LESSON BOOK



Near-Surface Wind Meteorology

Concepts, terminology and models for the wind between 10 and 150 metres

Thirty-one lessons · 701 pages

Ricardo Marquez, Ph.D.

ricardo-marquez.com

Near-Surface Wind Meteorology

Concepts, terminology and models for wind in the lowest part of the atmosphere

A lesson sequence, written up as a book

Ricardo Marquez

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How to read this book

This book is about one question: why the wind at a given location, height and time is what it is. Each chapter expands one lesson of a sequence that answers it by following the wind downward,

large-scale forcing → boundary layer → surface layer → terrain and roughness → local flow → measured wind.

The heights of interest are those of engineering applications, from the 10 m of a weather station to the 150 m of a wind turbine: the heights at which a transmission line is cooled by the wind and a wind farm is driven by it.

The book is written for a reader who builds or uses wind models and wants the concepts and the vocabulary of meteorology stated precisely. Every new quantity is therefore introduced in the same order: its name, its physical meaning, its units, its equation, and how it is measured or modeled. Where the field uses a shorthand term, a *Jargon* box expands it once and the text then uses the expanded meaning. Every symbol is defined in the notation chapter that follows.

One distinction recurs in every chapter: what the atmosphere does, and what a model or an instrument represents of it. A weather model reports a wind for a grid cell several kilometers wide; an anemometer reports the wind at one point; neither is "the wind" without a statement of place, height, averaging time and scale. Statements that depend on the hemisphere are given for both. The numbers in the examples are illustrative, chosen to make each relation concrete.

Each chapter closes with a statement of what to retain, a table of its terms, and exercises.

Contents

How to read this book **ii**

Notation and vocabulary **x**

1 The atmosphere as a wind system **1**
1.1 The question and the framework . . . 1
1.2 Wind and its origin . . . 3
1.3 Pressure and the forces on air . . . 7
1.4 The layer near the surface . . . 11
1.5 Scales of atmospheric motion . . . 12
1.6 Model and measurement, forcing and response . . . 14
1.7 What to retain . . . 18
1.8 Exercises . . . 19

2 Pressure, geopotential and the pressure-gradient force **20**
2.1 Pressure and hydrostatic balance . . . 20
2.2 Pressure surfaces and geopotential . . . 24
2.3 Height gradients as pressure gradients . . . 26
2.4 Thickness and temperature . . . 29
2.5 Weather maps and their vocabulary . . . 32
2.6 Pressure field and near-surface wind . . . 35
2.7 What to retain . . . 37
2.8 Exercises . . . 37

3 Coriolis force and geostrophic wind **39**
3.1 The direction of large-scale wind . . . 39
3.2 The Coriolis force . . . 41
3.3 Geostrophic balance . . . 45
3.4 Computing the geostrophic wind . . . 49
3.5 Friction and the surface wind . . . 52
3.6 Geostrophic wind as forcing . . . 56
3.7 What to retain . . . 59
3.8 Exercises . . . 59

4 Gradient wind and curved flow **61**
4.1 From straight to curved flow . . . 61
4.2 The gradient wind balance . . . 63
4.3 Flow around lows and highs . . . 66
4.4 The limit on anticyclonic flow . . . 70
4.5 Vorticity . . . 73
4.6 Curvature at different scales . . . 76

4.7 What to retain . . . 80
4.8 Exercises . . . 80

5 Synoptic, mesoscale and microscale flow 83
5.1 Scale and the three ranges . . . 83
5.2 The synoptic scale and the mesoscale . . . 86
5.3 The microscale and terrain . . . 90
5.4 What a model resolves . . . 93
5.5 Kinds of model . . . 96
5.6 Dimensionless numbers and the scale of a wind value . . . 99
5.7 What to retain . . . 101
5.8 Exercises . . . 101

6 The atmospheric boundary layer 104
6.1 The layer that feels the surface . . . 104
6.2 The daily cycle and the convective layer . . . 107
6.3 The stable and residual layers . . . 110
6.4 Layers within the boundary layer . . . 112
6.5 Potential temperature and static stability . . . 116
6.6 Shear, turbulence and the wind profile . . . 120
6.7 What to retain . . . 124
6.8 Exercises . . . 124

7 The surface layer 127
7.1 The surface layer and its constant flux . . . 127
7.2 From stress to shear . . . 131
7.3 The logarithmic wind profile . . . 134
7.4 Roughness length and displacement height . . . 137
7.5 Fetch and internal boundary layers . . . 141
7.6 Stability and the limits of the theory . . . 144
7.7 What to retain . . . 147
7.8 Exercises . . . 148

8 Turbulence 149
8.1 Turbulence and turbulent transport . . . 149
8.2 Eddies and the energy cascade . . . 153
8.3 Turbulent kinetic energy and its budget . . . 157
8.4 Kinds of turbulence near the ground . . . 160
8.5 Turbulence statistics in engineering . . . 162
8.6 Turbulence in models . . . 165
8.7 What to retain . . . 169
8.8 Exercises . . . 171

9 Reynolds decomposition and turbulent fluxes 172
9.1 The decomposition of a measured record . . . 172
9.2 Covariance and the momentum flux . . . 175
9.3 Fluxes of heat and moisture . . . 180
9.4 Measuring fluxes by eddy covariance . . . 182
9.5 Averaging rules and the closure problem . . . 187
9.6 Correlation, quadrants and the wind profile . . . 190
9.7 What to retain . . . 194
9.8 Exercises . . . 194

10 Friction velocity **197**
10.1 Friction velocity as a velocity scale . . . 197
10.2 Friction velocity and the wind profile . . . 200
10.3 The drag coefficient . . . 203
10.4 Measuring and inferring friction velocity . . . 206
10.5 Scaling, similarity and limits . . . 209
10.6 The stress vector and calibration . . . 212
10.7 What to retain . . . 215
10.8 Exercises . . . 215

11 Aerodynamic roughness **217**
11.1 The roughness length as an aerodynamic property . . . 217
11.2 Roughness, the wind profile and the surface drag . . . 221
11.3 Displacement height and canopy layers . . . 223
11.4 Roughness by direction, fetch and time . . . 226
11.5 Roughness and terrain . . . 230
11.6 Estimating and using the roughness length . . . 231
11.7 What to retain . . . 235
11.8 Exercises . . . 235

12 The logarithmic wind profile **238**
12.1 The logarithmic profile and its origin . . . 238
12.2 Extrapolation between heights . . . 242
12.3 The power law . . . 244
12.4 Stability and the region of validity . . . 248
12.5 Failures of the profile and what they indicate . . . 250
12.6 The profile in models and practice . . . 254
12.7 What to retain . . . 257
12.8 Exercises . . . 257

13 Atmospheric stability **260**
13.1 Stability and the lapse rates . . . 260
13.2 Stability and potential temperature . . . 265
13.3 Stability and the wind profile . . . 267
13.4 The Richardson numbers . . . 270
13.5 Static and surface-layer stability . . . 273
13.6 Stability in space, time and models . . . 276
13.7 What to retain . . . 280
13.8 Exercises . . . 280

14 Monin–Obukhov similarity theory **283**
14.1 The similarity hypothesis . . . 283
14.2 The Obukhov length . . . 287
14.3 Dimensionless gradients . . . 290
14.4 Integrated profiles . . . 294
14.5 Use in a weather model . . . 298
14.6 Limits of the theory . . . 301
14.7 What to retain . . . 304
14.8 Exercises . . . 304

15 Obukhov length and the stability parameter **307**
15.1 Reading the Obukhov length . . . 307
15.2 The stability parameter at a height . . . 312
15.3 Stability and the wind profile . . . 315
15.4 Wind speed, heat flux and feedbacks . . . 318
15.5 Free convection and the whole boundary layer . . . 322
15.6 Stability in forecasts . . . 324
15.7 What to retain . . . 327
15.8 Exercises . . . 327

16 The daytime convective boundary layer **329**
16.1 Thermals and the mixed layer . . . 329
16.2 Growth and entrainment . . . 334
16.3 Velocity scales and flux profiles . . . 339
16.4 Convection and the near-surface wind . . . 342
16.5 Organized convection and local theory . . . 345
16.6 The surface beneath the convective layer . . . 348
16.7 What to retain . . . 351
16.8 Exercises . . . 351

17 The stable nocturnal boundary layer **354**
17.1 Formation and structure of the stable boundary layer . . . 354
17.2 Turbulence under stable stratification . . . 357
17.3 Regimes, collapse and bursts . . . 360
17.4 Wind above and across the stable layer . . . 365
17.5 Terrain at night . . . 369
17.6 Models, measurements and diagnosis . . . 371
17.7 What to retain . . . 373
17.8 Exercises . . . 375

18 Low-level jets **376**
18.1 The jet profile . . . 376
18.2 The nocturnal inertial jet . . . 379
18.3 The jet through the night and morning . . . 385
18.4 Other kinds of low-level jet . . . 387
18.5 Observing and detecting jets . . . 391
18.6 Consequences for applications . . . 394
18.7 What to retain . . . 396
18.8 Exercises . . . 398

19 Thermal circulations **399**
19.1 Differential heating as a source of wind . . . 399
19.2 Sea and land breezes . . . 403
19.3 Slope, valley and drainage winds . . . 409
19.4 Cities, convergence and density currents . . . 412
19.5 Large-scale flow and stability . . . 415
19.6 Climatology, models and diagnosis . . . 417
19.7 What to retain . . . 419
19.8 Exercises . . . 420

20 Terrain-induced flow **423**
20.1 Flow over a hill . . . 423
20.2 The magnitude of the speed-up . . . 427
20.3 Separation and sheltering . . . 430
20.4 Channeling, gap flow and exposure . . . 433
20.5 Stratified flow over terrain . . . 435
20.6 Terrain in models and practice . . . 438
20.7 What to retain . . . 441
20.8 Exercises . . . 441

