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1 **Boundary Layer Meteorology-50th Anniversary Volume**

2 **Boundary-layer flow over complex topography**

3

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8 **Abstract** We review developments in the field of boundary-layer flow over complex
9 topography, focussing on the period from 1970 to the present day. The review follows two
10 parallel strands: the impact of hills on flow in the atmospheric boundary layer and gravity-
11 driven flows on hill slopes initiated by heating or cooling of the surface. For each strand we
12 consider the understanding that has resulted from analytic theory before moving to more
13 realistic numerical computation, initially using turbulence closure models and, more recently,
14 eddy-resolving schemes. Next we review the field experiments and the physical models that
15 have contributed to present understanding in both strands. For the period 1970-2000 with
16 hindsight we can link major advances in theory and modelling to the key papers that
17 announced them but for the last two decades we have cast the net wider to ensure that we
18 have not missed steps that eventually will be seen as critical. Two important new themes are
19 given prominence in the 2000-2020 period. The first is flow over hills covered with tall plant
20 canopies. The presence of a canopy changes the flow in important ways both when the flow
21 is nearly neutral and also when it is stably stratified, forming a link between our two main
22 strands. The second is the use of eddy-resolving models as vehicles to bring together hill
23 flows and gravity-driven flows in a unified description of complex terrain meteorology.

24 **Keywords** Boundary-layer meteorology . Turbulence . Complex topography. Gravity-
25 driven flows . Turbulence Modelling.

26

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27 **1 Introduction**

28 It is over 70 years since the similarity theory of Monin and Obukhov (MOST) was developed
29 in the USSR and over 50 years since experiments on the sweeping plains of South Eastern
30 Australia and the mid-West of the USA validated it for the first time (Monin and Obukhov
31 1954; Dyer, 1967; Kaimal and Wyngaard, 1990). Since then MOST has occupied a central
32 position in boundary-layer meteorology and forms the foundation for climate and weather
33 models used around the world. It is the point of departure for new theory and the benchmark
34 against which experimental data are assessed. However, at the most fundamental level
35 MOST applies only to flat homogeneous terrain and most of the earth's surface is not flat but
36 topographically complex on scales from hillocks to mountains so it is no surprise that the
37 study of airflow over hills and valleys has a history as long as MOST.

38 Everyday experience continually reminds us of the influence of topography on winds
39 and climate and the motivation for studying and quantifying these influences comes from all
40 areas of atmospheric science and engineering. The unique characteristics of mountain
41 weather and climate were well known by the early 20th century and the first general theories
42 of large scale hill and mountain flows in the 1940s were driven by the need to include
43 topographic drag in numerical weather models. Much of the work we will review in this
44 paper however concerns smaller scales. In agriculture and forestry we are concerned with
45 topographic influence on wind damage to crops and forests, on seed and pathogen dispersal,
46 water use and CO₂ uptake, and on heat stress and frost formation. In wind engineering we
47 must allow for locally enhanced forces on structures and at larger scale the impact of severe
48 hill and mountain flows on urban regions or airfields. Air quality is strongly conditioned by
49 steep terrain and pollutants can be concentrated in separated flow regions or by downslope
50 gravity currents. More recently, two new applications have guided the research effort. First,
51 in the rapidly growing wind energy industry, wind farm designers seek topographic
52 enhancement of their atmospheric feedstock but need to avoid the impact of increased
53 turbulence on their turbines. Second, in developments that link back to the earliest general
54 hill flow theories, the demand for greatly increased spatial resolution in weather and climate
55 predictions is now entraining small scale hill-flow dynamics into weather and climate
56 models.

57 Faced with the challenge of reviewing this field over the 50 year span that Boundary
58 Layer Meteorology has recorded it, an author naturally starts by looking at earlier reviews.
59 These in themselves form a useful chronology of the development of conceptual ideas and

60 theory as well as a history of the experiments that prompted and tested them. A list starting
61 with Hunt (1980) then Taylor et al. (1987), Finnigan (1988) and Blumen (1990), augmented
62 by the substantial preambles in the contemporary papers devoted to theoretical or
63 experimental advances and culminating in the splendid syncretic survey by Wood (2000),
64 allows us with hindsight to distinguish distinct stages in the development of our current
65 understanding.

66 From roughly 1940-1960, hill flow studies were dominated by the theoretical
67 advances of Queney (1948) and Scorer (1949). Their efforts were motivated by weather
68 forecasting and focussed on the larger scale effects of disturbances to the stratified
69 troposphere and, although Scorer's later papers acknowledged the importance of the turbulent
70 boundary layer, those early analyses in the main assumed laminar flow. Starting in the early
71 1970s, the next fifteen years were marked by an upsurge of interest in boundary layer flow
72 over hills initiated by the numerical models of Peter Taylor and his Canadian colleagues and,
73 almost simultaneously, by the analytic linear modelling led by Julian Hunt and announced by
74 the scene-setting analysis of Jackson and Hunt (1975). That first step by Jackson and Hunt
75 was quickly followed by extensions and alternative analytic approaches with critical
76 advances being made by Mason and Sykes (1979), Sykes (1980) and Hunt et al. (1988a). For
77 a time, the parallel developments in numerical modelling were strongly influenced by the
78 advances in analytic theory and both were provoked or guided by experiments over isolated
79 hills and in wind tunnels and towing tanks. Over a similar 1975-1990 period the study of
80 buoyancy-driven katabatic and anabatic flows on simple extensive hill slopes proceeded
81 alongside the study of isolated hill flows but with surprisingly little overlap.

82 From 1990-2010, as more accurate and mature numerical schemes became widely
83 available, driven in part by the needs of the growing wind energy industry, the community
84 began to address the problem of extending the by-now extensive corpus of knowledge
85 derived from simple hill configurations to complex topography on a larger scale, including
86 the use of eddy resolving models of coarse resolution as foreshadowed by Wood (2000). The
87 last two decades from 2000-2020 have also seen analytic and numerical advances in the study
88 of hills covered with tall plant canopies. New physics appears in this situation and has
89 provided insights into some existing theoretical problems as well as essential input to the
90 increasingly critical problem of measuring carbon exchange between the terrestrial biosphere
91 and atmosphere. Eddy-resolving numerical models have increasingly come to the fore, being
92 used both with coarse resolution to characterize regional flow over complex mountain
93 topography and at fine scales to model the detailed turbulence dynamics on boundary layer

94 hills. With major field campaigns like MAP-Riviera, Matterhorn-X and Perdigao (see
95 Section 6), this period has also seen a revival of the community's appetite for the large scale
96 cooperative field experiments that provided so much impetus in the 1980s. We hope that the
97 events that defined these stages will become clear in the sections below.

98 **1.1 Scope and Structure of this Review**

99 This review has several goals. First, it attempts to survey the developments in theory and
100 understanding of boundary layer flows over topography from the sudden spontaneous
101 upsurge of effort in the mid 1970s to around 2000. We will highlight key steps and the papers
102 in which they were recorded. Second, it will attempt to do the same for the parallel field of
103 gravity-driven flows on complex topography, although we note that experiments and
104 theoretical advances in this field lagged the study of flow over simple hills by around a
105 decade. Surprisingly most reviews thus far have treated only one or the other of these subject
106 areas or, when they are both studied as part of a single major campaign, they are discussed
107 separately (e.g., Blumen, 1990). Third, we will discuss the advances that have been made in
108 each area in the two decades 2000-2020. Over this period especially we will highlight new
109 areas of interest such as the radical changes to hill and slope flows that occur when a tall
110 plant canopy is present and the growing use of eddy resolving numerical models in both hill
111 and slope flow simulation.

112 Since time has not yet winnowed the large number of papers from this period down to
113 those that will eventually be seen to be key, we will provide a more comprehensive guide to
114 the publications that seem to us at this time to be most worthy of note. We apologize in
115 advance to authors, whom we should have included and didn't. Where we can, we will
116 assess the strengths of weaknesses of the state of the art with the aim of highlighting
117 shortcomings and new research directions. Finally we will briefly touch upon the latest
118 efforts at bringing together hill and slope flows as components of general complex terrain
119 meteorology. These involve bridging the gap between eddy resolving models of resolution
120 coarse enough to calculate the large scale flow patterns forced by topography and those with
121 grids fine enough to resolve the near-surface 3D turbulence, whose interaction with hills and
122 valleys produces those same emergent large scale responses. This latter is a subject that is
123 now ripe for a review of its own.

124 We have organized our review around a set of key topics rather than following a strict
125 chronology. The topic areas are, insofar as is possible, self-contained so that they can be used
126 as independent sources of reference but there is substantial cross referencing between
127 sections. The key topic areas we shall cover are:

159 affected by the influence of buoyancy on the turbulent stresses). The ratio U_0/N is typically
 160 1km so flow over hills of kilometre length scale may be free of strong buoyancy influences
 161 for much of the daylight hours and when winds are strong. In contrast, since the H/L ratios of
 162 natural topography are roughly 1:10, hills of much larger scale, say $L \sim 5\text{km}$, are subject to
 163 direct buoyancy influences all of the time.

164 A different set of scaling arguments obtains when we consider local, gravity-driven
 165 flows on topographic slopes. Katabatic or anabatic flows, respectively down or uphill, are
 166 driven by the component of the vertical hydrostatic pressure gradient resolved along the hill
 167 slope. This will dominate the hydrodynamic pressure gradient, generated by the deflection of
 168 the ABL flow over the hill, when the Froude number F_L formed from the ratio of these two
 169 pressure gradients is small i.e. $F_L \ll 1$. In this case F_L is defined by Belcher et al. (2012) as,

$$170 \quad F_L = \frac{U_0}{L} \left[\frac{g}{\Theta_0} \frac{\Delta\bar{\theta}}{L} \right]^{-1/2} \quad (3)$$

171 where $\Delta\bar{\theta}$ is the characteristic temperature difference between the thin hot or cool layer on
 172 the sloping hill surface and the temperature of the surrounding ABL at the same geopotential
 173 height. Cool surface layers are commonly generated by radiative cooling at night and the
 174 resulting katabatic flows are important in many practical circumstances. We discuss them in
 175 detail in Section 4. Anabatic flows are usually less important in daytime as strong surface
 176 heating usually generates convection cells spanning the ABL and the ‘footprints’ of these
 177 cells impose transient but relatively strong horizontal winds at the surface over scales of
 178 kilometres or more (e.g. Patton et al., 2016) but when synoptic winds are light, topography of
 179 appropriate scale can interact with convection to organize these convection cells (Dörnbrack
 180 and Schumann, 1993). Anabatic flows generated by combustion of vegetation of course are a
 181 critical issue in wildfire prediction and control but we do not address that issue here.

182 Large areas of significant topography such as the European pre-Alps, the Pyrenees or
 183 the foothills of the US Rocky Mountains and Sierra mix all scales promiscuously.
 184 Nevertheless, most of the theoretical development in hill flow studies thus far has
 185 concentrated on isolated hills in well-defined approach flows. In contrast, buoyancy-driven
 186 flows on simple slopes have been studied both in idealized conditions and in the context of
 187 large scale hill-valley wind systems. We will look first at the developments in the theory and
 188 understanding of boundary-layer hill flows and, in contrast to the somewhat restrictive
 189 criteria given above, we will adopt a more relaxed definition and include any hill-flow
 190 phenomena that are determined primarily by processes located within the boundary layer.

191

192 **1.3 Notation and Definitions**

193 For convenient reference, notation and definitions of parameters used throughout this
194 review are gathered here.

195 **Hill Flows-Coordinates and velocity components**

196 x, y, z denote components of a right handed coordinate frame, which can be Cartesian,
197 surface-following or streamline according to context. x is taken in the streamwise and z the
198 surface normal direction. u, v, w are the corresponding velocity components

199 $u = U + u'$; $w = W + w'$; $v = V + v'$, where capitals denote time means and primes turbulent
200 fluctuations. An overbar indicates time averaging so $U = \bar{u}$, etc. $\bar{\tau} = -\overline{u'w'}$ is the mean
201 kinematic turbulent shear stress.

202 θ is potential temperature and θ_v virtual potential temperature. Θ_0 is a reference
203 temperature in degrees K and g the acceleration of gravity so g/Θ_0 is the thermal expansion
204 coefficient of the atmosphere. $\overline{w'\theta'}$ is the eddy flux of θ in the z direction and
205 $\theta^* = -\overline{w'\theta'}/u^*$ is the temperature scale of the logarithmic temperature profile.

206

207 c represents a generic passive scalar with $C = \bar{c}$; $c' = c - C$ and $\bar{f}_c = \overline{w'c'}$ the eddy flux of c .

208 Δ denotes perturbations to mean quantities induced by the hill so $\Delta U, \Delta W, \Delta\bar{\theta}, \Delta\tau, \Delta\bar{f}_c$
209 are the perturbations in mean streamwise and surface normal velocity, potential temperature,
210 shear stress, and scalar flux, respectively.

211 $U_B(z)$ denotes the undisturbed mean wind profile approaching a hill. In analytic theory

212 $U_B(z)$ is usually assumed to follow the logarithmic law with displacement height, d and

213 roughness length, z_0 . $u^* = \sqrt{-\overline{u'w'}}$ is the friction velocity of the undisturbed upwind flow
214 and κ is von Karman's constant

215 U_0 is the characteristic mean streamwise velocity scale. In analytic hill flow models

216 $U_0 = U_B(h_m)$, where h_m is the middle layer height defined in Figure 2.

217 $\Delta S(x, y, z)$ denotes the mean velocity speed up over the hill and ΔS_{\max} is its maximum
218 value, usually located above the hill top.

219 **Hill length and Velocity Scales**

220 H is hill height and L hill length, defined as the horizontal distance from the summit to the
 221 half-height point. Hill slope is defined as H/L . In the analytic theory of HLR88, the depth of
 222 the shear stress layer is h_i and of the middle layer h_m (Fig. 2) and $U_0 = U_B(h_m)$.

223 Buoyancy parameters

224 The Brunt–Vaisala frequency is defined as $N = \sqrt{\frac{g}{\Theta_0} \frac{\partial \theta}{\partial z}}$ and the Froude Number based on

225 hill length as $F_L = \frac{U_0}{NL}$. We also use when appropriate a Froude Number based on hill height,

226 $F_H = \frac{U_0}{NH}$ and a Froude Number taken as the square root of F_p , the ratio of the hydrodynamic

227 and hydrostatic pressure gradients on a slope. $\sqrt{F_p} = F_L = \frac{U_0}{L} \left[\frac{g}{\Theta_0} \frac{\Delta \bar{\theta}}{L} \right]^{-1/2}$, which we identify

228 with F_L .

229 Canopy parameters

230 h_c is the canopy height, C_d the dimensionless drag coefficient of the foliage and a the foliage
 231 area per unit volume. The momentum absorption length scale of the canopy is formed from

232 average values of C_d and a , $L_c = (C_d a)^{-1}$.

233 Gravity-Driven Slope Flows-Coordinates and velocity components

234 s, n denote along-slope and slope-normal coordinates with s positive downslope. Along-
 235 slope and slope-normal velocity components are identified with a subscript s or n respectively
 236 as in u_s and w_n . The slope makes a positive angle α with the local geopotential surface.

237 H_s , L_s and U_s are the depth, length and velocity scales, respectively, of a gravity current.

238 Where appropriate, e.g. when the gravity current is on a hill, we assume $H_s = L$ and $U_s = U_0$.

239 $\Delta \bar{\theta}$ is the characteristic potential temperature difference between a gravity current and the
 240 ambient air. \bar{e} denotes the local turbulent kinetic energy; Λ is the local Obukhov length in
 241 the s - n coordinate frame.

242

243 2. Boundary layer flow over isolated hills- Theory and mathematical 244 modelling

245

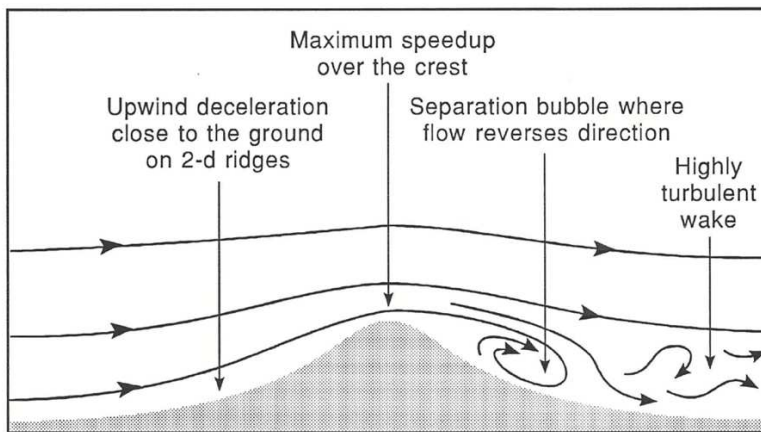
246 2.1 Analytic models

247 The general features of airflow over isolated 2D hills or ridges in strong winds have long
248 been known (Fig. 1a). Along streamlines close to the surface the wind decelerates slightly at
249 the foot of the hill before accelerating to peak at the crest then decelerates in the lee. If the
250 hill is steep enough downwind a ‘separation bubble’ forms with reversed mean flow and
251 enhanced turbulence levels. Whether a separation bubble forms or not, a turbulent wake with
252 a general velocity deficit extends downwind. On axisymmetric hills, the upwind deceleration
253 region disappears and is replaced by a region of flow divergence as the streamlines separate
254 to pass around the hill. These general features are shown in a different way as perturbations
255 to an upwind logarithmic velocity profile in Figure 1b. The task of research has been to
256 quantify these changes for the wide range of hill shapes and atmospheric conditions found in
257 nature.

258 Our conceptual understanding of boundary-layer flow over hills has been hugely
259 influenced by the seminal work of Jackson and Hunt (1975) (henceforth JH75) and a series of
260 follow up papers. JH75 developed an analytic model for neutral flow over a rough hill by
261 linearising the equations of motion about a background logarithmic wind profile. Their key
262 insight was that the flow can be divided into two layers: a thin inner layer of depth h_i near the
263 surface where perturbations to the turbulent Reynolds stress terms are important, and an outer
264 layer, where the flow perturbations are essentially inviscid (Fig. 2). Scaling analysis yields
265 different leading order terms in each layer and leads to a separate analytical solution in each
266 layer. These can then be matched asymptotically to give an overall solution. The various
267 assumptions made to approximate the equations limit the solutions to low hills where the
268 slope is small and where the roughness length z_0 is also small compared to the characteristic
269 width of the hill.

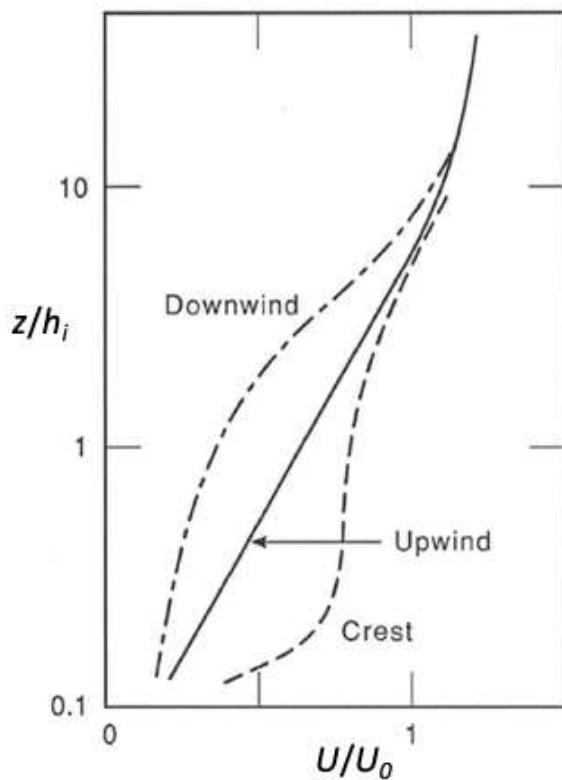
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281 a



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283 b



284

285 **Fig 1.** Main features of flow over low hills. a) flow over a 2D ridge. b) perturbation to an upstream logarithmic
286 mean velocity profile. See Figure 2 for definitions of h_i and U_0 . (After Kaimal and Finnigan, 1994)

287

288 The key result from the JH75 analysis was that there is much greater speed up than
289 the magnitude of the hill slope would suggest. For example, if the slope of the hill is 1/5 then
290 the wind speed at the crest is increased by a factor of about 1.5 (Belcher and Hunt, 1998)
291 This is because the hydrodynamic pressure gradient that deflects the wind over the hill is
292 proportional to the square of the approach velocity $U_B(z)$ well above the inner layer but the

293 resulting strong pressure gradient then acts on the much slower moving air layers close to the
 294 surface to produce a greater proportional acceleration. According to convention, the
 295 fractional speed-up is defined as,

$$296 \quad \Delta S(x,z) = \frac{U(x,z) - U_B(z)}{U_B(z)} \quad (4)$$

297 where the origin of the x coordinate is the hill top. The peak speed-up, ΔS_{\max} occurs some
 298 distance above the hill top (Fig. 1b) and predicting its location and magnitude was the focus
 299 of much early research and served as a convenient test of theory against measurements. In
 300 JH75 and other analytic theories, ΔS was intimately related to the thickness of the inner layer,
 301 h_i and the several competing definitions of h_i that emerged from different approaches to
 302 scaling the equations of motion were tested against field data (Taylor et al. 1987) and wind
 303 tunnel simulations. For example, Britter, et al. (1981) compared the JH75 theory to the
 304 numerical model of Taylor (1977a,b) and to wind tunnel experiments (Finnigan, 1988). For a
 305 hill with a slope of $H/L=0.26$ the mean velocity predicted by the analytic model compared
 306 well with experimental data and numerical results on the upwind slope and near the crest of
 307 the hill, but not in the wake region.

308 Taylor et al. (1987) reviewed the application of JH75 to low hills and recommended
 309 using the speed-up formula,

$$310 \quad \Delta S_{\max} = \left[\frac{U_B^2(L)}{U_B^2(h_i)} \right] \cdot \left[\frac{H}{L} \right] \zeta(x,z) \quad (5)$$

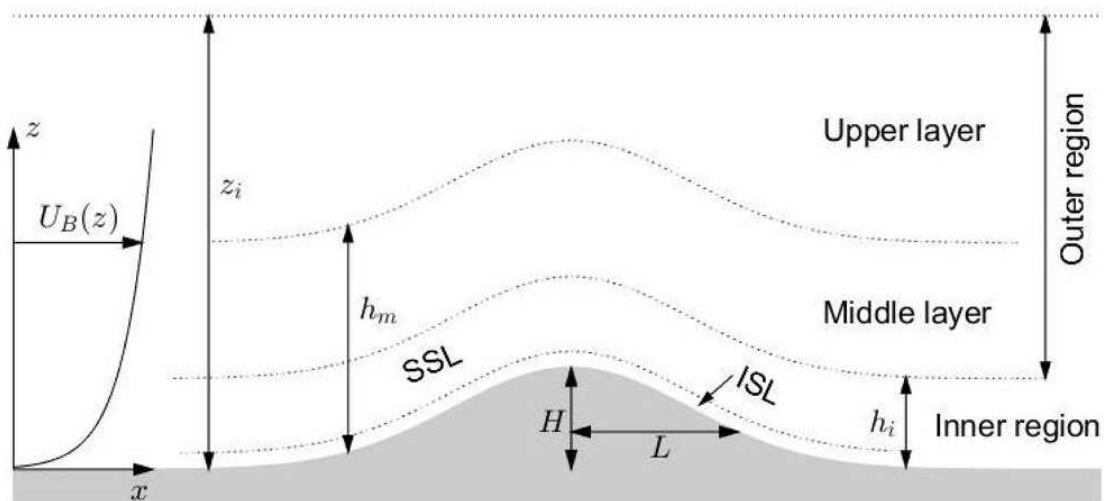
312 where ζ is a function that factors the precise shape of the hill into the formula. For values of
 313 H, L and z_0 typical of low hills with smooth contours, $\zeta \approx 1$ so that $\Delta S_{\max} \sim 2H/L$, the
 314 value recommended for 2D ridges in JH75. Taylor and Lee (1987) went on to formulate
 315 simple guidelines for wind speed changes over small scale topographic features. Over 3D
 316 axially symmetric hills they recommended $\Delta S_{\max} \sim 1.6H/L$, over 2D escarpments,
 317 $\Delta S_{\max} \sim 0.8H/L$ and over 2D ridges, $\Delta S_{\max} \sim 2H/L$. However, Finnigan (1988) noted that
 318 there exists an upper limit $\Delta S_{\max} \approx 1.25$, which is imposed by the appearance of a separation
 319 bubble on steeper hills. Once steady separation is established, the effective hill length, which
 320 determines the magnitude of the driving pressure perturbation through the hill steepness ratio
 321 H/L , is set by the geometry of the hill *plus* the bounding streamline of the separation bubble.
 322 We discuss experimental results on separation in Section 6 below.

324 Five years after the appearance of JH75, Sykes (1980) presented a rigorous
325 asymptotic analysis which pointed to some weaknesses in the earlier model. He showed that
326 the no-slip surface boundary condition, where the normal gradient of shear stress must match
327 the tangential pressure gradient, could not be accommodated in the single shear stress layer of
328 JH75. He also pointed out that the turbulent stresses in the JH75 upper layer should be well
329 described by rapid distortion theory. These insights were incorporated in the more complex
330 analytic model of Hunt et al. (1988a) (Henceforth HLR88), which refined JH75 by splitting
331 each layer into two sub-layers. The outer region of JH75 was subdivided into a middle layer,
332 where the mean vorticity is dynamically important, and an upper layer, which is inviscid and
333 irrotational. The inner region was divided into a shear stress layer as in JH75 and a thin inner
334 surface layer below, across which the shear stress varies only slightly but its gradient changes
335 rapidly to match the hill-induced streamwise pressure gradient at the no-slip surface
336 boundary condition (Fig. 2). Again, different terms in the momentum equation dominate in
337 each layer, giving different solutions which can each be matched asymptotically to the
338 adjacent layers.

