curve are due to limited vertical resolution in the
525 SCM. Both LESs show some of the same tendency to larger ratios in the morning and afternoon
526 transition periods. LES1 is from the U.S. Department of Energy LES Atmospheric Radiation
527 Measurement Symbiotic Simulation and Observation Workflow (LASSO) archive, simulation code
528 0030 of 6 June 2015 (Gustafson et al. 2016a; Gustafson et al. 2016b). It has a 30-m vertical grid
529 spacing. LES2 was created by Glenn et al. (2019) using the System for Atmospheric Modeling
530 (SAM), with a vertical grid spacing 20 m above 600 m.

531 **7 Outlook and Research Needs**

532 The progress in understanding ABL transitions over land in the last several decades has laid a firm
533 foundation for future research. Much could be learned from better-resolved LES modeling through
534 transitions, including all relevant processes, such as interactive radiation and an interactive ground

535 surface. These processes are already represented in mesoscale models. Better understanding of
536 cloudy boundary layers could also come from LES, and from mining existing datasets for cases that
537 may have been discarded from previous analyses. Entrainment is still an open topic, especially its
538 variation with wind shear, aerosol, and surface inhomogeneity. Non-column effects, such as
539 terrain, coasts, and surface heterogeneity, must be considered as model grid sizes decrease. The
540 proper representation of transitioning boundary layers in the grey zone or terra incognita (Wyngaard
541 2004; Efstathiou et al. 2016) requires exploration. The full energy balance at the surface and at the
542 top of the boundary layer should be taken into account, including soil, canopy, and vegetation
543 contributions to heat and moisture. Evolving measurement techniques can also contribute. Because
544 real transitional boundary layers are influenced by even small surface or terrain variations, multi-
545 scale networks of observations have the most promise to provide useful insight.

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