21 Vegetation and urban canopies **444**
21.1 The canopy as a distributed momentum sink . . . 444
21.2 Wind in and above a canopy . . . 447
21.3 Vegetation canopies . . . 452
21.4 Urban canopies . . . 454
21.5 Representing canopies in models . . . 460
21.6 Canopies and wind measurements . . . 462
21.7 What to retain . . . 464
21.8 Exercises . . . 464

22 Wind measurements **466**
22.1 What a wind sensor reports . . . 466
22.2 In-situ wind instruments . . . 469
22.3 Sampling, response and averaging . . . 474
22.4 Mounting and flow distortion . . . 479
22.5 Wind profiling instruments . . . 481
22.6 Measurement errors and quality control . . . 484
22.7 What to retain . . . 490
22.8 Exercises . . . 490

23 Measurement height and exposure **493**
23.1 Height and the standard measurement . . . 493
23.2 Obstacles, fetch and footprint . . . 497
23.3 Terrain exposure . . . 501
23.4 Representativeness . . . 503
23.5 Comparing a station with a model . . . 506
23.6 Metadata, homogeneity and networks . . . 511
23.7 What to retain . . . 514
23.8 Exercises . . . 516

24 Wind statistics **517**
24.1 Means and variability of the wind . . . 517
24.2 Statistics of wind direction . . . 521
24.3 The distribution of wind speed . . . 523
24.4 Exceedance, quantiles and extremes . . . 527
24.5 Structure in time . . . 530
24.6 Verification of modeled wind . . . 535
24.7 What to retain . . . 538
24.8 Exercises . . . 538

25 Numerical weather prediction **541**
25.1 What a weather model solves . . . 541
25.2 The model grid . . . 544
25.3 The surface and the boundary layer . . . 549
25.4 Starting and bounding the forecast . . . 553
25.5 Interactions and limits . . . 557
25.6 Reading model wind . . . 559
25.7 What to retain . . . 562
25.8 Exercises . . . 562

26 Boundary-layer parameterizations **565**
26.1 The task of a boundary-layer scheme . . . 565
26.2 Orders of closure . . . 569
26.3 Convection and nonlocal schemes . . . 573
26.4 Stable boundary-layer closures . . . 578
26.5 The scheme inside the model . . . 581
26.6 Resolution and scale awareness . . . 584
26.7 What to retain . . . 588
26.8 Exercises . . . 588

27 Terrain and land-surface representation **591**
27.1 The effective surface of a grid cell . . . 591
27.2 Terrain in the model . . . 594
27.3 Land cover in the model . . . 599
27.4 Subgrid terrain . . . 603
27.5 The thermal state of the surface . . . 606
27.6 Consequences for validation and correction . . . 608
27.7 What to retain . . . 612
27.8 Exercises . . . 612

28 Statistical and dynamical downscaling **615**
28.1 The downscaling problem . . . 615
28.2 Simple statistical corrections . . . 617
28.3 Regression and model output statistics . . . 620
28.4 Machine-learning downscaling . . . 624
28.5 Dynamical and microscale downscaling . . . 628
28.6 Choosing and combining methods . . . 630
28.7 What to retain . . . 633
28.8 Exercises . . . 633

29 CFD and microscale wind modeling **636**
29.1 Computational fluid dynamics for local wind . . . 636
29.2 Reynolds-averaged models . . . 639
29.3 Large-eddy simulation . . . 643
29.4 Domain, boundary conditions and mesh . . . 646
29.5 Separation, stability and large-scale forcing . . . 650
29.6 From simulation to site prediction . . . 653
29.7 What to retain . . . 658
29.8 Exercises . . . 658

30 Building a defensible site wind model **661**
30.1 The site wind problem and its target 661
30.2 Physical hierarchy and atmospheric regime 665
30.3 Site characterization and observations 667
30.4 Baselines, predictors and model complexity 670
30.5 Validation and physical diagnosis 673
30.6 Uncertainty, decisions and the model report 677
30.7 What to retain 685
30.8 Exercises 686

Notation and vocabulary

Every symbol below has the same meaning in every chapter. SI units are used throughout; pressure is quoted in hectopascals ($1\,\mathrm{hPa} = 100\,\mathrm{Pa}$), as is usual in meteorology. Signs that depend on the hemisphere follow from the sign of the Coriolis parameter f, positive in the Northern Hemisphere and negative in the Southern.

Symbol	Unit	Meaning
x, y, z	m	Eastward, northward and upward coordinates; z is geometric height.
u, v, w	m/s	Eastward, northward and upward components of the air velocity.
$\mathbf{V}$	m/s	Horizontal wind vector, (u, v).
U	m/s	Horizontal wind speed, $\sqrt{u^2 + v^2}$.
U_{10}	m/s	Wind speed at 10 m above the ground.
β	degree	Wind direction: where the wind comes from, clockwise from true north; $u = -U\sin\beta$, $v = -U\cos\beta$.
$\mathbf{V}_g$, U_g	m/s	Geostrophic wind and its speed: the wind that balances the pressure-gradient and Coriolis forces.
U_{gr}	m/s	Gradient wind speed: the balanced wind on a curved path.
$\mathbf{V}_a$	m/s	Ageostrophic wind, $\mathbf{V} - \mathbf{V}_g$.
p	Pa, hPa	Atmospheric pressure.
p_{stn}, p_{msl}	Pa, hPa	Station pressure (measured at the instrument's elevation) and mean sea-level pressure (reduced to sea level).
ρ	kg/m^3	Air density.
T, T_v	K	Absolute temperature; virtual temperature, the temperature at which dry air has the density of the moist air.
$\overline{T_v}$	K	Mean virtual temperature of a layer between two pressure levels.
R_d	J/(kg K)	Specific gas constant of dry air, 287 J/(kg K).
g, g_0	m/s^2	Gravitational acceleration; its standard value 9.80665 m/s^2.
H	m	Pressure scale height, $R_d T/g$.
Φ	m^2/s^2	Geopotential: the work needed to raise a unit mass from sea level to height z.
Z	m	Geopotential height, Φ/g_0.
∇_h, ∇_p	1/m	Horizontal gradient at constant height; horizontal gradient along a constant-pressure surface.
Ω, $\boldsymbol{\Omega}$	rad/s	Rotation rate of the Earth, 7.292×10^{-5} rad/s, and its vector.

Symbol	Unit	Meaning
ϕ	degree, rad	Latitude, positive north of the equator.
f	1/s	Coriolis parameter, $2\Omega \sin\phi$.
$\mathbf{k}$	–	Unit vector pointing upward.
$\frac{\mathrm{D}}{\mathrm{D}t}$	1/s	Rate of change following an air parcel.
R	m	Radius of curvature of an air parcel's path.
ζ, η	1/s	Relative vorticity, $\partial v/\partial x - \partial u/\partial y$; absolute vorticity, $f + \zeta$.
L, H	m	Horizontal and vertical length scales of a flow feature. (The pressure scale height is written H.)
Δx	m	Grid spacing of a numerical model.
ν	m^2/s	Kinematic viscosity of air.
Ro	–	Rossby number, $U/(\lvert f\rvert L)$: inertia relative to the Coriolis force.
Re	–	Reynolds number, UL/ν: inertia relative to viscosity.
Ri	–	Richardson number: buoyancy relative to shear.
$\mathbf{a}_{\mathrm{PGF}}$, $\mathbf{a}_C$	$\mathrm{m/s}^2$	Pressure-gradient and Coriolis forces per unit mass (accelerations).
$\mathbf{F}$	$\mathrm{m/s}^2$	Friction force per unit mass: near the ground, the effect of turbulent momentum transfer.
α	degree	Angle between the surface wind and the isobars (Chapter 3); elsewhere a profile exponent or a slope angle, as stated.
K, δ	m^2/s, m	Eddy viscosity and Ekman length, $\delta = \sqrt{2K/\lvert f\rvert}$ (Chapter 3).
λ	1/s, m	Linear drag rate (Chapter 3); wavelength (Chapter 5).
$\mathbf{t}$, $\mathbf{n}$	–	Unit vectors along the flow and to its left (natural coordinates, Chapter 4).
Ro_R	–	Curvature Rossby number, $U_g/(\lvert f\rvert R)$.
$\mathcal{T}$	s	Time scale of a flow feature; the advective time is L/U.
θ, θ_v	K	Potential temperature, $T(p_0/p)^{R_d/c_p}$ with $p_0 = 1000\,\mathrm{hPa}$; virtual potential temperature.
c_p	J/(kg K)	Specific heat of air at constant pressure, 1004 J/(kg K).
z_i	m	Depth of the atmospheric boundary layer.
$\overline{a}$, a'	as a	Mean of a quantity a over the averaging period; its fluctuation, $a - \overline{a}$.
σ_u, σ_v, σ_w	m/s	Standard deviations of the along-wind, cross-wind and vertical velocity.
$\overline{u'w'}$, $\overline{v'w'}$	$\mathrm{m}^2/\mathrm{s}^2$	Kinematic vertical fluxes of along-wind and cross-wind momentum (covariances).
$\overline{w'\theta'}$, $\overline{w'T'}$	K m/s	Kinematic vertical heat flux, with potential temperature or temperature.
τ, τ_0	Pa	Turbulent shear stress, $-\rho\,\overline{u'w'}$ in along-wind coordinates; its value at the surface.
u_*	m/s	Friction velocity, $\sqrt{\tau_0/\rho}$: the velocity scale of the surface stress, not a wind speed.
w_*	m/s	Convective velocity scale, for buoyancy-driven turbulence.
κ	–	von Kármán constant, about 0.4.
l	m	Mixing length; near the ground $l = \kappa z$.

