Mesoscale Dynamics

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[Lin, Y.-L., 2007: *Mesoscale Dynamics*. Cambridge University Press.]

Chapter 1 Overview

1.1 Introduction

Mesoscale dynamics uses a dynamical approach for the study of atmospheric phenomena with a horizontal scale ranging approximately from 2 to 2000 km. These mesoscale phenomena include, but not limited to, thunderstorms, squall lines, supercells, mesoscale convective complexes, inertia-gravity waves, mountain waves, low-level jets, density currents, land/sea breezes, heat island circulations, clear air turbulence, jet streaks, and fronts. Mesoscale dynamics may be viewed as a combined discipline of dynamic meteorology and mesoscale meteorology. From the dynamical perspective, mesoscale concerns processes with timescales ranging from the buoyancy oscillation ($2\pi / N$, where N is the buoyancy (Brunt-Vaisala) frequency) to a pendulum day ($2\pi / f$, where f is the Coriolis parameter), encompassing deep moist convection and the full spectrum of inertia-gravity waves but stopping short of synoptic-scale phenomena which have Rossby numbers less than 1. The Rossby number is defined as U / fL, where U is the basic wind speed and L the horizontal scale of the disturbance associated with the phenomenon. The study concerns with the analysis and prediction of large-scale weather phenomena, based on the use of meteorological data obtained simultaneously over the standard observational network is called *synoptic meteorology*. Synoptic scale phenomena include, but not limited to, extratropical and tropical cyclones, fronts, jet streams, and baroclinic waves. The synoptic scale is also referred to as the large scale, macroscale, or cyclonic scale in the literature and in this textbook. Traditionally, these scales have been loosely used or defined. For example, tropical cyclones have been classified as synoptic scale phenomena by some meteorologists but are classified as mesoscale phenomena by others (Table 1.1). The same is the case with the mesoscale. For example, a tornado has been classified as a mesoscale phenomenon by some meteorologists due to the scale of its environment for formation, while it is generally classified as a microscale phenomenon based on its scale of circulation. In the mean time, fronts have been classified as both large-scale and mesoscale phenomenon due to their different scales in the along-front and cross-front directions.

Before about 1980, due to the lack of observational data at the mesoscale, mesoscale meteorology had advanced less rapidly compared to synoptic meteorology. For example, some observed isolated, unusual values of pressure and winds shown on synoptic charts were suspected to be observational errors. Even though this may be true in some cases, others are now thought to represent true signatures of subsynoptic disturbances having spatial and temporal scales too small to be properly analyzed and represented on standard synoptic charts. Since the advancement of observational techniques and an overall increase in the number of mesoscale observational networks after about 1980 and rapid advancement in numerical modeling techniques, more and more mesoscale phenomena, as well as their interactions with synoptic scale and microscale flows and weather systems, have been revealed and better understood. In order to improve mesoscale weather forecasting, it is essential to improve our understanding of the basic dynamics of mesoscale atmospheric phenomena through fundamental studies by utilizing observational, modeling, theoretical, and experimental approaches simultaneously. Since mesoscale spans a wide horizontal range approximately from 2 to 2000 km, there is no single theory (such as the quasi-geostrophic theory for the large scale), which provides a unique tool for studying the dynamical structure of the variety of mesoscale motions observed in the Earth's atmosphere. In fact, the dominant dynamical processes vary dramatically from system to system, depending on the type of mesoscale circulation involved.

1.2 Definitions of atmospheric scales

Due to different force balances, atmospheric motions in fluid systems with distinct temporal and spatial scales behave differently. In order to better understand the complex dynamical and physical processes associated with mesoscale phenomena, different approximations have been adopted to help resolve the problems. Therefore, a proper scaling facilitates the choice of appropriate approximations of the governing equations.

