دانشگاه رازی

Fundamentals of Atmospheric Physics

Lecture 1

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References

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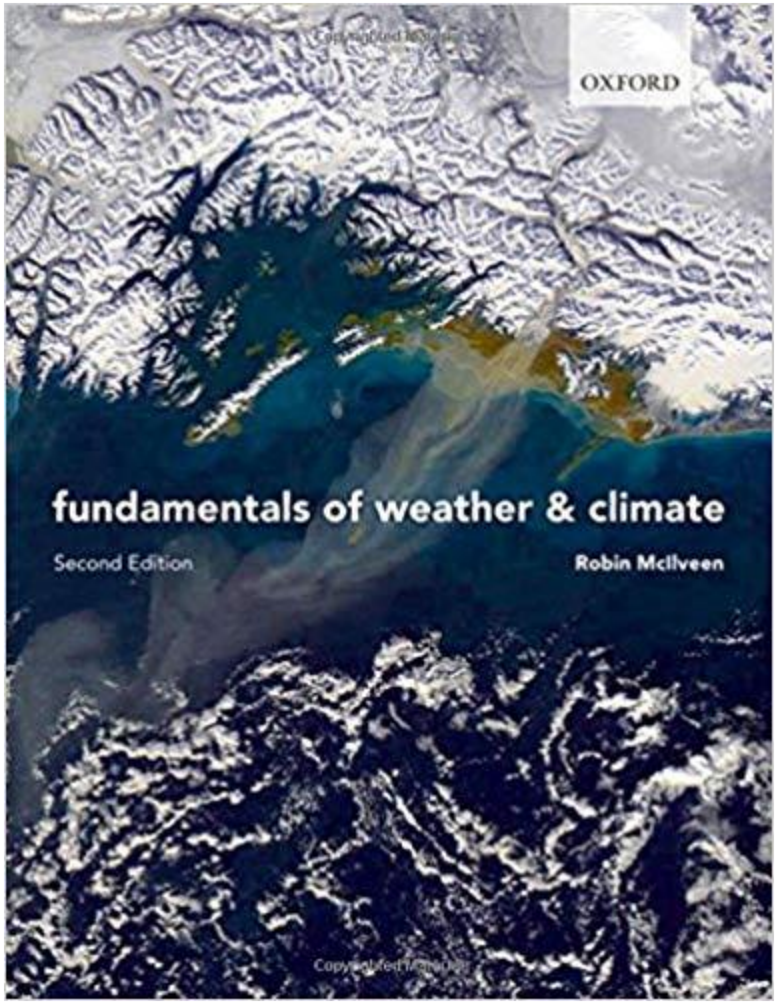

8- هواشناسی فیزیکی : بی. جی. رتالاک، مترجم : علیرضا صادقی‌حسینی

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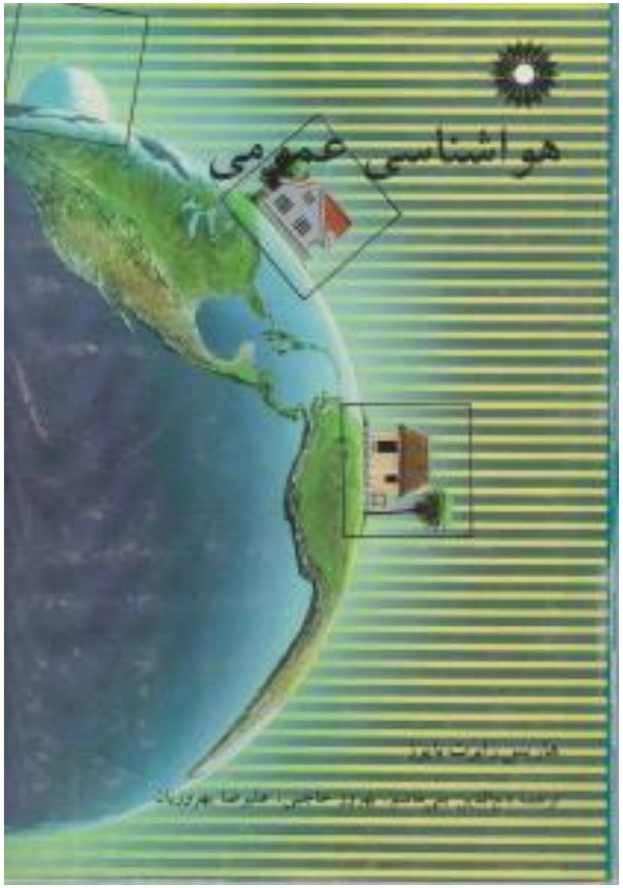


Grading

Homework- 20%

Providing Lecture - 20%

Final Exam - 60%

Chapter 1 - Introduction

Definitions

The atmosphere as a physical system

Atmospheric models

Two simple atmospheric models

Some atmospheric observations

Weather and climate Further reading

Some atmospheric observations include :

The mean temperature and wind fields, Gravity waves, Rossby waves, Ozone, Weather and climate

فصل اول- مقدمه

طبقات زمین، آب و هوا و اقلیم، ترکیب جو، تهی شدن ازن (ozone depletion) ، تحولات جو زمین، ارتفاع و ساختار جو.

فصل دوم- مشاهدات و اندازه گیریهای جوی

اهمیت اندازه گیری، اندازه گیریهای: دما، رطوبت، باد، فشار و بارش؛ سایر مشاهدات سطحی، شبکه های مشاهداتی شامل: شبکهٔ سطحی، شبکهٔ بالا و شبکهٔ ماهواره ای.

فصل سوم- تغییرات زمانی پارامترهای جوی در سطح زمین

تغییرات شبانه روزی، تغییرات فصلی، تغییرات اقلیمی.

فصل چهارم- مبانی تابش

مفاهیم، تعاریف و واحدهای تابش، تابش جسم سیاه، خطوط طیفی جذبی و گسیلی، جنبه های پایه ای انتقال تابش.

فصل پنجم- تابش خورشیدی و زمینی

ساختار خورشید، طیف خورشیدی، گردش زمین به دور خورشید، ثابت خورشیدی، توزیع جغرافیایی آفتابگیری در بام جو، پراکندگی، جذب و بازتابش تابش خورشیدی، تابش زمینی، اثر گلخانه ای، بودجهٔ گرمایی، توزیع مداری شارهای تابشی، توزیع فصلی شارهای تابشی، تغییرات شبانه روزی شارهای تابشی، ابزارهای اندازه گیری تابش.

فصل ششم- ترمودینامیک هوای خشک

قوانین گازها، معادلهٔ حالت برای هوای خشک، قانون اول ترمودینامیک، فرایندهای ویژه، دمای پتانسیلی، معادلهٔ پواسن، آنتروپی، فرایند بی در روی خشک، ترازمندی هیدروستاتیکی و پایداری ایستایی.