Symbol	Unit	Meaning
K_m, K_h	m^2/s	Eddy diffusivities for momentum and for heat: properties of the turbulence, not of the air.
z_0	m	Aerodynamic roughness length: the height at which the neutral logarithmic profile extrapolates to zero wind.
d, h_c	m	Zero-plane displacement height; height of the canopy or obstacles.
C_D	–	Drag coefficient, $\tau_0/(\rho U^2)$, for the wind at a stated height.
e	m^2/s^2	Turbulent kinetic energy per unit mass, $\frac{1}{2}(\overline{u'^2}+\overline{v'^2}+\overline{w'^2})$.
P_s, P_b, ε	m^2/s^3	Shear production, buoyancy production and dissipation rate of turbulent kinetic energy.
TI	–	Turbulence intensity, the standard deviation of wind speed divided by its mean.
H, LE	W/m^2	From Chapter 8 on: sensible and latent heat flux.
q, L_v	kg/kg, J/kg	Specific humidity; latent heat of vaporization of water.
r_{uw}	–	Correlation coefficient of the along-wind and vertical fluctuations, $\overline{u'w'}/(\sigma_u\sigma_w)$.
N	1/s	Buoyancy (Brunt–Väisälä) frequency, $N^2=(g/\theta)\,\mathrm{d}\theta/\mathrm{d}z$.
T_I	s	Inertial period, $2\pi/\lvert f\rvert$.
z_{jet}, U_{jet}	m, m/s	Height and wind speed of the core of a low-level jet.
B	–	Bowen ratio, H/LE: sensible over latent heat flux at the surface.
Fr	–	Froude number of stratified flow over terrain, $U/(Nh_m)$.
h_m, L_h	m	Height of a hill or mountain; its half-width, the horizontal distance from the crest to half the height.
S	–	Speed-up factor: wind at a point over terrain divided by the reference wind at the same height above ground upstream.
a, LAI, λ_f	1/m, –, –	Frontal area density of a canopy; leaf area index; frontal area index (frontal area of the obstacles per unit ground area).
G	–	Gust factor: gust speed divided by the mean speed of the same period.
f_s, f_N	Hz	Sampling frequency; Nyquist frequency, $f_s/2$.
c, k	m/s, –	Scale and shape parameters of the Weibull distribution of wind speed (in Chapter 24).
$\mathcal{T}_u$, Λ_u	s, m	Integral time scale and integral length scale of the along-wind turbulence.
RMSE, MAE	as the variable	Root-mean-square error and mean absolute error of a prediction.
ν_t	m^2/s	Eddy viscosity of a turbulence model (in Chapter 29; elsewhere K_m).
Ri_f	–	Flux Richardson number: buoyancy destruction of turbulent kinetic energy divided by its shear production.
$E(k)$, k	m^3/s^2, rad/m	Energy spectrum of turbulence and wavenumber.
η_K	m	Kolmogorov length, $(\nu^3/\varepsilon)^{1/4}$: the size of the smallest eddies.

Symbol	Unit	Meaning
L_{O}	m	Obukhov length, $-u_*^3\,\overline{\theta_v}/(\kappa\, g\,\overline{w'\theta_v'})$: the height scale at which buoyancy becomes as important as shear in producing or destroying turbulence. Negative in unstable, positive in stable conditions.
z/L_{O}	–	Stability parameter of the surface layer at height z (written ζ in much of the literature; this book keeps ζ for vorticity).
θ_*	K	Surface-layer temperature scale, $-\overline{w'\theta'}/u_*$.
ϕ_m, ϕ_h	–	Dimensionless wind shear, $(\kappa z/u_*)\,\partial U/\partial z$, and dimensionless potential-temperature gradient; functions of z/L_{O}, equal to one in neutral conditions.
ψ_m, ψ_h	–	Integrated stability corrections to the wind and temperature profiles.
Pr_t	–	Turbulent Prandtl number, K_m/K_h.
Γ_e, Γ_d	K/m	Environmental lapse rate, $-\partial T/\partial z$; dry adiabatic lapse rate, g/c_p.
Ri_g, Ri_B, Ri_c	–	Gradient and bulk Richardson numbers; critical value, about 0.25.
α, z_r	–, m	Exponent and reference height of the power-law wind profile, $U(z) = U(z_r)(z/z_r)^{\alpha}$.
α_c	–	Charnock parameter, in the roughness length of a water surface, $z_0 = \alpha_c\, u_*^2/g$.
δ	m	Also, in Chapter 7, the height of an internal boundary layer.

Vocabulary

Wind The motion of air relative to the Earth's surface; unless stated otherwise, its horizontal part.

Wind direction The direction the wind blows *from*: a southwesterly wind blows from the southwest toward the northeast.

Forcing and response A forcing is a physical driver imposed on the flow, such as a large-scale pressure gradient; the response is what the atmosphere does with it, which depends on friction, stability and terrain.

Free atmosphere The atmosphere above the layer that is directly influenced by the surface.

Atmospheric boundary layer (ABL) The part of the atmosphere that responds to the surface within about an hour or less.

Synoptic, mesoscale, microscale The three scale ranges of atmospheric motion: weather systems of about a thousand kilometers, circulations of one to a few hundred kilometers, and local flow below about one kilometer.

Resolved, parameterized A process a model's grid represents explicitly, against a process whose average effect the model computes from a formula.

Grid-cell value, point value The value a model reports for a whole grid cell, against the value an instrument measures at one place.

Measured, calculated, inferred A sensor reading; a quantity computed from readings by a known relation; a conclusion about a quantity that was not observed.

Chapter 1

The Atmosphere as a Wind System

1.1 The question and the framework

Every wind model, from a weather model with a grid of kilometers to the correction applied to a single anemometer, rests on a picture of why air moves. This chapter draws that picture in outline. Its question is

what causes wind, and why is the wind measured near the ground not simply the wind that the large-scale weather would produce?

The short answer is a chain of causes. The Sun heats the Earth unequally; unequal heating produces differences of air density; density differences produce differences of pressure; pressure differences accelerate the air; and the accelerated air is the wind. Wind is the atmosphere continuously working to remove pressure differences that unequal heating continuously creates. Near the surface, the result of that chain is then modified strongly by the rotation of the Earth, by friction against the ground, by turbulence, by terrain and by the stability of the air, which is whether vertical motions are damped or amplified by buoyancy. The rest of the chapter states each link of the chain once, with the depth needed to see the whole; Chapters 2 to 5 develop the first links in detail.

Jargon: free atmosphere, geostrophic forcing, stability, friction velocity, roughness length, roughness sublayer, fetch, nocturnal jet, parameterization

The *free atmosphere* is the atmosphere above the layer directly influenced by the surface (Section 1.4). *Geostrophic forcing* is the large-scale horizontal pressure gradient, expressed by the speed of the wind that would balance it against the Coriolis force, the geostrophic wind U_g in m/s (Chapter 3). *Stability* describes whether a parcel of air displaced vertically is pushed back (stable), left in place (neutral) or pushed further (unstable) by buoyancy; it is dimensionless in its usual measures and is treated in Chapter 13. The *friction velocity* u_*, in m/s, is a velocity scale built from the turbulent stress the surface exerts on the air, about a tenth of the mean wind near the ground over land. The *roughness length* z_0, in meters, is the height scale that characterizes how strongly a surface slows the wind; it ranges from about 10^{-4} m over calm water to about 1 m over forests and cities. The *roughness sublayer* is the layer within a few obstacle heights of the ground where the flow is dominated by individual trees and buildings. The *fetch* is the upstream distance over which the surface has been uniform, which sets how well the flow at a sensor has adjusted to that surface. A *nocturnal jet* (low-level jet) is a wind maximum a few hundred meters above the ground that forms at night when the air near the surface is stable. A *parameterization* is a formula with which a model computes the average effect of a process it cannot resolve on its grid, such as the turbulence of the boundary layer. Each of these terms receives its full treatment in a later chapter; they are named here because the framework below refers to them.

Key idea

Wind is the atmosphere's response to pressure differences created by unequal heating, turned by the Earth's rotation and slowed by friction; the wind measured at a point near the ground is the last link of a chain of transformations that starts in the large-scale pressure field and passes through the boundary layer, the surface and the terrain.

1.1.1 The organizing question

The question that organizes this book is: why is the wind at a given location, height and time what it is? The answer is built by following the wind downward through a sequence of layers and processes,

$$\text{synoptic forcing} \rightarrow \text{boundary layer} \rightarrow \text{surface layer} \rightarrow \text{terrain and roughness} \rightarrow \text{local flow} \rightarrow \text{measured wind},$$

in which each step receives the wind produced by the step above and transforms it. The first step is the large-scale pressure field of weather maps; the last is the reading of an instrument at a stated height above a particular piece of ground. The same decomposition can be written as a list of contributions,

$$\text{large-scale forcing} + \text{boundary-layer state} + \text{surface properties} + \text{terrain} + \text{local obstacles} = \text{local wind},$$

where the plus signs mean that each item contributes, not that the contributions add arithmetically. They do not: a rough surface reduces the wind by a fraction that depends on the stability, and a ridge accelerates whatever wind arrives at it. The precise statement is the functional relation of Section 1.5.1.

Every new quantity in the book is introduced in the same order: its name, its physical meaning, its units, its equation, and how it is measured or modeled. A second rule runs through every chapter: what the atmosphere does is kept apart from what a model or an instrument represents of it. Real turbulence is not the same object as a boundary-layer parameterization,

which is a formula for the average effect of that turbulence on a model grid. Real terrain is not the same object as the terrain of a model grid cell, which is an average elevation over several square kilometers. A wind value is therefore incomplete without a statement of where it applies, at what height, over what averaging time and at what spatial scale.