Scaling of atmospheric motions is normally based on observational and theoretical approaches. In the observational approach, atmospheric processes are categorized through direct empirical observations and the instruments used. Since observational data are recorded in discrete time intervals and the record of these data in the form of a standard surface or upper air weather map reveals a discrete set of phenomena, the phenomena are then also categorized into discrete scales. For example, sea breezes occur on a time scale of about 1 day and spatial scales of 10 to 100 km, while cumulus convection occurs on a time scale of about 30 min and encompasses a spatial scale of several kilometers. Figure 1.1 shows the atmospheric kinetic energy spectrum in the free atmosphere and near the ground for various time scales. In the free atmosphere, there are strong peaks at periods ranging from a few days (the synoptic scale) to a few weeks (the planetary scale - at which the β effect plays an important role). In addition, there are also peaks at 1 year and 1 day and a smaller peak at a few minutes, although this latter peak may be an artifact of the analysis. This energy spectrum therefore suggests a natural division of atmospheric phenomena into three distinct (but not wholly separable) scales: large-scale, mesoscale, and microscale. From the kinetic energy spectrum, the mesoscale therefore appears as the scale on which energy is allowed to transfer from the large-scale to the microscale and vice versa. Based on radar observations of storms, atmospheric motions can be categorized into the following three scales: (a) microscale: L < 20 km, (b) mesoscale: 20 km < L < 1000 km, and (c) synoptic (large) scale: L > 1000 km (Ligda 1951). The atmospheric motions have also been categorized into 8 separate scales: macro-α (L > 10,000 km), macro-β (10,000 km > L > 2000 km), meso-α (2000 km > L > 200 km), meso-β (200 km > L > 20 km), meso-γ (20 km > L > 2 km), micro-α (2 km > L > 200 m), micro-β (200 m > L > 20 m), and micro-γ (L < 20 m) scales (Orlanski 1975; Table 1.1). Based on observations, atmospheric phenomena have also been categorized into masocale, mesoscale, misoscale, mososcale, and musoscale (Fujita 1986).

Atmospheric motions may also be categorized using a theoretical approach. For example, for airflow over a mountain, the scale of the mechanically induced quasi-steady waves corresponds roughly to the scale of the imposed forcing. For such problems, adoption of the *Eulerian* (fixed in space) *time scale* is reasonable. For example, for two steady cumulus clouds being advected by a steady basic wind, the time scale for a stationary observer located on the ground is approximately the horizontal scale of the mountain divided by the basic wind speed. However, the above time scale has little to do with the physical processes associated with the cloud development. Instead, it is more meaningful physically to use the *Lagrangian Rossby number* R_o, which is defined as the ratio of intrinsic frequency and the Coriolis parameter ($\omega / f = 2\pi / fT$, where T is the *Lagrangian time scale*), because the Lagrangian time scale measures the time a fluid particle takes in following the motion. In the above example, the Lagrangian time scale is the time it takes an air parcel to rise to its maximum vertical displacement. Another example is the Lagrangian time scale for a cyclone, which is defined as $2\pi R / V_T$, where R is the radius of the circular motion and V_T is the tangential wind speed. The Lagrangian time scales and Rossby numbers for typical atmospheric systems are summarized below:

Phenomenon	**Time scale**	**Lagrangian R_o**

		$(\approx \omega / f = 2\pi / fT)$
Tropical cyclone	$2\pi R / V_T$	V_T / fR
Inertia-gravity waves	$2\pi / N$ to $2\pi / f$	N / f to 1
Sea/land breezes	$2\pi / f$	1
Thunderstorms and cumulus clouds	$2\pi / N_w$	N_w / f
Kelvin-Helmholtz waves	$2\pi / N$	N / f
PBL turbulence	$2\pi h / U*$	$U* / fh$
Tornadoes	$2\pi R / V_T$	V_T / fR

where:

R = radius of maximum wind scale

ω = frequency

T = time scale

V_T = maximum tangential wind scale

f = Coriolis parameter

N = buoyancy (Brunt-Vaisala) frequency

N_w = moist buoyancy (Brunt-Vaisala) frequency

$U*$ = scale for friction velocity

h = scale for the depth of planetary boundary layer.

