

Chapter 1

What is the mesoscale?

1.1 Space and time scales

Meteorological phenomena occur over a wide range of space and time scales. Phenomena having short time scales also tend to have small spatial scales, and vice versa (Fig. 1.1). Curiously, the ratio of (horizontal) space to time scales, which has units of velocity (m s^{-1}), is roughly the same order of magnitude for all features ($\sim 10 \text{ m s}^{-1}$). If a phenomenon's width is much greater than its depth, it may be inferred that the phenomenon is approximately hydrostatic (Fig. 1.2). This is a consequence of the first law of thermodynamics, mass continuity, and the equations of motion. We will revisit this issue at the end of this chapter.

Before defining what is meant by “mesoscale,” it may be easiest first to define what is meant by the “synoptic” scale. The adjective “synoptic” is defined in the American Meteorological Society’s *Glossary of Meteorology* as referring to meteorological data that are obtained simultaneously over a wide area in order to present a nearly instantaneous snap-shot of the state of the atmosphere. The early synoptic charts displayed the limited amount of data that could be collected routinely at the same times on a daily basis, and disturbances that could be resolved on these charts eventually were referred to as “synoptic-scale” disturbances. Thus, the term “synoptic,” though not initially intended to define a scale, ultimately became a term used to describe the scale of large-scale weather systems, which were the only types of meteorological phenomena that could be resolved regularly by the coarse resolution observing platforms of the middle 19th century.

The term “mesoscale” is believed to have been introduced by Ligda (1951) in an article reviewing the use of weather radar, in order to describe phenomena smaller than the synoptic scale but larger than the “microscale,” a term that was widely used at the time (and still is) in reference to phenomena having a scale of a few kilometers or less. Thus, the upper limit of the mesoscale can be regarded as being roughly near the limit of resolvability of a disturbance by an observing network approximately as dense as that present when the first synoptic charts became available, i.e., on the order of 1000 km. Today the mesoscale “officially” (per the American Meteorological Society’s *Glossary of Meteorology*) is defined as the 2–2000 km scale, with sub-classifications of meso- α , meso- β , and meso- γ scales referring to horizontal scales of 200–2000 km, 20–200 km, and 2–20 km, respectively. In this classification scheme, the microscale is reserved for horizontal scales smaller than 2 km.