1.1.2 One anemometer reading

The framework is used by asking a fixed set of questions of every wind value. Take an illustrative anemometer at 20 m above the ground that reads $7\,\mathrm{m\,s^{-1}}$. Read through the framework, the reading raises these questions, from the top of the chain to the bottom:

- What is the geostrophic forcing, the wind that the large-scale pressure gradient would drive in the absence of friction?
- What is the stability of the boundary layer, and how deep is the boundary layer at this hour?
- What are the friction velocity u_* and the roughness length z_0 of the upstream surface?
- Is 20 m inside the roughness sublayer of nearby trees or buildings?
- How long and how uniform is the upstream fetch?
- Is the sensor on a slope or a crest where the terrain accelerates the flow?
- Is there a nocturnal jet above the sensor that the boundary layer may at times mix downward?
- Does the weather model used for comparison resolve any of these features, or does it represent them only through parameterizations and grid-cell averages?

None of these questions has the same answer at noon and at midnight, or over a forest and over a lake, and each of them changes how the $7\,\mathrm{m\,s^{-1}}$ should be interpreted and transferred to another height or another place.

The applications that motivate the book are all of this kind. A transmission line is cooled by the wind that crosses its conductors at heights of 10 to 50 m, so the current it can carry safely, its dynamic line rating, depends on the speed and direction of the wind along every span. A wind turbine is driven by the wind at 80 to 150 m. A sensor network is only as useful as the siting of its instruments, and the flow over hills and valleys decides whether a measurement at one place says anything about another. In each case the wind of interest is near-surface engineering wind, and in each case the large-scale weather is necessary but not sufficient to know it.

1.2 Wind and its origin

This section defines wind as a vector, with the convention used to report its direction, and follows the first links of the causal chain: unequal heating, the density differences it produces through the ideal gas law, and the way temperature acts on the wind only through pressure.

1.2.1 Wind as air in motion

Wind is the motion of air relative to the Earth's surface. At a point it is a velocity, a vector with three components, (u, v, w), in meters per second: u is the eastward component (positive toward the east, negative toward the west), v the northward component (positive toward the north), and w the upward component. The horizontal wind vector is

$$\mathbf{V} = (u, v), \tag{1.1}$$

and the horizontal wind speed is its magnitude,

$$U = |\mathbf{V}| = \sqrt{u^2 + v^2}. \tag{1.2}$$

Near the surface, the vertical velocity averaged over minutes is usually much smaller than the horizontal components, $|w| \ll |u|, |v|$; vertical velocities of weather systems are of the order of centimeters per second against horizontal winds of meters per second. For this reason "wind" means the horizontal motion unless stated otherwise. The instantaneous turbulent fluctuations of w near the ground are not small, often tenths of a meter per second and more in daytime convection, and they carry momentum and heat vertically; Chapters 8 and 9 treat them.

Direction. Meteorology reports wind direction as the direction the wind blows *from*, measured in degrees clockwise from true north. Writing β for that direction, a north wind ($\beta = 0°$ or $360°$) blows from the north toward the south, an east wind ($\beta = 90°$) from the east, and a southwesterly ($\beta = 225°$) from the southwest toward the northeast. The components follow from the speed and direction as

$$u = -U \sin\beta, \qquad v = -U \cos\beta, \tag{1.3}$$

where the minus signs express that the air moves toward the direction opposite to β. Inverting, β is the angle whose sine is proportional to $-u$ and whose cosine is proportional to $-v$, computed as the two-argument arctangent $\beta = \operatorname{atan2}(-u, -v)$ and mapped to $[0°, 360°)$; it is undefined in calm air, $U = 0$. For an illustrative southwesterly of $U = 10\,\mathrm{m\,s^{-1}}$,

$$u = -10 \sin 225° = -10 \times (-0.7071) = +7.07\,\mathrm{m\,s^{-1}}, \qquad v = -10 \cos 225° = +7.07\,\mathrm{m\,s^{-1}},$$

both positive because the air moves toward the northeast.

Figure 1.1 draws this example. The circle is a compass seen from above, with north at the top and east to the right. The thick arrow is the wind vector of the southwesterly; the two thin arrows are its eastward component u and northward component v; the arc measures the direction β clockwise from north to the side the wind comes from. The arrow points away from the direction it is named after, which is the usual source of sign errors when converting between components and directions.

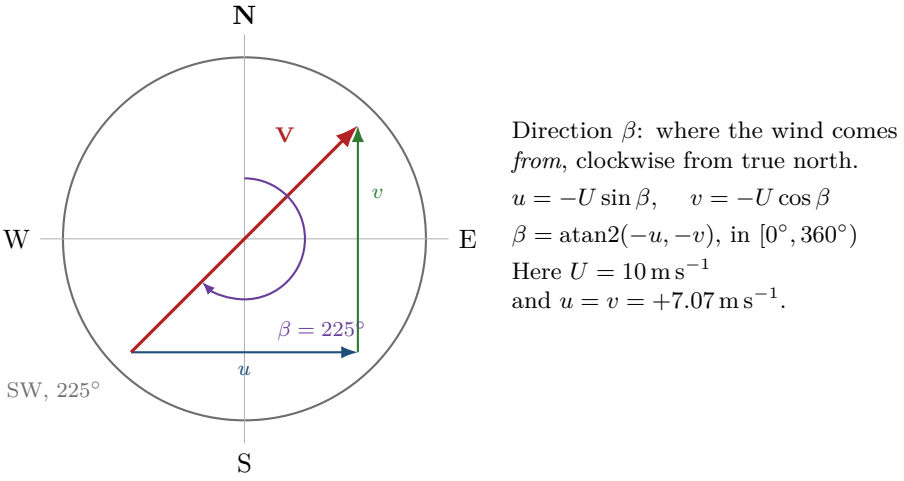


Figure 1.1: Components and direction of an illustrative southwesterly wind of $10\,\mathrm{m\,s^{-1}}$, seen from above, north at the top. The vector points toward the northeast; the direction $\beta = 225°$ names the southwest, where the wind comes from (equation (1.3)).

Measurement and modeling. A cup anemometer measures the speed U and a wind vane the direction β; a sonic anemometer measures the three components u, v and w directly, many times a second. A weather model carries u and v as variables at every grid point and level;

some models report them relative to the axes of their map projection, which then differ from east and north and must be rotated before equation (1.3) is applied. A direction measured with a magnetic compass must also be corrected to true north.

1.2.2 Unequal solar heating

The energy that drives the wind comes from the Sun, and the Sun does not heat the Earth evenly. The heating at a place depends on its latitude, which sets the angle of the sunlight; on the season and the time of day; on cloud cover; on the type of surface, since dark soil absorbs more than snow and a wet surface spends energy on evaporation that a dry one turns into heat; on soil moisture and vegetation; on elevation; and on whether the surface is ocean or land. Water heats and cools slowly because it mixes the absorbed energy through a deep layer and has a large heat capacity; land heats and cools quickly because only a thin layer takes part. A coast on a sunny afternoon therefore has warm air over the land beside cooler air over the sea, and a mountain slope in sunlight has warm air beside cooler air at the same height over the valley.

These differences act on every scale. Between the equator and the poles they drive the large-scale circulation and its weather systems; between land and sea they drive sea breezes; between a sunlit slope and the free air beside it they drive slope winds. In each case the heating is the original cause, but it is not the force that moves the air. The next two subsections follow how a difference of temperature becomes a force.

1.2.3 The ideal gas law and air density

Air behaves very nearly as an ideal gas, whose pressure, density and temperature are related by

$$p = \rho\, R_d\, T \qquad \Longleftrightarrow \qquad \rho = \frac{p}{R_d\, T}, \tag{1.4}$$

where p is the pressure in pascals, ρ the density in kilograms per cubic meter, T the absolute temperature in kelvin, and $R_d = 287\ \mathrm{J/(kg\,K)}$ the specific gas constant of dry air. Moist air is slightly less dense than dry air at the same pressure and temperature; Chapter 2 accounts for this with the virtual temperature, and here the dry form is enough. Equation (1.4) says that at equal pressure, warmer air is less dense: density is inversely proportional to absolute temperature.

For a worked example, take air at the standard sea-level pressure $p = 1013.25\,\mathrm{hPa} = 101\,325\,\mathrm{Pa}$. At $T = 15^\circ\mathrm{C} = 288.15\,\mathrm{K}$,

$$\rho = \frac{101\,325\,\mathrm{Pa}}{287\,\mathrm{J/(kg\,K)} \times 288.15\,\mathrm{K}} = \frac{101\,325}{82\,699} = 1.225\,\mathrm{kg/m^3}.$$

At $T = 35^\circ\mathrm{C} = 308.15\,\mathrm{K}$ and the same pressure, $\rho = 101\,325/(287 \times 308.15) = 101\,325/88\,439 = 1.146\,\mathrm{kg/m^3}$. The ratio of the two densities is $288.15/308.15 = 0.935$: a warming of 20 K makes the air 6.5 % less dense. A difference of this size between neighboring columns of air is ordinary over a sunny coast or a heated slope. Air temperature is measured by shielded thermometers; pressure by barometers; density is almost never measured directly but calculated from the two with equation (1.4).

1.2.4 Temperature acts through pressure

A temperature difference does not push air sideways. There is no horizontal force proportional to temperature in the equation of motion (Section 1.3); the only horizontal forces are pressure forces, the Coriolis force and friction. Temperature differences create wind indirectly, by changing how pressure is distributed in the vertical. The chain is

unequal heating → density differences → pressure differences → acceleration of the air → wind.