Based on this type of theoretical considerations, the following different scales for atmospheric motions can be defined: (a) synoptic (large or macro) scale, for motions which are quasi-geostrophic and hydrostatic, (b) mesoscale, for motions which are nonquasi-geostrophic and hydrostatic, and (c) microscale, for motions which are non-geostrophic, nonhydrostatic, and turbulent (Emanuel and Raymond 1984). Based on this interpretation, the mesoscale may be defined as that scale which includes atmospheric circulations that are large enough in horizontal scale to be considered hydrostatic but too small to be described as quasi-geostrophically. Note that hydrostatic assumption may not apply to some mesoscale weather systems, especially for those associated with convection.

Based on hydrostatic, convective, advective, compressible, and Boussinesq approximations of the governing equations - including temporal, horizontal and vertical spatial scales - in order to standardize existing nomenclature with regard to mesoscale phenomena - a more rigorous approach can be taken to define the atmospheric scales (e.g., Thunis and Bornstein 1996). This approach integrates existing concepts of atmospheric spatial scales, flow assumptions, governing equations, and resulting motions into a hierarchy which is useful in the classification of mesoscale motions. Horizontal and vertical scales of flow subclasses under unstable and stable stability conditions for deep and shallow convection are shown in Figs. 1.2 and 1.3, respectively. Table 1.1 summarizes some examples of the horizontal and temporal scales for typical atmospheric phenomena as proposed by different authors. In this book, we will adopt Orlanski's scaling, except where otherwise specified.

1.3 Energy generation and scale interactions

Although many mesoscale circulations and weather systems are forced by large scale or microscale flow, some circulations are locally forced at the mesoscale itself. Energy generation mechanisms for mesoscale circulations and weather systems may be classified into the following categories: (a) thermal or orographic surface inhomogeneities, (b) internal adjustment of larger-scale flow systems, (c) mesoscale instabilities, (d) energy transfer from either the large scale or microscale to the mesoscale, and (e) interaction of cloud physical and dynamical processes (Anthes 1986).

Examples of the first type of mesoscale weather systems are the land and sea breezes, mountain-valley winds, mountain waves, heat-island circulations, coastal fronts, dry lines, and moist convection. These mesoscale weather systems are more predictable than other types of systems that occur on the mesoscale. Examples of the second type of weather systems are fronts, cyclones, and jet streaks. These weather systems are less predictable since they are generated by transient forcing associated with larger-scale flows. Although instabilities associated with the mean wind or thermal structure of the atmosphere are rich energy sources of atmospheric disturbances, most atmospheric instabilities have their maximum growth rates either on the large scale through baroclinic, barotropic, and inertial instabilities or on the microscale through Kelvin-Helmholtz, conditional and potential (convective) instabilities. Symmetric instability appears to be intrinsically a mesoscale instability.

Energy transfer from small scales to the mesoscale also serves as a primary energy source for mesoscale convective systems. These mesoscale convective systems may start as individual convective cells that grow and combine to form thunderstorms and convective systems, such as squall lines, mesocyclones, and mesoscale convective complexes. **O**n the other hand, energy transfer from the large scale to the mesoscale also serves as an energy source to induce mesoscale circulations or weather systems. For example, temperature and vorticity advection associated with large-scale flow systems may help the genesis of mesoscale frontal systems through scale contraction. Another possible energy source for producing mesoscale circulations or weather systems is the interaction of clouds' physical and dynamical processes. Mesoscale convective systems may be generated by this interaction process through scale expansion.