فصل هفتم- بخار آب و آثار ترمودینامیکی آن

حالتهای مختلف آب، گرمای نهان، معادلهٔ کلازیوس-کلاپیرون (Clausius-Clapeyron) ، معادلهٔ حالت برای بخار آب، متغیرهای رطوبت، ترمودینامیک هوای مرطوب غیراشباع، دمای مجازی، محاسبهٔ دمای مجازی، روشهای رسیدن به اشباع (دمای نقطه شبنم، دمای تر، دمای هم ارز)، فرایند بی در روی اشباع شدهٔ برگشت پذیر، فرایند بی در رو وار برگشت ناپذیر هوای اشباع شده، محتوای آب مایع بی در رو، اثر رطوبت

فصل هشتم - نمودارهای ترمودینامیکی

کاربرد مؤلفه های فشار در راستای قائم، ویژگیهای مطلوب نمودارهای ترمودینامیکی، نمودار دما- آنتروپی، نمودار انرژی- جرم، نمودار اریبی T-LnP ، تفسیر نمودار جو شناختی ترسیمی، ارزیابی مقادیر گزارش نشده، دمای تر، دمای هم ارز، پایستاری ویژگیهای توده هوا، دمای پتانسیل تر وار، دمای پتانسیل هم ارز وار.

فصل نهم - ابر و بارش

ابر، اشباع، لایهٔ زیر ابر، لایهٔ زیر ابر خوب آمیخته، نمایهٔ نقطهٔ شبنم، مشاهدهٔ میعان، رشد قطرکهای ابر، توسعهٔ ابر، رده بندی ابرها، بارش، بارورسازی مصنوعی ابرها، الکتریسیتهٔ جوی، پدیده های نور شناختی.

Atmospheric physics – A definition

Atmosphere is the gaseous envelope of a celestial body that is confined due to gravitational attraction.

Physics of the Atmosphere - ie the study of all physical phenomena in the atmospheric system.

Meteorology

Atmospheric chemistry is a branch of atmospheric science in which the chemistry of the Earth's atmosphere and that of other planets is studied.

Atmospheric science is a relatively new, applied discipline that is concerned with the structure and evolution of the planetary atmospheres and with the wide range of phenomena that occur within them.

The role of physics

Thermodynamics
- phase transitions - condensation and evaporation
- adiabatic processes & T- gradients

Quantum Mechanics
- interaction of radiation & matter

Hydrodynamics: the branch of science concerned with forces acting on or exerted by fluids
- Navier-Stokes-Eq

Transport Phenomena
- turbulence
- diffusion
- matter & energy

DESCRIPTION

Composition and structure of the earth's atmosphere,

Atmospheric thermodynamics,

Atmospheric Radiation

Atmospheric dynamics,

Clouds and precipitation.

The atmosphere as a physical system

The Earth's atmosphere is a natural laboratory, in which a wide variety of physical processes takes place.

The purpose of this book is to show how basic physical principles can help us model, interpret and predict some of these processes.

This section presents a brief overview of the physics involved.

The atmosphere consists of a mixture of ideal gases: although molecular nitrogen and molecular oxygen predominate by volume, the minor constituents carbon dioxide, water vapour and ozone play crucial roles. The forcing of the atmosphere is primarily from the sun, though interactions with the land and the ocean are also important.

The atmosphere is continually bombarded by solar photons at infra-red, visible and ultra-violet wavelengths.

Some solar photons are scattered back to space by atmospheric gases or reflected back to space by clouds or the Earth's surface; some are absorbed by atmospheric molecules (especially water vapour and ozone) or clouds, leading to heating of parts of the atmosphere; and some reach the Earth's surface and heat it.

Atmospheric gases (especially carbon dioxide, water vapour and ozone), clouds and the Earth's surface also emit and absorb infra-red photons, leading to further heat transfer between one region and another, or loss of heat to space.

Solar photons may also be energetic enough to disrupt molecular chemical bonds, leading to photochemical reactions;

The atmosphere is generally close to hydrostatic balance in the vertical, except on small scales; that is, the weight of each horizontal slab of atmosphere is supported by the difference in pressure between its lower and upper surfaces.

An alternative statement of this physical fact is that there is a balance between vertical pressure gradients and the gravitational force per unit volume acting on each portion of the atmosphere.

On combining the equation describing hydrostatic balance with the ideal gas law we find that, in a hypothetical isothermal atmosphere, the pressure and density would fall exponentially with altitude.

In the real, non-isothermal, atmosphere the pressure and density variations are usually still close to this exponential form, with an *e-folding* height of about 7 or 8 km.

Gravity thus tends to produce a density stratification in the atmosphere

Thermodynamic principles are essential for describing many atmospheric processes.

For example, any consideration of the effects of atmospheric heating or cooling will make use of the First Law of Thermodynamics.

The concept of entropy (or the closely related quantity, potential temperature) frequently assists interpretation of atmospheric behaviour.

Atmospheric models

Unlike laboratory physicists, atmospheric researchers cannot perform controlled experiments on the large-scale atmosphere.

The standard 'scientific method', of observing phenomena, formulating hypotheses, testing them by experiment, then formulating revised hypotheses and so on, cannot be applied directly.

Instead, after an atmospheric phenomenon is discovered, perhaps by sifting through a great deal of data, we develop models, which incorporate representations of those processes that we hypothesise are most important for causing the phenomenon.

Two simple atmospheric models

It is a basic observational fact that the Earth's mean surface temperature is about 288 K

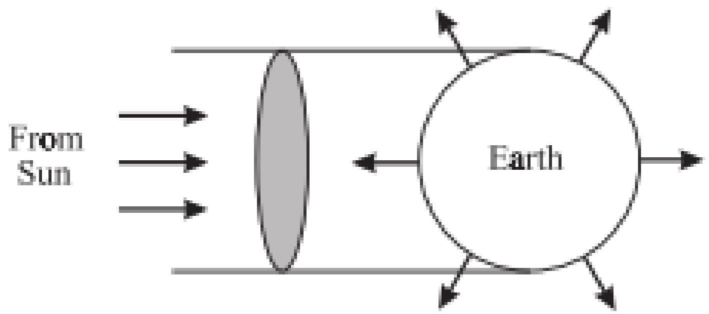
A model with a non-absorbing atmosphere

The solar power per unit area at the Earth's mean distance from the Sun (the total solar irradiance, TSI, formerly called the solar constant) is:

$$F_s = 1370 \text{ Wm}^{-2}$$

in a tube of cross-sectional area πa^2

The total solar energy received per unit time is $F_s \pi a^2$



We assume that the Earth-atmosphere system has a planetary albedo *A* equal to 0.3; that is, 30% of the incoming solar radiation is reflected back to space without being absorbed:

$$0.3 F_s \pi a^2$$

If the Earth is assumed to emit as a black body at a uniform absolute temperature T then, by the Stefan-Boltzmann law, Power emitted per unit area = σT^4

However, power is emitted in all directions from a total surface area

the total power emitted is: $4\pi a^2 \sigma T^4$

$$(1-A)F_s\pi a^2 = 4\pi a^2 \sigma T^4 \quad \rightarrow \quad T \cong 255^\circ K$$