The middle link works as follows. Pressure decreases with height because it is the weight of the air above (Section 1.3.1), and it decreases more slowly in a column of less dense, warmer air. For an idealized column at a uniform temperature, the hydrostatic balance of Section 1.3.2 combined with equation (1.4) gives $\mathrm{d}p/p = -g\,\mathrm{d}z/(R_d T)$, and integrating from the ground, where $p = p_0$,

$$p(z) = p_0\, e^{-z/H}, \qquad H = \frac{R_d\, T}{g}, \tag{1.5}$$

where H is the pressure scale height, the height over which the pressure falls by a factor e; Chapter 2 develops this relation. Take two illustrative columns with the same surface pressure $p_0 = 1000\,\mathrm{hPa}$, a cold one at 10°C (283.15 K, $H = 287 \times 283.15/9.81 = 8284\,\mathrm{m}$) and a warm one at 20°C (293.15 K, $H = 8576\,\mathrm{m}$). At $z = 1000\,\mathrm{m}$,

$$p_{\mathrm{cold}} = 1000\, e^{-1000/8284} = 886.3\,\mathrm{hPa}, \qquad p_{\mathrm{warm}} = 1000\, e^{-1000/8576} = 889.9\,\mathrm{hPa}.$$

At the same height, the pressure over the warm column is 3.7 hPa higher than over the cold one, although the two surface pressures are equal. Equivalently, each pressure level lies higher in the warm column: the 900 hPa level is at 873 m in the cold column and at 904 m in the warm one.

Figure 1.2 draws the two columns in a vertical cross section. The horizontal axis is distance, with the cold column on the left and the warm column on the right; the vertical axis is height. The gray lines are levels of equal pressure, which rise from the cold to the warm column, with their vertical displacement exaggerated so that it is visible. At a fixed height above the ground the pressure is higher over the warm column, so the horizontal pressure difference aloft pushes air from the warm column toward the cold one (upper arrow). As air leaves the warm column aloft, the mass above its surface decreases and its surface pressure falls; the cold column gains mass and its surface pressure rises. The surface pressure difference then drives air near the ground from the cold side toward the warm side (lower arrow), and the circulation closes with rising air over the warm surface and sinking air over the cold one. This is the mechanism of the sea breeze, in which the cold column is over the sea and the warm one over the land. Temperature set up the circulation, but every arrow in the figure is driven by a pressure difference.

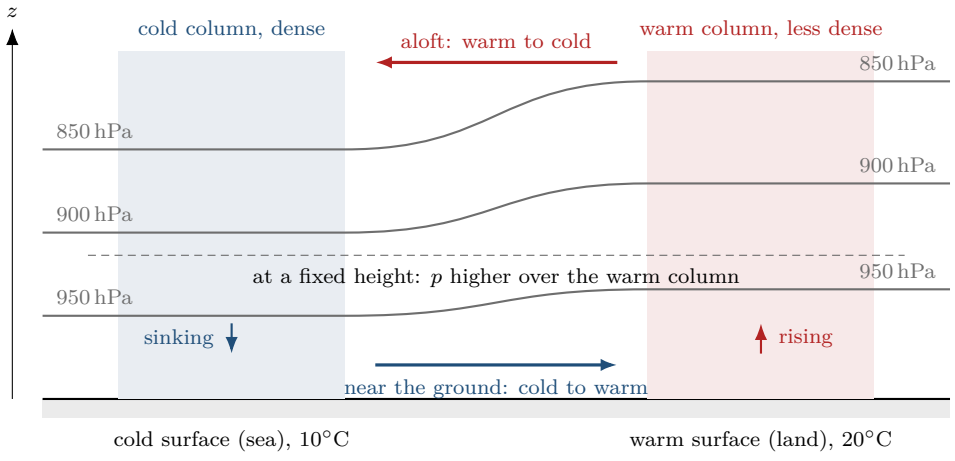


Figure 1.2: Vertical cross section of a cold and a warm column of air with equal surface pressure (schematic; vertical displacements of the pressure levels exaggerated). Gray lines: levels of equal pressure, higher in the warm column because pressure falls more slowly with height in less dense air. For uniform temperatures of 10 and 20°C and $p_0 = 1000\,\mathrm{hPa}$, equation (1.5) places the 900 hPa level at 873 and 904 m. Arrows: the circulation the resulting pressure differences drive.

1.3 Pressure and the forces on air

This section states what pressure is, writes the vertical balance that nearly always holds, derives the horizontal force that pressure differences exert, introduces the Coriolis force in outline, and assembles the three horizontal forces into one equation with the size of each term.

1.3.1 Pressure as the weight of the air above

Pressure is force per unit area, in pascals ($1\,\mathrm{Pa} = 1\,\mathrm{N/m^2}$); meteorology quotes it in hectopascals, $1\,\mathrm{hPa} = 100\,\mathrm{Pa}$. At any level in the atmosphere at rest, the pressure is very nearly the weight of the column of air above a unit area at that level. At sea level the standard pressure is $1013.25\,\mathrm{hPa} = 101\,325\,\mathrm{Pa}$, so the air above each square meter of sea surface has a mass of about

$$\frac{101\,325\,\mathrm{N/m^2}}{9.81\,\mathrm{m/s^2}} = 10\,330\,\mathrm{kg/m^2},$$

roughly ten metric tons per square meter. Going up, less air remains above, and the pressure decreases: by about half at 5.5 km, where the 500 hPa level is usually found.

Although pressure arises from the weight of the air above, it acts equally in all directions at a point: a small surface placed vertically feels the same pressure as one placed horizontally. This is why a pressure difference between two places at the same height exerts a horizontal force, even though the weight that produces the pressure acts vertically. Pressure is measured with barometers at stations (station pressure) and is reduced to a common reference, mean sea level, before horizontal comparisons are made; Chapter 2 treats that reduction.

1.3.2 Hydrostatic balance

Consider a thin horizontal slab of air of area A, between the heights z and $z + \mathrm{d}z$. Three vertical forces act on it: the pressure from below pushes up with $p(z)\,A$; the pressure from above pushes down with $p(z + \mathrm{d}z)\,A$; and gravity pulls the slab's mass $\rho A\,\mathrm{d}z$ down with the force $\rho g A\,\mathrm{d}z$. If the slab is not accelerating vertically, the three forces sum to zero,

$$p(z)\,A - p(z + \mathrm{d}z)\,A - \rho g A\,\mathrm{d}z = 0.$$

With $p(z + \mathrm{d}z) = p(z) + (\partial p/\partial z)\,\mathrm{d}z$, dividing by $A\,\mathrm{d}z$ gives

$$\frac{\partial p}{\partial z} = -\rho\, g, \tag{1.6}$$

the *hydrostatic balance*: the vertical pressure gradient balances gravity. Since ρ and g are positive, pressure decreases upward. Integrating equation (1.6) from a level z to the top of the atmosphere, where the pressure vanishes, gives $p(z) = \int_z^\infty \rho g\,\mathrm{d}z'$, which is exactly the weight per unit area of the air above z, the statement of Section 1.3.1.

Near sea level, with $\rho = 1.225\,\mathrm{kg/m^3}$ and $g = 9.81\,\mathrm{m/s^2}$, the pressure falls at

$$\rho g = 1.225 \times 9.81 = 12.0\,\mathrm{Pa/m}, \qquad \text{that is} \qquad 12.0 \times 100\,\mathrm{m} = 1202\,\mathrm{Pa} \approx 12\,\mathrm{hPa\ per\ 100\,m},$$

or about 1 hPa for every 8 m of ascent. (Over 100 m the density changes by about 1 %, so treating it as constant is accurate here.) A barometer carried from the base to the top of a 100 m tower reads about 12 hPa less at the top, which is three times the horizontal pressure difference across 100 km in a strong weather system (Section 1.3.3). The vertical pressure gradient is therefore thousands of times larger than the horizontal one, but it is almost exactly balanced by gravity, while the much smaller horizontal gradient is balanced by nothing of comparable size except the Coriolis force and friction. The atmosphere is usually very close to

hydrostatic balance; the balance fails in strong small-scale vertical motions, such as the updrafts of thunderstorms, and Chapter 2 builds on it.

Figure 1.3 shows the slab on the left. The box is the slab, between the heights z and $z + \mathrm{d}z$; the arrows on its lower and upper faces are the pressure forces, and the arrow from its center is its weight. In hydrostatic balance the upward arrow equals the sum of the two downward ones.

1.3.3 The pressure-gradient force

The same reasoning applied horizontally gives the force that starts the wind. Take a small box of air with sides $\mathrm{d}x$, $\mathrm{d}y$, $\mathrm{d}z$, and let the pressure vary in the x direction. The face at x is pushed in the $+x$ direction with the force $p(x)\,\mathrm{d}y\,\mathrm{d}z$; the face at $x + \mathrm{d}x$ is pushed in the $-x$ direction with $p(x + \mathrm{d}x)\,\mathrm{d}y\,\mathrm{d}z$. The net force in x is

$$\big[p(x) - p(x + \mathrm{d}x)\big]\,\mathrm{d}y\,\mathrm{d}z = -\frac{\partial p}{\partial x}\,\mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z,$$

and dividing by the box's mass $\rho\,\mathrm{d}x\,\mathrm{d}y\,\mathrm{d}z$ gives the force per unit mass, an acceleration, $-(1/\rho)\,\partial p/\partial x$. Doing the same in y and collecting the two horizontal components,

$$\mathbf{a}_{\mathrm{PGF}} = -\frac{1}{\rho}\,\nabla_h p, \qquad \nabla_h p = \left(\frac{\partial p}{\partial x}, \frac{\partial p}{\partial y}\right), \tag{1.7}$$

the *pressure-gradient force* per unit mass, in $\mathrm{m/s^2}$. The gradient $\nabla_h p$ points toward increasing pressure, so the minus sign makes the force point toward decreasing pressure: from high to low pressure, perpendicular to the isobars (lines of equal pressure on a map), and stronger where the isobars are closer together. The right panel of Figure 1.3 draws the box with its two pressure forces; the higher pressure on the left face wins and the net force points to the right, toward the lower pressure. The pressure-gradient force is the first force of wind generation: without a horizontal pressure difference, air at rest stays at rest.