Scale interaction generally refers to the interactions between the temporally and spatially averaged zonal flow and a fairly limited set of waves that are quantized by the circumference of the earth, while it refers to multiple interactions among a continuous spectrum of eddies of all sizes in turbulence theory (Emanuel 1986). However, scale interaction should not be viewed as a limited set of interactions among discrete scales because, on average, the mesoscale is much more like a continuous spectrum of scales. Scale interaction depends on the degree of relative strength of fluid motions involved. For example, for a very weak disturbance embedded in a slowly varying mean flow, the interaction is mainly exerted from the mean flow to the weak disturbance. If this disturbance becomes stronger, then it may exert an increasing influence on the mean flow, and other scales of motion may develop. In this case, scale interactions become more and more numerous, and the general degree of disorder in the flow becomes greater. At the extreme, when the disturbance becomes highly nonlinear, such as in a fully developed turbulent flow, then the interactions become mutual and chaotic, and an explicit mathematical or analytic description of the interaction becomes problematic. Examples of scale-interactive processes which occur at mesoscale include: (i) synoptic forcing of mesoscale weather phenomena, (ii) generation of internal mesoscale instabilities, (iii) interactions of cloud and precipitation processes with mesoscale systems, (iv) influence of orography, boundary layer, and surface properties on mesoscale weather system development and evolution, (v) feedback contributions of mesoscale systems to larger-scale processes, (vi) energy transfer associated with mesoscale systems, and (vii) mechanisms and processes associated with stratosphere-troposphere exchange (Koch 1997). Figure 1.4 shows the mutual interactions between a jet streak, inertia-gravity waves, and strong mesoscale convection that can occur on the mesoscale.

Figure 1.5 shows the energy transfer process through geostrophic adjustment in the response of the free atmosphere to a cumulus cloud, which radiates gravity waves that lead to a lens of less stratified air whose width is the Rossby radius of deformation. The process from state (a) to (b) in Fig. 1.5 represents a scale interactive process in which the system tends to reach geostrophic equilibrium. The above example of cumulus convection implies as least two distinct scales: (i) the cumulus scale ~ L_z, and (ii) the large scale ~ NL_z/f (Rossby radius of deformation in a stratified fluid). The Rossby radius of deformation is the horizontal scale at which rotational effects become as important as buoyancy effects. The Rossby radius of deformation can be understood as the significant horizontal scale fluid parcels experience when a fluid undergoes geostrophic adjustment in a homogeneous fluid, such as water, to an initial condition such as

$$\eta = \eta_o sgn(x) , \qquad (1.3.1)$$

where η is the vertical displacement of the fluid from the mean fluid depth, and η_o is the maximum η, x points eastward, and sgn is the sign function defined as $sgn(x) = 1$ for $x > 0$ and -1 for $x < 0$. In the earlier stage, the motion is dominated by the pressure gradient force, and the fluid particles move toward west ($-x$ direction). As time proceeds, the Coriolis force becomes more and more important and the fluid particles are deflected toward right (north) in the Northern Hemisphere. The Coriolis force eventually reaches a geostrophic balance with the pressure gradient force. In this final stage, the basic flow is northward (+y direction), and the Rossby radius of deformation is considered to be the horizontal distance from the location where the e-folding value of the vertical displacement is equal to the original average height of the homogeneous fluid.

In a shallow water fluid system, the Rossby radius of deformation is

$$\lambda_R = c_o / f \,, \tag{1.3.2}$$

where c_o is the shallow water wave speed ($\sqrt{gH}$, H is the fluid depth) induced by the gravitational (buoyancy) force.

1.4 Predictability

In mesoscale numerical weather prediction, the question of predictability concerns the degree to which a hydrodynamical model of the atmosphere will yield diverging solutions when integrated in time using slightly different initial conditions (e.g., Ehrendorfer and Errico 1995). Weather phenomena are considered to have limited predictability since there is an uncertainty associated with initial conditions determined from real observations. The question of predictability of mesoscale atmospheric phenomena was first investigated by using a simple model for the interaction of barotropic vorticity perturbations encompassing a number of diverse horizontal scales

(Lorenz 1969). Those results suggested that the mesoscale may be less predictable, i.e. yielding perturbed solutions that diverge faster, than the synoptic and planetary scales, essentially because the eddy timescale decreases on the horizontal scale. The predictability for synoptic scales is mainly limited by the nonlinear interactions between different waves with different wavelengths, i.e. different components of the wave spectrum. These interactions depend on the initial distribution of energy in the different wavenumbers and on the number of waves the model can resolve. Errors and uncertainties in the resolvable-scale waves and errors introduced by neglecting unresolvable scales grow with time and spread throughout the spectrum, eventually contaminating all wavelengths and destroying the forecast (Anthes 1986). The predictability for mesoscale motions is mainly limited by the rapid transfer of energy between the large scale and the microscale. In addition, the predictability for small scales is mainly limited by three-dimensional turbulence. Inevitable errors or uncertainties in initial conditions in the small scale of motion will propagate toward larger scales and will reach the mesoscale sooner than the large scale, therefore rendering the mesoscale less predictable.