339 From the point of view of weather models, one of the key motivations for studying
340 flow over hills is to understand and predict the drag they exert on the atmosphere. For
341 neutrally stratified flow this is entirely due to pressure or ‘form’ drag. In the JH75 and
342 HLR88 analytic models, the solutions in each layer are obtained by series expansions in
343 terms of a small parameter and terms of the same order are matched across the layers. The
344 leading order pressure term comes from the solution of the inviscid irrotational (potential)
345 flow in the upper layer and so is symmetrical about the hill and can exert no drag but it does
346 produce an asymmetrical velocity perturbation through the effect of the shear stress in the
347 inner layers. At second order, this asymmetrical velocity perturbation generates asymmetry in
348 the pressure field also and, consequently, pressure drag on the hill. In the prescient analysis
349 of Sykes (1980), his third major advance was to continue his asymptotic analysis to second
350 order, whence he was able to deduce an expression for the pressure drag on a small hill.

351 In a similar way, Belcher et al. (1993) were able to extend the HLR88 model to
352 second order to calculate the pressure drag. In contrast to the earlier calculation of Sykes
353 (1980), Belcher et al. (1993) were able to include the effect of upstream shear, which
354 significantly increased the form drag. They showed that the dominant mechanism for drag
355 over low hills is ‘separated sheltering’, the thickening of the boundary layer in the lee of the
356 hill. In an important clarification, Belcher et al. (1993) also showed that a traditional mixing
357 length eddy-viscosity formulation overestimated the turbulent stresses in the HLR88 outer

358 layers and consequently overpredicted form drag by a factor of two. As a result they altered
 359 their analytic formulation to use a truncated mixing length. This overprediction of the
 360 turbulent response had been implicit in the results of Sykes (1980) and HLR88 as well as in
 361 the wind tunnel measurements of Britter et al. (1981), who all agreed that the turbulent
 362 stresses above the shear stress layer should follow rapid-distortion rather than flux-gradient
 363 physics. Wood and Mason (1991, 1993) went on to investigate the sensitivity of the Belcher
 364 et al (1993) numerical solutions to the turbulence closure assumptions and also extended the
 365 analysis to three dimensions. They used their solutions to develop an approximate expression
 366 for the ‘effective roughness length’, the apparent roughness length $\langle z_0 \rangle$ one would obtain
 367 from fitting a logarithmic form to the horizontally-averaged mean velocity profile well above
 368 the hill. This concept is often used in numerical weather prediction (NWP) models to
 369 represent the effects of sub-grid scale topography on the low level flow.
 370



371
 372 **Fig. 2** Definitions of hill scales and HLR88 layers. SSL denotes the layer and ISL the inner surface layer. h_i is
 373 the shear stress layer depth and h_m the middle layer depth. $U_B(z)$ is the undisturbed approaching wind profile.
 374

375 Another important feature of neutral boundary layer hill-flow is the onset of
 376 separation (Fig.1). Although this is an inherently non-linear process, linear theory has proved
 377 effective in predicting the critical slope at which flow separation can be expected to occur.
 378 Nanni and Tampieri (1985) and Tampieri (1987) derived an expression for the slope needed
 379 to cause separation over a two-dimensional hill, using the original analytic JH75 solution.
 380 Wood (1995) extended this, using the more complete analytic solutions in Belcher (1990) and
 381 Belcher et al (1993) as well as generalising to three dimensions. The analytic expression of

382 Wood (1995) for the critical slope angle depends on the hill geometry through the inner and
383 middle layer heights, and also on the surface roughness. Wood also showed that these
384 estimates from linear theory compare well with the numerical simulations of Wood and
385 Mason (1993). Belcher and Hunt (1998) provide an excellent summary of the early
386 developments of the JH75 model for neutral flow. A comparative review of earlier studies of
387 flow separation, many from wind tunnel comparisons of hills of different geometry, can be
388 found in Finnigan (1988). That comparison showed that separation on rough 2D ridges
389 occurred at slope angles greater than ~15 degrees. But on axisymmetric hills, the critical
390 angle was ~20 degrees. In all the cases reviewed, the nature of the surface roughness at or
391 just upwind of the separation point was critically important.

392 Stratification is an all-pervading feature of the atmosphere and a number of studies
393 have attempted to incorporate this into the JH75 framework. Carruthers and Choularton
394 (1982) extended the JH75 solution to include an inversion layer between the neutral inner
395 layer and a stratified outer layer. The model of HLR88 was also extended by Hunt et al.
396 (1988b) to include the effects of several different upwind stratification profiles, although their
397 solution is limited to $F_H = (U_0/NH) \gg 1$, where F_H is a Froude Number based on hill
398 height. At such large Froude Numbers the airflow goes over the hill rather than being blocked
399 and forced to go around the hill by the stratification. A comprehensive treatment of these
400 ‘high Froude Number’ cases, where the influence of buoyancy is largely exerted through
401 changes to the shear in the approach flow, can be found in Carruthers and Hunt (1990).
402 Blocked flow is fundamentally non-linear and is not amenable to this type of linearised
403 analysis (Kaimal and Finnigan, 1994).

404 Belcher and Wood (1996) presented a more detailed analysis of flow over a hill with
405 stratification, based on the work of Belcher et al. (1993). In the presence of stable
406 stratification there is the possibility of generating internal gravity waves in the boundary
407 layer, and hence there may be wave drag as well as form drag acting on the hill. From their
408 solution Belcher and Wood (1996) derived expressions for both the pressure drag and the
409 wave drag and showed how the relative importance of these depends on the stratification and
410 on the hill shape, factors explored later both in field experiments (e.g. Vosper and Mobbs,
411 1997), numerical models and wind tunnel simulations (e.g. Ross et al. 2004; Allen and
412 Brown, 2006).

413

414 **2.2 Numerical models–Reynolds Averaged Navier–Stokes (RANS) Models**

415 We divide the discussion of numerical models into those that use closure assumptions to
416 relate the turbulent fluxes to the mean wind and temperature fields, collectively known as
417 Reynolds Averaged Navier–Stokes or RANS Models, and those that resolve the energy-
418 containing eddies of the turbulent flow directly, known as eddy-resolving or large eddy
419 simulations (LES). These are treated next in Section 2.3. We can trace two main threads of
420 RANS model development starting in the 1970s. One was initiated and led by Professor Peter
421 Taylor and involved his many students and colleagues from **Southampton and York**
422 **universities, and loosely associated with the Canadian Atmospheric Environment Service.**
423 Much of their later work was motivated by the nascent wind energy industry. A parallel
424 thread can be linked back to Dr Paul Mason at the United Kingdom Meteorological Office
425 (UKMO) and his colleagues there and at the Universities of Cambridge and later, Reading.
426 Both threads were strongly influenced by the analytical modelling ideas coming from Prof.
427 Julian Hunt and colleagues at Cambridge. A strong motivator of the UKMO work was the
428 effect of orographic drag on numerical weather prediction (Mason, 1985). It will be
429 convenient in this section to describe first the ‘Taylor family’ of models and then the ‘UKMO
430 family’ rather than swapping between the two chronologically although the many places
431 where the two threads tangled will be obvious.

432 Numerical models of turbulent flow over low hills actually preceded the appearance
433 of the JH75 analytic theory. In the mid 1970s Taylor and co-workers developed a non-linear
434 numerical model of flow over low hills (Taylor and Gent, 1974; Taylor et al., 1976; Taylor,
435 1977a,b), while Deaves (1975) and Clark (1977) independently presented their own non-
436 linear numerical scheme. These models employed more complex parameterizations of the
437 turbulent stresses than the simple mixing-length closures used in the analytic models. Most
438 used a so-called $k-l$ or $1\frac{1}{2}$ -order closure scheme, where an eddy diffusivity was formed from
439 the product of a velocity scale and a length scale. The velocity scale was taken as the square
440 root of the turbulent kinetic energy (TKE), referred to in this context as k , and a length scale
441 usually taken as the height above the surface at low levels merging smoothly to a constant
442 value at greater heights (e.g. Taylor, 1977a). $k-l$ closures require the solution of an equation
443 for the TKE in addition to the momentum and continuity equations and so increase the
444 demands on available computing power. Those demands could just about be satisfied in the
445 mid 1970s but even today, when computing resources are more abundant, $k-l$ schemes remain
446 the default closure in many RANS hill-flow models because of their robust and predictable
447 behaviour. They do have some significant shortcomings, however, which shall discuss below.

448 Following the appearance of JH75 and its clarification of the fundamental physics
449 governing flow over low hills, several numerical models were developed that adopted the
450 JH75 formalism as the basis for computationally efficient numerical schemes e.g. those of
451 Walmsley et al. (1982) and Taylor et al. (1983). These models were based on linearised flow
452 equations and made direct use of analytic formulations with various layers in the flow
453 asymptotically matched to provide an approximate solution for hills of small slope. Within
454 the limits to accuracy set by simply adding the impacts of surface roughness to those of
455 topography, as permitted by their linear structure, these models allowed predictions of flow
456 over terrain with changing surface roughness and elevation (Walmsley et al. 1986). In the
457 wind energy industry the models most used were WAsP (Troen and Petersen, 1989) and to a
458 lesser degree MS3DJH, originating from the work of Taylor et al. (1983). Both models are
459 still used today in the wind energy industry.

460 These models structured on linear theory were very useful in that they allowed rapid
461 calculation of complex flow fields over arbitrary topography but they were limited to gentler
462 slopes. Because of their linear approximations they lost accuracy when used over terrain with
463 slopes above ~ 0.2 . This was mainly a result of their ignoring the strongly non-linear part of
464 the advection terms in the model formulation and the weakly non-linear terms in the turbulent
465 closure. Another area of weakness was in their vertical representation of the flow. The
466 analytic theories of JH75 and HLR88 divided the flow into layers, in the case of HLR88, two
467 inner and two outer, and this type of vertical decomposition tended to be overly restrictive,
468 often not representing the real atmospheric boundary layer particularly well. At least partly
469 in response to this, Beljaars et al. (1987) formulated a model that was less analytical in that it
470 did not use explicit vertical layers. Though still using linearisations, the model used a mixed
471 spectral-finite difference (MSFD) formulation, in which the terrain-following dimensions
472 were spectral (Fourier modes) and the multi-layered analytic formulation was replaced by a
473 finite-difference boundary-value solution of the equations for the mean and turbulent flow.
474 Rather than explicitly imposing a layered structure on the solution, in which various
475 analytical solutions held, this type of model accomplished a natural and smooth transition
476 between the regions in the flow where different force balances dominated. Upstream mean
477 and turbulent flow parameters (zero-order parts in the linearisation used within the model
478 formulation) were provided by analytic forms of the boundary layer equations or, more
479 commonly, by integrating the full set of boundary layer equations to a steady state over a
480 horizontally homogeneous domain.

481 In addition to removing restrictions on the vertical structure accommodated within the
482 model, this approach allowed more complex and arguably more realistic representations of
483 turbulent flow. Examples of this were the use of higher-order turbulence closures (Newley,
484 1985; Zeman and Jensen, 1987; Ayotte and Taylor, 1994), full boundary layer formulations
485 including rotation (Ayotte and Taylor, 1995) and adjoint data assimilation of upstream flows
486 containing more realistic measured wind profiles (Ayotte, 1997). In the context of wind
487 energy or wind engineering applications, which were key drivers for these developments,
488 these changes to the model formulations moved away from the analytic restrictions noted
489 above but still did not address the errors introduced by linearisation. Those ranged in type
490 and magnitude from the more obvious misrepresentation of mean separation in the lee of
491 steep slopes to the more gradual errors associated with the linearisation of modelled
492 advection. The latter errors scaled with the hill slope and were manifested in an over-
493 estimation of speed-up at the hill crest and an under-estimation of deceleration in the lee of
494 hills of even modest slope. Evidence of this for atmospheric boundary layer flows was
495 present in the literature quite early, for example in comparisons of modelled flow and field
496 measurements over Askervein Hill (see Walmsley and Taylor, 1996). It was also explored
497 more systematically in the wind tunnel measurements of flow over axisymmetric hills of
498 varying steepness and roughness performed by Ayotte and Hughes (2004). Within the wind
499 energy industry, these errors were understood to exist but were poorly quantified in flow over
500 real terrain (Bowen and Mortensen, 2004) and were often compensated for in a practical
501 sense by the use of such things as the Ruggedness Index (Mortensen et al., 2008), which was
502 meant to allow the user to adjust solutions for these errors or at least be aware of them in
503 particular modelling applications.

504 Another source of error when it came to the practical application of these models was
505 their separation of the roughness and topographic influences on the flow. As noted above, the
506 linear models of Walmsley et al. (1986) or Troen and Petersen, (1989) did combine both
507 topography and roughness changes but allowed no feedback between the two influences on
508 the flow. However, HLR88 had revealed the important influence of the shear in the middle
509 layer of the upstream flow on the hill-top acceleration and so, because surface roughness
510 changes clearly affect shear at mid-levels, the coupling between orographic accelerations and
511 upstream surface roughness is unavoidable. Ayotte (1997) showed how upstream roughness
512 changes and the associated changes in vertical shear in the approach flow can result in
513 significant changes in the hill-top speed-up.

514 Although the simple linearised models such as WAsP, MS3DJH and MSFD in one of
515 its many variants were subject to a number of the potential errors listed above, to the extent
516 that these models were used over quite low slopes (H/L less than ~ 0.2) and in terrain where
517 the surface cover did not change dramatically as well as in the near-neutrally stratified
518 conditions of most interest to windfarm developers, the errors were acceptable and of the
519 same order as other uncertainties in wind turbine siting. Recalling the available computing
520 power of the day it is easy to see the attraction of these simplified models. They were easy to
521 use (particularly the WAsP model), fast enough to give an answer quickly and accurate
522 enough to give commercially acceptable uncertainties when used within the parameter space
523 for which they were designed. However, they were often used outside this parameter space
524 (sometimes unwittingly) and as such gained an undeserved reputation for being inaccurate.
525 Some effort was made to address the shortcomings of linear models with weakly non-linear
526 extensions to the MSFD model (Xu and Taylor, 1992, Xu et al., 1994). However, by the
527 early 2000's with computing power becoming much cheaper and more available, the focus
528 had shifted away from linear to non-linear formulations and to models in which the full set of
529 Reynolds Averaged Navier–Stokes (RANS) equations were being solved.

530 Returning now to the 1970s and the UKMO modelling effort, Mason and Sykes
531 (1979) had extended the 2D JH75 model to 3D and used a numerical implementation of their
532 theory to successfully model flow around an isolated hill, Brent Knoll. There then followed a
533 series of publications where numerical models with $k-l$ closure were applied to successive 2D
534 ridges (Mason and King, 1984), to an isolated hill (Mason and King, 1985) and, comparing
535 the predictions of several numerical models, to the drag of small scale topography (Taylor et
536 al. 1989). As part of his PhD thesis at Reading University, where Mason was one of his
537 supervisors, Wood developed the non-linear BLASIUS model, which incorporated many of
538 these earlier ideas and also employed $k-l$ closure. This model was used to study orographic
539 drag in neutrally stratified flow by Wood and Mason (1993) and later by Brown and Wood
540 (2001) and then extended to stably stratified flow to calculate both pressure and wave drag on
541 3D hills by Brown et al. (2003). More advanced closure schemes also began to be
542 implemented in BLASIUS and Ross et al. (2004) were able to compare $1\frac{1}{2}$ order and 2nd
543 order closure simulations of neutrally stratified and stable hill flow with wind tunnel
544 modelling while Lewis et al. (2008a) used a series of mixing length simulations to explore
545 the impact of surface heating on flow separation.

546 Although non-linear formulations were standard by the mid 1990s, we can see with
547 hindsight that non-linear modelling began as early as the mid-1970's (Taylor and Gent,

1974, Taylor et al., 1976, Taylor ,1977a,b, Mason and King, 1984, Newley, 1985) but was greatly hampered by insufficient computing power. Models of this type usually had an incompressible pressure formulation and used a non-orthogonal, regular grid, loosely following the formulation of Clark (1977). Ayotte (2008) has described the progression of non-linear modelling that followed, leading to the non-linear RANS calculations now used by many within the wind energy industry. Employing the wind tunnel measurements of Ayotte and Hughes (2004) as a benchmark, Ayotte (2008) highlighted the improvement in accuracy afforded by the use of a fully non-linear model formulation. The parameter space addressed by RANS simulations has also expanded to include stability-affected flows as routine. See for example Lewis et al. (2008a), who were able to model the effect of surface heating on separation or Bleeg et al. (2015), who modelled the effect of stable stratification on the wind resource of a windfarm, and many other examples can now be found in the literature. In addition widely used commercial or freeware modelling packages like WaSP or BLASIUS offer a variety of closures to suit specific problems. However, despite their widespread use, RANS models remain severely limited in their ability to simulate the abrupt increase in turbulent scale and intensity that follows separation, a phenomenon that is of great practical importance in a range of applications from wind turbine siting to dispersion of seeds and pollutants in steep topography. As a result, the exponentially increased computational power that has now become available has shifted the focus of effort to eddy-resolving calculations and we discuss these approaches next.

568

569 **2.3 Turbulence Resolving Large Eddy Simulations**

570 Although, as we have seen in the previous Section, RANS approaches have contributed
571 enormously to the community's ability to usefully model turbulent flow over complex terrain,
572 capturing the interplay between atmospheric turbulence and the terrain-induced pressure
573 distribution in RANS models largely hinges upon turbulence parameterizations derived from
574 horizontally-homogeneous boundary layers. While such techniques perform reasonably well
575 for unseparated flow over hills of low slope, they struggle to simulate flow separation and
576 recirculation and the accompanying large-scale high-intensity turbulence. The intimate non-
577 linear coupling between turbulent velocity and pressure in steep terrain almost guarantees
578 erroneous solutions when RANS is used in such situations. In RANS models, such errors can
579 certainly be reduced to some extent by applying tuning parameters to the model closures (as
580 noted in the previous Section) but these tunings are 'situation specific' and rarely transfer to
581 novel conditions.

582 The advent of turbulence-resolving simulation in the 1960's and 1970's introduced a
583 new paradigm for studying turbulence. Using numerical methods to solve the non-linear
584 Navier–Stokes equations directly was pioneered at the National Center for Atmospheric
585 Research (see review by Fox and Lilly, 1972). The researchers involved quickly realized that
586 direct numerical simulation (DNS), where a simulation resolves the flow down to the
587 Kolmogorov microscale, is impractical for studying the atmospheric boundary layer because
588 of the vast range of scales of motion to be resolved (mm's to km's), which greatly exceeded
589 the capacity of computers at that time (and in fact still does), and the varying character and
590 roughness of the underlying surface. However, the researchers also recognized that
591 discretizing the flow field on a finite grid represents a low-pass filtering operation and that, if
592 they placed that filter at a scale well into the inertial subrange, they could calculate the time
593 evolution of the largest energy-containing scales of motion directly while using theoretical
594 arguments to parameterize the smallest scales, which act to dissipate the energy in the
595 resolved scales. This technique is known as large-eddy simulation (LES). We note here that
596 the engineering and geophysical fluid dynamics cultures use the label 'LES' to mean subtly
597 different things. In the engineering communities, LES implies parameterising the sub filter-
598 scale turbulent motions while still retaining sufficient resolution to capture the near-surface
599 viscous wall layer, i.e. they retain the viscous terms in the equations. This is sometimes
600 referred to as 'wall-resolved' LES. In contrast, in the geophysical community, viscous terms
601 at solid surfaces are usually ignored because of the very large disparity between the filter
602 scale and the Kolmogorov microscale. In those models, therefore, momentum transfer to the
603 surface must be completely handled by a wall model (i.e. an imposed drag law with
604 corrections for atmospheric stability and roughness). Engineering LES is sometimes referred
605 to 'finite Reynolds number LES', and geophysical LES as 'infinite Reynolds number LES'.

606 Turbulence-resolving simulations transformed the community's understanding of
607 atmospheric turbulence over horizontally homogeneous terrain, showing for example, that
608 instabilities create organization or coherent eddy structures in turbulent flows, whose
609 character depends on the balance of driving forces, see Deardorff (1970a,b, 1972a,b). It also
610 quickly became apparent (e.g. Gal-Chen and Somerville, 1975) that numerically solving for
611 the time evolution of flow over an irregularly shaped surface requires application of one of
612 two broad concepts. Either a Cartesian grid framework is retained and special techniques
613 applied to represent the lower boundary, a concept that now has numerous names, such as
614 immersed boundaries or cut-cell methods. Alternatively coordinate transformation is used to
615 convert complicated domains into a Cartesian representation (e.g. Phillips 1957). Although

616 they allow one to match boundary conditions explicitly, coordinate transformations generate
617 additional terms in the equations, which, depending on the method chosen, can complicate
618 the pressure solution and can introduce unphysical singularities.

619 To our knowledge, Clark (1977) was the first to apply a turbulence-resolving
620 simulation to flow over aerodynamically-rough hills using the grid-transformation method
621 and demonstrating the importance of properly treating the surface boundary and turbulent
622 inflow conditions when studying orographic drag. In the early 1990's, Schumann (1990),
623 Walko et al. (1992) and Dörnbrack and Schumann (1993) began exploring atmospheric LES
624 as a tool to interrogate the influence of buoyancy on turbulent flow over sloping terrain or
625 interacting with two-dimensional topography. Schumann (1990) determined that the
626 equilibrium state for heated turbulent flows over a sloping boundary depends solely on the
627 slope angle and the ratio of the surface roughness length (z_o) to a buoyancy length scale
628 related to the amplitude of the surface buoyancy flux and stratification in the free-atmosphere
629 above the boundary layer. However, he acknowledged some uncertainty in his findings
630 associated with his having applied MOST in heterogeneous terrain. Walko et al. (1992)
631 showed that buoyancy in the zero mean wind boundary layer over repeating hills generates
632 circulations with updraft plumes focused at the hill crests while Dörnbrack and Schumann
633 (1993) found that the scale of the orography alters the character of the hexagonal turbulent
634 convection cells that dominate free convection boundary layers. They also found that
635 convectively driven turbulence eliminates recirculation zones that would otherwise persist in
636 the lee of steep terrain at the wind speeds they studied. Gong et al (1996) applied Dörnbrack
637 and Schumann's LES to neutrally-stratified flow over two dimensional ridges to explain the
638 longitudinally aligned roll-like structures that they found in their wind tunnel measurements.
639 These probably arise from a secondary instability mechanism induced by the hills such as
640 a Craik-Leibovich type-2 instability (See Section 7.1). Using 'wall-resolved' LES over low-
641 amplitude two-dimensional hills, Henn and Sykes (1999) confirmed the presence of an inner
642 layer, about 5% of the hill wavelength in depth, in which velocity variances vary rapidly.
643 They also noted dramatic increases in the lateral velocity variance on the windward side of
644 their hills that they also speculate are associated with an un-determined terrain-induced
645 instability mechanism.

646 In his seminal review, Wood (2000) outlined the difficulties associated with
647 producing "well-resolved" LES of neutrally-stratified turbulent flow over hills. His key
648 findings were that: the simulations must be 3-D; grids must be isotropic; the domain needs to
649 be large enough to permit the largest eddies anticipated yet fine enough to resolve the small

650 eddies within the inner layer; the lowest grid point needs to be sufficiently close to the
651 surface to allow application of law-of-the-wall boundary condition; and finally, simulations
652 need to be run for long enough to generate stationary turbulence statistics. While such
653 requirements made accurate simulations difficult to implement in 2000, computing resources
654 available today make them attainable (e.g. Bou-Zeid, 2014; De Bruyn Kops and Riley, 2019).
655 Nevertheless, Brown et al. (2001) circumvented Wood's fourth requirement by recognizing
656 that, when using such high resolution, one begins to resolve some of the roughness elements
657 such as tall trees and accounting for momentum loss to the canopy elements, which occurs
658 primarily through pressure drag, minimizes the importance of near-surface processes beneath
659 the canopy. Interestingly, Allen and Brown (2002) found that applying a canopy model rather
660 than surface roughness over their simulated hill increased the depth and overall extent of their
661 predicted separation bubble compared with observations. We shall see in the next section that
662 this phenomenon is caused by the interaction between fundamental features of canopy flow
663 and hill flow.

664 Because of its control over and influence on near-surface mean wind and momentum
665 flux profiles, the sub-filter scale model formulation used to close the equations and represent
666 the momentum transport can affect flow separation predictions in the lee of steep hills (e.g.
667 Satoru and Kondo, 2004, 2006; Wan et al., 2007; Chow and Street, 2009, Wan and Porte-
668 Agel, 2011). Silva-Lopes et al. (2007) noted that resolution can also impact the size, shape,
669 intensity, and intermittency of the separation bubble. Many of the early LES applications
670 simulated wind tunnel rather than field measurements in an attempt to reduce terrain
671 complexity and maintain some control over the driving parameters (e.g. Gong et al, 1996;
672 Henn and Sykes, 1999). However, researchers fully recognized that transferability of that
673 knowledge to geophysical applications outdoors would require an understanding of whether
674 Reynolds number similarity applies, how stratification alters the relationships between hill-
675 induced pressure and turbulent momentum stress, and how terrain containing a broad
676 spectrum of scales alters the relationships and theories developed for idealized terrain (see
677 Section 7 below).