For an illustrative strong gradient of 4 hPa per 100 km and $\rho = 1.2\,\mathrm{kg/m^3}$,

$$|\nabla_h p| = \frac{400\,\mathrm{Pa}}{100\,000\,\mathrm{m}} = 4 \times 10^{-3}\,\mathrm{Pa/m}, \qquad |\mathbf{a}_{\mathrm{PGF}}| = \frac{4 \times 10^{-3}}{1.2} = 3.3 \times 10^{-3}\,\mathrm{m/s^2}.$$

Gradients of this size occur around vigorous storms; around 1 hPa per 100 km, giving $8 \times 10^{-4}\,\mathrm{m/s^2}$, is more common in midlatitude weather. The acceleration looks small, but acting alone on air initially at rest it would produce $3.3 \times 10^{-3} \times 3600 = 12\,\mathrm{m\,s^{-1}}$ after one hour and $72\,\mathrm{m\,s^{-1}}$ after six hours. Winds near the ground do not grow like this, because two other forces intervene: the Coriolis force turns the moving air, and friction slows it. The gradient is computed, not measured directly: it is the difference between barometer readings reduced to a common level, divided by their distance, or it is read from the analyzed pressure field of a weather model.

1.3.4 The Coriolis force

The wind is described in a frame of reference that rotates with the Earth, because that is the frame of every anemometer and every weather map. In a rotating frame, Newton's second law holds only if apparent forces are added. One of them, the centrifugal force, is folded into the observed gravity g. The other is the *Coriolis force*, which per unit mass is

$$\mathbf{a}_{\mathrm{C}} = -2\,\boldsymbol{\Omega} \times \mathbf{u}, \tag{1.8}$$

where $\boldsymbol{\Omega}$ is the Earth's rotation vector, of magnitude $\Omega = 7.292 \times 10^{-5}$ rad/s and pointing along the axis toward the North Pole, and $\mathbf{u} = (u, v, w)$ is the three-dimensional velocity relative to

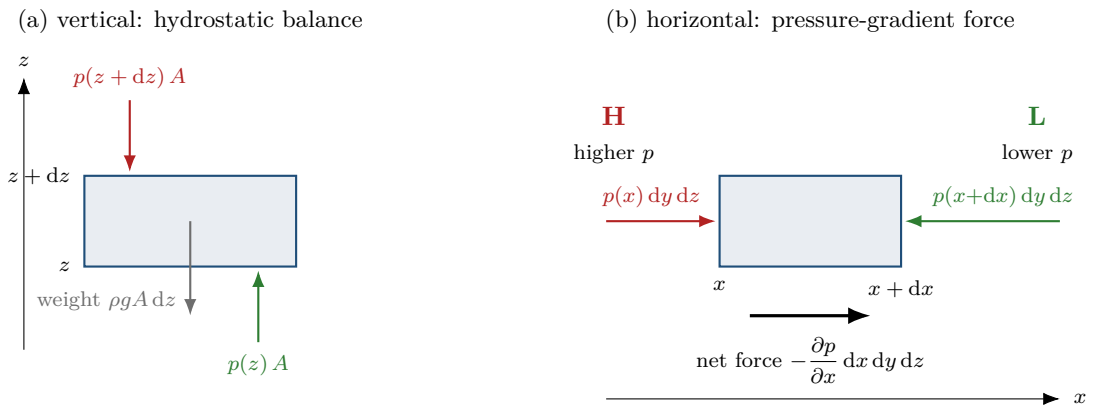


Figure 1.3: Pressure forces on a small volume of air. (a) A horizontal slab between z and $z+\mathrm{d}z$: the pressure force from below balances the pressure force from above plus the weight, which gives $\partial p/\partial z = -\rho g$ (equation (1.6)). (b) A box in a horizontal pressure gradient: the higher pressure on the left face exceeds the lower pressure on the right face, and the net force, per unit mass $-(1/\rho)\,\partial p/\partial x$, points from high to low pressure (equation (1.7)).

the Earth. At latitude ϕ, in the local east-north-up coordinates, $\boldsymbol{\Omega} = \Omega\,(0, \cos\phi, \sin\phi)$. Working out the cross product,

$$-2\,\boldsymbol{\Omega} \times \mathbf{u} = \big(2\Omega \sin\phi\, v - 2\Omega\cos\phi\, w,\ -2\Omega\sin\phi\, u,\ 2\Omega\cos\phi\, u\big).$$

For large-scale horizontal motion the term in w is negligible because $|w| \ll |u|$, and the vertical component is negligible against gravity. What remains is the horizontal Coriolis force,

$$\mathbf{a}_{\mathrm{C},h} = (fv, -fu) = -f\,\mathbf{k} \times \mathbf{V}, \qquad f = 2\Omega\sin\phi, \tag{1.9}$$

where $\mathbf{k}$ is the upward unit vector and f the *Coriolis parameter*, in s^{-1}. At 45°N, $f = 2 \times 7.292 \times 10^{-5} \times \sin 45^\circ = 1.03 \times 10^{-4}\,\mathrm{s}^{-1}$; at 23.5°S, near the Tropic of Capricorn in southern Brazil, $f = 2 \times 7.292 \times 10^{-5} \times \sin(-23.5^\circ) = -5.82 \times 10^{-5}\,\mathrm{s}^{-1}$; at the equator $f = 0$.

Three properties follow from equation (1.9). The force is perpendicular to the wind, so it changes the direction of the moving air but not its speed, and it does no work. Its magnitude is $|f|\,U$: zero for air at rest, $1.03 \times 10^{-3}\,\mathrm{m/s^2}$ for a $10\,\mathrm{m\,s^{-1}}$ wind at 45°N and $5.8 \times 10^{-4}\,\mathrm{m/s^2}$ at 23.5°S, comparable to the pressure-gradient force of Section 1.3.3. And its direction depends on the sign of f. For an eastward wind, $u > 0$ and $v = 0$, the force is $(0, -fu)$: toward the south when $f > 0$, which is to the right of the motion, and toward the north when $f < 0$, which is to the left. In the Northern Hemisphere moving air is deflected to the right of its direction of motion; in the Southern Hemisphere, to the left.

Figure 1.4 illustrates the consequence. Each panel is a map seen from above, north at the top, with straight east-west isobars and the low pressure to the north; the left panel is in the Northern Hemisphere and the right panel in the Southern. A parcel starts from rest at the dot, and the pressure-gradient force (black arrow) accelerates it northward, toward the low. As it gains speed the Coriolis force (purple arrow) pushes it sideways, to the right in the north and to the left in the south, and the path (red) turns until the air moves along the isobars rather than across them: toward the east with the low on its left in the Northern Hemisphere, toward the west with the low on its right in the Southern. Large-scale wind therefore does not blow straight from high to low pressure; it turns until it flows nearly parallel to the isobars. Chapter 3 derives that balanced state, the geostrophic wind, and shows that the exact path of a parcel released from rest oscillates about it rather than turning smoothly as drawn.

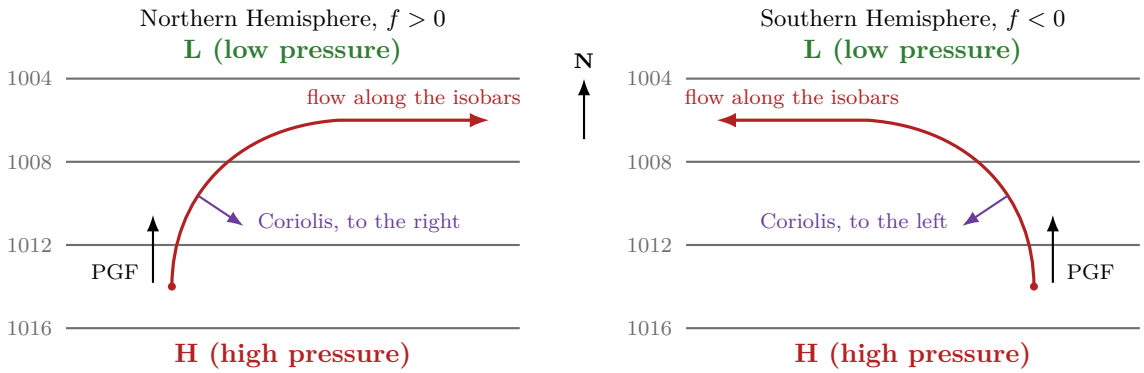


Figure 1.4: Deflection by the Coriolis force (schematic map view, north at the top, no friction). Gray: isobars, in hPa, with the low to the north. A parcel released from rest at the dot is accelerated toward the low by the pressure-gradient force (PGF) and deflected by the Coriolis force, equation (1.9), to the right in the Northern Hemisphere and to the left in the Southern, until it moves along the isobars. The smooth turn is schematic; the exact path oscillates about the final direction.

1.3.5 The horizontal momentum equation

The forces of this section combine into one equation for the horizontal wind following an air parcel,

$$\frac{\mathrm{D}\mathbf{V}}{\mathrm{D}t} = -\frac{1}{\rho}\nabla_h p \;-\; f\,\mathbf{k}\times\mathbf{V} \;+\; \mathbf{F}, \tag{1.10}$$

which reads, term by term,

acceleration following the parcel = pressure forcing + rotation + friction and turbulence.