The response of a fluid system to a steady forcing tends to fall into one of the following four categories: (1) steady for a stable system, perfectly predictable, (2) periodic for a weakly unstable system, perfectly predictable, (3) aperiodic with a "lumpy" spectrum for a moderately unstable system, less predictable, and (4) aperiodic with a monotonic spectrum for a fully turbulent system, rather unpredictable (Emanuel and Raymond 1984). The atmospheric system falls into category 3. Monotonicity of the kinetic energy spectrum (Fig. 1.1) through the mesoscale implies that energy is mainly transferred from larger (large scale) to smaller (microscale) scales, although it can be generated intermittently at the mesoscale. This tends to limit the predictability at the mesoscale.

Beside the natural constraints imposed by forcing and physical processes, predictability of mesoscale phenomena is also affected by the initial conditions set up in a mesoscale numerical weather prediction model. If a mesoscale phenomenon does not exist at the beginning of the numerical simulation, then the predictability is less influenced by the accuracy of the initial conditions used in a mesoscale numerical weather prediction model. Under this situation, the mesoscale circulations are normally forced by surface inhomogeneities (thermal or orographic), internal adjustment of larger-scale flow systems, mesoscale instabilities, energy transfer from either the larger scale or the microscale, or the interaction of cloud physical and dynamical processes, as discussed earlier. For mesoscale circulations induced by larger scale motion, the time scale for predictability of these types of mesoscale systems could exceed the actual time scale of the mesoscale systems themselves. On the other hand, if a mesoscale phenomenon exists at the beginning of the numerical prediction, then it is necessary to include the observed and analyzed motion and thermodynamic variables in the initial conditions in order to make an accurate numerical prediction. The accuracy of the numerical prediction relies more on observations in the beginning and less on the model because it takes time for the model to spin up. For example, if a numerical model starts its integration at a time where a tropical cyclone already exists, then the vortex and moisture fields associated with the tropical cyclone must be initialized in the model. Thus, observations and their implementation in the numerical model are very important in the beginning of the numerical prediction. Naturally, contributions of the model to numerical weather prediction become more and more important as time proceeds.

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Table 1.1 Atmospheric scale definitions. (Adapted after Thunis and Borstein 1996)

Horizontal Scale	Lifetime	Stull (1988)	Pielke (2002)	Orlanski (1975)	Thunis and Bornstein (1996)	Atmospheric Phenomena
10 000km	1 month	Macro	Synoptic Regional	Macro-α	Macro-α	General circulation, long waves
2000 km	1 week	Macro	Synoptic Regional	Macro-β	Macro-β	Synoptic cyclones
200 km	1 day	Macro	Synoptic Regional	Meso-α	Macro-γ	Fronts, hurricanes, tropical storms, short cyclone waves, mesoscale convective complexes
20 km		Meso	Meso	Meso-β	Meso-β	Mesocyclones, mesohighs, supercells, squall lines, inertia-gravity waves, cloud clusters, low-level jets thunderstorm groups, mountain waves, sea breezes
2 km	1 h	Meso / Micro	Micro	Meso-γ	Meso-γ	Thunderstorms, cumulonimbi, clear-air turbulence, heat island, macrobursts
200 m	30 min	Meso / Micro	Micro	Micro-α	Meso-δ	Cumulus, tornadoes, microbursts, hydraulic jumps
20 m	1 min	Micro	Micro	Micro-β	Micro-β	Plumes, wakes, waterspouts, dust devils
2 m	1 s	Micro / Micro δ	Micro	Micro-γ	Micro-γ	Turbulence, sound waves
		Micro δ	Micro	Micro-γ	Micro-δ	Turbulence, sound waves