678 Through the 1990's and 2000's, the Askervein hill experiment (Section 5) served as the
679 most complete field data set available, hence many LES studies began targeting data from
680 Askervein as computing power became more readily accessible (e.g. Undheim et al., 2006;
681 Chow and Street, 2009; Golaz et al. 2009). It was soon realized that, while Askervein hill
682 was excellent for testing linear theories, the shallow slopes were insufficient to study the
683 application of non-linear models to fully separated flow and that the Askervein hill

684 experiment lacked enough turbulence measurements to characterise the flow complexity
685 adequately. To address these issues with newly emerging remote sensing capabilities, the
686 wind energy community initiated a new set of field experiments that began with what is now
687 known as the Bolund experiment (Section 6). Bechmann et al. (2011) presented a blind
688 intercomparison of many RANS codes and a few LES codes upon which the wind energy
689 community relies with the most interesting result being that the highly-tuned RANS models
690 outperformed the few engineering-style LES codes that were included, however, the poor
691 performance of the LES codes was largely attributed to modeller skill. Follow-on efforts
692 using LES to target the Bolund experiment have shown substantially improved skill (e.g.
693 Diebold et al., 2013). Recent field campaigns targetting flow over hills for wind energy
694 resource assessment (e.g., Mann et al., 2017) should enable transformative new
695 understanding and turbulence resolving simulation capabilities; for example, Dar et al. (2019)
696 used LES to demonstrate hill-induced flow separation influences on the downstream
697 evolution of turbine wakes. Several modelling packages now widely used in the community
698 such as the previously mentioned WAsP and BLASIUS as well as the more mesoscale
699 orientated WRF (<https://www.mmm.ucar.edu/weather-research-and-forecasting-model>) offer
700 LES capability but these codes need to be treated with due appreciation of the caveats listed
701 above.

702 **2.4 Modelling Scalar Transport over Rough Hills**

703 Thus far we have focussed on momentum transport and flow patterns but, as noted at the
704 beginning of this section, many of the applications of hill flow research require information
705 on scalar transport. The examples we have already given, where the effects of buoyancy on
706 the windfield were calculated, implicitly required knowledge of heat as well as momentum
707 transport, however, modelling of scalar transport over hills dealt first passive scalars. Padro
708 (1987) used a quasi-analytical approach to deduce the perturbations to the field of a generic
709 scalar flowing over a 3D hill. He derived the wind field was from the Walmsley (1980)
710 numerical implementation of the JH75 theory as discussed in Section 2.2. Padro and
711 Walmsley (1990) extended this approach by including the concentration field calculations in
712 the formalism of the linearised MSFD, model (Beljaars et al. 1987), which employed an
713 improved $K-l$ closure scheme. This combination provided a generally useful tool for
714 application. Hunt et al. (1998c) reviewed the general principles governing scalar fluxes to
715 hills in the light of the scaling arguments encapsulated in HLR88 and its stratified extension
716 (Hunt et al. 1998b), however, the first direct application of the then new linear analytic
717 HLR88 theory to scalars was the elegant study by Raupach et al. (1992), who analysed the

718 exchange of a generic scalar, $c(x,z)$ with a rough 2D ridge and then applied their model to a
 719 classic problem in micrometeorology-the partition of solar energy into sensible and latent
 720 heat at the surface.

721 As in HLR88, Raupach et al. (1992) (henceforth RWCH92) divided the flow-field
 722 into an outer layer and a shear stress layer of depth h_i (Fig. 2). In the outer layer, scalar
 723 perturbations are governed by inviscid dynamics while in the shear stress layer, changes to
 724 the scalar flux also play a role at first order. The linearised mass conservation equation in the
 725 outer layer is,

$$726 \quad U \frac{\partial \Delta C}{\partial x} + \Delta W \frac{\partial C}{\partial z} = 0 \quad (6)$$

727 which implies that in the outer layer $\Delta C(x,z)$, the perturbation in the time mean
 728 concentration of the generic scalar c , is entirely the result of distortion of the isopycnals of
 729 $C(x,z)$ as streamlines converge and diverge. In the inner layer, where $\overline{\Delta f_c}$, the perturbation
 730 in the eddy flux of the scalar is important, the linearised mass balance is

$$731 \quad U \frac{\partial \Delta C}{\partial x} + \Delta W \frac{\partial C}{\partial z} = - \frac{\partial \overline{\Delta f_c}}{\partial z} \quad (7)$$

732 RWCH showed that in this region two other mechanisms become important in determining
 733 $\Delta C(x,z)$. The first is the changes induced by the hill in the eddy flux field $\overline{\Delta f_c}(x,z)$ itself
 734 and hence in its divergence. The second is the change to the flux of C from the surface that
 735 occurs because the surface shear stress perturbation, $\overline{\Delta \tau}(x,0)$ varies as the hill is traversed.
 736 The mechanism for this in the RWCH92 model is the representation of the surface source by
 737 a flux-gradient expression, $\overline{\Delta f_c}(x,0) = -K_c \partial(C + \Delta C)/\partial z$, involving the scalar diffusivity,
 738 $K_c = \kappa u^* z (1 + \overline{\Delta \tau}/2)$. In the shear stress layer this modulation of the surface flux boundary
 739 condition is the dominant influence on $\Delta C(x,z)$. The RWCH92 theory was later used to
 740 investigate the general influence of topography on meteorological variables at whole
 741 landscape scale by Raupach and Finnigan (1997) and see also Huntingford et al. (1998).

742 It was not long before the leading numerical models were incorporating the effects of
 743 buoyancy by carrying an equation for heat transport as well as momentum transport fully
 744 incorporated in their analysis schemes. Belcher and Wood (1996) used such a version of the
 745 BLASIUS model in their study of form and wave drag over low ridges while Weng (1997)

746 presented a version of the MSFD model called MSFD-STAB in a comparison with buoyancy
747 affected field measurements over a low ridge (Coppin et al. 1994). This approach was then
748 generalised to consider the generation of internal gravity waves over generalised topography
749 Weng et al. (1997). Since then the BLASIUS model has been applied to more complex
750 questions such as the impact of surface heating on separation (Lewis et al. 2008a) while
751 buoyancy calculations are now standard features of most hill flow model applications (e.g.
752 Bleeg et al. 2014).

753

754 **3. New Physics-Canopies on hills**

755 By the early 1990's a new and powerful motivation for studying boundary layer flow over
756 topography had appeared: the quantification of the exchange of carbon and energy between
757 the atmosphere and the biosphere through direct measurement of the turbulent exchange flux.
758 Organised into the international FLUXNET program (<https://fluxnet.fluxdata.org>), this
759 methodology built on a 25 year legacy of direct eddy flux measurement by fast response wind
760 and scalar sensors. With today over 900 'flux towers' measuring exchange from different
761 biomes around the world, particularly from tall forests where the major living carbon stores
762 reside, FLUXNET comprises one of the largest geophysical experiments ever mounted.
763 However, from its earliest days it was clear that significant problems in interpretation were
764 posed when the flux towers were situated in hilly terrain, even when steeper slopes were
765 avoided. Raw 24 hour totals of net carbon exchange were often wildly at variance with
766 independent stoichiometric limits on photosynthesis and respiration (Finnigan, 2008). Since
767 most flux tower sites were located in relatively gentle topography, the extension of linear
768 theory to accommodate a tall canopy rather than a rough surface was an obvious step towards
769 understanding what might be going on. Interestingly, the study of canopy flow on hills
770 brought together two fields which over the previous two decades had independently
771 undergone rapid development to each arrive at a generally accepted understanding of their
772 essential dynamics but their interaction revealed some new and surprising physical
773 phenomena.

774 We have already described the development of the basic theory of hill flows but by the
775 mid 1990's a set of questions that had taxed canopy flow researchers for at least 25 years
776 were also being answered, particularly the origin and nature of the characteristic large
777 coherent turbulent eddies that dominate canopy flow. By then, these were known to originate
778 from a hydrodynamic instability of the inflected mean velocity profile that inevitably

779 develops when momentum is absorbed over a finite height range rather at a plane surface
780 (Raupach et al., 1996; Finnigan, 2000; Finnigan et al. 2009). An early indication of the new
781 phenomena that appeared when topography was added beneath a deep canopy came from the
782 wind tunnel experiment of Finnigan and Brunet (1995). By placing a well-studied model
783 canopy on a 2D ridge they observed strong modulation of the critical inflected velocity
784 profile with the inflection point disappearing on the upwind hill slope and being exaggerated
785 and squeezed up into the upper canopy on the crest. They also observed a large continuous
786 separation bubble although the hill steepness was only at the margin of what would trigger
787 separation on a rough hill.

788

789 **3.1 Basic Theory and Analytic Models**

790 A first step in extending linear theory was taken by Finnigan and Belcher (2004) (henceforth
791 FB04), who effectively replaced the inner surface (shear stress matching) layer of the HLR88
792 model by a two-layer canopy scheme. In the upper canopy the flow is linearised and the hill-
793 induced pressure gradient, which is calculated as in HLR88 from the outer layer inviscid
794 flow, is balanced by a combination of perturbations in the turbulent shear stress and in the
795 aerodynamic drag of the canopy elements. In the lower canopy layer, which is inherently
796 non-linear the pressure gradient is balanced predominantly by the non-linear canopy drag,
797 $F_D = U^2/L_c$, which is represented as proportional to the square of the mean velocity divided
798 by a momentum absorption length scale, $L_c = (C_d a)^{-1}$, which is the reciprocal of the
799 dimensionless foliage drag coefficient C_d times the foliage area per unit volume, a (Finnigan
800 and Brunet, 1995). The non-linearity of the lower canopy layer cannot be avoided as the
801 background mean velocity decays exponentially with depth into the canopy (see Figure 3a
802 and Finnigan, 2000). As a result, in a deep or dense canopy, the background mean velocity in
803 the lower canopy is smaller than the perturbations driven by the pressure gradient, which
804 passes through the foliage effectively undiminished.

805 The coupling of the lower and upper canopy and shear stress layer solutions by
806 turbulent stress divergence leads to significant differences between the magnitude and
807 vertical distribution of the velocity perturbations in the FB04 and HLR88 models. The
808 differences can be seen in Figure 3a, where the FB04 and HLR88 models are applied to a 2D
809 hill of Gaussian cross section and where the surface z_0 in the HLR88 solution matches the
810 canopy z_0 in the FB04 solution. The differences are even clearer in Figure 3b, where only the
811 normalised velocity perturbations are shown. The HLR88 solution only goes down to a level

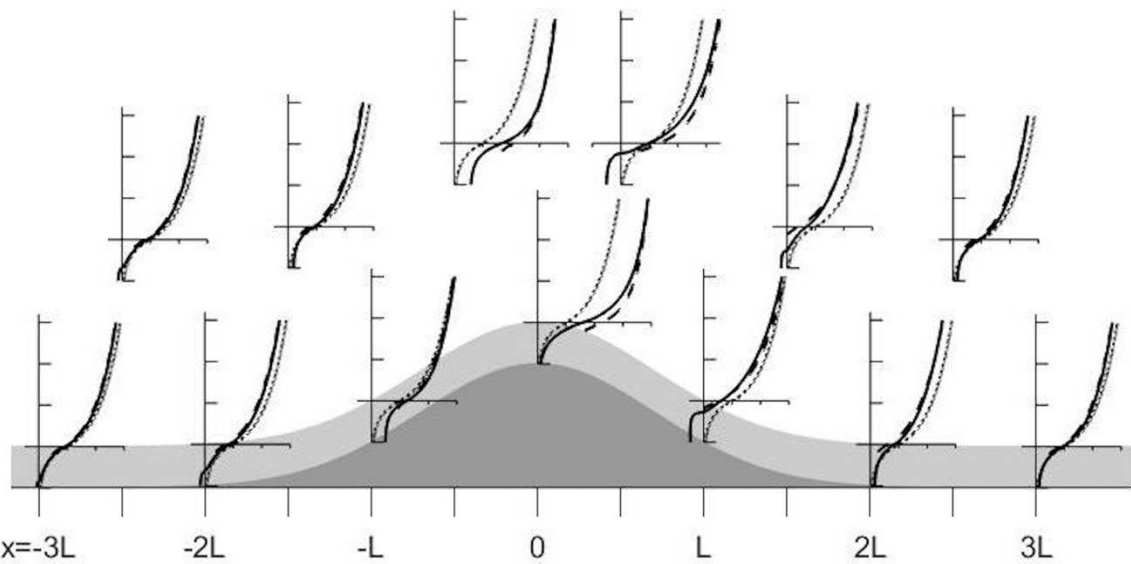
812 $d+z_0$, where d is the displacement height of the upwind logarithmic wind profile, but it is
813 evident that the nature of the solution is changed significantly through the depth of the shear
814 stress layer above the canopy. An immediate consequence of the differing forms of the
815 leading approximation to the momentum equation in the upper and lower canopy layers is
816 that the velocity perturbation, ΔU in the upper canopy, just as in the shear stress layer above,
817 is proportional to the pressure perturbation whereas in the lower canopy, ΔU follows the
818 streamwise gradient of the pressure perturbation. This can be clearly seen in Figure 4a, where
819 the velocity perturbations along streamwise transects above and within the canopy on the hill
820 are compared with the forcing pressure perturbation and its gradient.

821 The differences are even clearer in Figure 4b, where the velocity perturbations ΔU
822 from both models are plotted along streamwise transects through the shear stress, upper and
823 lower canopy layers. In the shear stress layer at $z=3h_c$ the two solutions are close and both
824 models have ΔU approximately in phase with the pressure perturbation but at the canopy top,
825 $z=h_c$ the FB04 perturbation is significantly smaller than the HLR88 solution and peaks further
826 upwind. As we descend deeper into the canopy, the phase of ΔU changes to match the
827 pressure gradient in the lower canopy solution and the phase difference between ΔU in the
828 upper canopy and in the shear stress layer approaches $\pi/2$, following the phase difference
829 between the leading order pressure perturbation and its gradient. We reemphasize that it is the
830 coupling of the non-linear lower canopy solution with the linear upper canopy and shear
831 stress layer solutions by the turbulent shear stress that leads to these global differences.

832 The fundamental non-linearity of the lower canopy solution in the FB04 model leads
833 to several features that are not seen in fully linear models. First it allows the hill flow solution
834 to affect the background flow, at least in principle, unlike models of the JH75 and HLR88
835 type, where the background flow remains unchanged. Wood (2000) pointed out that from the
836 NWP perspective, where we are interested in global changes to topographic drag, this is an
837 important limitation on linear models. Second, FB04 predicts the appearance of a separated
838 region within the canopy on the lee slope and, when that separation bubble occupies the full
839 canopy depth, we expect the flow above to separate also. Indeed, flow separation occurs for
840 hills of lower slope (by a factor of 2-3) than if the hill were a rough wall. This has been
841 confirmed in wind tunnel (Harman and Finnigan 2013), flume (e.g. Poggi et al. 2008) and
842 large eddy simulations (e.g. Dupont et al. 2008, Patton and Katul 2009) and is also reported
843 in field observations (Zeri et al. 2010).

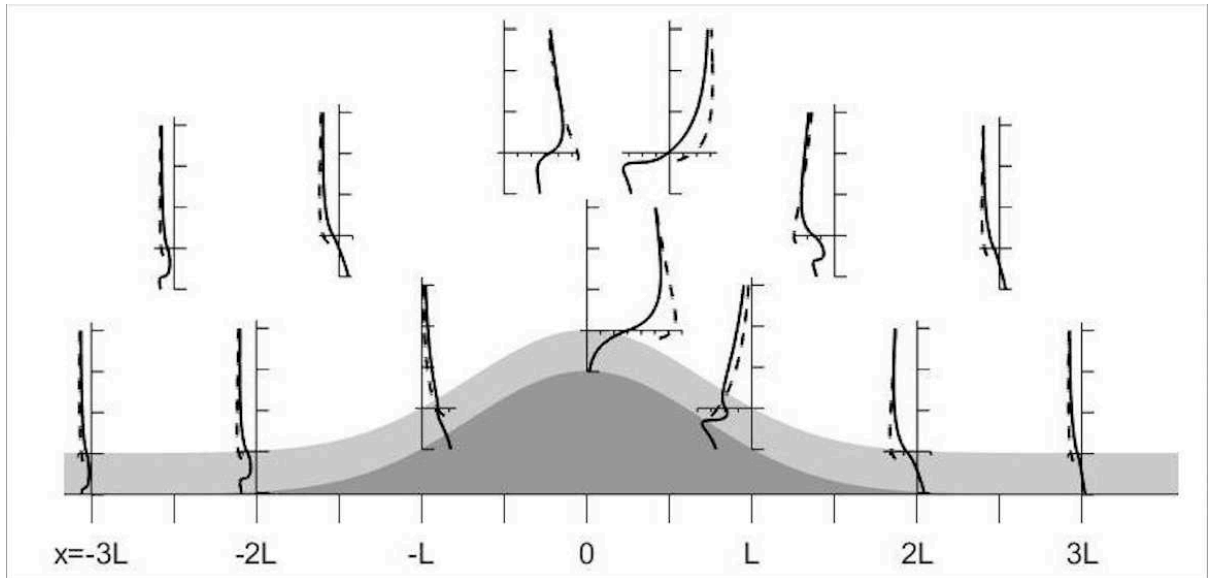
844 Third, the non-linearity leads to convergence of the perturbation flow in the lower
 845 canopy towards the hill crest (see Figure 4b). The associated vertical motion is in anti-phase
 846 with the leading order pressure perturbation and $\pi/2$ out of phase with the vertical motion
 847 induced by the deflection of the mean flow by the hill, which peaks on the upwind slope. For
 848 hills of short length scale, L , and tall but sparse canopies, the two components of the vertical
 849 motion are comparable, leading to significant changes in the magnitude and phasing of the
 850 perturbation pressure response and associated form drag (Poggi and Katul 2008, Poggi et al.
 851 2008). Simple scaling suggests that the pressure response decreases by factor of $1 + F$ where
 852 $F \sim (h_c^2 L_c / HL^2)^{1/2}$, with more complex variations in phasing. As we shall see in Section 3,
 853 this feature has a significant impact on scalar transport.

854 **a**



855
 856 **b**

857



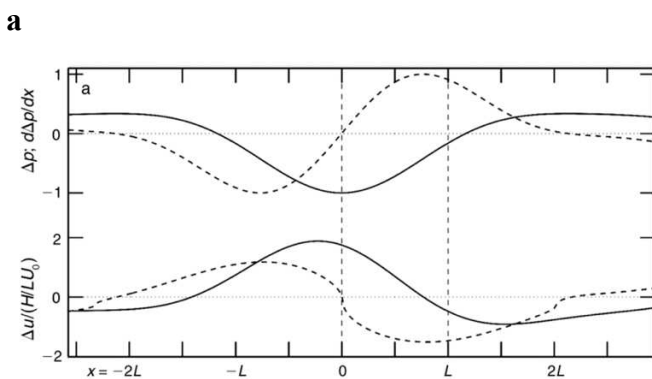
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859

860 **Fig. 3** Profiles of wind speed above a Gaussian hill covered with a canopy calculated with FB04 and HLR88. **3a**
 861 shows total mean wind profiles. The dotted lines show the wind profile in the absence of a hill; solid lines the
 862 RB04 solution and dashed lines the HLR88 solution for the same effective z_0 . Note the HLR88 solution extends
 863 down only to $z=d+z_0$, where d is the displacement height and z_0 the roughness length of the upwind logarithmic
 864 wind profile.

865 **3b**, as for Figure 3a but showing just the streamwise velocity perturbations for RB04 (solid line) and HLR88
 866 (dashed line).

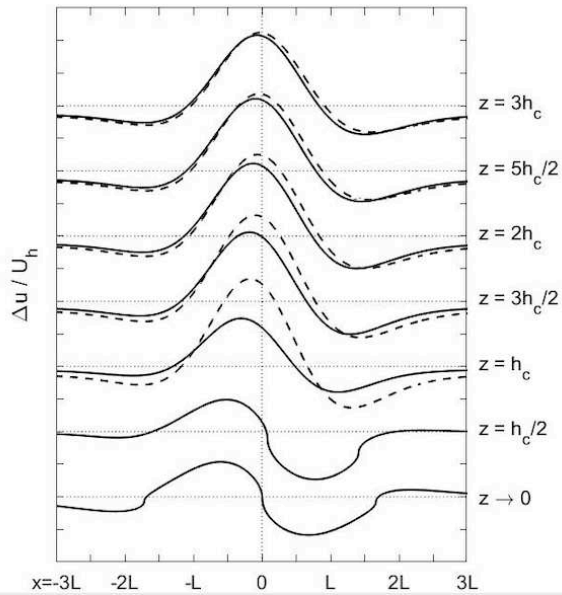
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868

869

b



870

871 **Fig. 4. (a)** Phasing of forcing terms across the hill. Upper part of the panel shows variation across the hill of the
 872 pressure perturbation (solid line) and the pressure perturbation gradient (dashed line); perturbations have been
 873 normalized by the maximum value taken across the hill. The lower part of the panel shows variation of the
 874 normalized wind speed perturbation across the hill. Solid line is at $z=3h_c$ above the canopy, which is nearly in
 875 phase with the pressure perturbation. The dashed line is at $z=h_c/2$ within the canopy, which is in phase with the
 876 pressure perturbation gradient.

877 **(b)** velocity perturbations along streamwise transects through the shear stress layer and down into the canopy.
 878 Solid line denotes RB04 solution and dashed line HLR88.

879

880 The original FB04 model made a series of simplifying assumptions and these restrict
 881 its application to a relatively small region of a parameter space spanned by H/L , h/L_c and
 882 L_c/L , variables which provide a useful classification of experiments and model results (see
 883 Belcher et al., 2012). If we restrict our attention to hills where analytic theories like FB04
 884 strictly apply, that is $H \ll L$, Patton and Katul (2009) have collapsed the 3D parameter space
 885 of Belcher et al. (2012) into a 2D plane, which the competing influences of canopy height,
 886 canopy density and hill length divide into five regimes (Fig. 5). The envelope

887 $h_c/L_c = 2(H/L)(L/L_c)^2$ marked by the slanting line in Figure 5 delineates regions in which

888 the mean vertical velocity inside the canopy is expected to be large enough to affect the
 889 outer-layer pressure. Patton and Katul (2009) refer to flow regimes above this envelope to as
 890 ‘interactive’ pressure regimes, and flow regimes below this envelope as ‘fixed’ pressure
 891 regimes. FB04 is valid in the shaded area (Regime 1). Formalised extensions to the FB04
 892 analysis, to cover a wider range of hill shapes and canopy densities have been presented by
 893 Poggi et al. (2008) and Harman and Finnigan (2010, 2013). The extensions include the effect

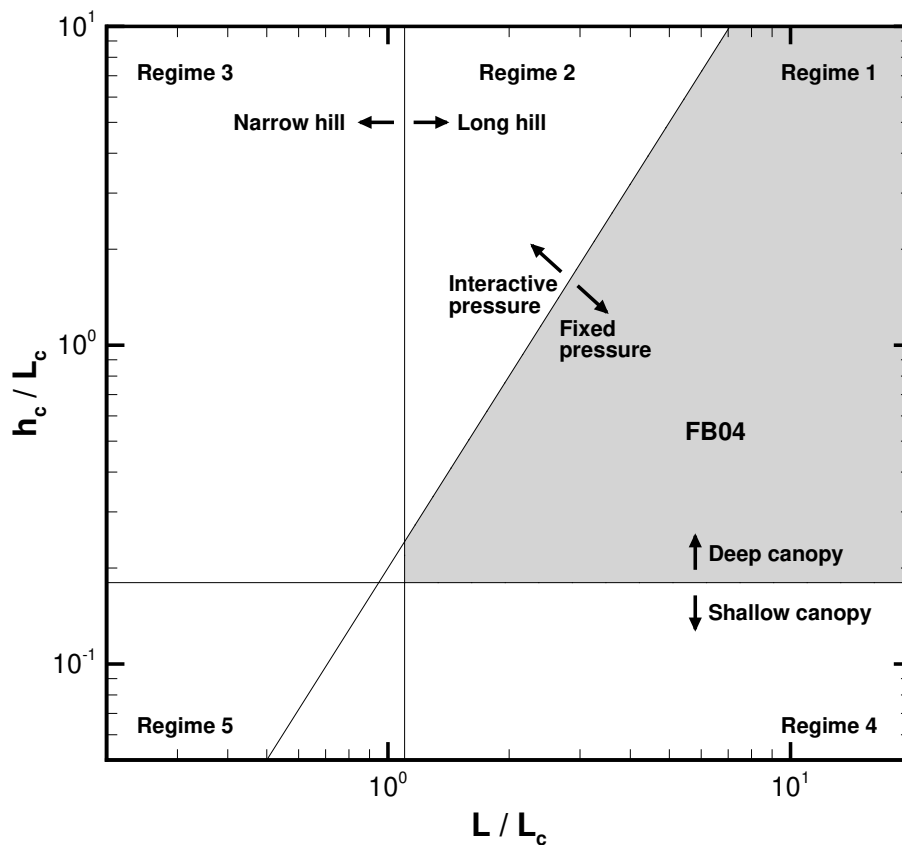
894 of advection in the upper canopy and better representation of the effect of the coupling
895 between the lower canopy flow and the upper canopy and shear stress layers on the pressure.

896 Finally, and of both theoretical and practical importance, FB04 demonstrated that,
897 even in the asymptotic limit of hills of very low slope, the canopy flow solution does not
898 converge to the HLR solution for a rough wall; a constant roughness length z_0 is not the
899 formal limit of a shallow canopy flow. This should not be too surprising since Harman and
900 Finnigan (2007, 2008), analysing flow in homogeneous canopies on flat ground, have
901 demonstrated that treating z_0 as a parameter determined by the surface geometry rather than
902 as a flow variable leads to significant errors in predicting the momentum and scalar fields in
903 the roughness sublayer. Since rough walls are themselves simply canopies of roughness
904 elements, they should be dynamically similar to tall canopies when correctly scaled (Raupach
905 et al., 1991).