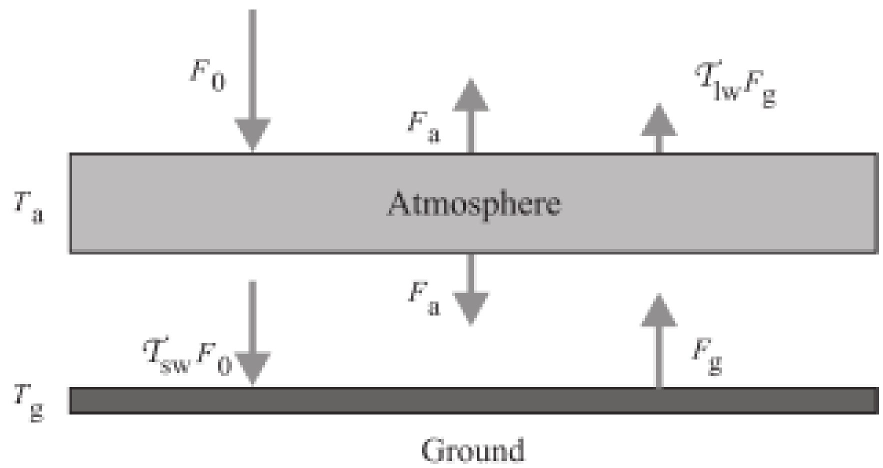
The temperature obtained from this calculation is called the effective emitting temperature of the Earth:

$$T_e \equiv \left(\frac{F_0}{\sigma}\right)^{1/4} \approx 255K$$

Its value is significantly lower than the observed mean surface temperature of about 288 K

A simple model of the greenhouse effect

We now consider the effect of adding a layer of atmosphere, of uniform temperature T_a



The atmosphere is assumed to transmit:

a fraction of any incident solar (short-wave) radiation τ_{sw}

fraction of any incident thermal (infra-red, or long-wave) radiation τ_{lw}

These fractions are called transmittances

We assume that the ground is at temperature T_g

Taking account of albedo effects and the difference between the area of the emitting surface $4\pi a^2$

and the intercepted cross-sectional area of the solar beam πa^2

The mean unreflected incoming solar irradiance at the top of the atmosphere is:

$$(1-A)F_s \pi a^2 = 4\pi a^2 \sigma T^4 \qquad F_0 = \frac{1}{4}(1-A)F_s \approx 240\ Wm^{-2}$$

F_0

$\tau_{sw} F_0$ is absorbed by the ground

$(1-\tau_{sw})F_0$ is absorbed by the atmosphere.

The ground is assumed to emit as a black body

an upward irradiance $F_g = \sigma T_g^4$

$\tau_{lw} F_g$ reaches the top

The atmosphere is not a black body, but emits irradiances

$$F_a = (1-\tau_{lw})\sigma T_a^4$$

Kirchhoff's law: the emittance – the ratio of the actual emitted irradiance to the irradiance that would be emitted by a black body at the same temperature – equals the absorptance

$$\text{emittance} = \frac{\textit{the actual emitted irradiane}}{\textit{the irradiance that would be emitted by a black body}} = 1 - \tau_{lw}$$

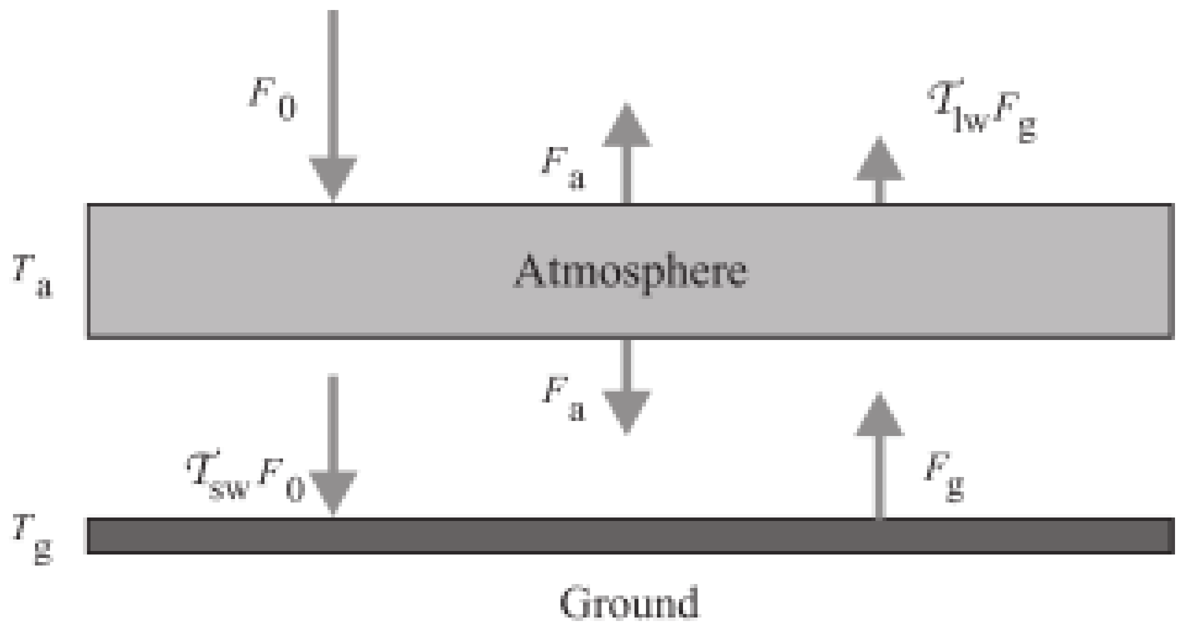
We now assume that the system is in **radiative equilibrium: that is, energy transfer** takes place only by the radiative processes described above, and the associated irradiances are in balance everywhere;

we neglect any energy transfers due to non-radiative processes such as fluid motions. Equating irradiances, we have

$$F_0 = F_a + \tau_{lw} F_g$$ above the atmosphere, and

$$F_0 = F_a + \tau_{lw} F_g \quad *$$

above the atmosphere



$$F_g = F_a + \tau_{sw} F_0 \quad *'$$

between the atmosphere and the ground

By eliminating F_a from equations we obtain

$$F_g = \sigma T_g^4 = F_0 \frac{1+\tau_{sw}}{1+\tau_{lw}}$$

In the absence of an absorbing atmosphere, we would have

$$\tau_{sw} = \tau_{lw} = 1 \qquad F_g = F_0 \qquad T_g \approx 255K$$

Taking rough values for the Earth's atmosphere to be

$\tau_{sw} = 0.9$ strong transmittance and weak absorption of solar radiation

$\tau_{lw} = 0.2$ weak transmittance and strong absorption of thermal radiation

$$T_g \approx 286K$$

This close agreement is somewhat fortuitous, however, since in reality non-radiative processes also contribute significantly to the energy balance.

We can also find the atmospheric emission from * equations:

$$F_a = (1-\tau_{lw})\sigma T_a^4 = F_0 \frac{1-\tau_{sw}\tau_{lw}}{1+\tau_{lw}}$$

and this gives the temperature of the model atmosphere,

$$T_a \approx 245K$$

One way to quantify the 'greenhouse effect' of an absorbing gas is in terms of the amount $Fg\text{-}F_0$ by which it reduces the outgoing irradiance from its surface value: in the case discussed above this reduction is 140 Wm^{-2}

This equals the difference between the amount

$$(1-\tau_{lw})\sigma T_g^4 = 304\ Wm^{-2}$$

of the thermal emission from the 'warm' surface that is absorbed by the 'cool' atmosphere

and the smaller amount

$$F_a = (1-\tau_{lw})\sigma T_a^4 = 164\ Wm^{-2}$$

that the atmosphere re-emits upwards.