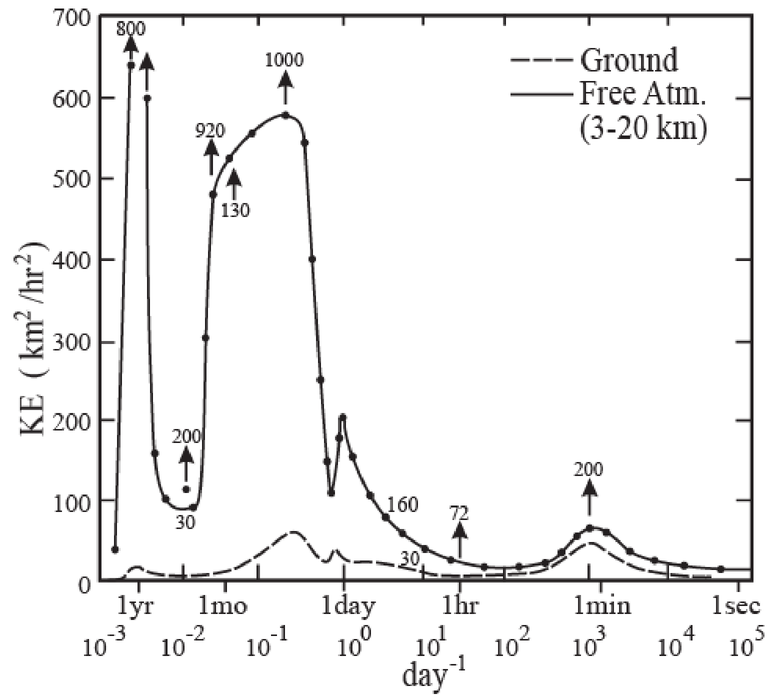

Fig. 1.1: Average kinetic energy of west-east wind component in the free atmosphere and near the ground. (Adapted after Vinnichenko 1970, reproduced with permission from Blackwell Publishing.)

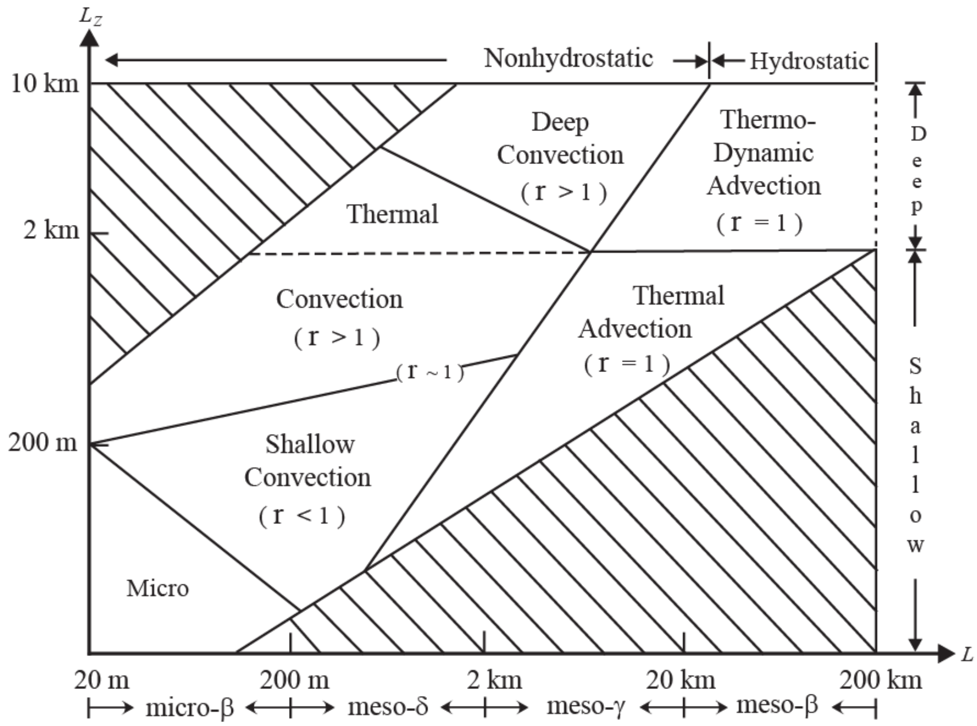

Fig. 1.2: Schematic of flow classification under unstable stability conditions, where hatched zones indicate nonphysical flow regimes, the dotted line indicates merging of thermodynamic advection with macroscale, *r* represents scaled ratio of buoyancy and vertical pressure gradient forced perturbations, and the dashed line represents division of thermal convection into its deep and shallow regimes. (Adapted after Thunis and Bornstein 1996)