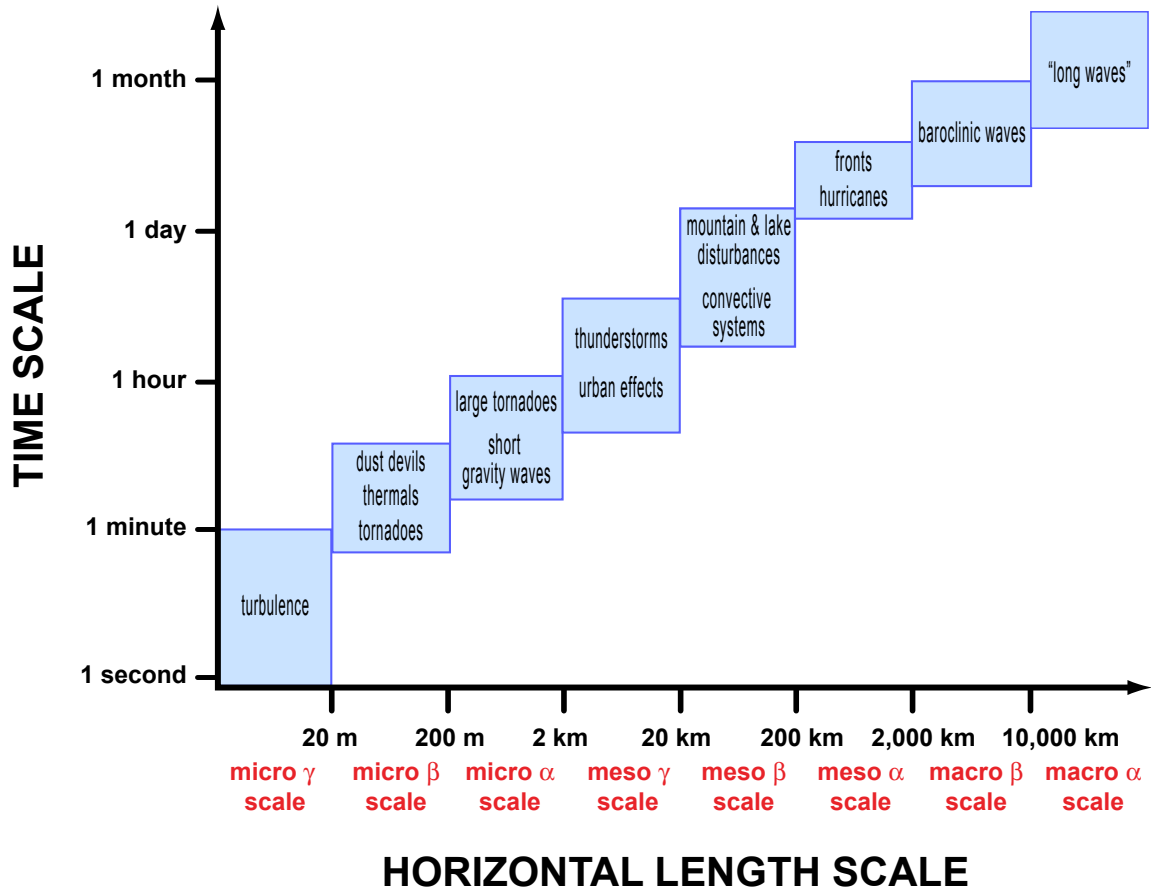


Figure 1.1. Scale definitions and the characteristic time and horizontal length scales of a variety of atmospheric processes. [Adapted from Orlandi (1975).]

The specification of the upper and lower limits of the mesoscale does have some dynamical basis, although perhaps only coincidentally. It can be argued that the synoptic scale is driven largely by a single, dominant instability—baroclinic instability. Baroclinic instability is most likely to be realized by disturbances having a horizontal wavelength on the order of a few 1000 kilometers. On the other hand, on the mesoscale there does not seem to be a similar dominant instability that drives motions. Apparently a variety of instabilities, many of which will be examined in this textbook, can operate on the mesoscale, and the dominant instability on a given day depends on the local state of the atmosphere on that day (which may be heavily influenced by synoptic-scale motions). Thus, one could argue that the mesoscale should refer to scales that generally are too small for baroclinic instability to be dominant.

Perhaps another way of defining the mesoscale is by the terms we will need to retain in the dynamical equations that govern atmospheric phenomena. For example, on the synoptic scale, several terms in the governing equations safely can be disregarded due to their relative unimportance on that scale (e.g., vertical accelerations of air, advection by the ageostrophic part of the wind). Likewise, on the microscale, different terms in the governing equations often can be neglected (e.g., the Coriolis force, even the horizontal pressure gradient force