Since the atmosphere is in equilibrium, it also equals the difference between the downward emission F_a from the atmosphere and the small proportion

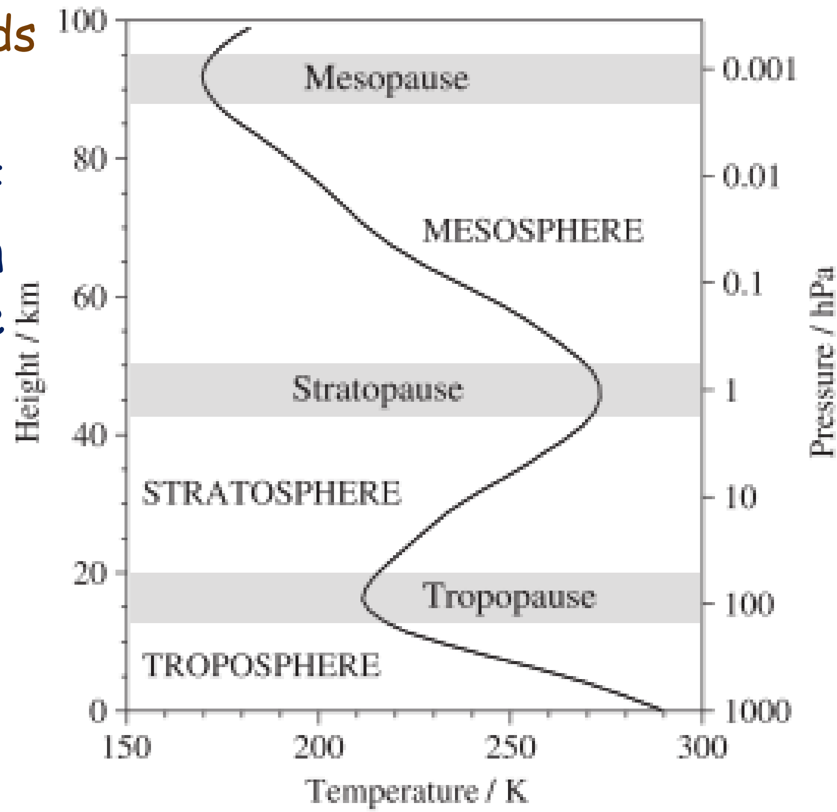
$$(1-\tau_{sw})F_0 = 24\ Wm^{-2}$$

of the solar irradiance that it absorbs.

Some atmospheric observations

In this section we present a selection of examples of basic atmospheric observations and give some indication of their physical explanation.

The mean temperature and wind fields



Typical vertical structure of atmospheric temperature (K) in the lowest 100 km of the atmosphere.
Based on data from Fleming *et al.* (1990).

Layer One (Troposphere)

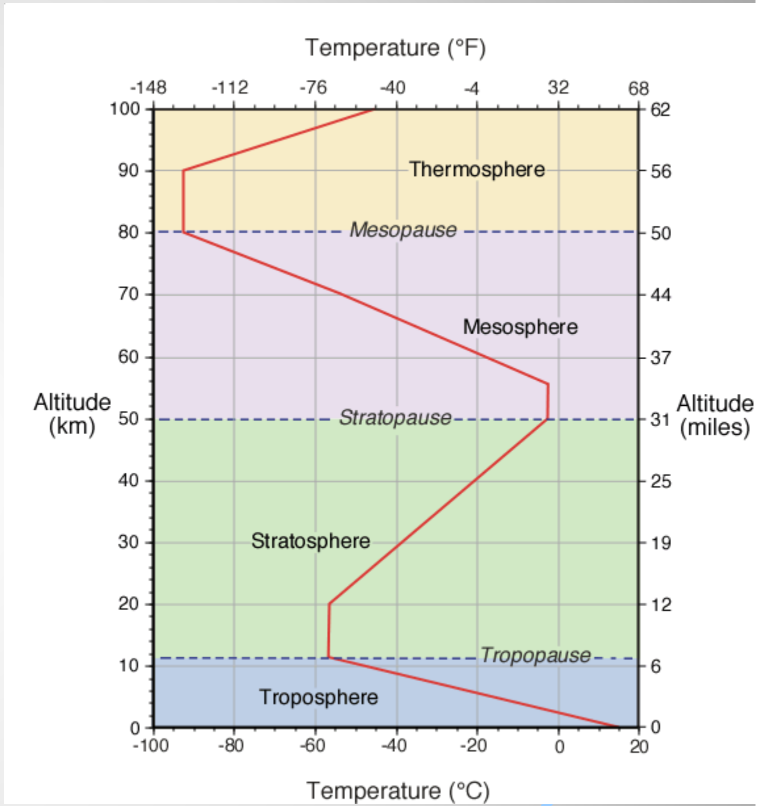
The layer closest to the Earth

Often referred to as "The Weather Layer"

Rain snow and wind stick to this layer.

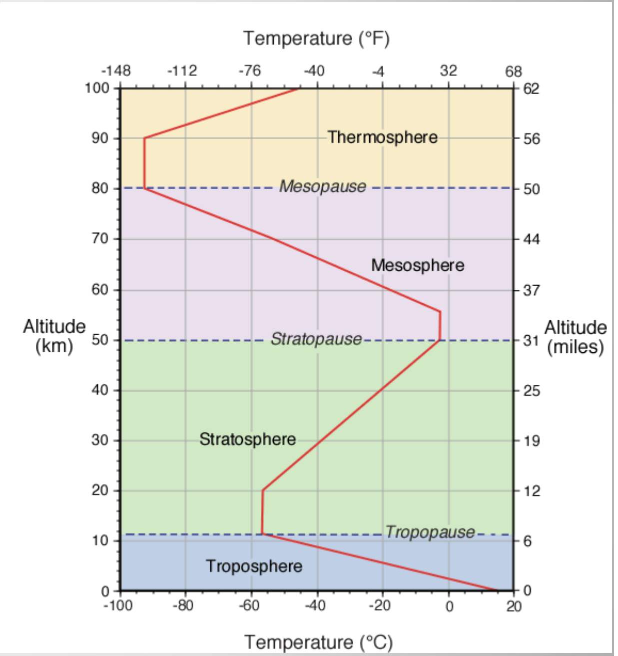
Planes fly in this layer

Below an altitude of 12 km the temperature decreases from 15 °C to -55 °C as altitude increases.



Layer Two (Stratosphere)

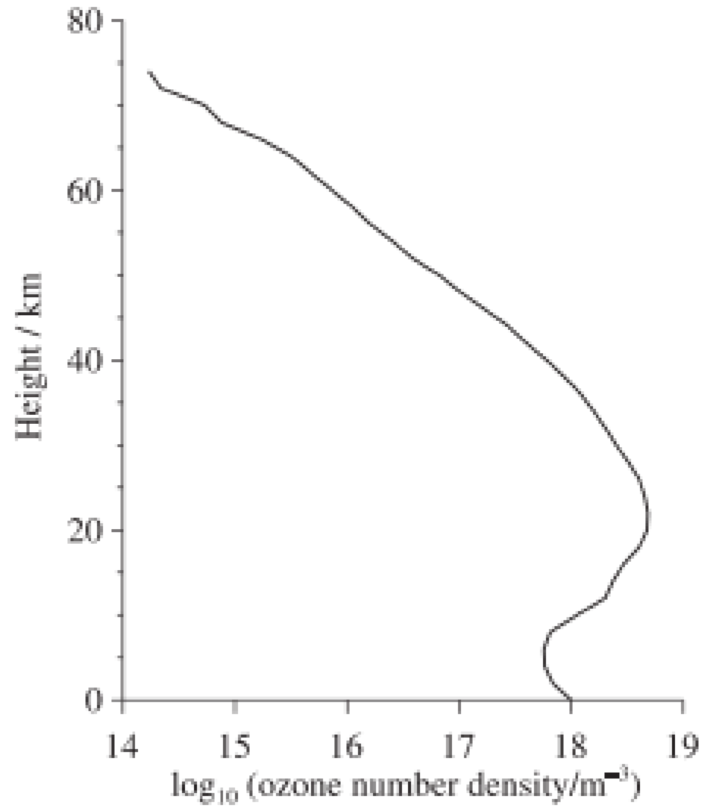
In the stratosphere (12 km - 50 km) the temperature increases with altitude from -55 °C to -2 °C.