906

907 **3.2 Effects of stability**

908 A further set of fundamental changes in the hill-canopy flow physics occurs when the
909 boundary layer becomes stably stratified, which typically occurs through radiative cooling at
910 night. These are discussed in detail in Belcher et al. (2008). Interestingly, these global
911 changes are the emergent result of differences in the microphysics of exchange processes at
912 leaf level. The efficiency of transport of a scalar between the foliage surface and the canopy
913 airspace is determined by the molecular conductivity of the scalar whereas, at typical natural
914 Reynolds Numbers, momentum transfer to the foliage is dominated by pressure drag. The
915 ratio of the efficiency of scalar to momentum transfer is expressed by the Stanton Number,
916 which is $O[0.1]$ in natural canopies. As a result, the gradient of temperature in a radiatively
917 cooling canopy is much smaller than the gradient of mean windspeed produced as the foliage



918

919 **Fig. 5** Length scale regimes imposed by hill geometry and canopy morphology for low hills where $H \ll L$, after
 920 Poggi et al. (2008). Hills are classified as narrow or long and canopies as deep or shallow

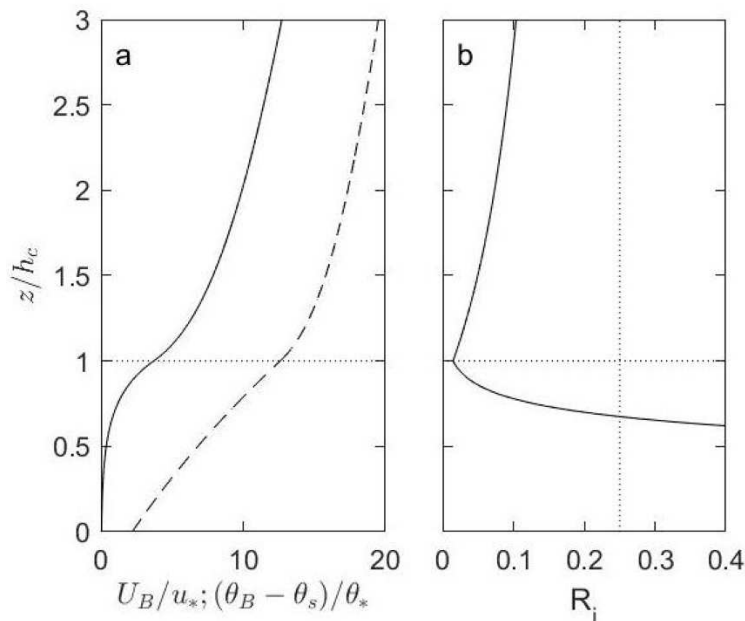
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922 absorbs momentum. Figure 6a shows profiles of temperature and windspeed computed using
 923 a simple mixing length model in a canopy with constant L_c and constant foliage surface
 924 temperature. Figure 6b shows the associated gradient Richardson Number $Ri = N^2 / (\partial U / \partial z)^2$,
 925 which has a minimum at $z = hc$, where the shear is largest, but quickly exceeds the critical
 926 value of $Ri = 0.25$ within the canopy (Belcher et al., 2008).

927

928 This rapid change in Ri typically leads to the existence of a subcritical region just
 929 above the canopy top within which $0.25 > Ri > 0$ and internal gravity waves can be supported.
 930 These are often observed at flux sites at night-time (van Gorsel et al., 2011) and add a level
 931 of complexity to the hill flow that has not yet been studied in any detail. When the within-
 932 canopy Ri becomes large enough, we see the situation modelled in a wind tunnel by Finnigan
 933 and Hughes (2008), where flow above the canopy was stably stratified but fully turbulent
 934 with $Ri \sim 0.1$ while within the canopy $Ri \sim 10$ and rotational turbulence was quenched,
 935 effectively decoupling the above- and within-canopy airflow. Within the canopy Finnigan
 936 and Hughes observed downslope gravity currents on both up and downwind hill slopes while

936 flow above the canopy behaved like flow over a rough hill (see Section 7.). Both the
 937 existence of large Ri in the night-time canopy, the decoupling of the within- and above-
 938 canopy flow and the consequent strong downslope gravity current seen in the idealised wind
 939 tunnel study were also observed in the field observations of Yi et al. (2005) at a FLUXNET
 940 site in the Front Range of the Rocky Mountains in Colorado. The gravity wave case (Van
 941 Gorsel et al., 2011; Lee and Barr, 2006) and the fully decoupled case are both of great
 942 importance to the carbon flux modelling community and we will return to them in Sections 4
 943 and 5.



944
 945 **Fig. 6** Profiles of (a) windspeed (solid line) and temperature (dashed line) and (b) gradient Richardson Number,
 946 Ri through a canopy on level terrain. Note that for the moderate stratification of these plots, Ri remains sub-
 947 critical above the canopy, implying turbulent flow but supercritical within the canopy, implying collapse of the
 948 turbulence.

949
 950 **3.3 Numerical RANS models**

951 Just as was the case over rough hills, numerical RANS models have played an important role
 952 when canopies are added to complex terrain. In order to represent canopy flow in a RANS
 953 model, an approach similar to that in analytic models like FB04 is typically taken, first by
 954 including a term in the momentum equation to represent the foliage drag, and second by
 955 modifying the turbulence closure to represent the different processes controlling turbulence in
 956 canopies. The canonical model is that the turbulence is dominated by eddies generated at the
 957 inflected shear layer near canopy top (analogous to a mixing layer). This motivates the choice
 958 of a constant mixing length l_m in the canopy. For $1\frac{1}{2}$ order closure models with a prognostic

959 equation to compute the turbulent kinetic energy, there may also be an additional dissipation
960 term to represent the short-circuit in the energy cascade resulting from the small canopy
961 elements rapidly breaking up large eddies into smaller scale eddies (Balducchi and Meyers,
962 1988; Wilson et al, 1998; Finnigan, 2000). Wilson et al (1998) tested the closure by
963 comparing with wind tunnel data from the Finnigan and Brunet (1995) “Furry Hill”
964 experiment and showed that the scheme worked as well as other more complicated closures
965 e.g. Sogachev and Panferov (2006) or Sogachev (2009).

966 Following the steps outlined above, Ross and Vosper (2005) modified the *K-l* closure
967 version of the BLASIUS model for canopy applications and inter-alia used this to validate the
968 linear analytic FB04 solution. They also used the numerical simulations to study the onset of
969 flow separation in a canopy and confirmed that a canopy on the hill surface indeed promoted
970 earlier separation than a roughness length parametrisation. Similarly they showed that the
971 canopy enhanced the form drag on the hill due to a shift in the pressure field as predicted by
972 FB04. Their model has subsequently been used to study the effects of canopy heterogeneity,
973 both variable canopy density (Ross, 2012) and hills partially covered by canopies (Ross and
974 Baker, 2013). Ross (2011) and Ross and Harman (2015) investigated the impact of source
975 distribution on tracer transport using the RANS model as a way of addressing the impact of a
976 canopy on flux variability over complex terrain even in neutral flow. As well as studying
977 idealised problems, the model has also been run over realistic terrain by Grant et al, (2016) to
978 compare to observations from the Arran canopy experiment described in Grant et al (2015).
979 Various RANS CFD models have also been applied to canopy flows. Yi et al (2005) used a
980 CFD model to study nocturnal drainage flows in forested complex terrain. More recently the
981 importance of canopy effects for assessing wind energy resources and for wind engineering
982 has led to a number of CFD studies including canopy effects (e.g. Chávez Arroyo et al 2014;
983 Desmond et al, 2017).

984 There has been debate in the literature over the applicability of simple RANS mixing-
985 length closure models in canopy flows. Eddy covariance observations do show counter-
986 gradient turbulent fluxes in some canopies, suggesting that local parameterisation is not
987 appropriate, however various studies have shown that in practice first order schemes are
988 useful (Grant et al, 2016). The assumptions and limitations of first order canopy closure
989 schemes are analysed by Finnigan et al (2015). There are a number of reasons for the
990 surprising success of first order closure schemes. First, analytic canopy models such as FB04
991 show that the leading order dynamics governing flow perturbations in canopy flows over hills
992 are inviscid (see Belcher et al., 2012), reducing the impact of the turbulence closure scheme.

993 Second, at least for momentum fluxes, the strongest shear is at the canopy top where the
994 constant mixing length assumption works fairly well as demonstrated for example by LES
995 studies like Ross, 2008 and in comparison with observations such as Katul et al, 2004 and
996 Grant et al, 2015. The LES studies suggest that the mixing length may not be constant with
997 height deeper in the canopy although the shear is generally low there and so this does not
998 significantly impact on the momentum fluxes. However, it may be more important for scalar
999 fluxes being emitted from the surface, although the impact of this has not yet been fully
1000 assessed.

1001

1002 **3.4 LES Models**

1003 Although Brown et al. (2001) in their attempt to resolve competing requirements in LES
1004 modelling of neutral turbulent flow over a rough hill adopted Shaw and Schumann's (1992)
1005 strategy, which was developed to allow the incorporation of canopy physics in LES, canopy-
1006 resolving LES over hills really began with a pilot effort by Patton et al. (2006) who identified
1007 some of similarities and differences between flow over isolated and sinusoidally repeating 2D
1008 ridges. They demonstrated that resolving the canopy increases turbulence levels at the hill
1009 crest and confirmed the prediction of FB04 that, because a canopy primarily interacts with
1010 the flow through pressure drag, modelled flow over hills with resolved canopies will separate
1011 at much lower slopes than would flow over the same hill with an unresolved rough surface
1012 with the same z_0 . A series of LES studies of hill-canopy flow followed. Tamura et al. (2007)
1013 simulated flow over an isolated three-dimensional canopy-covered hill and found that the
1014 flow took longer to recover from separation in the canopy case than in a smooth hill case.
1015 Next, Ross (2008) used LES to study the influence of 2D ridges on exchange processes at the
1016 canopy level and found that the turbulence is dominated by sweep/ejection events just as in
1017 homogeneous canopies but that the structure changes across the hill according to hill-induced
1018 modification of the mean flow.

1019 Dupont et al. (2008) confirmed some additional flow characteristics predicted by the
1020 FB04 linearized theory, particularly that within-canopy flow accelerations on the upwind
1021 slope resulted in reduced canopy-top mean shear, that enhanced canopy-top shear at hill crest
1022 was responsible for increased turbulence kinetic energy production there and that, although
1023 canopy exchange occurs through similar mechanisms on either side of the hill, structures on
1024 the windward side of the hill are not correlated with those in the lee. From this they infer that
1025 such structures initiated on the windward slope end up being advected downstream of the hill
1026 in the region above the separation zone.

1027 Patton and Katul (2009) investigated vegetation density influences on second-order flow
1028 statistics over gentle sinusoidal 2D ridges with key findings that included the fact that
1029 restricted domain heights can influence phase relationships between hill-induced
1030 perturbations in mean velocity and velocity variances, that an order-of-magnitude increase in
1031 canopy density does not significantly alter the broad phase relationship between pressure and
1032 the topography but can shift the pressure minimum downstream sufficiently to increase the
1033 hill-induced pressure drag by about 15%, and that hill-induced regions of increased turbulent
1034 momentum flux create regions of high amplitude pressure fluctuations. Ross (2011) used
1035 LES to study the transport of scalars emitted at a specified rate by a resolved canopy on a hill
1036 and found that those hill-induced pressure forces act like a pump to efficiently remove scalars
1037 from the canopy space, thereby reducing mean within-canopy scalar concentrations overall.
1038 However scalar concentrations exhibit high spatial variability with respect to location over
1039 the hill (see Section 4. below).

1040

1041 **3.5. Scalar transport in canopies on Hills**

1042 Once direct measurement of carbon and energy exchange over hills using fast response
1043 sensors on ‘flux towers’ had become widespread significant problems began to appear.
1044 Uncorrected eddy flux measurements were often biologically unrealistic (Finnigan, 2008) and
1045 so to address this problem systematically in the early 2000s, detailed field experiments
1046 commenced at a number of sites in Europe (later formalized in the ADVEX initiative), where
1047 topographic complexity and canopy structure varied across sites (e.g. Lee and Hu, 2002;
1048 Feigenwinter et al., 2004). These studies showed that much of the imbalance between
1049 absorption or release of CO₂ and its vertical eddy flux is caused by the advective terms,
1050 especially under near neutral and unstable atmospheric stratification Aubinet et al. (2010).
1051 To further guide these field experiments and contribute to understanding the genesis of such
1052 imbalance, several model investigations of scalar transport in tall canopies on gentle hills
1053 were initiated.

1054 Just as the FB04 model was developed as an extension of HLR88, and motivated by
1055 the same questions posed by direct eddy flux measurements of carbon exchange, a canopy
1056 extension of RWCH88 was developed for the case of a concentration boundary condition on
1057 the foliage (Finnigan, 2006). Like the momentum field in FB04, the scalar perturbation in the
1058 canopy was divided into a linearised upper-canopy solution and a non-linear, lower-canopy
1059 solution, which were matched asymptotically. In the upper canopy, on scaling grounds, the

1060 scalar conservation equation reduces to a balance between the scalar eddy flux divergence
 1061 and the perturbation to the canopy scalar source or sink.

1062 In the lower canopy the scalar flux divergence becomes small, however, a sensible
 1063 velocity perturbation ΔU continues to drive the scalar source term so that the conservation
 1064 equation becomes a balance between advection of c along streamlines and the scalar source
 1065 strength. Like its momentum equivalent in FB04 the lower canopy equation is non-linear but
 1066 for a different reason. At leaf level, the leaf boundary layer conductance g_b , which for a
 1067 constant concentration boundary condition controls the source strength, depends only on the
 1068 magnitude of the wind velocity and not on its direction so that we must write $g_b = A|\bar{u}|^n$,
 1069 where A is a constant depending on leaf morphology and n is an exponent between 0.5 and
 1070 0.8 (Finnigan and Raupach, 1987). In the upper canopy, where $U_b > \Delta U$ this dependence on
 1071 absolute velocity need not be made explicit, but in the lower canopy, where $U = \Delta U$, it is
 1072 critical. Setting $n=1$ allows the equations to be solved analytically with results that are
 1073 qualitatively the same as for the fractional values of n .

1074 Two useful results follow from the analytic solution. First, the typical magnitudes of the
 1075 velocity and scalar perturbations within the canopy, U_c and C_c , are respectively,

$$1076 \quad U_c = \frac{U_0^2 H L_c}{U_h L^2}; \quad C_c = r U_c \frac{C_h}{U_h} \quad (8)$$

1077 The magnitude of the velocity perturbation depends on the driving pressure gradient, which is
 1078 $O[U_0^2 H / L^2]$, where U_0 is defined as the background velocity at the middle layer height,
 1079 (Fig. 2) i.e. $U_0 = U_b(h_m)$ and so is determined by the outer layer flow as well as by the
 1080 momentum absorption in the canopy, which is characterized by L_c and U_h . Note that
 1081 subscript h refers to values at the canopy top, $z=h_c$. With the choice of a constant
 1082 concentration boundary condition on the foliage surface, the scalar perturbations are caused
 1083 entirely by the windfield and not by variations in the source/sink strength and we see that
 1084 they are relatively smaller than the velocity perturbations that drive them, the proportionality
 1085 factor being the leaf-level Stanton number, r , which was introduced in Section 3.2. Second,
 1086 changes to the concentration and flux fields above the canopy in the region where eddy flux
 1087 towers are usually located, lead to changes in the relative phases of horizontal and vertical
 1088 advection sufficiently large that a measurement of the vertical eddy flux at a point can differ

1089 from the area average flux by around +/-50% (Finnigan, 2006). To test this result further we
1090 need to model the more complicated surface boundary conditions that control biologically
1091 active scalars like CO₂ and water vapour and for this, numerical implementations of the
1092 linearised scheme or fully numerical models are necessary.

1093 At roughly the same time as the linear analytic model development, Katul et al.
1094 (2006) produced a study where the analytic velocity field of FB04 was used to drive a scalar
1095 transport model with realistic parameterisation of energy and CO₂ exchange at the leaf
1096 surfaces and which was applied to a hill sufficiently steep that a separation bubble almost
1097 filled the canopy in the lee of the crest. The effects of topography and canopy on radiation
1098 attenuation were also considered and included in the leaf gas exchange equations. This more
1099 realistic but still simple model confirmed that streamwise and vertical advection are
1100 individually much larger than the biological sinks (leaf-area weighted photosynthesis) at
1101 many positions within the canopy and across the hill. As in the fully analytic model of
1102 Finnigan (2006), the two advective terms are usually opposite in sign but do not precisely
1103 cancel each other locally even when averaged across the hill. The imbalance between them is
1104 sufficiently large to decouple the local canopy photosynthesis from the local turbulent flux,
1105 implying that linking tower-based eddy-flux measurements to local biological sources and
1106 sinks on hilly terrain is difficult to impossible without knowledge of both advective terms.

1107 The flow convergence caused by the non-linearity of the momentum and scalar solutions in
1108 the lower canopy described in Section 3.1, forces a plume of concentrated scalar to be ejected
1109 just behind the hill crest. In Figure 7 this is illustrated by a numerical solution of the scalar
1110 equations driven by the FB04 model for a very gentle hill (maximum slope 3 de.g.) covered
1111 with a tall dense canopy. The solution assumes a constant scalar source of unit strength and
1112 the appearance of a plume of low vertical eddy flux $\overline{w'c'}$ behind the hill crest is clear as is
1113 the compensating horizontal and vertical advection terms and a smaller contribution from the
1114 divergence of the horizontal turbulent flux of c .

1115 More accurate quantitative modelling by Ross (2011), using a Lagrangian transport
1116 scheme embedded in the 1.5 order closure implementation of the BLASIUS model (Ross and
1117 Vosper, 2005) revealed the first-order effect of canopy source distribution on the overall
1118 transport and its spatial distribution and confirmed the qualitative results of the analytic
1119 modelling described above. In addition, Ross (2011) was able to assess the contribution of
1120 higher order terms in the turbulent transport of the scalar relative to advection more

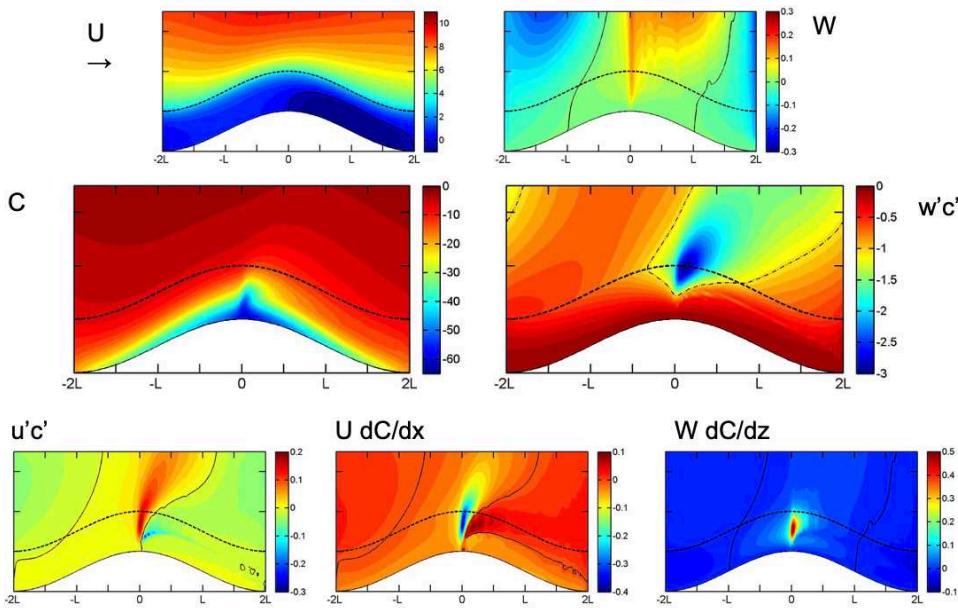
1121 accurately. Ross and Harman (2015) used the same RANS modelling framework to
1122 investigate the ecologically important case of a ground respiration source of CO₂ combined
1123 with an upper-canopy assimilation sink, which is the typical state of a photosynthesising
1124 canopy during daytime. They modelled a very gentle hill, which would not a priori be
1125 expected to generate significant advective errors in any daytime eddy flux measurements, but
1126 showed that the differential advection in the lower and upper canopy would lead to
1127 significant underestimation of daytime carbon assimilation from a flux tower placed on a hill
1128 top, which is where towers are usually located.

1129 Most of these examples involved relatively gentle hills but the presence of a
1130 recirculation zone within the canopy, which occurs even on quite gentle hills, if the canopy is
1131 deep and dense enough, has a large impact on scalar transport. This was the focus of recent
1132 LES work by Chen et al. (2019), who generalised the concept of differential advection by
1133 exploring passive and reactive scalar dispersion within canopies on gentle and steep hills,
1134 using Lagrangian particle tracking. In neutral flow conditions, two main pathways were
1135 identified for parcels of air to be transported out of the canopy volume: a “local path-way”,
1136 corresponding to nearly vertical transport out of the canopy by turbulent ejection events with
1137 some lateral displacement associated with finite mean velocity, and an “advection path way”,
1138 corresponding to parcels that travel horizontally from the source towards the recirculation
1139 zone on the lee side of the hill and reside at the separation point until they are transported out
1140 of the canopy by turbulence. The dominance of one pathway over the other was primarily
1141 determined by the relative time-scales for vertical transport by turbulence (dictated by $\overline{w'^2}$)
1142 and mean-wind advection and the height of scalar release.

1143 For a wide range of hill and canopy conditions, it was shown that the local pathway is
1144 dominant for scalar releases in the upper part of the canopy whereas the advection pathway is
1145 dominant in the lower part. A major consequence of the advection pathway is that almost all
1146 source locations contribute air parcels to the total escape at the separation point, resulting in
1147 the ‘chimney’ effect that can be clearly seen in Figure 7. Sources near the ground will
1148 contribute more than sources in the upper canopy, but the collection of parcels from all
1149 source locations leads to a large total escape at the separation point on the lee of the hill crest.
1150 This pathway results in a probability density function of escape locations displaying a strong
1151 peak at the separation point, supporting the observations of elevated concentrations and
1152 fluxes in other models (Katul et al., 2006; Ross, 2011; Ross and Harman, 2015).

1153 The LES results also show that vertical transport in the recirculation region is
 1154 performed predominantly by turbulence, giving rise to the intermittent accumulation–ejection
 1155 cycles observed in the flume experiments of Poggi and Katul (2007a) (see section 7 below).
 1156 However, near the ground, where vertical velocity fluctuations are damped, mean vertical
 1157 advection contributes significantly to transporting air parcels upward to levels of more
 1158 intense turbulence, where they can then be readily transported out of the canopy by ejections.
 1159 This vertical advection is responsible for a reduction in residence times for gases emitted in
 1160 the bottom part of the canopy, as compared to flat terrain conditions, which in turn results in
 1161 a larger escape fraction for reactive compounds. Thus, the increased out-of-canopy transport
 1162 efficiency observed over topography in several studies (Ross, 2011) appears to be caused by
 1163 the small but important effect of mean vertical advection near the ground in the vicinity of
 1164 the separation point or recirculation region.

1165



1166 Red, high values, blue low values, black contour marks the zero line

1167 **Fig. 7** Numerical calculation of scalar transport driven by the RB04 model. Cosine hill
 1168 profile with $H=20\text{m}$, $L=400\text{m}$, $hc=20\text{m}$, $Lc=30\text{m}$, $u^*/U(hc)=0.3$, $S(z)=S_0=\text{constant}$ for
 1169 $hc > z > 0$. $S_0=1$. Figures from Dr I. N Harman.

1170

1171 It is clear that the fundamental non-linearity of scalar (as well as momentum transport)
 1172 that is introduced when even a gentle hill is covered with a tall canopy opens the possibility
 1173 of emergent changes to the mean transport efficiencies of scalars between biosphere and
 1174 atmosphere at landscape scale. For some scalar fluxes, such as the photosynthetic

1175 assimilation of CO₂ by vegetation or the evaporation of water, energy supply provides a
1176 global constraint on the landscape scale flux (Katul et al., 2006; Raupach and Finnigan, 1997)
1177 but for other scalars, net changes in the rate of ventilation of the canopy could lead to
1178 sensible large scale changes (Ross, 2011). As a final comment it is necessary to point out that
1179 almost all these studies of scalar transport involving canopies have been carried out on 2D
1180 ridges. The magnitude of the driving velocity perturbation field is generally smaller on
1181 axisymmetric hills but recent unpublished measurements on 3D hills covered with canopies
1182 indicate much more complex flow patterns can occur in such cases according to unpublished
1183 results by Dr I. N. Harman and Dr E. G. Patton. The impact of these flow patterns on local
1184 exchange is probably significant but has not yet been quantified.

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1186 **4. Gravity Driven Flows**

1187 Studies of slope flows, driven by gravity currents, have also been an important focus of
1188 complex-terrain meteorology, but have developed somewhat independently of boundary-
1189 layer hill flow studies. This disconnect has occurred in part through a separation in the
1190 dominant scales of motions but has been mainly driven by different scope and objectives.
1191 Where hill flow studies aim to understand how orographic features modulate boundary layer
1192 winds, slope flow studies ask how buoyant forcing from surface cooling or heating interacts
1193 with orography to drive slope-scale mean winds and turbulence. In practice, the resulting
1194 winds usually interact with larger scale terrain forcing causing those same near-surface
1195 temperatures and surface fluxes to evolve so that the problem is intrinsically non-linear.
1196 Nevertheless, just as hill flow studies have advanced through analysis of the simple cases of
1197 isolated 2D and 3D hills, many of the fundamental slope-flow studies have focussed on
1198 gravity flows over extensive uniform mountain slopes, over steep valley sides or over ice
1199 sheets and glaciers, treating these ‘simple’ flows as the building blocks of the more complex
1200 interactions that drive mountain wind patterns. Similarly, theory initially concentrated on
1201 idealised situations, where the flow did not affect the forcing that generated it, but as
1202 knowledge has advanced, more interactive situations are being modelled.