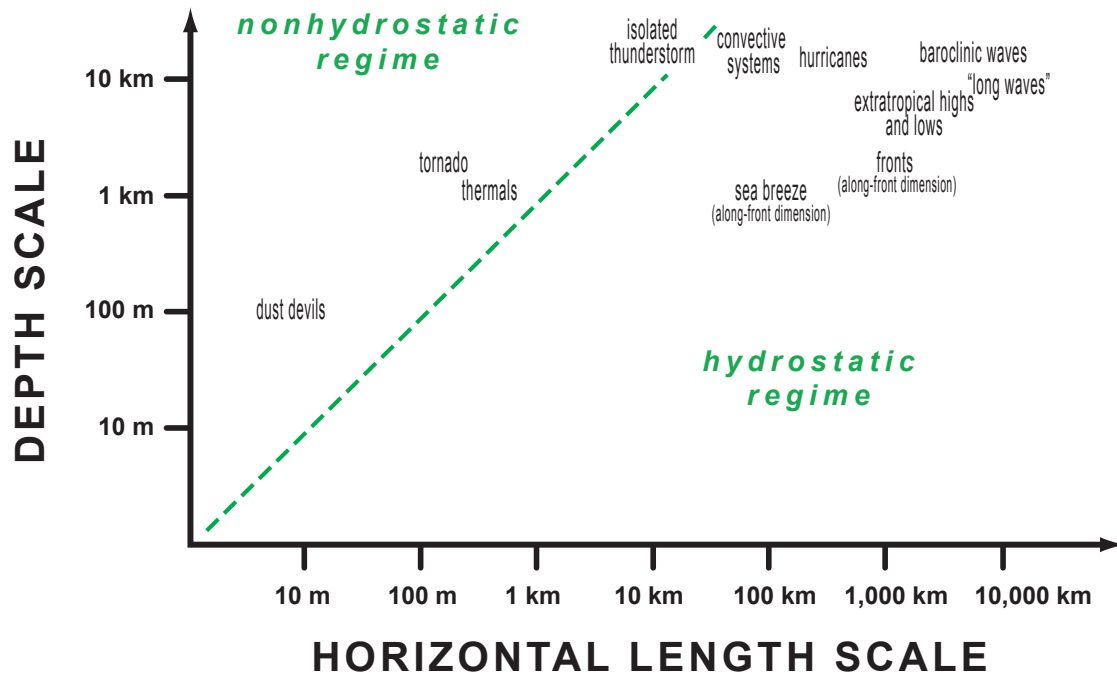


Figure 1.2. It can be inferred that when the horizontal length scale of a phenomena is significantly larger than its vertical depth scale, then the phenomenon is hydrostatic. Shown above are some examples of nonhydrostatic and approximately hydrostatic phenomena plotted as a function of depth versus horizontal length (i.e., width) scale.

on occasion). But on the mesoscale—in between the synoptic scale and microscale, by definition—we frequently cannot disregard *any* of the terms in the dynamical equations. For example, in a long-lived mesoscale convective system, there are large pressure gradients and horizontal and vertical accelerations of air, tremendous latent heat release and absorption and associated regions of positive and negative buoyancy, and important influences exerted by radiative effects. Yet even the Coriolis force also has been shown to influence the structure and evolution of these systems.

1.2 Dynamical distinctions between the mesoscale and synoptic scale

1.2.1 Geostrophic balance

Geostrophic balance is a poor approximation to the air flow on the mesoscale. Thus, another good definition of mesoscale is the scale on which ageostrophic motions are important. On the synoptic scale, phenomena tend to be characterized by a near balance of the Coriolis and pressure gradient forces, so accelerations of air parcels and ageostrophic motions tend to be very small. On the mesoscale, pressure gradients can be considerably larger than on the

synoptic scale, whereas the Coriolis acceleration (proportional to wind velocity) is of similar magnitude on the synoptic scale. Thus, mesoscale systems often are characterized by large wind accelerations and large ageostrophic motions.

On scales below ~ 1000 km, the Coriolis acceleration becomes fairly unimportant compared to the pressure gradient force, and on scales much larger than ~ 1000 km, ageostrophic motions become decreasingly significant. A nondimensional number, called the Rossby number, quantitatively accounts for the relative importance of the Coriolis force and ageostrophic motions (more precisely, the relative importance of the Coriolis force and actual air parcel accelerations). The Rossby number can be used to distinguish synoptic-scale weather systems from subsynoptic-scale phenomena. The Rossby number, Ro , is defined as

$$Ro = \frac{O(d\mathbf{v}/dt)}{O(-f\mathbf{k} \times \mathbf{v})} \sim \frac{V/T}{fV} \sim \frac{V^2/L}{fV} \sim \frac{V}{fL} \quad (1.1)$$

where f is the Coriolis parameter, d/dt is the Lagrangian derivative, V is the magnitude of a characteristic vector wind, \mathbf{v} , and L is a characteristic length scale. On the synoptic scale, where the quasi-geostrophic approximation usually can be made, $Ro \ll 1$. For mesoscale systems, $Ro \gtrsim 1$.

It may be worthwhile to do a scale analysis of the horizontal momentum equation (the x equation, without loss of generality):

$$\frac{du}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + fv + F_u, \quad (1.2)$$

where fv is the Coriolis acceleration in the x direction to a good approximation, F_u represents viscous effects on u , and all other terms have their conventional meanings. We will neglect F_u for now, but we will find later that effects associated with the F_u term often *are* important.

On the synoptic and mesoscale, for $O(v) \sim 10 \text{ m s}^{-1}$, the Coriolis acceleration is of order

$$O(fv) \sim (10^{-4} \text{ s}^{-1}) (10 \text{ m s}^{-1}) \sim 10^{-3} \text{ m s}^{-2}.$$

On the synoptic scale, the pressure gradient force has a scale of

$$O\left(-\frac{1}{\rho} \frac{\partial p}{\partial x}\right) \sim \frac{1}{1 \text{ kg m}^{-3}} \frac{10 \text{ mb}}{1000 \text{ km}} \sim 10^{-3} \text{ m s}^{-2};$$

thus, the Coriolis and pressure gradient forces approximately balance one another and it can be inferred that accelerations (du/dt) are small (“quasigeostrophic balance”). Furthermore, because $v = v_g + v_a$ and $v_g = \frac{1}{\rho f} \frac{\partial p}{\partial x}$, it is easily shown that $fv_a = -\frac{1}{\rho} \frac{\partial p}{\partial x} + fv$. Therefore, ageostrophic motions also are small on the synoptic scale, owing to the approximate balance between the Coriolis and pressure gradient forces.