Ozone layer found in the stratosphere

Ozone is a gas that absorbs harmful UV rays and protects us from too much solar radiation. Pollution has created a hole in the Ozone layer over the South Pole.

The ozone molecules in the atmosphere



Typical vertical structure of the mean midlatitude ozone number density (molecules m^{-3}). Basedon data from US Standard Atmosphere (1976).

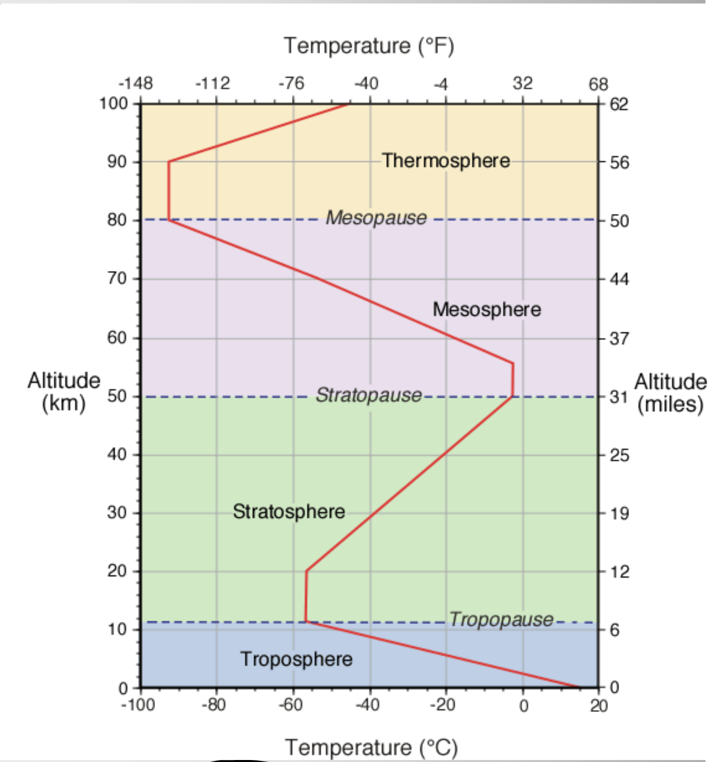
Layer Three (Mesosphere)

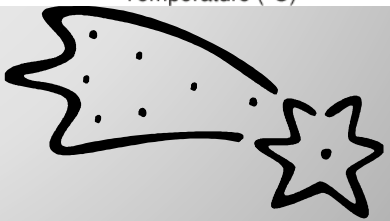
In the mesosphere (50 km - 80 km) the temperature decreases (-2 °C to -90 °C) .

Atmosphere reaches it's coldest temperature of about -90° C.

A lot of meteors disintegrate in this layer from friction from entering the Earth's atmosphere.

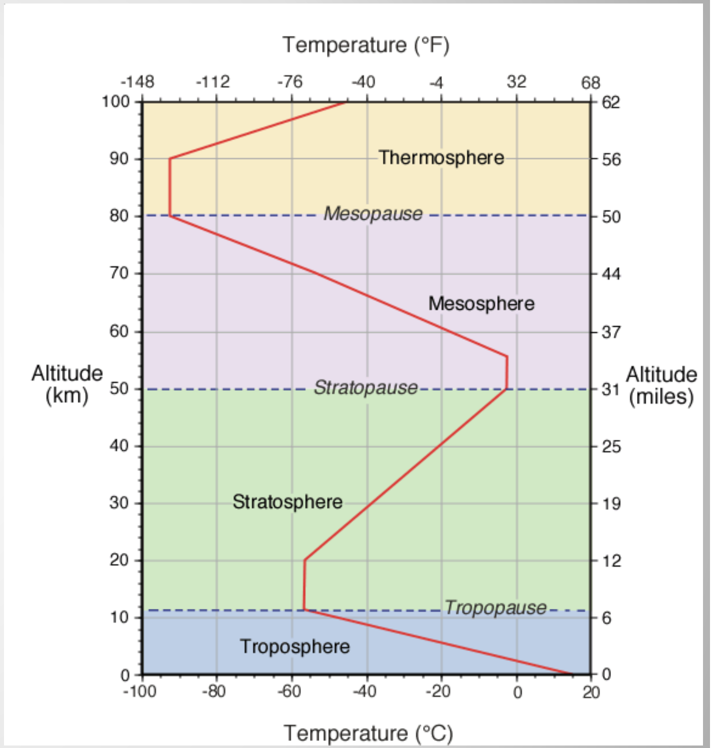
The stratosphere and mesosphere together are sometimes referred to as the *middle atmosphere.*