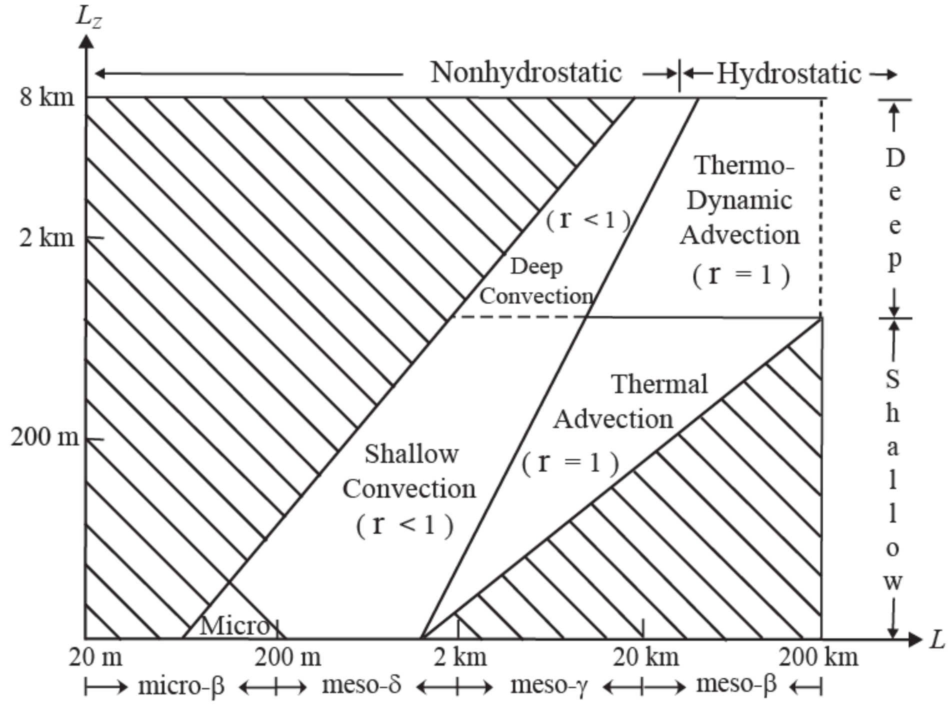

Fig. 1.3: As in Fig. 1.2 except for stable stability conditions. (Adapted after Thunis and Bornstein 1996)