On the mesoscale, the horizontal pressure gradient, $\partial p/\partial x$, may range from 5 mb/500 km (e.g., in quiescent conditions) to 5 mb/5 km (e.g., beneath a thunderstorm). At the small end of this range, the Coriolis and pressure gradient forces may be approximately in balance, but at the large end of this range, the pressure gradient force is two orders of magnitude larger than on the synoptic scale (i.e., 10^{-1} m s^{-2} versus 10^{-3} m s^{-2}). On these occasions, the pressure gradient force dominates, the Coriolis force is relatively unimportant, and accelerations and ageostrophic motions are large.

1.2.2 Hydrostatic balance

In many atmospheric applications (e.g., synoptic meteorology, large-scale dynamics), it is assumed that the atmosphere is in hydrostatic balance. In other words, we assume that the vertical pressure gradient and gravitational acceleration are nearly balanced, resulting in negligible vertical accelerations. The vertical momentum equation can be written as

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial z} + 2\Omega u \cos \phi - g + F_w, \quad (1.3)$$

where $2\Omega u \cos \phi$ is the component of the Coriolis force in the z direction, F_w represents viscous forces acting on w , and all other terms have their conventional meanings. The scale of g is 10 m s^{-2} and the scale of $|\frac{1}{\rho} \frac{\partial p}{\partial z}|$ also is $\sim 100 \text{ mb}/1000 \text{ m} \sim 10^4 \text{ Pa}/10^3 \text{ m} \sim 10 \text{ m s}^{-2}$. We'll neglect F_w in this simple analysis, although F_w can be important, particularly near the edges of clouds and in rising thermals. Also, it is easily shown that the vertical component of the Coriolis force is negligible compared to the size of the gravitational acceleration and the vertical pressure gradient, with a scale of 10^{-3} m s^{-2} . Thus, we'll neglect this term as well, which gives

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial z} - g. \quad (1.4)$$

It is obvious that if dw/dt is negligible, then (1.4) becomes the so-called hydrostatic approximation,

$$\frac{\partial p}{\partial z} = -\rho g. \quad (1.5)$$

Here's the million dollar question: in which types of phenomena can we assume (1.5), i.e., that dw/dt is negligible compared to $|\frac{1}{\rho} \frac{\partial p}{\partial z}|$ and g ? In other words, what determines whether a phenomenon can be regarded as a "hydrostatic phenomenon" versus a "nonhydrostatic phenomenon?"

It turns out we cannot simply scale the terms in (1.4) to determine which are significant because the two on the right-hand side are nearly equal in magnitude but opposite in sign and, thus, we must compare dw/dt to their residual. We do this by defining a base state (e.g., an average over a large horizontal area) density and a base state pressure defined to be in hydrostatic balance with it. We then express the total pressure and density as

$$p = \bar{p}(z) + p'(x, y, z, t) \quad (1.6)$$

$$\rho = \bar{\rho}(z) + \rho'(x, y, z, t), \quad (1.7)$$

which we require to satisfy

$$0 = -\frac{\partial \bar{p}}{\partial z} - \bar{\rho} g. \quad (1.8)$$

Multiplying (1.4) by ρ , subtracting (1.8), and dividing by ρ yields

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p'}{\partial z} - \frac{\rho' g}{\rho}. \quad (1.9)$$

The relative importance of dw/dt compared to $|\frac{1}{\rho} \frac{\partial p'}{\partial z}|$ (and $\rho'g/\rho$) is

$$\frac{O\left(\frac{dw}{dt}\right)}{O\left(-\frac{1}{\rho} \frac{\partial p'}{\partial z}\right)}. \quad (1.10)$$

The scale of w can be obtained from the continuity equation (2D Boussinesq approximation used for simplicity),

$$\frac{\partial w}{\partial z} \approx -\frac{\partial u}{\partial x}; \quad (1.11)$$

thus,

$$O(w) \sim \frac{UH}{L}, \quad (1.12)$$

where $O(w)$ is the scale of w , and U , H , and L are the characteristic horizontal velocity scale, depth scale, and length scale of the phenomenon. From (1.12), the scale of dw/dt is therefore

$$O\left(\frac{dw}{dt}\right) \sim \frac{UH}{LT}, \quad (1.13)$$

where T is the characteristic time scale of the phenomenon.