The left-hand side, $\frac{\mathrm{D}\mathbf{V}}{\mathrm{D}t}$, is the rate of change of the wind of one parcel of air as it moves, in $\mathrm{m/s^2}$; it is not the rate of change at a fixed anemometer, which also changes when a different parcel arrives. The first term on the right is the pressure-gradient force of equation (1.7); the second the Coriolis force of equation (1.9); the third, $\mathbf{F}$, is the force per unit mass exerted by friction, which near the ground is the drag of the surface transmitted upward by turbulence. The equation omits small terms that arise from the curvature of the Earth and the vertical velocity, which do not matter at the scales of this book.

The importance of each term depends on where the air is. Table 1.1 gives illustrative orders of magnitude for a midlatitude weather system above the boundary layer, with a wind of $U = 10\,\mathrm{m\,s^{-1}}$ that changes over a distance of $L = 1000\,\mathrm{km}$, and for the air near the ground. Above the boundary layer, the pressure-gradient and Coriolis forces are both of order $10^{-3}\,\mathrm{m/s^2}$ and nearly cancel; the acceleration is about ten times smaller, of order $U^2/L = 10^{-4}\,\mathrm{m/s^2}$; friction is negligible. Their ratio $U/(|f|L) = 10^{-4}/10^{-3} = 0.1$ is the Rossby number, which measures how much the acceleration matters against rotation and which Chapters 4 and 5 develop. Near the surface the friction term becomes as large as the other two, so all three forces matter: the wind is slower, and it crosses the isobars toward low pressure instead of flowing along them.

Principle 1.1. *Air is pushed by the horizontal pressure gradient, from high toward low pressure; rotation turns the moving air, to the right in the Northern Hemisphere and to the left in the Southern; and friction near the surface slows it. Aloft the first two nearly balance; near the ground all three are of the same size.*

Table 1.1: Terms of the horizontal momentum equation, equation (1.10), with illustrative orders of magnitude. Above the boundary layer: $U = 10\,\mathrm{m\,s^{-1}}$, $L = 1000\,\mathrm{km}$, $f = 10^{-4}\,\mathrm{s^{-1}}$, a gradient of about 1 hPa per 100 km. Near the surface: the same gradient, a slower wind, and turbulent drag.

Term	Expression	Physical meaning	Above the boundary layer ($\mathrm{m/s^2}$)	Near the surface ($\mathrm{m/s^2}$)
Acceleration	$\frac{\mathrm{D}\mathbf{V}}{\mathrm{D}t}$	change of a parcel's wind; scale U^2/L	10^{-4}	10^{-4} to 10^{-3}
Pressure-gradient force	$-(1/\rho)\nabla_h p$	push from high to low pressure	10^{-3}	10^{-3}
Coriolis force	$-f\,\mathbf{k}\times\mathbf{V}$	deflection by rotation; scale $\lvert f\rvert U$	10^{-3}	a few 10^{-4}
Friction	$\mathbf{F}$	surface drag carried by turbulence	negligible	10^{-3}

1.4 The layer near the surface

The engineering heights of this book lie in the lowest part of the atmosphere, where the surface acts on the air directly. This section describes what the surface does to the flow, defines the atmospheric boundary layer and the surface layer, and places the heights of common instruments within them.

1.4.1 Drag and turbulence over a real surface

Above the boundary layer, the moving air is not in contact with any surface. Near the ground, it meets the surface and everything on it: trees, buildings, hills and mountains, crops, lakes and the sea, and patches of ground at different temperatures. Every obstacle exerts a drag on the flow, and the flow behind obstacles breaks into eddies of many sizes: turbulence. Turbulence mixes the air vertically, carrying slow air from near the ground upward and fast air from above downward, and in doing so it spreads the effect of the surface drag through a layer much deeper than the obstacles themselves.

The surface also heats and cools the air. In the afternoon a sunlit surface warms the air in contact with it, which rises in thermals and stirs the layer vigorously; at night the ground cools by radiation and chills the air above it, which then resists vertical motion and damps turbulence. The same surface drag can therefore be spread through two kilometers of air in the afternoon and through a hundred meters at night. This dependence on time of day and weather is what makes near-surface wind different from the wind aloft, and it is why the friction term $\mathbf{F}$ of equation (1.10) cannot be given a single value.

1.4.2 The atmospheric boundary layer

The *atmospheric boundary layer* (ABL), also called the planetary boundary layer (PBL), is the lowest part of the atmosphere, the part that feels the surface directly and responds to changes at the surface, such as heating, cooling and drag, within about an hour or less. Above it is the free atmosphere, where the surface is felt only indirectly. The ABL depth, written z_i and measured in meters, is not set by an equation but by a criterion, such as the height where turbulence vanishes or where a temperature inversion caps the mixing; Chapter 6 states the criteria. It is measured from profiles of temperature, humidity and wind (radiosondes, towers),

and remotely with lidars and ceilometers that detect the top of the turbid, mixed layer. In a weather model it is diagnosed by the boundary-layer parameterization.

The depth varies by more than an order of magnitude. Under strongly stable conditions, on a clear night with light wind, the ABL can be about 100 m deep or less; in vigorous daytime convection it can exceed 2 km; commonly it lies between about 500 and 2000 m. Over land it follows the daily cycle of heating, growing through the morning and collapsing near sunset; over the sea, whose temperature changes little during the day, it varies much less. A tower of 150 m is therefore deep inside the ABL in the afternoon and may reach above it at night.

1.4.3 The surface layer

The *surface layer* is the lowest part of the ABL, roughly its lowest 5 to 10 %, in which the vertical turbulent fluxes of momentum and heat are close to their surface values, varying with height by less than about ten percent. In this layer the surface controls the flow most directly, the wind speed increases rapidly with height, and the Coriolis force is unimportant against friction. Its depth follows from that of the ABL: 5 to 10 m under a 100 m deep stable layer, 50 to 100 m under a 1000 m layer, 100 to 200 m under a 2000 m convective layer. Below it, within a few obstacle heights of the ground, lies the roughness sublayer of the jargon box of Section 1.1, where the flow is shaped by the individual obstacles. Chapters 7, 9 and 12 develop the surface layer, its fluxes and its wind profile.

The engineering heights of this book fall in or near the surface layer: weather stations measure at 10 m, transmission-line sensors at about 20 to 50 m, wind-turbine hubs are at about 80 to 150 m, and meteorological towers span these heights. Whether a given height is inside the classical surface layer is not a property of the height. It depends on the conditions, and the same instrument can be inside it in the afternoon and above it at night.

1.4.4 Engineering heights in the boundary layer

Figure 1.5 places the engineering heights against the layers. The vertical axis is height above the ground on a logarithmic scale, from 1 m to 3 km. The three columns are three illustrative boundary layers: a strongly stable night with $z_i = 100$ m, a typical case with $z_i = 1000$ m (bracket: the common range of 500 to 2000 m), and vigorous daytime convection with $z_i = 2500$ m. In each column the light fill is the ABL and the darker fill the surface layer, drawn to 10 % of z_i with a dashed line at 5 %. The horizontal bands across all three columns are the instrument heights.

The figure shows the point of the previous subsection directly. On the convective afternoon, every engineering height up to the turbine hub lies in the surface layer. On the typical day the transmission-line sensors are inside it and the turbine hubs at or above its top. On the stable night only the 10 m station is at the top of the surface layer; the line sensors are above it, and the turbine rotor reaches the top of the ABL, where the wind can be decoupled from the surface and a nocturnal jet can form. One instrument can thus sample different layers at different hours, and a relation between heights fitted on daytime data does not transfer to the night.

1.5 Scales of atmospheric motion

The wind at one point is the sum of motions of very different sizes and durations. This section writes the near-surface wind as a function of the influences that act on it at once, and then sorts the motions of the atmosphere into the three scale ranges that meteorology uses.

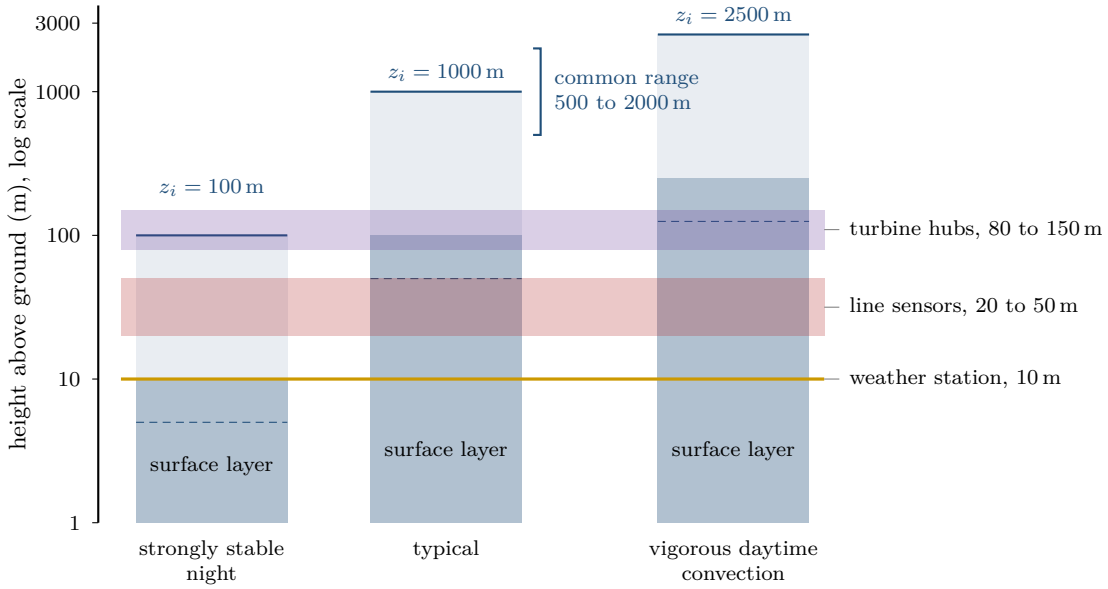


Figure 1.5: Engineering heights against three illustrative boundary layers, height on a logarithmic scale. Light fill: atmospheric boundary layer of depth z_i; darker fill: surface layer to 10 % of z_i, with a dashed line at 5 %. Bands: weather station (10 m), transmission-line sensors (20 to 50 m), wind-turbine hubs (80 to 150 m). The same height lies in different layers under different conditions.