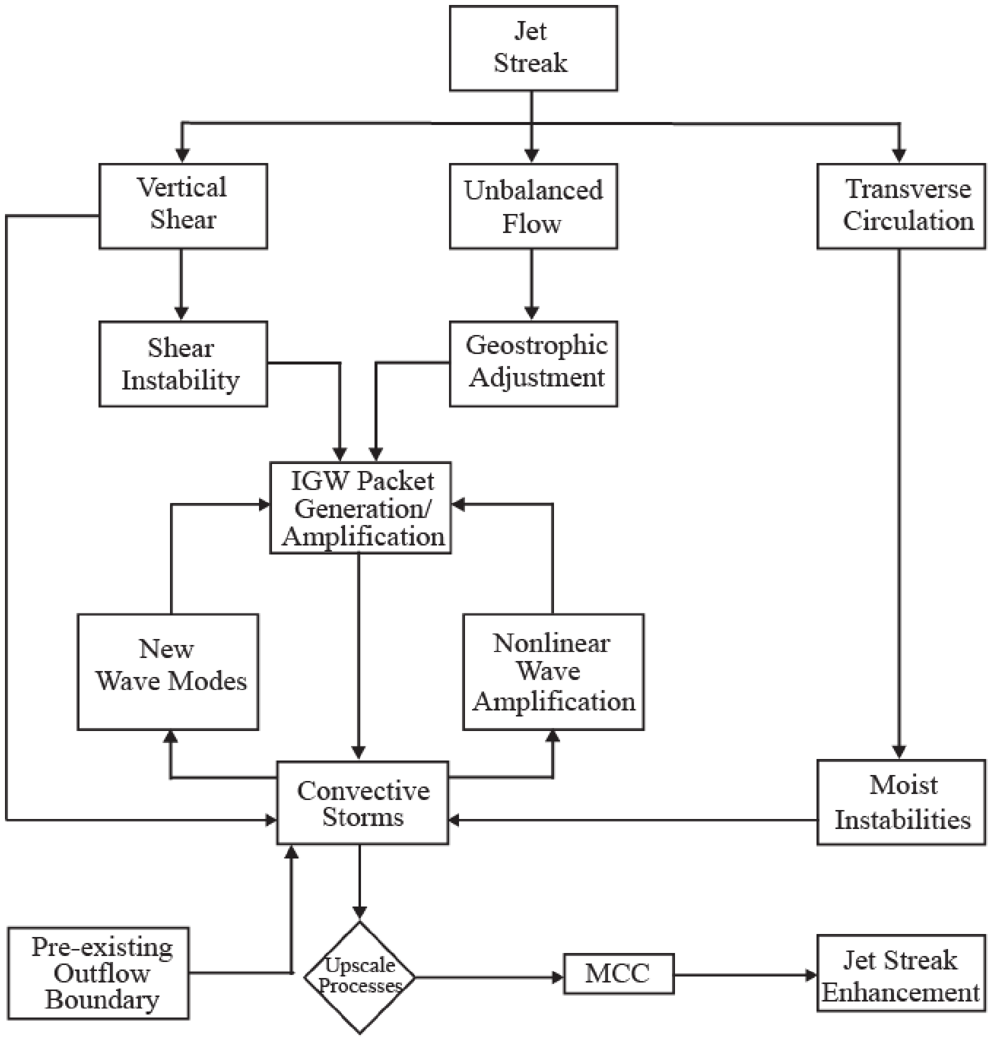

Fig. 1.4: Sketch of mutual interactions between the jet streak, inertial-gravity waves (IGW), and moist convection. (Adapted after Koch 1997)

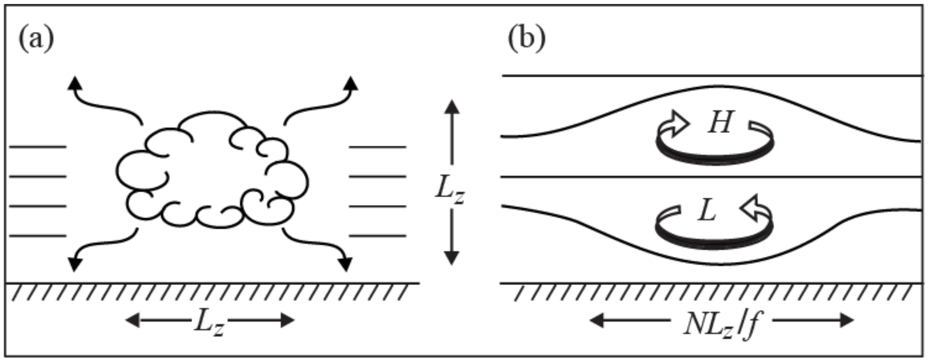

Fig. 1.5: Energy transfer process through geostrophic adjustment in the response of the free atmosphere to a cumulus cloud, (which radiates gravity waves that lead to a lens of less stratified air whose width is the Rossby radius of deformation – repetition). (a) The response of the free atmosphere to a cumulus cloud is the formation of gravity waves away from the cloud, which, in turn, lead to the formation of (b) a lens of less stratified air whose width is the Rossby radius of deformation (NL_z/f). (Adapted after Emanuel and Raymond 1984)