The scale of $|\frac{1}{\rho} \frac{\partial p'}{\partial z}|$ may be written as

$$O\left(-\frac{1}{\rho} \frac{\partial p'}{\partial z}\right) \sim \frac{\delta p'}{\rho H}, \quad (1.14)$$

where $\delta p'$ is the characteristic pressure difference across the phenomenon. We want to eliminate $\delta p'$ and ρ in favor of scales that we used to approximate the dw/dt scale (e.g., U , H , L , and T). We do this by using

$$\frac{du}{dt} \approx -\frac{1}{\rho} \frac{\partial p'}{\partial x} \quad (1.15)$$

$$O\left(\frac{du}{dt}\right) \sim O\left(-\frac{1}{\rho} \frac{\partial p'}{\partial x}\right) \quad (1.16)$$

$$\frac{U}{T} \sim \frac{\delta p'}{\rho L} \quad (1.17)$$

$$\frac{UL}{T} \sim \frac{\delta p'}{\rho}; \quad (1.18)$$

therefore, using (1.14),

$$O\left(-\frac{1}{\rho} \frac{\partial p'}{\partial z}\right) \sim \frac{UL}{TH}. \quad (1.19)$$

Using (1.13) and (1.19), (1.10) becomes

$$\frac{O\left(\frac{dw}{dt}\right)}{O\left(-\frac{1}{\rho} \frac{\partial p'}{\partial z}\right)} \sim \frac{\frac{UH}{LT}}{\frac{UL}{TH}} = \left(\frac{H}{L}\right)^2. \quad (1.20)$$

The quantity H/L is known as the aspect ratio of the phenomenon—the ratio of the characteristic depth scale of the phenomenon to the horizontal length scale (or width) of the phenomenon. When a phenomenon is much wider than it is deep ($H/L \ll 1$), dw/dt is relatively small compared to the vertical pressure gradient (and gravity) and the phenomenon can be considered a “hydrostatic phenomenon;” i.e., the hydrostatic approximation can be justifiably invoked. When a phenomenon is approximately as wide as it is deep ($H/L \sim 1$), dw/dt is relatively large compared to the vertical pressure gradient (and gravity) and the phenomenon is considered a “nonhydrostatic phenomenon;” i.e., the hydrostatic approximation should not be invoked. See Fig. 1.2.

On the synoptic scale, $H/L \sim 10 \text{ km}/1000 \text{ km} \sim 1/100 \ll 1$. On the mesoscale, H/L can be ~ 1 or $\ll 1$, depending on the phenomenon. For example, in a thunderstorm updraft, $H/L \sim 10 \text{ km}/10 \text{ km} \sim 1$ (i.e., the thunderstorm updraft can be considered to be a nonhydrostatic phenomenon). But for the rain-cooled outflow that the thunderstorm produces, $H/L \sim 1 \text{ km}/10 \text{ km} \sim 1/10 \ll 1$ (i.e., the outflow can be considered to be an approximately hydrostatic phenomenon).

In a hydrostatic atmosphere, pressure can be viewed essentially as being proportional to the weight of the atmosphere above a given point. Pressure changes in a hydrostatic atmosphere arise from changes in the density of air vertically integrated over a column extending from the location in question to $z = \infty$ ($p = 0$). For a nonhydrostatic phenomenon, we cannot relate pressure fluctuations solely to changes in the weight of the overlying atmosphere. Instead, significant “dynamic” effects may contribute to pressure changes. Examples are the low pressure found in the core of a tornado and above the wing of an airplane in flight, and the high pressure found beneath an intense downburst and on the upwind side of an obstacle. The relationship between the pressure field and wind field is discussed in much greater depth later in section 2.5.

In the next chapter we review some of the basic equations and tools that will be relied upon in the rest of the book. The experienced reader may wish to skip ahead to chapter 3.

Further reading

Emanuel, K. A., 1983: The dynamical definitions of “mesoscale.” *Mesoscale Meteorology: Theories, Observations, and Models*, D. K. Lilly and T. Gal-Chen, Eds., Reidel, 1-11.

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