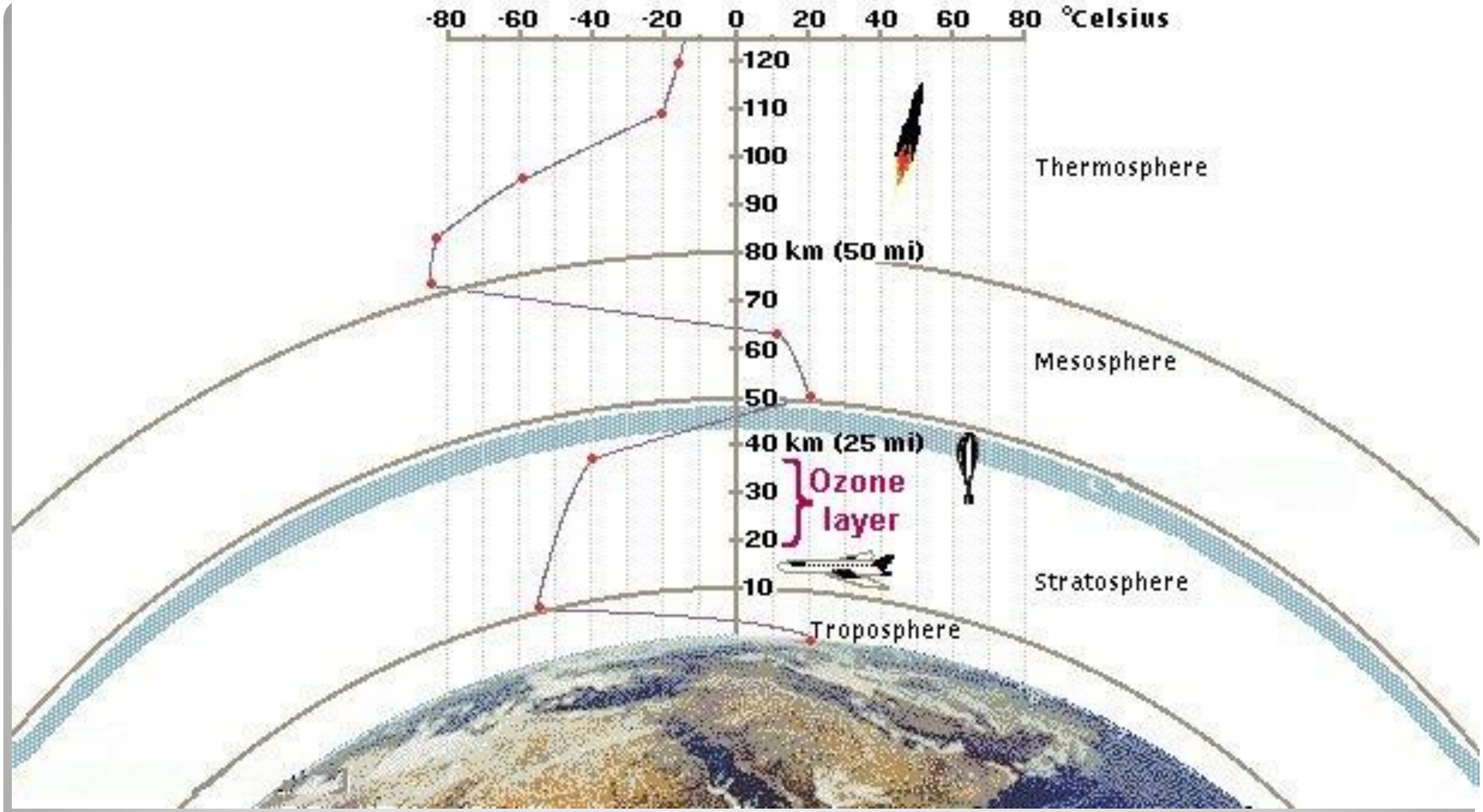


Layer Four (Thermosphere)

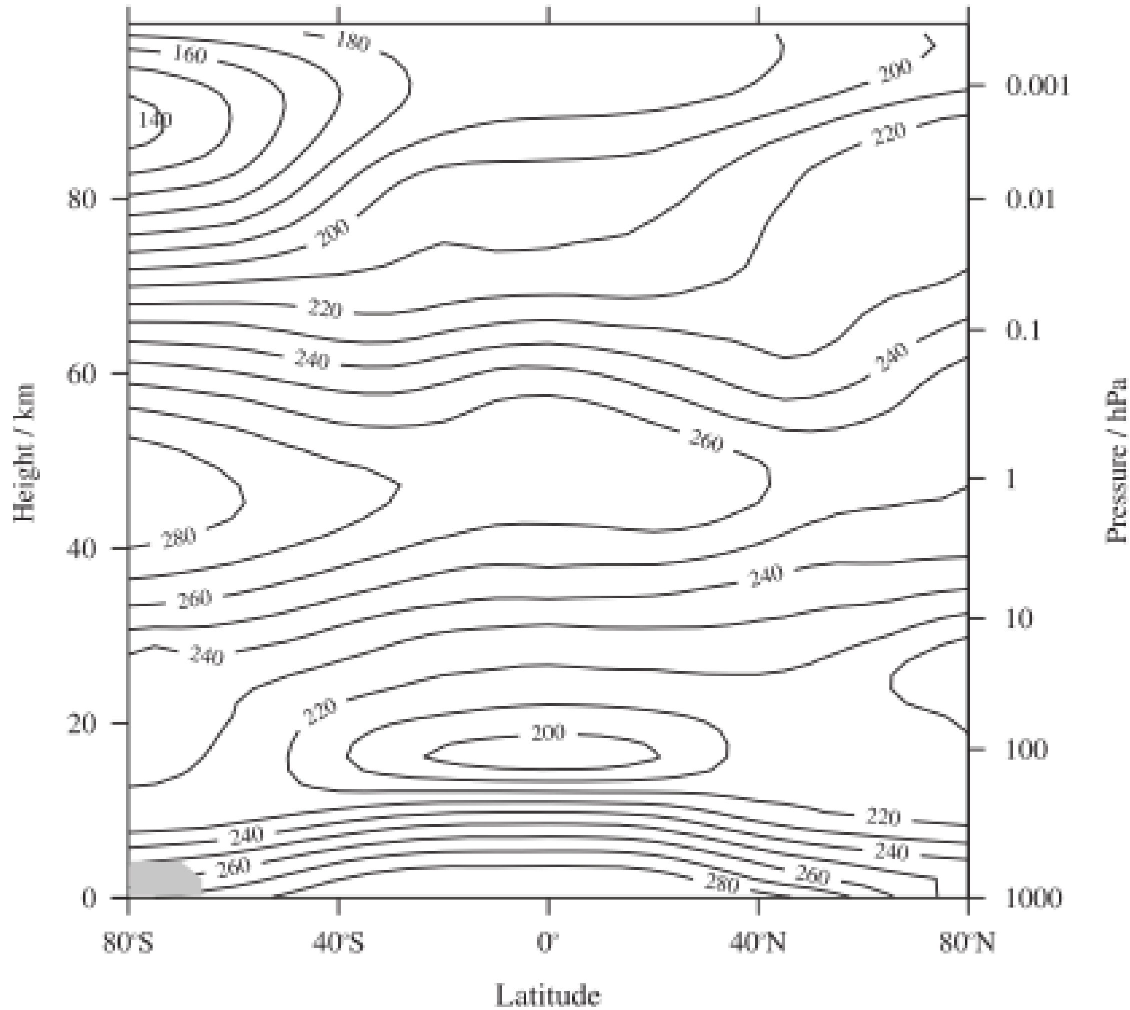
The thermosphere begins 80km above the earth.
Temperatures in the thermosphere go up when moving farther away from ground level due to the sun's energy.



- Hottest of all layer, Temperatures in this layer can get as high as 1300-1800°C.
- Includes the Ionosphere, a region of the atmosphere, which is filled with charged particles.



-80 -60 -40 -20 0 20 40 60 80 °Celsius
120
110
100
90
80 km (50 mi)
70
60
50
40 km (25 mi)
30
20
10
Thermosphere
Mesosphere
Ozone layer
Stratosphere
Troposphere



Zonal-mean temperature (K) for January, from the CIRA (COSPAR International Reference Atmosphere) dataset. A small region at low levels over Antarctica is omitted.

The pressure vs temperature vs density relates like this:

Ht. (M)/Press.--------Temp (in K)---------- Density/kg m-3

0m, 1013.2mb-------288.2----------------------------1.23

5km, 540.5mb-------255.7---------------------------0.736

10km, 265.0mb-------223.2---------------------------0.414

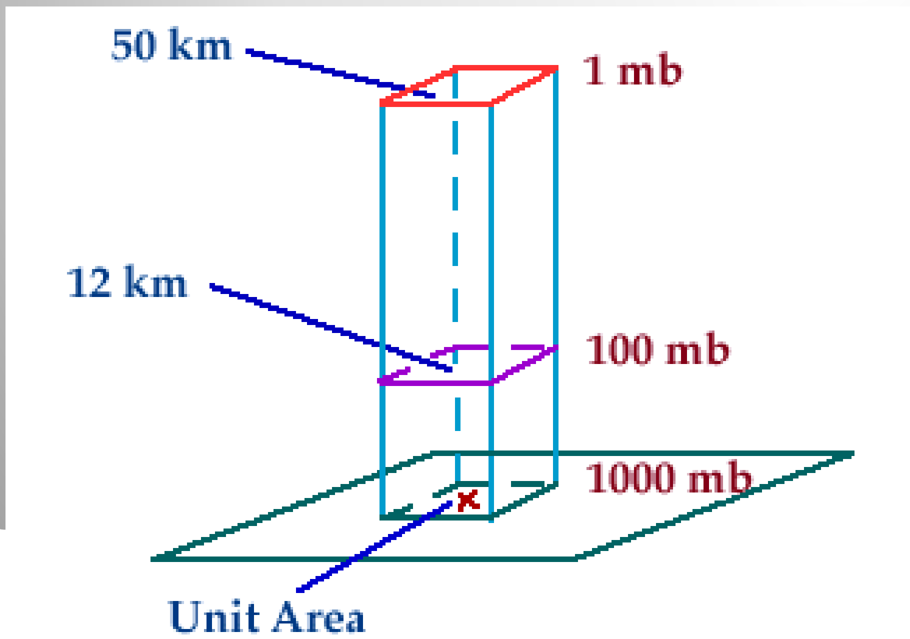
15km, 121.1mb--------216.7---------------------------0.195

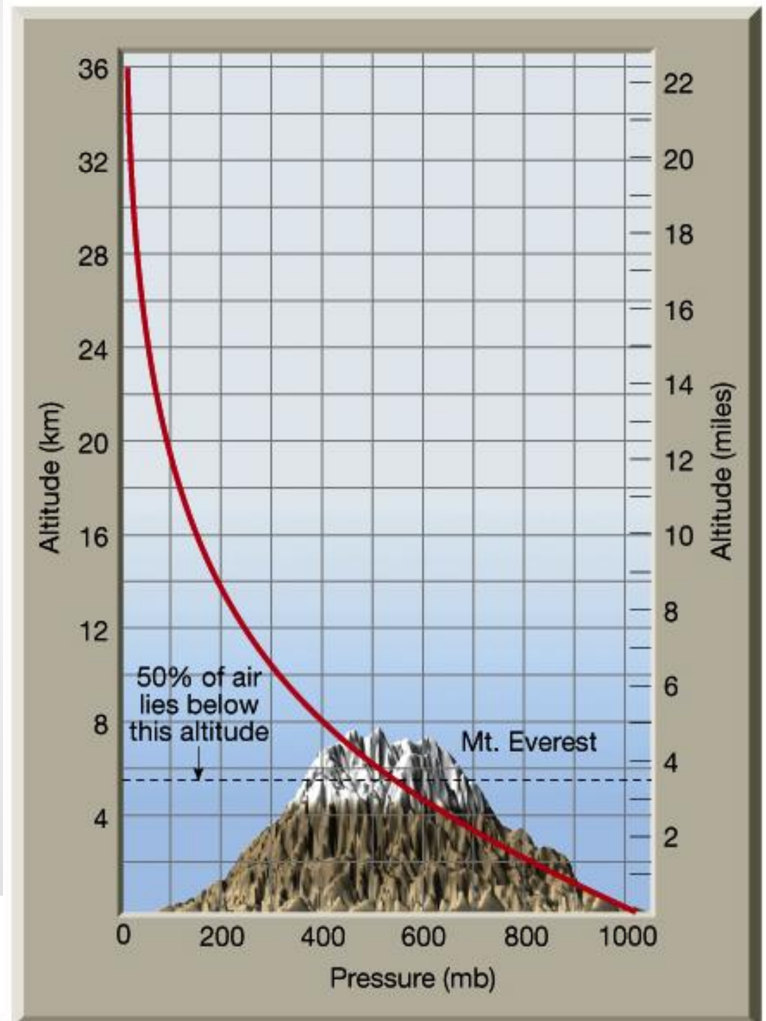
20km, 55.3mb--------216.7----------------------------0.0889

Pressure with Height

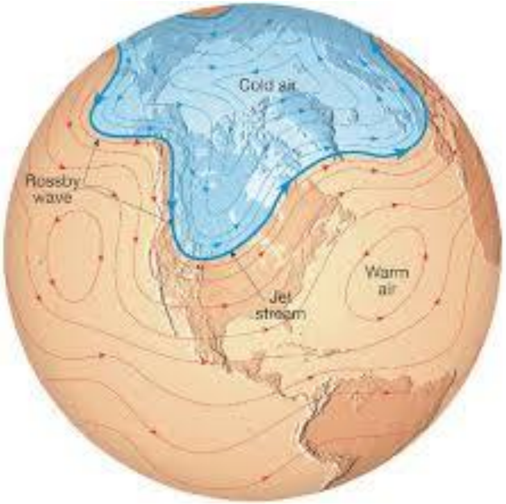
pressure decreases with increasing altitude

The number of air molecules above a surface changes as the height of the surface above the ground changes.