1203 Modelling and prediction of the characteristic features of katabatic and anabatic slope
1204 flows dates back nearly eighty years, to the observations and theories of Prandtl (1942).
1205 However, earlier meteorology and weather reports had documented diel slope- and valley-
1206 wind patterns, mostly in the 1920s and 1930s in the European Alps (e.g., Wagner 1938), and
1207 their existence must have been known much earlier to explorers, mountaineers and

1208 communities living in mountain regions. Slope flows occur most often with weak synoptic
1209 forcing under clear skies when virtual potential temperature differences between the surface
1210 and the adjacent air are greatest. When the sloping surface is cooler than the ambient air, for
1211 example at night as a result of radiative cooling, the downhill component of the hydrostatic
1212 pressure gradient generates downslope density currents or drainage flows whereas daytime
1213 surface heating generates upslope, or anabatic flow (Hahn 1981; Catalano and Moeng 2010).

1214 In transitional periods during mornings and evenings, flow reversals can occur that
1215 last from a minute to an hour or more (e.g., Bader and Mckee 1983; Papadopoulos and
1216 Helmis 1999; Nadeau et al. 2012, 2018; Fernando et al. 2013; Zardi and Whiteman 2013;
1217 Jensen et al. 2017). Transitional regimes are typically characterized by quiescent winds prior
1218 to flow reversal and very weak turbulence. Over mountainous terrain, these periods can be
1219 difficult to compare or translate between sites because they can be triggered locally by
1220 topographical shading effects, propagating shadow fronts or localized insolation
1221 (Papadopoulos and Helmis 1999; Nadeau et al. 2012; Jensen et al. 2017). In general,
1222 transitional flows can exhibit a range of behaviours, driven by a variety of multi-scale
1223 mechanisms and more observational, theoretical and numerical studies will be necessary to
1224 understand these regimes better.

1225 An excellent introduction to the basic physics of katabatic slope flows is provided by
1226 Mahrt (1982). Starting from the assumption that the flows are driven almost entirely by the
1227 surface energy balance, he was able to clarify the different assumptions implicit in earlier
1228 modelling and analysis. He used a rigorous scale analysis to classify different regions of the
1229 parameter space spanned by the downslope velocity scale, U_s , the slope length scale, L_s , the
1230 characteristic depth of the gravity current, H_s , the characteristic potential temperature deficit
1231 of the cool layer, $\Delta\bar{\theta}$, and α , the (positive) angle the slope makes with the local geopotential
1232 surface. By further restricting attention to ‘nearly stationary’ flows, where inertial
1233 acceleration and advective effects roughly balanced the hydrostatic forcing and by also
1234 ignoring large scale Ekman-Gravity flows, where Coriolis effects were important (e.g.,
1235 drainage flows over large ice sheets in Greenland and Antarctica; see Parish and Cassano,
1236 2003), he was able to cover much of the parameter space corresponding to the scales we have
1237 considered in boundary layer hill flows. In this ‘nearly stationary’ regime, gravity currents
1238 could be roughly divided into ‘Tranquil Flows’, where the Froude Number based on the
1239 gravity current depth, $F_{H_s} = U_s^2 / \left[H_s \left(g \Delta\bar{\theta} / \Theta_0 \right) \right]$ is small and ‘Shooting Flows’, where F_{H_s}
1240 is large. In ‘Tranquil Flows’ the thermal wind term, which is the contribution to the

1241 hydrostatic pressure gradient caused by changes of gravity current depth or temperature
1242 deficit along the slope, is important and tends to oppose the other component of hydrostatic
1243 forcing, the vertical hydrostatic pressure gradient resolved down the slope. This trade-off
1244 tends to keep F_{Hs} low. In general, the thermal wind term can be ignored when $H_s/L_s \ll 1$, a
1245 condition that is satisfied in many or most of the large scale mountain- and hill -slope gravity
1246 currents that have been studied. Shooting flows in contrast are those for which $F_{Hs} \gg 1$,
1247 which implies that $U_s \gg \left[H_s (g\Delta\bar{\theta}/\Theta_0) \right]^{1/2}$. Shooting flows can be further subdivided into
1248 ‘advective-gravity flow’, where the buoyancy term leads to acceleration down the slope and
1249 ‘equilibrium flow’, where the buoyancy terms are approximately balanced by frictional
1250 effects on the ground and at the top of the gravity current. Marht (1982) goes on to derive
1251 useful idealised solutions for further subdivisions of these flow classes but to obtain models
1252 that apply to the more complex boundary conditions found in real life, more realistic
1253 representations of the flow dynamics are required.

1254 **4.1 Localized katabatic slope flows**

1255 A characteristic feature of buoyancy-driven slope flows is a jet-shaped mean velocity profile,
1256 exhibiting an elevated velocity maximum as first described by Prandtl (1942) and observed in
1257 a wide range of subsequent studies (Fig. 8). The jet shape develops as the air layer cooled by
1258 interaction with the cold surface accelerates down the slope but is decelerated by surface
1259 friction from below and the mixing of warmer air from above. Entrainment of the warmer
1260 ambient air also tends to deepen the gravity flow layer in the downslope direction (e.g.,
1261 Manins and Sawford 1979b; Princevac et al. 2005; Grachev et al. 2016). Katabatic flows on
1262 open slopes tend to be extremely shallow so a current extending kilometres or more in the
1263 downwind direction will be only ~10-100-m deep with jet peaks as low as 1m (e.g., Horst
1264 and Doran, 1986; Oldroyd et al. 2014). These shallow flow depths facilitate tower-based
1265 observations, but are extremely difficult to resolve sufficiently in numerical models
1266 (Söderberg and Parmhed 2006). For example, the lowest grid cell in a typical NWP model
1267 could contain the entire katabatic layer. Gravity-driven flows over very large ice sheets, for
1268 example in Greenland and Antarctica, are stronger and deeper than those over isolated
1269 mountain slopes as they develop over long stretches of sloping terrain and hence, Coriolis
1270 forcing also becomes significant (King 1989).

1271 Despite understanding some of the key features of slope flows and how they develop,
1272 systematically predicting their onset, the depth of the flow layer, the height of the jet peak
1273 and its strength remain a challenge. Currently, the biggest barrier to model improvement is
1274 devising better parameterizations of turbulent mixing. This is especially the case for wall
1275 models, which are critical for translating the surface forcing to the overlying atmosphere.
1276 Prandtl's (1942) slope-flow model, which assumed laminar flow and so a constant molecular
1277 kinematic viscosity, ν was initially used as an analogue for simple turbulent flow models that
1278 replaced ν with a constant eddy-diffusivity (e.g., Defant 1949). This so-called 'Prandtl
1279 model' is still used (e.g., Burkholder et al. 2011; Shapiro et al. 2012; Shapiro and Fedorovich
1280 2014) but has also been extended using variable eddy-diffusivities to account for more
1281 complicated turbulence structure (Rao and Snodgrass 1981; Grisogono and Oerlemans 2001;
1282 Parmhed et al. 2004; Giometto et al. 2017a). Additional parameterizations can be included to
1283 model the momentum retardation effects of entrainment at the upper boundary of the cool
1284 layer and its effects on flow depth modulation (e.g., Manins and Sawford 1979b; Princevac et
1285 al. 2005, 2008).

1286 A wide variety of other turbulence parameterizations have also been used, ranging
1287 from two-layer slab models (Manins and Sawford 1979b; Fitzjarrald 1984; Kondo and Sato
1288 1988), through flux-gradient parameterizations based on local MOST closure schemes (Lee
1289 and Kau 1984; Ye et al. 1990; Oldroyd et al. 2014) to RANS closures of $1\frac{1}{2}$ and higher order
1290 (Horst and Doran 1988; Denby 1999; Goger et al. 2018). However, most of these closures
1291 and associated empirical constants were originally derived for horizontal, homogeneous
1292 terrain and cannot, a priori, be expected to apply to wall jet slope flows, where simple surface
1293 layer scaling does not apply (Mahrt 1999; Skillingstad 2003).

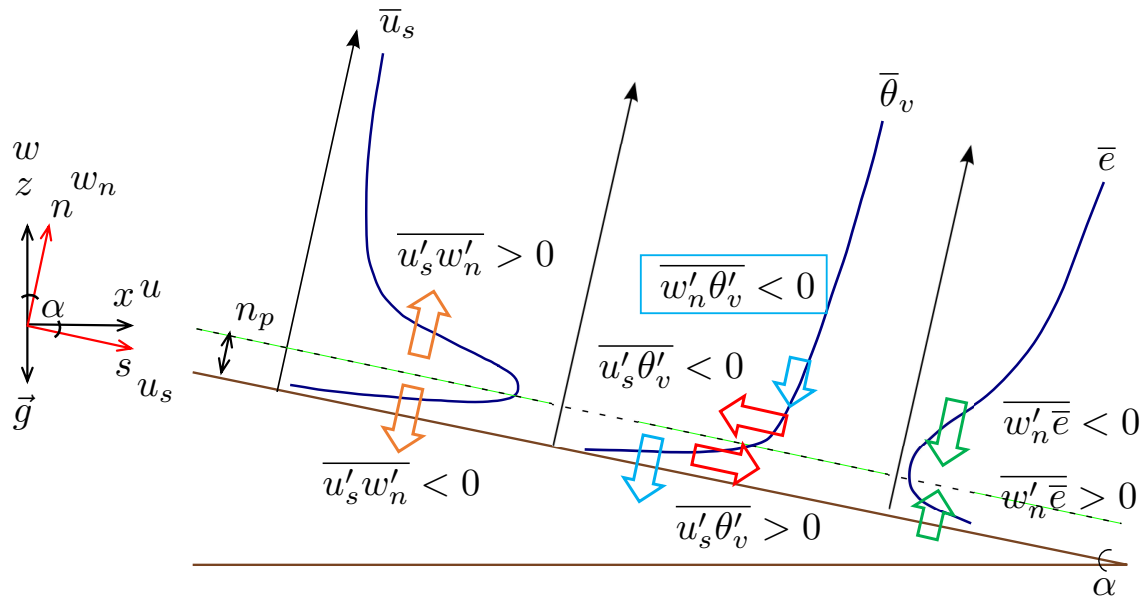
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Fig. 8 Slope flow coordinate systems (black is aligned with gravitational acceleration and red is terrain following) and schematic of mean and turbulent flux quantities for katabatic flow (Horst and Doran 1988; Denby 1999; Grachev et al. 2016). The dashed line indicates the height of the katabatic jet peak, n_p , u_s and w_n are velocities in the downslope (s) and slope-normal (n) directions, respectively, θ_v is virtual potential temperature, and \bar{e} is the mean turbulent kinetic energy (TKE) per unit mass. Block arrows indicate the relative directions of momentum fluxes (orange), buoyancy fluxes (red and blue) and slope-normal turbulent transport of TKE (green).

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The earliest experimental studies of turbulence in katabatic flows had insufficient resolution to reveal the flow structure in complete detail. Despite this limitation, significant deviations from MOST turbulence theory such as large values of eddy flux divergence were observed, even below the jet peak (Horst and Doran 1986). More recent observational studies with higher vertical resolution both below and above the jet peak have confirmed some of the earlier findings, conjectures and theories but also highlight novel features of the turbulence structure and the buoyancy driven dynamics (Nadeau et al. 2012; Oldroyd et al. 2014; Grachev et al. 2016). The following list highlights some of these key features of the turbulence structure as we now understand them:

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- Significant surface-normal momentum flux divergence is characterised by negative momentum fluxes below the jet peak and positive above, crossing zero near the peak, where the streamwise velocity gradient also changes sign (Horst and Doran 1986, 1988; Denby 1999; Smeets et al. 2000; Oldroyd et al. 2014; Grachev et al. 2016).
- Significant surface-normal divergence of the kinematic heat and buoyancy fluxes is characterised by stronger gradients from the ‘bulge’ of the jet down to the surface and a

1323 weaker divergence above (Grachev et al. 2016). Radiative surface cooling causes surface
1324 normal heat fluxes to be negative (or close to zero) throughout the slope-flow layer.

- 1325 • The slope-parallel kinematic heat and buoyancy fluxes also have a tendency to change
1326 sign near the jet peak (Horst and Doran 1988; Denby 1999; Grachev et al. 2016). This
1327 behaviour is expected when the shear and gradient production terms dominate in the rate
1328 equations for the eddy fluxes. Using a slope-aligned coordinate system with positive u_s
1329 directed down the slope (Fig. 8), the slope-parallel buoyancy fluxes are positive below the
1330 jet peak, indicating a warming flux down the slope, whereas they are negative above the
1331 peak, indicating a cooling flux down the slope (Horst and Doran 1988; Denby 1999;
1332 Grachev et al. 2016). The physical implications of this sign change are summarized in
1333 the next point.
- 1334 • Buoyant production or suppression of TKE by the vertical buoyancy flux, $g/\Theta_0 \overline{w'\theta'_v}$,
1335 which on horizontal ground affects only $\overline{w'^2}$ the vertical component of velocity variance,
1336 affects both streamwise $\overline{u_s'^2}$ and surface normal $\overline{w_n'^2}$ components when the variance
1337 equations are rotated into the slope-aligned coordinate frame of Figure 8. Hence, the net
1338 vertical buoyancy term in the TKE equation contains contributions from the slope-normal
1339 buoyancy fluxes, which are negative and act to suppress TKE, and the slope-parallel
1340 buoyancy fluxes, which can be either negative or positive. As a result, the net vertical
1341 buoyancy flux can produce TKE when, $\overline{u_s'\theta'_s}/\overline{w_n'\theta'_n} > \cot \alpha$ leading Horst and Doran
1342 (1988) and Denby (1999) to estimate that buoyant TKE production will occur over slopes
1343 with angles greater than 30° and 25° , respectively. Those estimates assume an
1344 approximately constant ratio of slope-parallel to slope-normal buoyancy fluxes, however,
1345 Oldroyd et al. (2016) show that the buoyancy flux ratio can be highly variable and that
1346 buoyant TKE production may occur over much shallower slopes. This has important
1347 implications for how various stability parameters can be used for turbulence modelling, as
1348 well as for how they should be used (i.e., in the surface-normal versus vertical coordinate
1349 frame) and interpreted (i.e., as a representation of stability versus as a scaling parameter)
1350 (Oldroyd et al. 2016).
- 1351 • Profiles of mean TKE exhibit a local minimum near the jet peak, where shear production
1352 and the slope-parallel buoyancy flux approach zero (Fig. 8) (Horst and Doran 1986;
1353 Denby 1999; Grachev et al. 2016). The sign changes in momentum and slope-parallel

1354 buoyancy fluxes may indicate turbulence decoupling from the surface near the jet peak
 1355 (Horst and Doran 1988; Denby 1999; Grachev et al. 2016). Hence, turbulent transport
 1356 acts to transfer TKE into the bulge of the jet both from below and above, serving as an
 1357 important turbulence coupling mechanism maintaining non-zero TKE at the peak (Arritt
 1358 and Pielke 1986; Smeets et al. 2000; Söderberg and Parmhed 2006; Giometto et al.
 1359 2017b). This is analogous to the situation in the lower part of a plant canopy flow, where
 1360 shear production is small and TKE is maintained by transport from the region of strong
 1361 shear production at canopy top (Finnigan, 2000). The critical inference is that, if both
 1362 TKE and eddy fluxes near the jet peak are maintained by third moment transport terms,
 1363 local eddy diffusivity-type closures are bound to fail there.

- 1364 • In contrast to the shear production term for TKE, the gradient of mean virtual potential
 1365 temperature and the slope-normal temperature flux are both large at the jet peak so the
 1366 dominant gradient production term in the rate equation for the variance of virtual
 1367 potential temperature, $\overline{\theta_v'^2}$, remains large and profiles of $\overline{\theta_v'^2}$ exhibit a local maximum
 1368 near the jet peak (Denby 1999; Grachev et al. 2016). Compared to TKE, $\overline{\theta_v'^2}$ and its
 1369 relation to turbulent potential energy (Zilitinkevich et al. 2009; Łobocki 2017) has
 1370 received much less attention with the exception of variance similarity scaling efforts as
 1371 discussed next.

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1373 **4.2 Similarity scaling for katabatic flows**

1374 Clearly, the turbulence structure of slope flows does not conform to traditional horizontally
 1375 homogeneous surface-layer behaviour. This poses several modelling challenges. Most
 1376 critical, is that traditional MOST (Monin and Obukhov 1954; Obukhov 1971), which is used
 1377 in nearly all NWP models in some form (Foken 2006), and the associated empirical
 1378 parameterizations developed over idealized terrain (e.g., Businger et al. 1971; Dyer 1974)
 1379 break down. As an alternative to MOST, local similarity scaling (Nieuwstadt 1984a, b) has
 1380 been applied to slope flows but with mixed results for turbulent momentum fluxes and
 1381 variances and very poor results for heat fluxes, especially at higher stabilities (Forrer and
 1382 Rotach 1997; Smeets et al. 2000; Heinemann 2004; Nadeau et al. 2013; Stiperski and Calaf
 1383 2017; Sfyri et al. 2018; Stiperski et al. 2019). These mixed results with local scaling have
 1384 prompted the use of other characteristic length scales, such as the height of the jet peak (van
 1385 der Avoird and Duynkerke 1999; Smeets et al. 2000; Söderberg and Parmhed 2006) or local

1386 z -less (or n -less) scaling, which been shown to work relatively well for dimensionless
1387 velocity gradients $(\kappa z/u^*)\partial U/\partial z$ (or equivalently $(\kappa n/u^*)\partial U_s/\partial n$) above the jet peak,
1388 where turbulence is somewhat decoupled from the surface (Forrer and Rotach 1997; Grachev
1389 et al. 2016). However, a major challenge when using the height of the jet peak (or z -less
1390 scaling for regions above the peak) as the characteristic length scale is that this height is
1391 unknown a priori and can vary with time, stability, jet strength and distance along the slope.

1392 Summarising, while local similarity scaling has had mixed success, nearly all attempts
1393 at applying horizontal-terrain scaling parameterizations for the heat fluxes underestimate the
1394 turbulent mixing of the dimensionless temperature gradient $(\kappa z/\theta^*)\partial\bar{\theta}/\partial z$ that occurs with
1395 relatively high stability ranges (typically $n/\Lambda > 1$, where Λ is the local, surface-normal
1396 Obukhov length). Furthermore, there is some evidence that turbulent mixing associated with
1397 the heat fluxes can, in some cases, actually increase with stability (Forrer and Rotach 1997;
1398 Smeets et al. 2000), which is most probably due to an accompanying increase in the slope-
1399 parallel buoyancy fluxes and subsequent reduction of buoyant TKE suppression or even TKE
1400 production, as discussed above; however, this has not yet been rigorously established from
1401 observational studies. Consequently, ‘universally’ appropriate turbulence length scales and
1402 velocity scales for katabatic flow are still being actively debated. Finally, a scaling
1403 framework for other scalars such as water vapour or CO_2 in katabatic flows has rarely been
1404 studied over bare or sparsely vegetated slopes (Forrer and Rotach 1997; Nadeau et al. 2013).
1405 Hence there is a set of critical open questions for katabatic flows, whose answers could have
1406 significant positive impact impacts on our modelling capabilities for complex topography
1407 flow in general.

1408

1409 **4.3 LES modelling of gravity flows**

1410 We discussed the development of turbulence closure parameterisations suitable for RANS
1411 modelling approaches in Section 4.1. However, just as in the case of hill flows (Section 2.3),
1412 in recent years LES has been applied to slope flows to try to resolve some of the difficulties
1413 listed above. We have already referenced (Section 2.3) the early efforts of Schumann (1990),
1414 Walko et al. (1992) and Dörnbrack and Schumann (1993), who addressed different aspects of
1415 buoyancy driven flow on slopes. Of more general application is the finding of Burkholder et
1416 al. (2011), who determined that, although their simulated mean fields were insensitive to the
1417 choice of sub-filter-scale model, the model choice did substantially impact the simulated
1418 second-order moments, especially the buoyancy fluxes and vertical velocity variances.

1419 Skyllingstad (2003) used LES to demonstrate the role turbulence plays in controlling
1420 the strength and depth of katabatic flows while Smith and Skyllingstad (2005) studied the
1421 influence of multi-angle slopes on katabatic flows. They found that on a steep
1422 upper slope followed by a gentler lower slope, a rapid acceleration was
1423 generated on the upper slope followed by a transition to a slower evolving structure
1424 characterized by an elevated jet over the lower slope. In contrast, a case with uniform slope
1425 having the same total height change yielded a more uniform slope flow profile with stronger
1426 winds at the slope bottom. As well as these two modelling efforts, Axelsen and van Dop
1427 (2009a, b) simulated katabatic flows observed on glaciers while Grisogono and Axelsen
1428 (2012) compared their LES with the classic Prandtl modelling approach and ranked the
1429 reasons for the departures between them. All these models relied on MOST-based wall
1430 models, and so needed high vertical resolution below the jet peak to reduce the relative
1431 dependence on the MOST scaling assumptions. Over rough surfaces, this can generate
1432 problems with prescribing a ‘surface-layer’ model within the roughness sublayer (Basu and
1433 Lacser 2017).

1434 Although generally limited to low Reynolds numbers, direct numerical simulations
1435 have also been performed (Shapiro and Fedorovich 2014; Umphrey et al. 2017). These tend
1436 to overpredict jet strength and under-predict the height of the jet peak. However, a recent set
1437 of DNS data at very high Grashof Number has delivered detailed information on both
1438 katabatic and anabatic flows of real relevance (Giometto et al. 2017b). That said, better
1439 understanding of the turbulence structure in katabatic flows and how best to model it remain
1440 critical open questions and we expect LES and even DNS to play an increasingly important
1441 role in answering them.

1442

1443 **4.4 Gravity-driven flows in Plant Canopies**

1444 As emphasised in Sections 2 and 3, measurement of carbon exchange from flux towers has
1445 been a major driver of boundary layer hill flow research for the last two decades and the
1446 study of katabatic flows in canopies on relatively gentle complex terrain has been a necessary
1447 component of this. These flows tend to decouple fluxes of CO₂ from the soil surface at night
1448 from the eddy flux measured on towers above the canopy and can lead to significant errors in
1449 24 hour carbon budgets (Goulden et al. 2006; van Gorsel et al, 2011). Steeper canopy-
1450 covered slopes are also widespread in mountainous areas and have been studied as
1451 components of larger complex terrain field campaigns (e.g. van Gorsel et al., 2003). Unlike
1452 katabatic flows on open slopes, where the depth of the flow is set by a delicate balance

1453 between cooling from the ground and the entrainment of warmer air from above, tall closed
1454 canopies like forests that are radiatively cooling at night develop very stable buoyancy
1455 profiles in the crown space that quench and decouple the canopy turbulence from the
1456 boundary layer above (see Figure 6 and the accompanying discussion in Section 3.2 *et seq.*).

1457 Modelling of these flows has proceeded along several fronts. Watanabe (1994)
1458 calibrated one- and two-layer slab models of radiative and convective heat and momentum
1459 transfer in a canopy against a multi-layer model and then applied it to drainage flow but the
1460 vertical integration necessary to define his slabs precluded a detailed treatment of some of the
1461 important physics. Hatcher et al., (2000) studied a physical model of a turbulent gravity
1462 current through a canopy of obstacles in a flume, producing similarity solutions that can be
1463 used to describe the initiation of the slope flow in the atmospheric case. Here we will discuss
1464 the more directly relevant analysis of Belcher et al. (2008), who extended the FB04 canopy-
1465 on-hill model by adding a hydrostatic pressure gradient term, $g\bar{\theta}/\Theta\sin\alpha$ to the streamwise
1466 momentum equation. Because their canopy was assumed to be of constant height h_c and this
1467 was also assumed to be the top of the gravity current, $H_s = h_c$, the depth of the cool layer is
1468 constant so the thermal wind term was ignored. In their analysis Belcher et al. (2008) also
1469 equated the gravity current lengthscale L_s with the hill length scale L .

1470 In forests, the time scale for the night-time radiative cooling of the canopy air
 1471 layers is typically about four hours (Watanabe 1994). Within the canopy, the
 1472 characteristic timescales of turbulent mixing h_c/u^* , and advection, $L/U(h_c)$, are much
 1473 shorter, and so the flow caused by the cooling can be treated as if it is steady over time
 1474 scales short compared with the radiative cooling time. As shown in Figure 6, the cooling
 1475 leads to high Richardson numbers, and collapse of the turbulent mixing within the canopy
 1476 and between the canopy and overlying boundary layer. Hence the canopy flow can be
 1477 estimated by considering only the pressure forces acting on the canopy airspace. For
 1478 dense canopies radiative losses occur predominantly from the top of the foliage, while the
 1479 soil remains warmer, resulting in an unstable temperature profile between soil surface
 1480 and crown (Kaimal and Finnigan 1994: Chapter 3). The air in the trunk space is
 1481 convectively mixed, therefore, but entrainment of cooler air from above causes this
 1482 mixed region to cool progressively. As a result, while the lower canopy may be locally
 1483 unstably stratified, it is cooler than air at the same geopotential height outside the canopy
 1484 and so subject to downslope buoyancy forces. on both sides of the hill crest.

1485 As we saw in Section 2, neutral flow over a hill covered with a forest canopy
 1486 generates a hydrodynamic pressure gradient, which also drives flow within the canopy.
 1487 For small positive stability, the hydrodynamic pressure gradient changes little from that
 1488 found for neutral flow over a hill and so within the canopy it drives flow perturbations
 1489 towards the hill top on both sides of the crest. Hence, the hydrodynamic pressure gradient
 1490 opposes the hydrostatic pressure gradient. Under these conditions the air within the
 1491 canopy flows up the hill slopes toward the crest if the hydrodynamic pressure dominates,
 1492 but flows down the slopes away from the crest if the hydrostatic pressure gradient
 1493 dominates. The competition between these two processes, and the onset of drainage flows
 1494 within the canopy, is measured by constructing their ratio.