1.5.1 The near-surface wind as a function of many influences

Consider the wind at 20 m at an illustrative site on a given afternoon. Several influences act on it at once. The large-scale weather sets the background: a synoptic system, say a high to the southeast and a low to the northwest in the Northern Hemisphere, drives a regional southwesterly wind. The boundary-layer structure decides how much of that wind reaches the ground: daytime convection mixes faster air from aloft downward. The surface roughness upstream slows the mean wind: a forest a kilometer upwind has reduced it. The terrain accelerates or decelerates the flow: a ridge on which the sensor stands speeds it up. And a local obstacle disturbs it: the wake of a building upwind adds turbulence and reduces the mean speed in certain directions.

The wind at 20 m is therefore not simply the synoptic wind, and no single one of these influences determines it. In functional form,

$$U_{20\mathrm{m}} = \mathcal{F}(\text{large-scale forcing},\ \text{stability},\ \text{roughness},\ \text{terrain},\ \text{obstacles},\ \text{height}), \qquad (1.11)$$

where $U_{20\mathrm{m}}$ is the wind speed at 20 m and $\mathcal{F}$ stands for the whole chain of transformations of Section 1.1.1, and the arguments interact: the speed-up over a ridge depends on the stability, and the slowing by a forest depends on the height and on the wind direction. Each argument belongs to a different scale of motion, which is the subject of the next subsection.

1.5.2 Synoptic, mesoscale and microscale

Meteorology sorts atmospheric motions into three ranges of horizontal size, each with a typical lifetime. The *synoptic scale*, of about 1000 km and more, contains the midlatitude cyclones and anticyclones (lows and highs) and the fronts between air masses; these systems last days, and they provide the background pressure forcing. The name comes from synoptic charts, on which observations made at the same time over a large area are plotted together. The *mesoscale*, from about 1 to 1000 km, contains sea breezes, mountain-valley circulations, thunderstorms, low-level

jets and urban circulations; these last minutes to hours. The *microscale*, below about 1 km, contains building wakes, the speed-up of the wind over hills, the turbulence inside and above forest canopies, local gusts and the flow around obstacles; these last seconds to minutes. The boundaries are conventions and differ between authors, particularly for the microscale. Table 1.2 lists the three ranges with their sizes, durations, examples and their role in near-surface wind.

Table 1.2: The three scales of atmospheric motion. Sizes and durations are approximate; the boundaries are conventions.

Scale	Horizontal size	Duration	Examples	Role for near-surface wind
Synoptic	about 1000 km and more	days	midlatitude cyclones, highs, fronts	background pressure forcing; the regional wind direction
Mesoscale	about 1 to 1000 km	minutes to hours	sea breezes, mountain-valley circulations, thunderstorms, low-level jets, urban circulations	regional modification and local forcing; the daily cycle
Microscale	below about 1 km	seconds to minutes	building wakes, hill speed-up, canopy turbulence, gusts, obstacle flow	the wind at one sensor or one span; gusts and turbulence

Figure 1.6 draws the same ranges as regions in a plane of size and duration. The horizontal axis is the horizontal size of a motion and the vertical axis its lifetime, both on logarithmic scales; the three shaded rectangles are the scale ranges, and the dots are examples placed at their typical size and lifetime. Larger motions live longer, so the examples lie along a band from the lower left to the upper right. Engineering wind needs all three ranges at once: the synoptic scale gives the forcing for days ahead, the mesoscale the regional and daily modification, and the microscale the wind that actually reaches a conductor, a rotor or an anemometer. Chapter 5 develops the scale ranges and the dimensionless numbers that decide which forces matter in each.

1.6 Model and measurement, forcing and response

This section separates two pairs of objects that the rest of the book keeps apart: the wind a weather model reports and the wind an instrument measures, and the large-scale forcing and the response of the near-surface air to it. It closes with the downward chain that organizes the book.

1.6.1 Grid-cell wind and point wind

A weather model divides the atmosphere into grid cells and computes one value of each variable per cell and level. Its 10 m wind, written U_{10}, is a *grid-scale wind*: the wind the model assigns to a whole cell, at 10 m above the cell's effective terrain elevation, over the cell's effective roughness, with turbulence and surface fluxes supplied by parameterizations. A model value of $U_{10} = 6.0\,\mathrm{m\,s^{-1}}$ is therefore not the wind measured by an anemometer 10 m above the ground at a particular spot inside that cell. With a grid spacing of $\Delta x = 3\,\mathrm{km}$ the cell covers $3\,\mathrm{km} \times 3\,\mathrm{km} = 9\,\mathrm{km}^2 = 9 \times 10^6\,\mathrm{m}^2$; with $\Delta x = 9\,\mathrm{km}$ it covers $81\,\mathrm{km}^2$. Features smaller than a

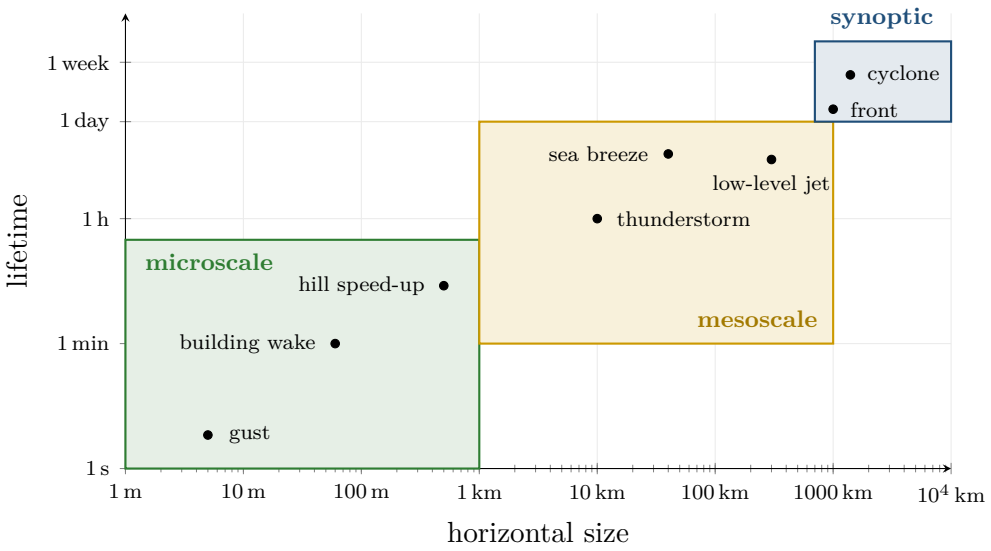


Figure 1.6: The three scales of atmospheric motion in a plane of horizontal size and lifetime, both logarithmic (schematic). Green: microscale, below about 1 km, seconds to minutes; gold: mesoscale, about 1 to 1000 km, minutes to hours; blue: synoptic, about 1000 km and more, days. Dots: illustrative examples at their typical size and lifetime.

few grid spacings, such as a single ridge or a patch of forest, are not represented at all, only their average effect on the cell.

A *point observation* is the value an instrument measures at one place: the anemometer samples the air passing through a volume of a few liters. Averaging over time widens what it sees: over a ten-minute average at $6\,\mathrm{m\,s^{-1}}$, about $6 \times 600 = 3600\,\mathrm{m}$ of air passes the instrument, but that air is a line upstream of the sensor along the wind direction, not the $9\,\mathrm{km^2}$ of the cell, and it carries the imprint of the obstacles and terrain right next to the sensor. The difference between the two is a difference of objects, not only an error of the model: one is a grid-scale atmospheric state, the other a point measurement. The part of their disagreement that comes from this difference of objects is called the *representativeness error*; it remains even when the model is perfect at its own scale and the instrument is perfect at its point.

Figure 1.7 draws the difference in a vertical cross section through three cells of a model with $\Delta x = 3\,\mathrm{km}$. The horizontal axis is distance in kilometers and the vertical axis elevation in meters, with a vertical exaggeration of about ten. The black line is the real terrain, with a ridge, a valley and a forest; the blue steps are the model terrain, one mean elevation per cell. The anemometer stands on a knoll in the middle cell, at 296 m, while the model ground of that cell is at its mean elevation of 272 m. The model's 10 m wind refers to 282 m above sea level over a flat cell with averaged roughness; the anemometer's 10 m is 306 m above sea level on a crest, where the terrain accelerates the flow. The two values describe different air.

1.6.2 Forcing and response

A *forcing* is a physical driver imposed on the flow from outside the layer of interest; the *response* is what the air does with it. For near-surface wind, the large-scale horizontal pressure gradient is the forcing, and terrain, turbulence, surface drag, heating and stability determine the response. The same forcing can produce very different winds near the ground. In an illustrative case, a synoptic pressure gradient sustains a wind of $15\,\mathrm{m\,s^{-1}}$ above the boundary layer. At night, with strong surface drag and a stable boundary layer that suppresses vertical mixing, the wind at 10 m is $5\,\mathrm{m\,s^{-1}}$, one third of the wind aloft. On a sunny afternoon with the same forcing, strong convective mixing brings fast air down from aloft, and the wind at 10 m is $9\,\mathrm{m\,s^{-1}}$, sixty

END OF THE PREVIEW

Near-Surface Wind Meteorology

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Ricardo Marquez, Ph.D.

ricardo-marquez.com