1495 The hydrodynamic pressure gradient generated by the flow over the hill can be
 1496 estimated to be $d\Delta p/dx \approx U_0^2 H/L^2$ (Belcher et al., 2012) whereas the hydrostatic
 1497 pressure gradient is estimated as $(g\Delta\bar{\theta}/\Theta_0)\sin\alpha \approx (g\Delta\bar{\theta}/\Theta_0)H/L$ so their ratio can be

1498 written as $R_p = \frac{U_0^2 \Theta_0}{g L \Delta\bar{\theta}} \approx F_L^2$ which is effectively the square of the Froude Number F_L .

1499 The big surprise here is that, while the slope of the hill H/L enters the estimate of both the
1500 hydrostatic and hydrodynamic pressure gradients, it cancels from their ratio. The
1501 outcome of the competition between the tendency for the hydrodynamic pressure gradient
1502 to force flow up the slopes against the tendency for the hydrostatic pressure gradient to
1503 force flow down the slopes is determined by the characteristic wind speed in the HLR88
1504 outer layer U_0 , the average temperature deficit of the air within the canopy, $\Delta\bar{\theta}$, and the
1505 length of the slope, L . Drainage flows occur when R_p drops below a threshold, which we
1506 expect to be about 1. For more stable conditions, the magnitude and form of the
1507 hydrodynamic pressure gradient also changes. The progressive impact of this on the
1508 velocity field that develops over a hill has been extensively studied in the context of
1509 mountain flows (Queney 1948; Scorer 1949; Kaimal and Finnigan 1994, Ch. 5).

1510 Figure. 9 shows the variation of the hydrodynamic and hydrostatic pressure
1511 gradients in flow over 2D sinusoidal ridges and valleys, for a range of Froude Numbers,
1512 F_L . At neutral stability ($F_L \sim \infty$) the hydrodynamic pressure gradient is antisymmetric
1513 about the hill, driving upslope flow within the canopy on both sides of the crest. As the
1514 stratification increases, F_L decreases and the hydrodynamic pressure forcing becomes
1515 smaller, while the hydrostatic pressure forcing increases so that by $F_L=1.2$, they almost
1516 cancel. For, F_L close to but just above 1 (Fig. 9c) the critical value of R_p has been
1517 exceeded and the net pressure gradient forces drainage currents down both the slopes.
1518 The hydrodynamic pressure becomes symmetric about the crest at $F_L \approx 1$ (Fig. 9d). As
1519 the Froude number reduces further, the flow above the canopy is stable enough to support
1520 gravity waves. The resulting hydrodynamic pressure gradient now augments the
1521 hydrostatic forcing and tends to increase the strength of the drainage flows (Fig. 9e).

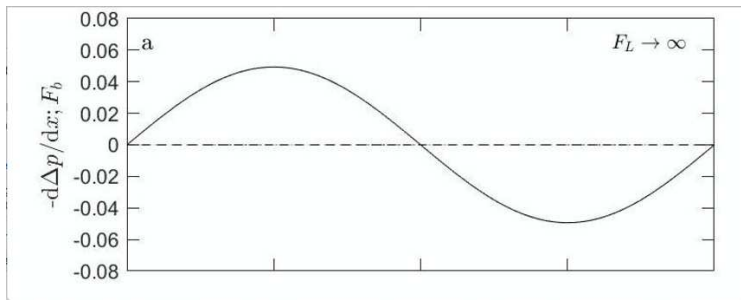
1522 The effect of these opposing pressure gradient forces on the flow within and just
1523 above the canopy is illustrated in Figure 10. For near neutral flow ($F_L \geq 1.5$) we see
1524 convergence of the flow perturbations to the hill top within the canopy while above the
1525 canopy the maximum stream-wise perturbation is just above the crest as seen in Figures 3
1526 and 4 of Section 3. As the influence of stability starts to make itself felt, the hydrostatic
1527 pressure gradient dominates near the crest and the convergence at the crest splits and
1528 moves up and downwind ($1 \leq F_L \leq 1.5$) until by $F_L = 1$, the gravity current dominates

1529 within the canopy and the flow perturbations above are starting to weaken. When the
1530 hydrodynamic pressure gradient flips around ($0.66 \leq F_L \leq 1$), the two pressure gradient
1531 forcings are in the same direction over part of the hill, and we see that the perturbation
1532 above the canopy is in the same direction as within. For even stronger stabilities
1533 ($F_L \leq 0.66$) the gravity current is dominant within and above the canopy but we should
1534 be cautious about pushing the fundamentally linear FB04 analysis to Froude numbers any
1535 smaller than this.

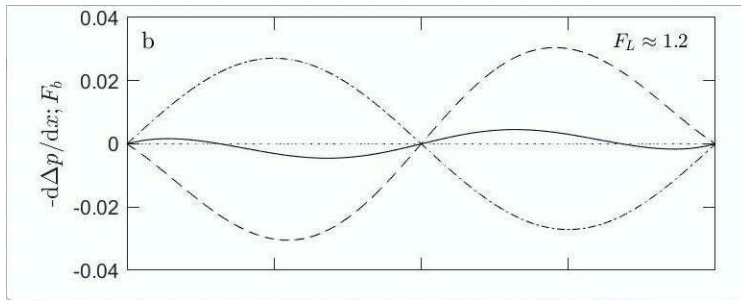
1536 The small timescale of turbulent adjustment compared to that of radiative cooling
1537 has permitted us to describe the evolution of the gravity current in the canopy as a
1538 sequence of steady states, however, in nature, other effects can confound this simple
1539 picture. For example, once the within-canopy flow near the crest starts to diverge
1540 (Fig. 10; $1 \leq F_L \leq 1.5$), conservation of mass demands that warmer air must be entrained
1541 into the canopy at the hill top, reducing the temperature deficit of the canopy airspace,
1542 $\Delta\bar{\theta}$ and so decreasing the hydrostatic pressure gradient. Anecdotally it is often observed
1543 that forest gravity currents accelerate after initiation but then weaken. Sometimes this
1544 effect can lead to oscillations throughout the night. As a final comment, it is also
1545 regularly observed that the gravity current can propagate upwind on level ground for
1546 many hill lengths as long as the within-canopy Richardson Number is large and turbulent
1547 entrainment of warmer air into the current is small. This phenomenon was observed in
1548 the wind tunnel simulation of Finnigan and Hughes (2008) (see Section 7 below) and has
1549 been noted at a flux tower site in the Amazon (Prof. Y. Malhi. *pers. comm.*).

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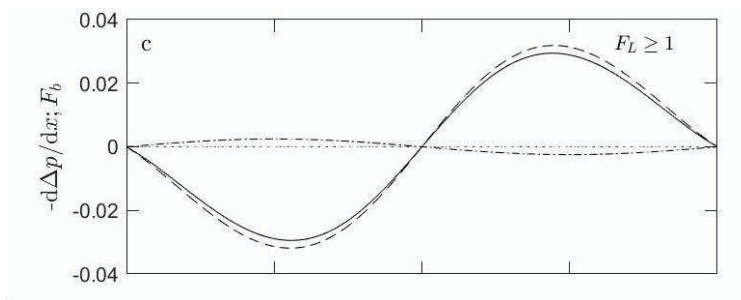
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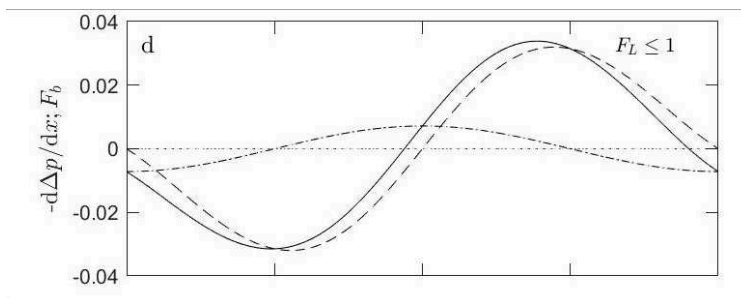
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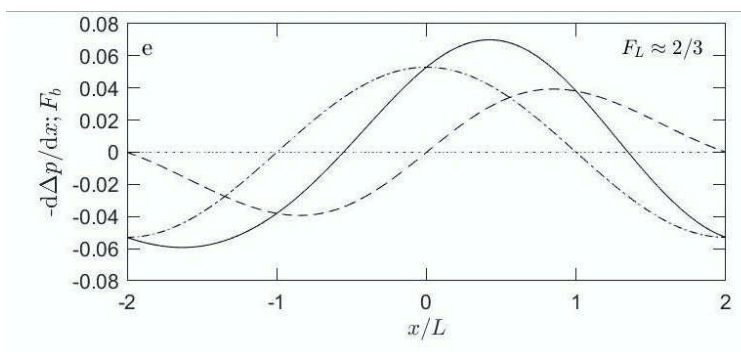
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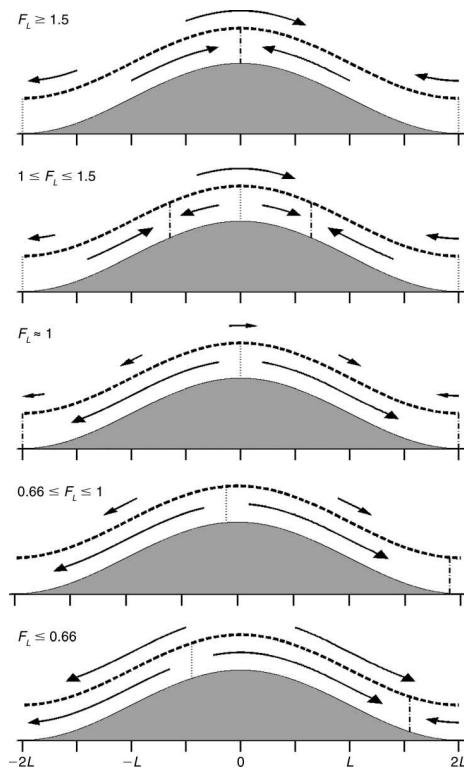


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1565 **Fig. 9** Hydrodynamic $-d\Delta p/dx$ and hydrostatic Fb pressure gradients within a canopy on a sinusoidal
 1566 2D ridge as the stability of the boundary layer varies.

1567 (a) $F_L \sim \infty$, (b) $F_L = 1.2$, (c) $F_L \geq 1.0$, (d) $F_L \leq 1.0$, (e) $F_L \approx 2/3$ The dashed line is the
 1568 hydrostatic pressure gradient, Fb, the dash dotted line (-) the hydrodynamic gradient, $-d\Delta p/dx$ and the
 1569 solid line shows the net forcing. Indicative pressure gradients in units of Pa/m for a hill of the scale shown
 1570 in Figures 3 and 4 are given on the y axis.

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1575 **Fig. 10** Schematic showing forces on flow deep within and just above the canopy on a sinusoidal ridge
 1576 from Figure 9. For a range of boundary layer stabilities as quantified by the Froude Number F_L . Above the
 1577 canopy top (dashed line) arrows indicate the hydrodynamic pressure gradient only. Within the canopy
 1578 (below the dashed line) arrows indicate the balance between the hydrodynamic and hydrostatic pressure
 1579 gradients. Arrow lengths indicate the magnitude of the forcings. Dotted lines indicate regions of divergence
 1580 within the canopy (and descent at canopy top). Dash-dotted line indicate regions of convergence in the
 1581 canopy and ascent at canopy top.

1582

1583 **4.5 Gravity driven flows in a wider context**

1584 In mountainous regions, buoyancy-driven valley and slope flows (Whiteman, 1990;
1585 Schmidli and Rotunno, 2010) often interact and significantly impact one another (Manins
1586 and Sawford 1979a; Mahrt and Larsen 1982; Fitzjarrald 1984; Arritt and Pielke 1986;
1587 Horst and Doran 1988; Doran et al. 1990; Mahrt and Larsen 1990; Doran 1991;
1588 Amanatidis et al. 1992; Smeets et al. 1998; Mahrt et al. 2001; Lehner et al. 2015; Mahrt
1589 et al. 2018). Valley flows and ambient winds can deflect slope flows from their natural
1590 fall lines, inducing directional shear (Manins and Sawford 1979a; Kottmeier 1986; Horst
1591 and Doran 1988) or can be strong enough to smear or overtake the slope-scale flow
1592 entirely (Mahrt and Larsen 1990; Litt et al. 2015). Down valley flows tend to limit the
1593 depth of the slope side wall inversions (Doran et al. 1990) and generate a ‘skin-flow’, or
1594 shallower, sheltered drainage flow, as opposed to a deeper, ‘pure’ katabatic flow (Manins
1595 1992; Mahrt et al. 2001).

1596 Katabatic flows can also contribute to the development of cold pools, in which cold
1597 air stagnates in basins and valleys (Gryning et al. 1985; Mahrt et al. 2010; Burns and
1598 Chemel 2014; Geiss and Mahrt 2015; Foster et al. 2017). Subsequently in the presence
1599 of a cold pool, katabatic flows can ‘peel off’, or intrude, into the cold pool (Mahrt et al.
1600 2010; Whiteman et al. 2010; Haiden et al. 2010; Soler et al. 2014), generating waves
1601 (Burns and Chemel 2014, 2015) and instigating sloshing (Lehner et al. 2015) or seiches
1602 (Lareau and Horel 2015). Additionally, when approaching cold pools or adjacent
1603 horizontal surfaces (plains or oceans), a katabatic jump (analogous to a hydraulic jump)
1604 may develop, generating turbulence and strong vertical motions (Gallée and Schayes
1605 1992; Yu et al. 2005; Yu and Cai 2006, Mayr et al. 2007. Interactions between slope and
1606 valley flows can generate significant vertical transport, cross-valley circulations
1607 (Hennemuth 1986; Kuwagata and Kimura 1997; Weigel et al. 2007; Choukulkar et al.
1608 2012; Arduini et al. 2016) and meso-scale heat transport (Noppel and Fiedler 2002).
1609 Turbulent fluxes in valleys have been shown to scale better with the local slope-scale
1610 variables than with larger scale terrain features (Rotach et al. 2008), while the katabatic
1611 flows generated on sloping valley walls can extend a considerable distance into the
1612 horizontal terrain below, akin to an internal boundary layer (Mahrt et al. 2018). In
1613 extreme cases hydrodynamic and hydrostatic forcing can combine to produce dangerous
1614 phenomena like rotors and extreme downslope winds (Sheridan et al. 2004; Mobbs et al.

1615 2005; Grubisic et al. 2008) and we can also observe hydrodynamically driven flows over
1616 hill tops and ridges effectively decoupled from the valley flows on either side (Lewis et
1617 al. 2008b)

1618 At first site, understanding the interaction between the complex flows generated as
1619 synoptic winds encounter extensive steep topography and the multiplicity of valley-scale
1620 and slope-scale flows generated by solar heating and cooling seems an intractable
1621 problem. Almost all the theory and models we have surveyed so far address flows over
1622 simple idealised topographies but in nature hydrodynamic forcing and heterogeneous
1623 heating and cooling can generate multiple flows evolving at multiple scales and the
1624 number of possible combinations grows geometrically (Soler et al. 2002; Trachte et al.
1625 2010; Martínez et al. 2010; Serafin et al. 2016). However, surprisingly, recent syntheses
1626 of field experiments are beginning to reveal useful simplifications and paths forward; see
1627 for example Rotach and Zardi (2007). We shall return to consider this paradox in the
1628 Discussion Section 7.

1629

1630 **5. Field Experiments**

1631 **5.1 Boundary Layer Hill Flows**

1632 Now we want to discuss in some detail the largest hill flow field campaign of the 1980s,
1633 Askervein in Scotland, and contrast it with the most recent, Perdigao in Portugal to
1634 provide a reference frame for the many other experiments, which have underpinned the
1635 theoretical developments that we have concentrated on so far. Many of the early theories
1636 and models from the 1970s and 1980s, discussed in Sections 2.1 and 2.2, were linked to
1637 field experiments on conveniently located hills of relatively simple geometry, for
1638 example Brent Knoll (Mason and Sykes, 1979), Black Mountain (Bradley, 1980), Ailsa
1639 Craig (Jenkins et al., 1981), Kettles Hill (Taylor et al., 1983; Salmon et al. 1988),
1640 Bungendore ridge (Bradley, 1983), and Nyland Hill (Mason, 1986). Indeed, Bradley's
1641 laboratory in Canberra Australia was literally at the foot of Black Mountain. Although
1642 some of the earlier experiments only measured windspeeds on an upwind and a hilltop
1643 mast, others added more measurement stations to track the flow development over the
1644 hill. Nevertheless there was an emphasis on measuring the speed up, ΔU as the
1645 touchstone for modelling success (Section 2.1). As new theory developed, however, the

1646 search for field sites to test it unequivocally became more urgent as was the realisation
1647 that a field campaign able to deploy sufficient sensors to gather the information theory
1648 now demanded was probably beyond the capabilities of a single laboratory.

1649 In the early 1980s, therefore, the search for an ‘ideal’ 2D ridge with uniform
1650 surface cover, where the predictions of theory would not be confounded by other
1651 influences like roughness changes or complex approach flows, sent two research teams to
1652 the Outer Hebrides of Scotland. The 1982 and 1983 Askervein campaigns (Taylor and
1653 Teunissen, 1985, 1987) on South Uist were a combined effort between researchers from
1654 the Atmospheric Environment Service, Canada, the Risø National Laboratory, Denmark,
1655 the University of Hannover, Germany, the University of Canterbury, New Zealand and
1656 both the Building Research Establishment, and ERA Technology Ltd. in the United
1657 Kingdom. It had been initiated at an International Energy Agency meeting with a view to
1658 supporting research related to wind farm siting. At the same time, in 1982 the UKMO
1659 mounted a campaign on another isolated hill, Blashaval on North Uist (Mason and King,
1660 1985). In their review of field experiments up to that time Taylor et al. (1987) classified
1661 hills in terms of two parameters H/L and L/z_0 . Most experiments up to that time had been
1662 on isolated hills or ridges of moderate slope and smooth profile, ($H/L < 0.7$) and, except
1663 for Black Mountain, which was forested, on hills with grass or scrub surfaces so that L/z_0
1664 $\geq \sim 10^4$. The focus was on moderate to strong winds, near neutral stratification and overall
1665 terrain length scales, $4L \sim 1$ km. The prevailing westerly winds and uniform sheep-
1666 cropped turf on North and South Uist certainly conformed to those conditions.

1667 The Askervein experiments generated by far the largest data set of the 1980’s
1668 projects and the results were made widely available, being distributed as scanned copies
1669 of the Atmospheric Environment Service reports (available at
1670 <https://www.yorku.ca/pat/research/Askervein/index.html>). The data set has been used by
1671 many modellers as a test case-see for example Chow and Street (2009) who list seven
1672 other papers using Askervein-and it is one of the data sets still recommended for use in
1673 the wind energy community-<https://windbench.net/askervein>.

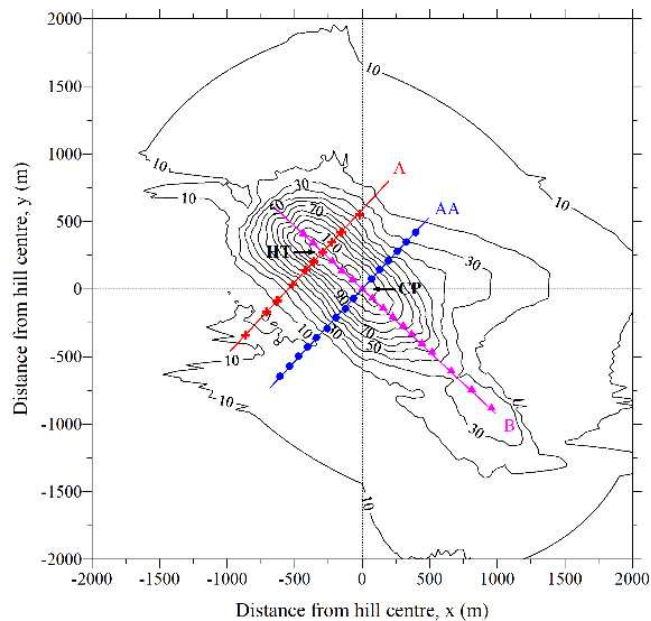
1674 Askervein Hill (Fig. 11a) can be considered as a segment of a 2D ridge of total
1675 length 2km, oriented NW-SE and with width ($4L$) about 1 km. There is a good, relatively
1676 uniform, flat fetch to the SW beyond which lies the Atlantic Ocean, about 3 km to the

1677 West. There are other hills to the East of Askervein but the predominant winds were from
1678 the West. As can be seen in the figure, there are no trees on the hill and the roughness
1679 length was relatively low, estimated as $z_0=0.03\text{m}$ from reference mast profiles. The
1680 surface on the hill itself appeared similar to the reference site apart from a few mobile
1681 roughness elements (sheep) and it came as a great surprise when near-surface wind
1682 profiles at the hilltop appeared to have $z_0 = 0.001\text{m}$ (Mickle et al, 1988). Walmsley and
1683 Taylor (1996) argued that this could be increased to 0.005m if cup anemometer over-
1684 speeding was accounted for while Niels-Otto Jensen of the Danish Riso laboratory, who
1685 spent a lot of time at the hilltop, estimated that the hilltop area surface would have $z_0 =$
1686 0.01m . Based on an analysis in streamline coordinates, Finnigan has calculated that the
1687 anomalously small roughness length could result from ignoring the stabilising effect of
1688 the convex streamline curvature at the hill crest, the so-called ‘curvature Richardson
1689 Number’ effect (Bradshaw, 1969; Finnigan, 1988, Finnigan et al. 1990). Nevertheless, it
1690 remains something of a puzzle.

1691 **a**



1692
1693 **b**



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Fig. 11 Askervein. (a) Photo, by Hans Teunissen, taken from the 50m reference mast. (b) topographic contours (m) and mast positions

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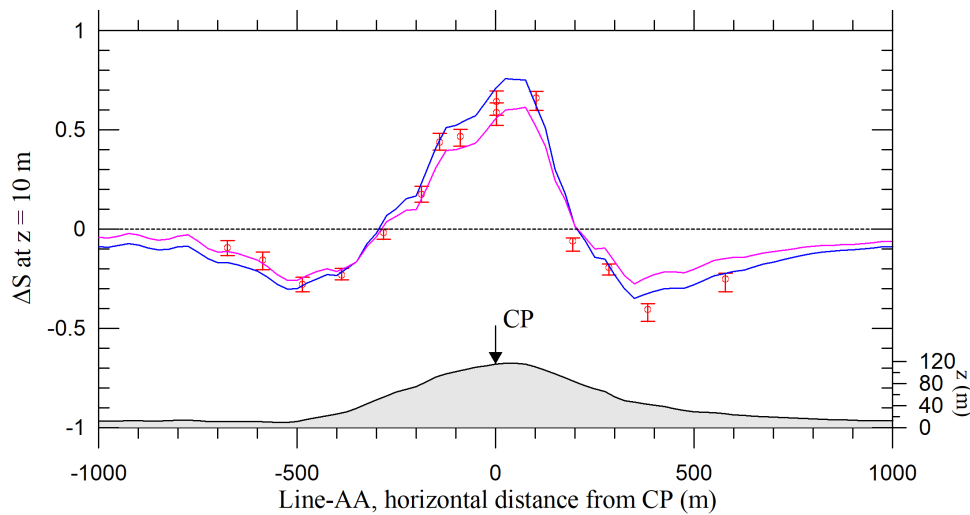
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The Askervein experiment deployed a far more extensive sampling network than anything attempted to that time. This comprised 10m wind speeds along two lines (A, AA) of 10m masts (50 in total) across and one line (B) along the hill ridge (Fig. 11b). There were also 50m masts on the hilltop (HT) near the NW end of the ridge and at an “upwind” reference location about 3 km SSW of the hilltop, a 30m mast W of the hilltop and a 16m mast at a second ridge top location (CP) at the “centre point” of the ridge. Some 10m masts and the 16m CP mast had Gill UVW anemometers while others were making measurements with cup anemometers. The 50 m masts had sonic anemometers, Gill UVW anemometers and cup anemometers and in addition TALA kites were flown and line tensions measured to determine wind speeds to greater heights, including profiles up to 500m with multiple kites near the reference tower. Some radiosondes were released and confirmed near-neutral stratification. Wind directions were measured continuously with several systems. Mean wind speed profiles at the hilltop and reference sites provide speed-up information for different wind directions while data from the 10-m tower lines provide information on relative wind speeds at different positions on the hill, again for different background wind directions.

1714 In the 1983 experiment, an extensive set of Mean Flow data was collected with
 1715 wind speeds in the 5-15 ms⁻¹ range. These were used to guide model development and
 1716 showed that the earlier linearized RANS modelling approach of Beljaars et al. (1987)
 1717 could be extended to a deeper and non-linear planetary boundary layer approach, which
 1718 could better deal with flow reductions in the lee of the hill and better match hilltop
 1719 profiles. (Weng and Taylor, 2011) as shown in Figure 12.

1720 Turbulence measurements at Askervein were less extensive but still important.
 1721 Two of the participating groups installed sonic anemometers, three groups used Gill
 1722 UVW anemometers, one had gust anemometers and the standard deviations from cup
 1723 anemometers were also used. The most reliable sonic anemometer data (Taylor and
 1724 Teunissen, 1985, Fig 4.5) showed reductions in hilltop profiles of $\overline{u'^2}$, $\overline{w'^2}$, and $-\overline{u'w'}$
 1725 relative to upwind values and an increase in $\overline{v'^2}$ in a middle layer (20m > z > 5m), in
 1726 accord with Rapid Distortion Theory (Hunt and Carruthers, 1990). The $\overline{u'^2}$ reductions
 1727 were also evident in the cup anemometer wind speed variance data. Mason and King
 1728 (1985) and Mason (1986) had also found evidence of Rapid Distortion impacts in hill top
 1729 turbulence profiles at Blashaval and Nyland Hill (Section 2).

1730 **a**



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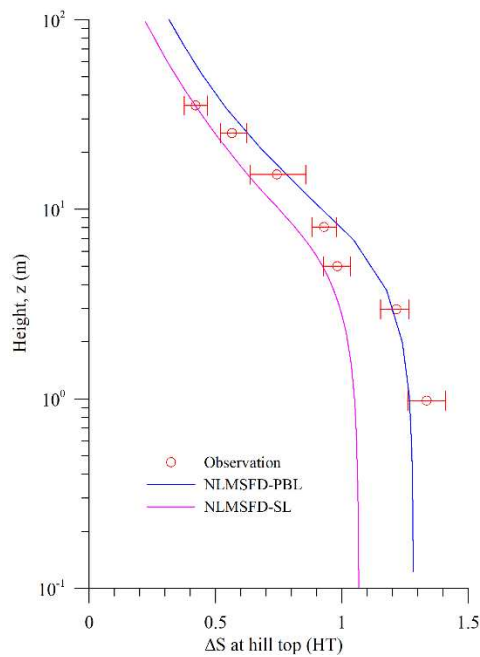
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1736 **b**



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1738 **Fig. 12** Comparisons between model calculations (Weng and Taylor, 2011) and Askervein 1983 field data
1739 (Taylor and Teunissen, 1985) (a) 10m mean windspeeds along transect AA. (b) hilltop speed up profiles.

1740

1741 Since the Taylor et al (1987) review there have been other reviews and several notable
1742 field program reports on flow over topographic features. Coppin et al (1994) analyzed
1743 1984/1985 measurements over Cooper's Ridge, a grass covered feature by the side of the
1744 Wollondilly river in New South Wales, Australia. The emphasis was on the effects of
1745 stratification. A previously unencountered source of anomalous data from Coopers Ridge
1746 proved to be the result of the mobile roughness elements on that hill (cows) finding that
1747 the sensor power cables provided a tasty alternative to chewing grass. Vosper and Mobbs
1748 (1997) directly measured the pressure drag on a large hill in NW England-as we noted in
1749 Section 2. topographic drag was an issue of major concern to the NWP community and
1750 there were few if any field data available up to that time with which to compare theory.
1751 The same scientific team based at Leeds University went on to perform a multi-season
1752 measurement campaign over Tighvein, a large ($H \approx 500m, L \approx 2000m$) hill in the SW
1753 corner of the Isle of Arran, which is 22 km off the SW coast of the Scottish mainland
1754 (Vosper et al. 2002) As well as surface pressure they made detailed mean windspeed and
1755 turbulence measurements at multiple locations over the hill . As we would expect from

1756 the scaling criteria discussed in Section 1., Tighvein was subject to buoyancy influences
1757 and they found that at times when $F_L \geq 0.25$ there was a pressure minimum over the hill
1758 top, which was also the position of maximum near-surface speed-up, while for $F_L \leq 0.25$
1759 the pressure field is more asymmetric and the lee-slope flow is generally stronger than on
1760 the windward slope.

1761 A more complex terrain configuration but at much smaller scale was described by
1762 Berg et al. (2011) in their Bolund hill study. This low ($H=12\text{m}$) coastal hill involves a
1763 steep cliff and a transition from water (Roskilde fjord) to grass covered land and so added
1764 a roughness change and an abrupt upwind step to complicate modelling or interpretation
1765 but the relatively small scale of Bolund hill had advantages in characterizing the flow in
1766 detail. Sonic anemometers, plus some cup anemometers and thermometers were used to
1767 measure winds and turbulence at levels up to 9m on eight masts on the hill, with two
1768 other masts providing reference profiles. A companion paper (Bechmann et al, 2011)
1769 reports on an extensive modelling inter-comparison exercise (Section 2.3).

1770 An equally complex site on a larger scale than Bolund was studied by Grant et al.
1771 (2015). They made measurements over a partially forested hill on the NE coast of the isle
1772 of Arran. The hill height varied from 160m to 260m asl. over its 1.5km length and both
1773 its NE slope, which falls to the sea, and its SW slope are steep enough to ensure
1774 downwind separation. Measurements were made using sonic and cup anemometers on
1775 three 23m masts, one upwind of the SW slope and two on the crest. In addition 12
1776 automatic weather stations were deployed on SW-NE transects and recorded data at 2m
1777 height. As might be expected at such a complex site with patchy forest clearings and an
1778 abrupt roughness change from sea to forest cover in North Easterly airflows, the wind
1779 and turbulence field is complex with strong directional shear between the masts. The data
1780 were modelled by Grant et al. (2016) using a 1.5 order closure RANS model and quite
1781 satisfactory agreement was obtained under neutral conditions as long as the horizontal
1782 variability in canopy structure was explicitly represented. Data to allow such structural
1783 detail to be incorporated in the model are now readily obtained from airborne lidar
1784 measurements.

1785 On a larger scale than Askervein, both physically and in terms of the number of
1786 scientists involved, the Perdigoão - 2017 field campaign in Portugal (Fig. 12) represents a

1787 major step forward in characterising a complex large scale flow field. Fernando et al
1788 (2019) provide the background and some initial results while a series of papers have and
1789 continue to appear in EGU journals. A preliminary study, Perdigoão – 2015 had used the
1790 site to develop scanning multiple lidar methodologies (Vasiljević et al, 2017). The UCAR
1791 and University of Porto web sites https://www.eol.ucar.edu/field_projects/perdigao,
1792 <https://perdigao.fe.up.pt/> have multiple links, including public access to the data and a
1793 promotional video. There is also access to slide presentations from Perdigoão workshops
1794 showing the latest progress with processing and interpreting the data.

1795 The Perdigoão project covers two parallel ridges, orientated approximately NW-SE
1796 with a broad valley between (Fig. 13). The scale is comparable to Askervein ($L \sim 250$ m,
1797 $H \sim 150$ m) but one big difference is in the land cover, which is mostly forest with trees of
1798 mean height 10 m. There is also a large wind turbine, the focus of some wake studies.
1799 The surrounding terrain is complex, and thermal and larger scale topographic effects are
1800 very present and have a major impact. Fernando et al (2019, Figure 4) shows that while
1801 the dominant winds 60m above the ridge top are from W to SW or NE directions, i.e
1802 approximately normal to the ridge, winds in the valley are generally from the NW or SE
1803 or calm ($< 1\text{ms}^{-1}$). There is a slight SE to NW gradient along the valley contributing to
1804 these low velocity valley winds, downslope at night and upslope during the afternoons.
1805 While the details may be site-specific the valley wind effect is a common feature in
1806 moderately complex terrain (see Section 4.) and Perdigoão will provide an important data
1807 set for testing models that can resolve and accurately predict this feature. Palma et al
1808 (2019) demonstrate this possibility with models that nest from global (GFS) through
1809 mesoscale (WRF) to the micro-scale (VENTOS[®]/M) and successfully reproduce local
1810 flow behaviour at Perdigoão over a 24-hour period.

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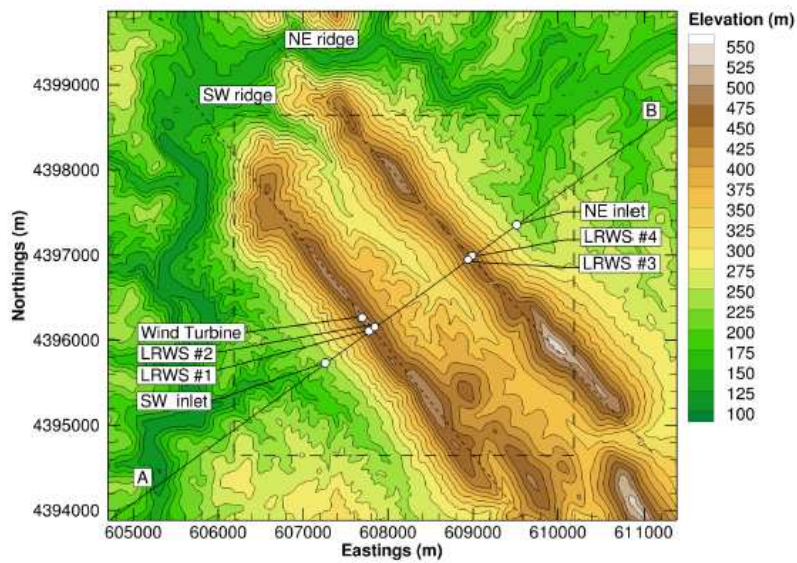
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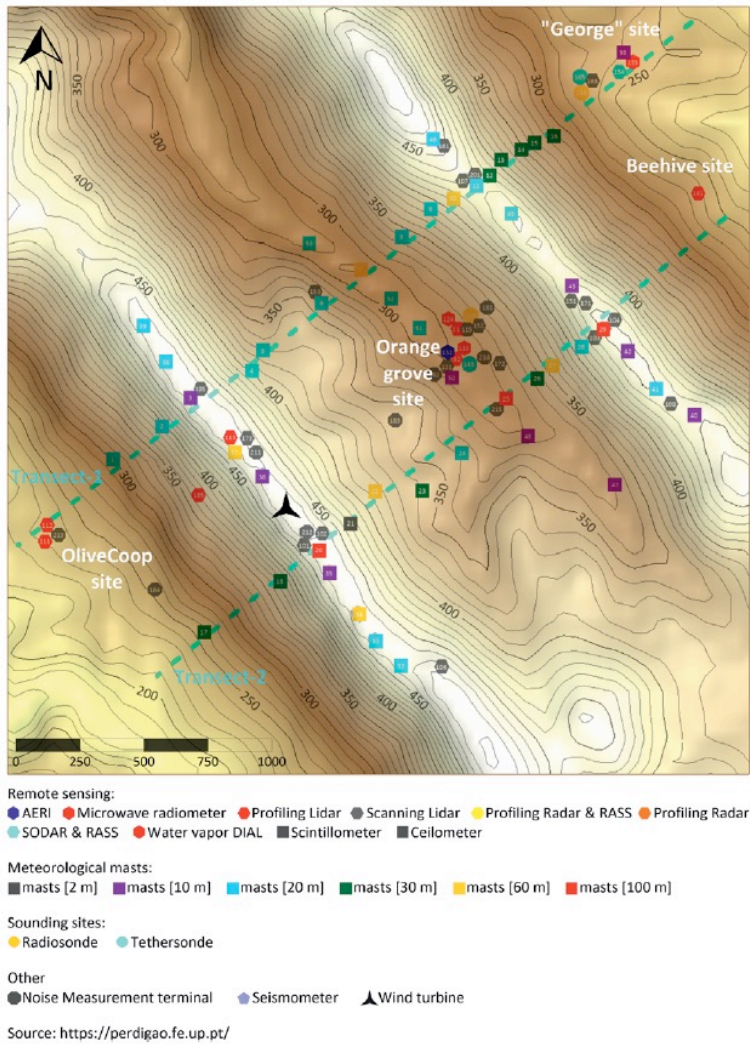
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(b) Photo by Mr. Nuno Soares, Smart Box, Comunicação Audiovisual, Portugal



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(c) Image reprinted from Palma et al. (2019)



1821

1822 **Fig. 13** Perdigão Field site, Portugal. (a) photo, (b) contour map of the region, (c) sensor locations

1823

1824 5.2 Gravity-Driven flows

1825 Most thermodynamically driven slope flows develop both in space and time so

1826 characterising them properly requires measurements at multiple heights at multiple

1827 positions along a slope over an extended period. Only a few field experiments have

1828 achieved this and this has contributed to the difficulty of deriving simple general rules for
 1829 the structure of gravity flows with the generality of the JH75-type linear hill flow models.

1830 As we shall see in the next Section 6, it is also the case that physical modelling has not

1831 played as large a role in gravity flow studies as it has in hill flows so what we do know

1832 comes primarily from the interplay of field observations and theory.

1833 It is possible to roughly divide the field studies into three groups. First, dedicated
1834 studies of katabatic flows driven by radiative cooling on simple slopes usually at shallow
1835 angles (<5 degrees). Second, studies of flow on valley sides driven by radiative
1836 heating/cooling as part of larger scale campaigns targetting mountain-valley systems.
1837 These are often on much steeper slopes (~30-40 degrees) and also deal with transitional
1838 flows and shadow effects. Third, flows over glaciers, often focussing on the interaction
1839 between larger scale and katabatic forcing. As we shall see, many of the simple slope
1840 studies were performed as elements of larger scale mountain meteorology campaigns
1841 such as ASCOT (Blumen et al., 1990), MAP-RIVIERA (Rotach et al., 2004), VTMS
1842 (Monti et al., 2002), CASES-99 (Mahrt et al., 2001) and MATERHORN (Fernando et al.
1843 2015) and so the local slope measurements were often supported by extensive
1844 climatological data.

1845

1846 **5.2.1 Flows on simple slopes.**

1847 Although drainage winds (particularly in valleys) had been studied in the past, few
1848 detailed measurements had been made of katabatic flows on simple slopes prior to the
1849 late 1970s. Mahrt (1982) for example was able to list only eight cases from six
1850 observational studies that measured profiles of both wind and temperature. The study of
1851 Manins and Sawford (1979a) on a long gentle slope using a combination of balloon
1852 profiling equipment and virtual potential temperature measurements was one of the first
1853 to produce detailed data. They concluded that the 1D models then current were
1854 inadequate to describe their observations. Their major findings were that even on a
1855 simple slope, the flow was significantly 3D and that the main cause of flow retardation
1856 was entrainment from the boundary layer above rather than surface friction. Their study
1857 prompted the development of their two-layer slab model (Manins and Sawford, 1979b).

1858 A series of fruitful experiments over simple slopes followed through the 1980's
1859 with, in several cases, multiple towers or observation points equipped with fast response
1860 sensors, which allowed the spatio-temporal development of the flow to be recorded as
1861 well as the turbulent fluxes that maintained it. An early example was the study by
1862 Clements, and Nappo (1983), employing two towers and 2 weather stations on a 1 km
1863 slope but more influential were the measurements of Horst and Doran (1988), who

1864 measured at two sites as part of the US DoE ASCOT field campaign. Their two sites
1865 differed markedly in surface roughness; the first slope was covered by 1-2m bushes and
1866 10-30m trees while the second consisted of 10-30cm grass and scattered bushes. At the
1867 second site, which more closely approximated a simple tilted plane, they were able to
1868 follow flow development using four 20m towers and fast response turbulence sensors and
1869 produced valuable data on turbulent fluxes above and below the observed low level jet
1870 peak.

1871 A gap of over a decade intervened before more simple slope studies appeared
1872 with the measurements of Monti et al. (2002) in the Salt Lake Basin of Utah as part of the
1873 Vertical Turbulence and Mixing (VTMX) campaign. Their measurements showed the
1874 influence of internal waves on entrainment and the data were also used by Princevac et al.
1875 (2005) in their parametrisations of entrainment. More recently, the study by Whiteman
1876 and Zhong (2008) at the same Salt Lake Valley site and also as part of VTMX used four
1877 tethered balloon profiles along a 1km downslope transect and reported a much thicker
1878 and stronger katabatic current than the earlier studies had found for similar slope and
1879 thermodynamic configurations. Most recently, the detailed sampling of small scale
1880 turbulence structure by Grachev et al. (2016), used four comprehensively instrumented
1881 towers on a slope on Granite Mountain during the 30-day long MATERHORN field
1882 campaign in 2012 to produce a data set with unprecedented temporal and spatial
1883 resolution of turbulence structure (see Section 4.1).

1884 Moving away from simple slopes, Mahrt and various collaborators (Mahrt et al.
1885 2001; Soler et al. 2002) had focussed on the interaction of shallow drainage flows with
1886 both the ambient wind and with larger scale drainage flows developing over the major
1887 landscape features. Using mainly data from the CASES99 experiment, he was able to
1888 develop a general understanding of this complex situation which is important in many
1889 biometeorological applications.

1890

1891 **5.2.2 Flow on valley sides**

1892 The development of both anabatic and katabatic flows on valley sides has been studied
1893 primarily in the context of campaigns to characterise mountain-valley wind systems more
1894 generally. As noted by Whiteman (1990) such flows can be expected to differ from those

1895 on simple slopes because of two factors. First, the formation and growth of a surface-
1896 based inversion over the valley floor will mean that the ambient stratification in which
1897 the valley side flow develops will change as the inversion deepens. Second along-valley
1898 circulations will affect the structure and evolution of the shallow slope flows so that these
1899 are fundamentally 3D and unsteady. Despite these complications we can contrast two
1900 experiments both of which have provided valuable data. First, as part of the large scale
1901 MAP-RIVIERA campaign (Rotach et al., 2004) measurements were made from a single
1902 tower equipped with six 3D sonic anemometers and other sensors in a 13m high mixed
1903 deciduous forest on a 35 degree slope on the side of the Riviera valley in the Italian Alps.
1904 We have already noted these results in the context of slope flows in canopies (van Gorsel
1905 et al. 2011; Section 4.2) for which they provide valuable data.

1906 A useful contrast is provided by the measurements of Oldroyd et al. (2014) who
1907 also made detailed wind and turbulence measurement from a single tower on a slope also
1908 of 35 degrees in Val Ferret, Switzerland. However, their measurements were made over a
1909 surface of short grass and a classic wall jet type wind profile was observed. Their data
1910 have provided a valuable input to our understanding of turbulent structure in this ‘classic’
1911 situation (Oldroyd et al. 2016) but also of more complex flow dynamics during
1912 transitional periods (Nadeau et al. 2018). The experiment of Oldroyd et al. (2014)
1913 followed the earlier study at the same ‘SLOPE’ field site by Nadeau et al. (2013) who
1914 used a comprehensive instrument array, comprising two turbulence towers, two weather
1915 stations, five surface temperature measurement stations and a tethered balloon system
1916 deployed down a 400m slope transect to study the transitional flow generated by
1917 advancing shadow fronts.

1918

1919 **5.2.3 Flow over ice and snow surfaces**

1920 The strong katabatic winds in Antarctica have long been a subject of study and analysis
1921 (e.g. King, 1989; Parrish and Cassano, 2003) but the difficulty of doing tower based
1922 measurement in that hostile environment has meant that studies have mainly been at the
1923 synoptic scale. At a smaller but still whole-of-mountain scale, the KABEG’97
1924 experiment (Heineman, 1999) combined aircraft and surface measurements to disentangle
1925 the effects of synoptic and katabatic forcing on a tundra and ice sheet in West Greenland.

1926 On the smaller slope-flow scale, there has been continuing interest in katabatic flows
1927 developing over European glaciers. To some degree this has been motivated by general
1928 research into the structure of stable equilibrium boundary layers (e.g. Smeets et al. 1998)
1929 but the turbulence structure of katabatic winds developing over an extensive Austrian
1930 glacier was recorded by Smeets et al., (2000), who showed the important role played by
1931 the turbulent energy fluxes in the mass balance of the glacier ablation zone.

1932 Reviewing this necessarily incomplete survey of slope flow field experiments, we
1933 are struck by several things. First, the very large range of different hill and valley
1934 combinations and the similarly large range of ‘typical’ flow responses, has meant that
1935 many of the field data sets cannot be directly compared. Even in the most obvious case of
1936 simple, thermodynamically driven slope flows, finding data free of secondary synoptic
1937 forcing so that they can directly inform theory is difficult. Second, the wide range of
1938 important configurations to be studied has meant that the effort has been spread thin with
1939 only a few useful data sets for each archetypal situation-slope flow, valley flow, glacier
1940 flow, etc. This is in quite stark contrast to boundary layer hill flows where, even if ideal
1941 hills are hard to find, most field experiments have contributed to a single theoretical
1942 framework. The wide spread of local slope flow data can be best understood, therefore, in
1943 the context of the larger scale dynamics of hill-valley flows as addressed for example in
1944 the 1990 meteorological monograph edited by Blumen (1990) and in the summaries of
1945 the more recent large scale field campaigns such as MAP-RIVIERA (Rotach and Zardi,
1946 2007). These problems are well recognized by workers in the field, see for example
1947 Stiperski and Rotach, (2016). We will return to consider this context in the Discussion,
1948 Section 7.

1949 **6. Physical modelling: Wind Tunnel and Flume Studies**

1950 Partly as a result of the elusive nature of ‘simple isolated’ hills in nature and partly
1951 because it is very difficult in field experiments to sample with sufficient resolution and
1952 range to fully characterize the flow, physical modelling has played an important part in
1953 developing theory and understanding. This has particularly been the case for separated
1954 flow, where field exploration of the separation bubble was almost always restricted to the
1955 layer that towers could reach, whereas the bubble depth is $O[H]$. The extensive

1956 deployment of lidars and boundary layer profilers at Perdigão is the first time that the
1957 complex dynamics of the separation bubble have been fully characterized in a large scale
1958 hill flow experiment (Palma et al. 2019). In this section we first discuss the benefits and
1959 limitations of physical modelling before looking at the more recent efforts in modelling
1960 canopy covered hills. For a comprehensive review of earlier simulations of flow over
1961 rough hills, the reader is referred to Finnigan (1988).

1962 The scaling laws that govern physical modelling set limits on what can be
1963 modelled and also determine how we must interpret results. For neutrally stratified flow,
1964 the key dimensionless group to match between real life and the model experiment is the
1965 Reynolds Number, $Re = U_0 L / \nu$. However, characteristic Re values for boundary layer
1966 hills in the atmosphere are $10^8 - 10^9$ whereas the largest Re values achievable in
1967 boundary layer wind tunnels or flumes, where topography is reduced in size by factors of
1968 $10^3 - 10^4$, are only about 10^5 . Operating windspeeds in boundary layer wind tunnels are
1969 typically 10-30 m/s while water flumes run at a tenth or less of that velocity but, since the
1970 kinematic viscosity of water is about ten times that of air, they achieve similar Re values
1971 to wind tunnels. This reduction in model Re implies significant differences in the balance
1972 of viscous and inertial forces between real and simulated flows but experience has shown
1973 that, if the modelled flow is ‘aerodynamically fully rough’, then acceptable turbulent
1974 boundary layer characteristics can be reproduced.

1975 Fully rough flows occur when the momentum absorption at the model surface is
1976 almost entirely through pressure drag on the surface roughness elements. This requires
1977 the roughness Reynolds Number, $Re^* = (u^* z_0 / \nu)$ to be 5 or greater (Raupach et al.,
1978 1991). Re^* almost always exceeds 5 in atmospheric flows but in a wind tunnel or flume,
1979 it requires the model surface to be far rougher than strict geometric scaling would imply
1980 unless the prototype hill is covered with a tall plant canopy or buildings. If the roughness
1981 elements of a real hill covered with turf or rocks were scaled down in the same ratio as
1982 the gross hill dimensions, H and L , then the model surface would be aerodynamically
1983 smooth or transitional ($Re^* \sim 1$) and momentum would be absorbed predominantly as
1984 viscous rather than pressure drag. As a result the turbulence dynamics of the near-surface
1985 model layer can be significantly different from that over the real hill. Conversely, if we

1986 make the model surface fully rough, the asymptotic scaling laws used to define the inner
1987 layer depth l often imply that the inner layer is entirely within the model roughness
1988 sublayer.

1989

1990 **6.1 Flow over hills**

1991 As well as changing the turbulent structure of the inner layer by exaggerating the surface
1992 roughness, boundary layer wind tunnels, especially smaller ones, typically thicken the
1993 boundary layer artificially before the working section and this usually means that the
1994 inertial or logarithmic layer in the approach flow occupies a much larger fraction of the
1995 boundary layer depth than in nature (Hunt and Fernholz, 1975; Gong and Ibbetson, 1989;
1996 Finnigan et al. 1990). The effect of this on the shear in the approach flow impacts the
1997 modelled speed up and drag because, as we discussed in Section 2.1 (Belcher et al. 1993),
1998 both these features are sensitive to the upwind shear. Conversely, if the approach flow
1999 boundary layer is allowed to grow naturally, then the hill may occupy a significant
2000 fraction of the total boundary layer depth. For example in the multiple ridge simulations
2001 of Athanassiadou and Castro (2001) the model hills were $0.2z_i$ and $0.33z_i$ respectively,
2002 implying that the prototype hills stretched the strict criterion for boundary layer hills set
2003 out by Belcher and Hunt (1988) (see Section 1, above). Another ploy used by some
2004 researchers has been to generate a fully rough approach flow but then relax to geometric
2005 similarity on the model hill itself. Unfortunately, this means that interpretation has to
2006 contend with the complex effects of both hill and roughness change (e.g. Pearse et al.,
2007 1981; Takahashi et al. 2005; and see the discussion in Section 2.2 above). Indeed, abrupt
2008 combinations of roughness changes and hill effects are only just now being modelled
2009 successfully (e.g. Grant et al., 2016).

2010 Despite all these caveats, wind tunnel studies provided a great deal of guidance as
2011 theory and understanding was being developed in the 1970s and 80s. The experiment of
2012 Britter et al. (1981) clarified the nature of turbulence changes in the rapid distortion
2013 region above the inner layer while Finnigan et al (1990) analysed turbulence dynamics
2014 over a fully rough model in streamline coordinates, which illuminated the competing
2015 effects of shear, plane strain and curvature on turbulence development over the hill. The
2016 comprehensive measurements of Gong and Ibbetson (1989) tested the predictions of

2017 competing theories for the HLR88 outer layers but not those features that depended in an
2018 essential way on the inner layer dynamics, while the wind tunnel simulations of
2019 Askervein by Teunissen and Shokr (1985a,b) and Teunissen et al. (1987), where the
2020 model surfaces were smooth or transitional, were invaluable in interpreting the field data
2021 from that campaign.

2022 The second area where the nature of the surface drag mechanism can be critical is
2023 modelling flow separation. Wind tunnel studies have been very valuable in defining the
2024 gross geometric determinants of separation on 2D and 3D hills (see Finnigan, 1988) but,
2025 when the hill steepness is close to the critical angle that promotes separation, the surface
2026 roughness has a large effect. A hill that does not separate when the surface is smooth can
2027 fully separate when it is rough. This feature is particularly striking when the hill is
2028 covered by a tall canopy as we have already noted. Finnigan (1988) discusses the
2029 mechanism of rough wall turbulent separation at some length and also compares the large
2030 number of wind tunnel studies of flow separation performed through the 1970s and 1980s
2031 to derive general guidelines for the steepness required for separation when hills are 2D
2032 versus 3D and rough versus smooth.

2033 One area that has received relatively less attention is the effect of repeated hills.
2034 There have only been a few attempts to model this situation in boundary layer wind
2035 tunnels (Gong et al., 1996; Athanassiadou and Castro, 2001) and water flumes (Poggi et
2036 al. 2007a,b). The configuration in each case has been one of repeated 2D ridges and a
2037 particular focus has been the development of the flow as it comes into equilibrium with
2038 the periodic distortion and whether linear theory still applies. Repeated ridges come
2039 closer to real complex topography than most isolated hills but even so, multiply repeated
2040 2D ridges are scarce in nature, the field study of Mason and King (1984) in South Wales
2041 or even Perdigao, where only two ridges are present, being rare examples. Intriguingly,
2042 both wind tunnel simulations show the development of secondary flows, which can be
2043 attributed to large scale streamwise vortices and which produce significant spanwise
2044 variation in the mean flow and statistics. A similar phenomenon was observed in an
2045 earlier unpublished study of flow over repeated ridges in the CSIRO Australia boundary
2046 layer wind tunnel by Dr W. Gong, Prof. P. Taylor and Dr K. Ayotte (K Ayotte. *pers.*
2047 *comm.*), suggesting that such secondary flows, possibly generated by a Craik-Leibovich

2048 Type-2 instability, may be an important feature of atmospheric flow over real complex
2049 terrain.

2050 The flume experiments of Poggi et al. (2007) over a train of gentle smooth-
2051 surfaced cosine hills explored the interplay between the viscous sublayer and the inner
2052 layer. Recognising our earlier caveats about fully rough simulations, that experiment
2053 revealed that many aspects of the inner layer, including the 2D shape of the mean
2054 velocity and Reynolds stress profiles, follow the JH75, theory provided the approaching
2055 mean velocity profile is appropriately specified. The experiment also showed that the
2056 hydrostatic pressure approximation, the linearization of the longitudinal mean advection
2057 term and the minor contribution from the perturbed vertical velocity to the overall
2058 advection in the mean momentum balance-simplifications all adopted in JH75-are
2059 acceptable.

2060

2061 **6.2 Stability effects**

2062 So far we have discussed neutrally stratified flow. A different set of modelling
2063 considerations apply when the flow is diabatically stable or unstable. In such cases, as
2064 well as trying to match the Reynolds Number between real life and model, we need to

2065 match the Froude Number, $F_L = \frac{U_0}{NL} = \frac{U_0}{L} \bigg/ \left(\frac{g}{\Theta_0} \frac{\Delta\bar{\theta}}{H} \right)$ For significant buoyancy effects in

2066 the atmosphere, we require $|F_L| < 1$ but to achieve this in the wind tunnel where both L
2067 and H are scaled down by 10^3-10^4 , U_0 must be as small and $\Delta\bar{\theta}$ as large as possible.

2068 There are practical limitations on how large $\Delta\bar{\theta}$ can be made while the need for small
2069 U_0 conflicts with the need to have Re as large as possible and realistic simulations of
2070 stable and unstable flows over hills have only been possible in a few specialist wind
2071 tunnels around the world, e.g. Ross et al. (2004), Takahashi et al. (2005), Loureiro and
2072 Silva Freire (2005), and Pospisil et al. (2017). Of these, only the results of Ross et al.,
2073 who compared neutral and stable flow over steep hills with an emphasis on how well
2074 flow separation can be modelled, are sufficiently well characterized to easily contribute
2075 to a general understanding of hill flow dynamics. The other studies mentioned, while of

2076 relevance to their specific wind engineering applications are difficult to generalize or in
2077 some cases to interpret in terms of atmospheric stability measures..

2078 Strong stable stratification in the atmosphere leads to phenomena like flow
2079 blocking in front of 2D ridges or, on 3D hills, the dividing streamline effect, where above
2080 some level the air flows over the hill while below it, the air flows around the hill. This
2081 phenomenon was studied in a towing tank where the water was stratified with salt
2082 solution (Hunt and Snyder, 1980). This allowed values of F_L as low as 0.1 to be achieved
2083 but at the expense of abandoning Re similarity. This is another example of flows that do
2084 not satisfy the criteria for boundary layer hills set out in Section 1.

2085

2086 **6.3 Physical modelling of canopy covered hills**

2087 The wind tunnel study of Finnigan and Brunet (1995) has already been referenced. They
2088 placed an aeroelastic canopy model over a 2D cosine shaped ridge. The model
2089 configuration had dimensions, $H=150\text{mm}$, $L=500$, $h_c=50\text{mm}$, $z_i=700\text{mm}$. Since the hill
2090 and canopy occupied such a large fraction of the model boundary layer depth, most of the
2091 conditions of the analytic theories like JH and HLR were voided but the striking changes
2092 to the canopy and shear stress layer flow fields that were observed (see Section 3.) led to
2093 a wave of interest including *inter alia* the theoretical study of FB04. The predictions of
2094 FB04 in turn prompted a series of experimental investigations.

2095 In the mid to late 2000s, detailed flume experiments were undertaken to measure
2096 the mean flow and turbulence structure inside canopies covering the train of gentle cosine
2097 hills already studied with smooth surfaces by Poggi et al. (2007). Since the initial focus
2098 was on the assumptions and results of FB04, the components of the mean longitudinal
2099 momentum balance were investigated first (Poggi and Katul, 2007b). That revealed that
2100 the presence of a recirculation region within the canopy is sufficiently large to distort the
2101 hydrostatic pressure perturbation assumption and to negate the mixing length hypothesis
2102 of FB04 entirely in the recirculation zone. Moreover, an alternative to the FB04 model
2103 for the deeper layers of the canopy was proposed that maintained the same interplay
2104 between the drag force and the mean pressure gradient but included a linearized mean
2105 advection term. The revised model was shown to delineate the onset of the recirculation
2106 zone better (Poggi et al., 2008) and out-performed FB04 in the deeper layers of the

2107 canopy, suggesting that mean advection remains a leading order term there. Advection in
2108 the upper canopy was subsequently included in the extensions to FB04 by Harman and
2109 Finnigan (2010, 2013) and increased the range of applicability of the model.

2110 The flume experiments also explored the shape of the higher-order statistics up to
2111 triple moments and the properties of the ejection-sweep cycle contributing to momentum
2112 transport (Poggi and Katul, 2007c). These experiments showed that sweeps dominate
2113 momentum transport within the roughness sublayer whereas ejections dominate
2114 momentum transport above the roughness sublayer and showed that the interface where
2115 sweeps and ejections are in approximate balance proved to be a pragmatic definition of
2116 the roughness sublayer thickness on hilly terrain. The study also showed that third-order
2117 cumulant expansion methods (i.e. correcting for asymmetry only in the joint probability
2118 density function of w' and u') as used by Nakagawa and Nezu (1977) and Raupach
2119 (1981) in smooth and rough-wall boundary layers, can also be used to link the ejection-
2120 sweep ratio to the flux transport term $\overline{(w'w'u')}$ in the more complex hill-canopy flow,
2121 thereby establishing a bridge between conditional sampling methods, quadrant analysis,
2122 and RANS closure schemes. Finally, the flume experiments also contrasted the shapes of
2123 the turbulent intensities and turbulent spectra over bare hills and hills covered by a
2124 canopy (Poggi and Katul, 2008). The key finding from those studies is that rapid
2125 distortion theories proved successful at predicting the turbulent intensities in the outer
2126 layer but not the inner or roughness sublayers, confirming the fundamental assumption in
2127 the JH75, HLR88 and FB04 analyses.

2128 Returning to the wind tunnel, Harman and Finnigan (2013) made measurements
2129 over a model in which the well-studied 'tombstone' model canopy (Raupach et al, 1986),
2130 whose equilibrium characteristics were well characterised, was placed on a 2D ridge in
2131 the CSIRO Australia boundary layer wind tunnel. The objective of the experiment was to
2132 test extensions made to the FB04 model by Harman and Finnigan (2010), which
2133 incorporated advection in the upper canopy and a more sophisticated coupling of the
2134 lower and upper canopy solutions as they affected the overall pressure perturbation on the
2135 hill. These extension were motivated by the RANS modelling of Ross and Vosper (2005),
2136 the flume experiments of Poggi and Katul (2007b,c) and the LES modelling of Patton and
2137 Katul (2009). The experiment focussed on the mean flow and pressure fields but was also

2138 able to show that some of the impacts on scalar transport, particularly those relevant to
2139 measuring carbon exchange from eddy flux towers, as discussed in Section 3, might be
2140 smaller than predicted by FB04.

2141 As discussed at length in Section 3.2, the fundamental mechanics of momentum
2142 and scalar transport at leaf level ensure that a very stable layer develops in forest or crop
2143 canopies at night, when radiative cooling is active. Finnigan and Hughes (2008) modelled
2144 this situation in a novel way by placing the hill-canopy model of Harman and Finnigan
2145 (2013), whose surface and canopy elements could be electrically heated, on the roof of
2146 the tunnel. This reversed gravity so the heated surface and canopy elements generated a
2147 strongly stable layer within the canopy. A model stably stratified flow produced in this
2148 way is self-limiting in its depth and has the advantage that the stable layer is generated
2149 naturally by surface heating. This is in contrast to stable flows that are ‘the right way up’,
2150 where a stable temperature gradient has to be generated in the approach flow by a grid of
2151 heating and cooling elements (e.g. Ross et al. 2004). The Harman and Finnigan
2152 experiment covered a range of Froude numbers and for $F_L \leq 0.3$, it was observed that,
2153 while flow over the hill above the canopy was fully turbulent with shear stress profiles
2154 identical to the neutral case ($F_L = \infty$), a downslope gravity current filled the canopy layer
2155 on both upwind and downwind slopes (see Section 3.2). Surprisingly, the downslope
2156 current on the upwind face of the hill, which was in the opposite direction to the flow
2157 above the canopy, penetrated upwind a distance of $11L$ at which point the surface and
2158 canopy were no longer heated. This confirmed that as long as the canopy remains cool
2159 enough, the flows above and within canopy can be completely decoupled.

2160 Physical modelling is one of the three pillars on which our understanding of
2161 atmospheric flow has traditionally rested, the others being theory and field experiments.
2162 Today, increases in computing power have advanced large eddy simulations (Section 2.4)
2163 to the point that they can be considered a fourth pillar. Each approach has its strengths
2164 and weaknesses and they are deployed most powerfully when they are used together,
2165 each filling in gaps in the other techniques and suggesting fruitful new lines of attack.
2166 Physical modelling in particular sometimes reveals important new physics. The collapse
2167 of turbulence and establishment of a resilient stable layer in a radiatively cooling night-

2168 time canopy was first seen in a wind tunnel experiment. It was later shown to be a
2169 consequence of basic transfer physics at leaf level and so a ubiquitous feature of canopy
2170 flow. Furthermore the theoretical result that followed showing that the strength of a
2171 resulting gravity current depended on the temperature deficit and slope length rather than
2172 slope angle (Section 5.4) explained many perplexing observations from flux towers on
2173 very gentle terrain.

2174 As we have seen in this section, wind tunnel and flume simulations have played a
2175 major role in shaping our understanding of hill flow dynamics and should continue to do
2176 so in the future. Modern instrumentation, particularly the ready availability of laser
2177 doppler and particle image velocimetry has now removed many of the difficulties of
2178 measuring flows of very high turbulent intensity or which regularly reverse direction such
2179 as inevitably occurs in canopies and in hill separation regions. The ability to stratify the
2180 flow in a limited number of specialised tunnels or by resorting to the subterfuge of
2181 Finnigan and Hughes for the stable case suggests that the very large investment in field
2182 campaigns like Perdigao could benefit enormously from concomitant physical modelling
2183 as was the case in the 1970s and 1980s.

2184

2185 **6.4 Gravity Driven Flows**

2186 Gravity driven slope flows do not seem to have been modelled in wind tunnels (except in
2187 the example just discussed above). Scaling considerations indicate that gravity currents
2188 generated on a rough or smooth plate of wind tunnel dimensions would be extremely
2189 thin. Most physical modelling of gravity currents has therefore been of density currents in
2190 water in an oceanographic context (see Baines, 2001 and references therein). However,
2191 the study of a katabatic water current flow through a canopy of bluff obstacles by Hatcher
2192 et al. (2000) has already been noted (see Section 4.3) as it is of direct relevance to gravity
2193 driven flows through plant canopies in the atmosphere. Similarly, the study of turbulent
2194 entrainment into atmospheric slope flows by Princevac et al. (2005) used a set of
2195 laboratory experiments on density current flows in water tanks as benchmarks against
2196 which their atmospheric measurements could be assessed. The references in Princevac et
2197 al. (2005) and in Hatcher et al. (2000) are also a good guide to laboratory work that is of
2198 direct relevance to atmospheric gravity currents.

2199

2200 **7 Bringing it all back Home: New Challenges and new scientific** 2201 **directions**

2202

2203 As we foreshadowed in the introduction, we have seen how the study of boundary layer
2204 flow over complex topography has gone through several phases. The 1970s and 1980s
2205 saw the development of analytic theory that clarified the fundamental physics. Basic
2206 questions such as, why the speed-up in the mean wind over hill crests was so much larger
2207 than the hill slope would suggest, were answered qualitatively and, to a large degree,
2208 quantitatively (Section 2.1). Numerical RANS models developed in parallel with analysis
2209 and, for a while, used the simplifications suggested by theory to produce fast
2210 computations for gentle terrain but then turned to more complex formulations to deal
2211 with steeper hills and changing surface cover (Section 2.2). These refinements and
2212 developments of RANS models continue apace driven by a host of practical applications,
2213 especially wind energy and NWP. However, for the first application especially, the
2214 weaknesses of RANS approaches in dealing with separation and the turbulent structure of
2215 the separated and near-wake flow are clear. RANS approaches work best when the model
2216 equations are effectively parabolic but the feedback on the driving pressure field that
2217 occurs with separation makes the equations unavoidably elliptic. In addition, the
2218 challenge of simultaneously parameterising the fine-scale surface processes that
2219 determine the separation point and the $O[H]$ size eddies in the separation bubble and
2220 wake, stretch the abilities of most RANS closure schemes past breaking point.

2221 As predicted by Wood (2000) in his prescient review, the last two decades has
2222 seen the increased use of LES approaches to capture this wide dynamic range and,
2223 perhaps as importantly, to give insights into the turbulence structure that were difficult to
2224 deduce from the sparse sampling of field experiments (Section 2.3; Section 5.). Wind
2225 tunnel simulations gave more detailed information about turbulence but over rough hills,
2226 the need to have aerodynamically fully-rough models precluded proper simulation of the
2227 shear stress dominated inner layer. Moreover, until the ready availability of Laser-
2228 Doppler- and Particle-Image-Velocimetry systems over the last decade, use of the
2229 standard wind tunnel turbulence sensor, the hot wire anemometer or even the pulsed wire

2230 sensor, precluded the proper characterisation of the 3D reversing flow in the separation
2231 region (Section 6.1).

2232 While this development of hill flow dynamics relied on a constant interplay
2233 between theory and dedicated field experiments on ‘ideal’ topography, supported in
2234 critical areas by wind tunnel simulations, the analogous study of gravity-driven flows, on
2235 hill and valley slopes, which began in earnest at the end of the 1970s, faced different
2236 challenges (Section 4). Even the simplest examples of these flows, such as katabatic
2237 winds down extensive uniform slopes under weak synoptic forcing, develop in both space
2238 and time and require a correspondingly extensive instrument deployment to characterise
2239 them properly in the field. At the same time, insights from physical modelling have not
2240 been as directly applicable to atmospheric gravity currents as they were to hill flows.
2241 Laboratory simulations generally involved dense water currents in less dense ambient
2242 fluid and were biased towards simulating large scale oceanographic or geophysical
2243 phenomena (Section 6.4). Finally, the many possible topographical configurations that
2244 generate gravity flows are difficult to represent by a single archetype equivalent to an
2245 isolated hill. Instead, field studies have ranged from ‘ideal’ simple slopes to the sides of
2246 steep valleys, where the gravity flows themselves coalesce to form hill-valley wind
2247 systems. These in turn add complexity to the downslope currents (Section 4.5). Despite
2248 these difficulties, considerable progress has been made in characterising the structure and
2249 dynamics of gravity flows and the main obstacles to theoretical progress have been
2250 identified as achieving better scaling laws and parameterisations of the fundamentally
2251 inhomogeneous turbulence of these wall-jet type flows.

2252 In the last two decades also, a good deal of effort has been devoted to
2253 understanding flow over hills covered by tall plant canopies (Section 3). As well as their
2254 important application to quantifying the terrestrial carbon cycle, analytic models that
2255 resolve the flow in the canopy and dedicated simulations in wind tunnels and flumes have
2256 resulted in a deeper understanding of hill-flow dynamics more generally such as the
2257 mechanics of flow separation on rough hills, surface roughness being in effect, simply a
2258 shallow canopy. The coupling of hill-flow and canopy dynamics has also revealed
2259 important new physical processes that were previously unsuspected such as the inevitable
2260 collapse of turbulence in a radiatively cooling canopy at night and the generation of

2261 robust and persistent downslope flows (Section 4.4). Moreover, research teams are
2262 finding still more surprising consequences of combining canopies and hills; currently
2263 unpublished data from wind tunnel experiments and LES indicate that the flow patterns
2264 that appear on 3D hills covered with tall canopies are significantly more complex than on
2265 2D ridges (Dr I. G. Harman, Dr E. G. Patton *pers. comm.*). There is clearly still much
2266 more to be discovered about the elementary building blocks of complex terrain flows.
2267 One point that has not been emphasised enough in the literature is the feedback between
2268 canopy processes at the individual hill scale and topographic drag at the landscape scale.
2269 The mechanisms of ‘separated sheltering’ in flow over low hills, which is exaggerated
2270 when a canopy is present (Belcher et al. 1993; Finnigan and Belcher, 2004) and of earlier
2271 separation caused by the canopy can easily double the topographic drag of relatively
2272 gentle topography if a rough surface is replaced by a canopy.

2273 Despite these remaining knowledge gaps, it seems clear that we have reached a
2274 point where hill flow and gravity flow studies are poised to coalesce with valuable
2275 information to be exchanged between the two fields. The urgent driver for this is the need
2276 for information on the impacts of global heating on local climate at scales where
2277 operational decisions must be made. This means that climate and weather models must be
2278 resolved at the 1-10km scale and implies that the sub boundary-layer scale dynamics we
2279 have been discussing in this review are critical. These imperatives are driving two
2280 responses. First we increasingly see resources and collaboration being donated to
2281 multinational measurement campaigns in regions of complex topography such as MAP-
2282 Riviera (Rotach et al. 2004), MATERHORN-X (Fernando and Pardyjak 2013; Fernando
2283 et al. 2015; Di Sabatino 2016), Perdigão (Fernando et al. 2019 and see Section 5) and
2284 others in the planning stage today. Second, we are seeing serious attempts to bridge the
2285 gap between eddy-resolving models on coarse meso-scale grids that can reproduce
2286 terrain-forced wind and temperature patterns around larger topographic features and the
2287 ‘classic’ LES models discussed in Section 2.3, which resolve the dynamics of boundary
2288 layer turbulence as it responds to topographic forcing. As this coalescence of modelling
2289 scales and the results from new field experiments that will test its success are likely to
2290 shape the research efforts of the coming decade, it is appropriate to end this review with
2291 some detailed remarks on where we now stand.

2292 For some years, one theme of hill-flow research has been the use of LES to
2293 represent the coupling between terrain and turbulence in meso-scale regional and climate
2294 modelling systems, for example the widely used weather research and forecasting system
2295 (WRF) <https://www.mmm.ucar.edu/weather-research-and-forecasting-model>. At the
2296 same time, researchers using those larger-scale models have been pushing their
2297 simulations to finer and finer scales as they realized that unresolved processes associated
2298 with surface interactions directly impact their predictive skill, especially over longer time
2299 scales. The advent of grid nesting capabilities has enabled many larger-scale numerical
2300 modelling codes to now refine their meshes sufficiently to include an LES-mode in the
2301 innermost domain (e.g., Chow et al. 2006; Golaz et al., 2009). While the idea of nesting
2302 an LES within a larger-scale model might seem straightforward, doing so accurately
2303 remains an ongoing challenge (e.g., Wyngaard, 2004; Talbot et al., 2012; Muñoz-Esparza
2304 et al., 2014; Shin and Hong, 2015; Honnert, 2016; Rai et al., 2016; Muñoz-Esparza et al.,
2305 2017; Bao et al., 2018; Muñoz-Esparza and Kosovic, 2018; Hald et al., 2019; Arthur et
2306 al., 2020). Indeed, Cuxart (2015) argues for a more stringent definition rather than simply
2307 describing this style of calculation as LES. The fundamental interplay between larger-
2308 scale dynamics and mountain boundary-layer turbulence drives these efforts. To take just
2309 two recent examples: Kirshbaum (2017) used a nested-LES framework to identify
2310 regimes and scaling associated with the influence of hill-induced thermal forcing on
2311 boundary-layer turbulence and how the hill-induced pressure distribution results in
2312 upstream blocking of larger scale dynamical processes, while Babić and de Wekker
2313 (2019) found that complex terrain reduces the aspect ratios of boundary-layer rolls and
2314 cells compared to what is expected over horizontally-homogeneous terrain.

2315 Wyngaard (2004) pointed out that one of the key differences between micro-scale
2316 models used to study turbulence and meso- to larger-scale models used to study weather
2317 and climate is the ratio of the energy-containing scales of turbulence to the filter-scale of
2318 the model. In ‘classic’ LES the filter-scale falls deep in the inertial subrange, whose
2319 statistical properties are simple and well understood and can be parametrised with
2320 confidence. Conversely, to simulate large domains, larger-scale models filter the flow at
2321 scales notably larger than the largest scales of boundary-layer turbulence, which range
2322 from tens of meters to a few kilometres, so the physics that a model of the unresolved

2323 processes needs to represent must change with the scale at which the equations are
2324 filtered, a concept now called ‘scale-aware parameterization’.

2325 These challenges were clearly expressed in the penetrating analyses of boundary
2326 layer structure in mountainous terrain by Rotach and Zardi (2007) and Weigel et al.
2327 (2007). Based on experience from many large scale measurement campaigns they were
2328 able to make two crucial observations. First, the exchange of scalars between these
2329 deeply corrugated regions and the free troposphere was predominantly by the ventilation
2330 of hill-valley systems, whose flow patterns were dominated by local thermally-driven
2331 ridge-valley flows. These flows themselves were modulated by the interaction of synoptic
2332 winds and the topography. Under certain conditions this ‘topographic venting’ could be
2333 several times larger than the turbulent exchange (Henne et al. 2004; Weigel et al. 2007).
2334 Second, was the observation that despite the highly complex and heterogeneous structure
2335 of the hill-valley flows, characteristic and transferable patterns in the turbulence
2336 structures and larger scale flow patterns could be found.

2337 It might be instructive to compare this state of affairs with that which faced
2338 researchers into canopy turbulence thirty years ago. There too it was necessary to
2339 integrate the complex interacting patterns of mean flow and turbulence around canopy
2340 elements, which, if we include urban canopies, varied in size from leaves to buildings.
2341 There were even many studies of flow around individual leaves or twigs or branches in
2342 the hope that these could be combined into an average ‘canopy flow’. Although flow
2343 patterns around individual canopy elements were heterogeneous and distinct, three
2344 conceptual advances eventually allowed real progress to be made. The first was the use of
2345 formal spatial averaging to deduce conservation equations for the whole canopy
2346 (Raupach and Shaw, 1982; Finnigan and Shaw, 2008). The second was arriving at a
2347 detailed understanding of the differences between the spatially-averaged properties of
2348 turbulence in the canopy airspace and that in the free air above (e.g. Finnigan, 2000). The
2349 third was the realisation that at the whole-canopy scale, ‘emergent properties’, which
2350 could not be deduced by analysis of the element flows alone, determined the interaction
2351 of the canopy with the overlying boundary layer (Raupach et al. 1996).

2352 In mountainous regions individual hill and hill-valley flow systems play the part
2353 of canopy elements. The definition of form drag around hills is exactly analogous to the

2354 form drag around canopy elements that is produced in canopy-flow momentum equations
2355 by spatial averaging. The distinctly different natures of within-canopy and boundary-
2356 layer turbulence manifests itself in different scaling laws, different ratios of turbulence
2357 moments and different spectral dynamics. The analogy to this in flow in and around
2358 topography may be a new approach to flow scaling that takes the irreducible anisotropy
2359 of boundary layer turbulence in complex topography at face value (e.g. Stiperski, 2017;
2360 Stiperski et al. 2019). This approach may be what is needed to write sub-filter scale
2361 parameterisations for eddy-resolving models that resolve Wyngaard’s dilemma by a
2362 physically-based statistical treatment of the unresolved motions in complex topography.
2363 Presumably, this would use the statistics of the topography to inform the statistics of the
2364 space-time structure of the airflow. An analogy to the third element in the modern
2365 description of canopy turbulence-the emergent nature of large scale energy-containing
2366 eddies that result from the inflexion point in the mean velocity profile-is harder to
2367 predict. It may be resolved by the application of the modern understanding of the
2368 thermodynamic constraints on the energy and entropy balances of topographical flow
2369 (e.g. Kleidon, 2016) or by approaches currently hidden somewhere in the large span of
2370 work we have surveyed here.

2371

2372 **References**

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