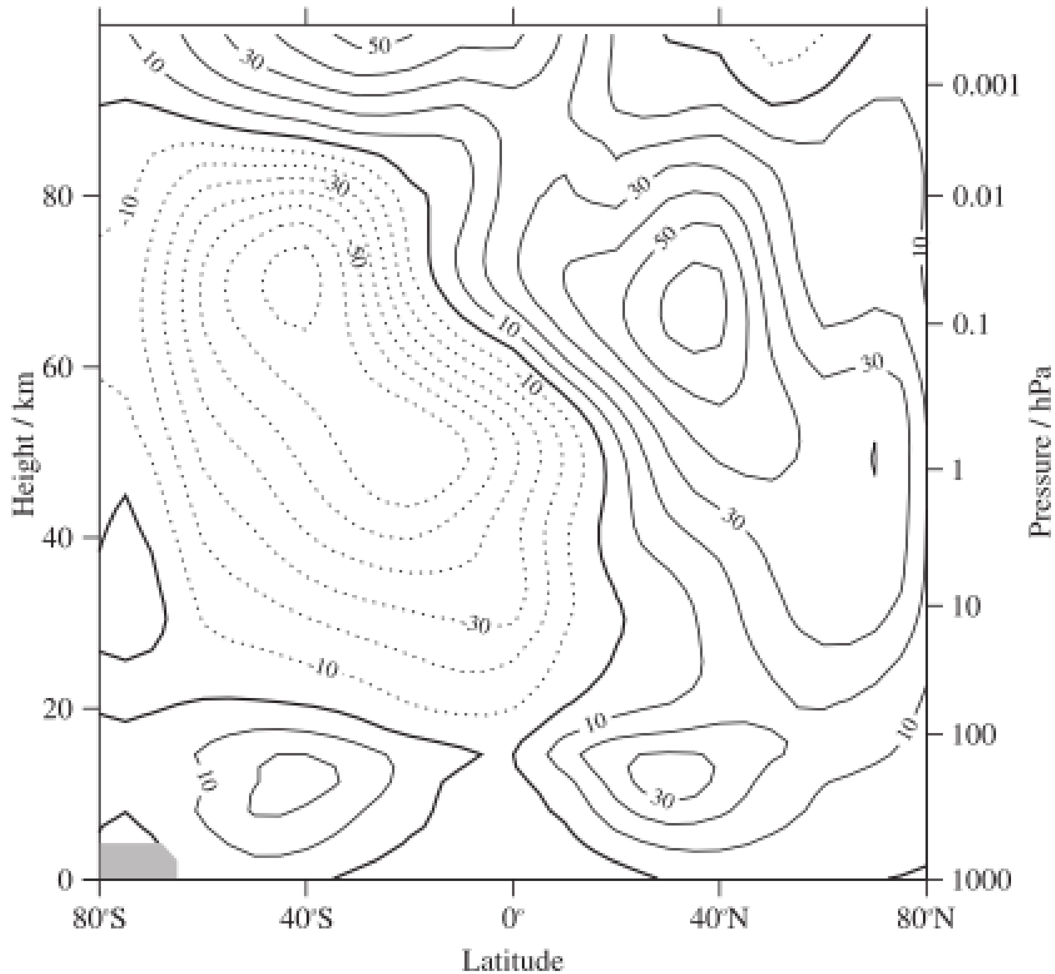




Rossby waves

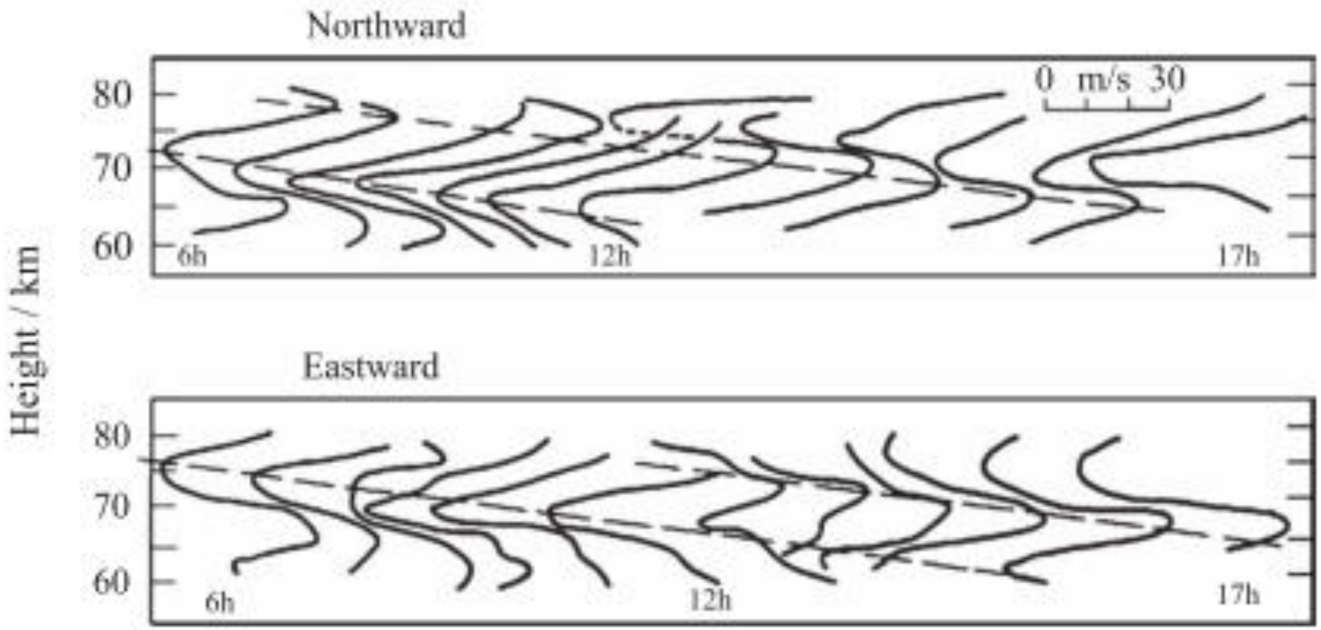


The troposphere is also called the **lower atmosphere. It is here that most ‘weather’**
phenomena, such as cyclones, fronts, hurricanes, rain, snow, thunder and lightning, occur.
The stratosphere and mesosphere together are called the **middle atmosphere. A notable**
feature of the stratosphere is that it contains the bulk of the ozone molecules in the atmosphere;
see Figure 1.4. The neighbourhood of the ozone maximum in the lower stratosphere
is loosely known as the **ozone layer. The production of ozone (O**
) molecules occurs
through photochemical processes involving the absorption of solar ultra-violet photons by
molecular oxygen (O2
) in the stratosphere, three O23 molecules eventually forming two O
molecules. The equilibrium profile of ozone depends also on chemical ozone-destruction
processes and on the transport of ozone by the winds (see Chapter 6).



Height / km
Pressure / hPa
Latitude
80°S
40°S
0°
40°N
80°N
0
20
40
60
80
0.001
0.01
0.1
1
10
100
1000
10
30
50
-10
-30
-50

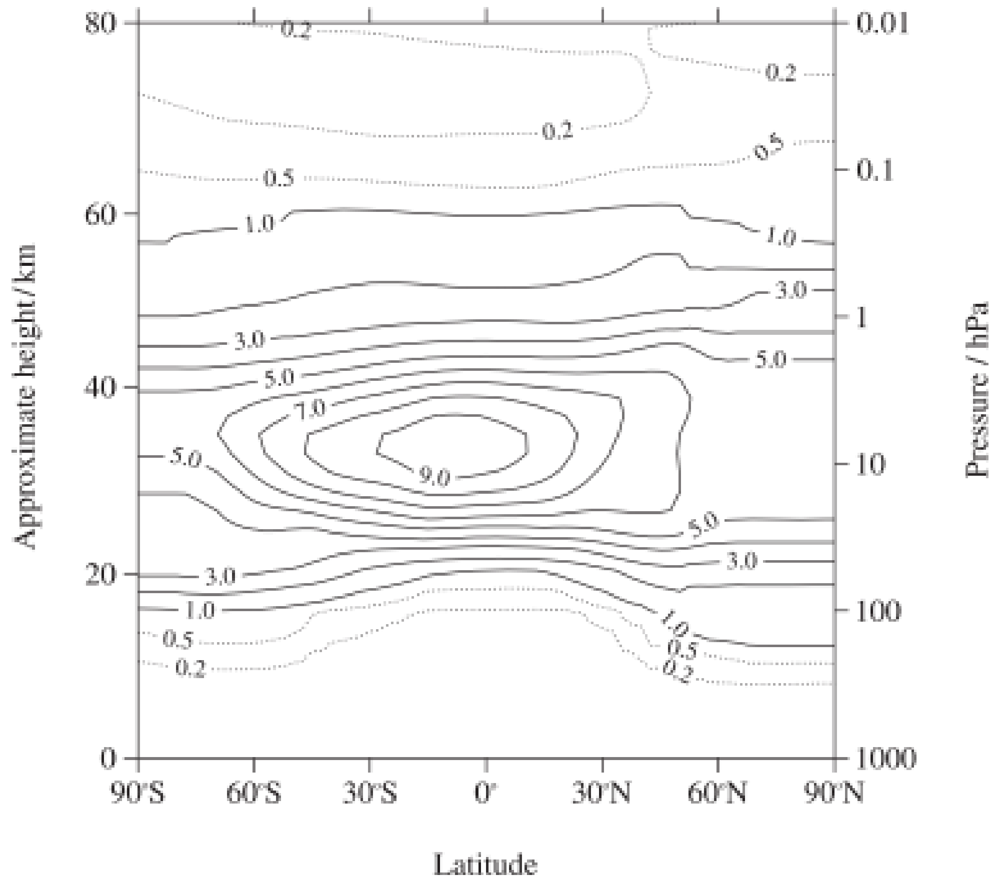
Gravity waves



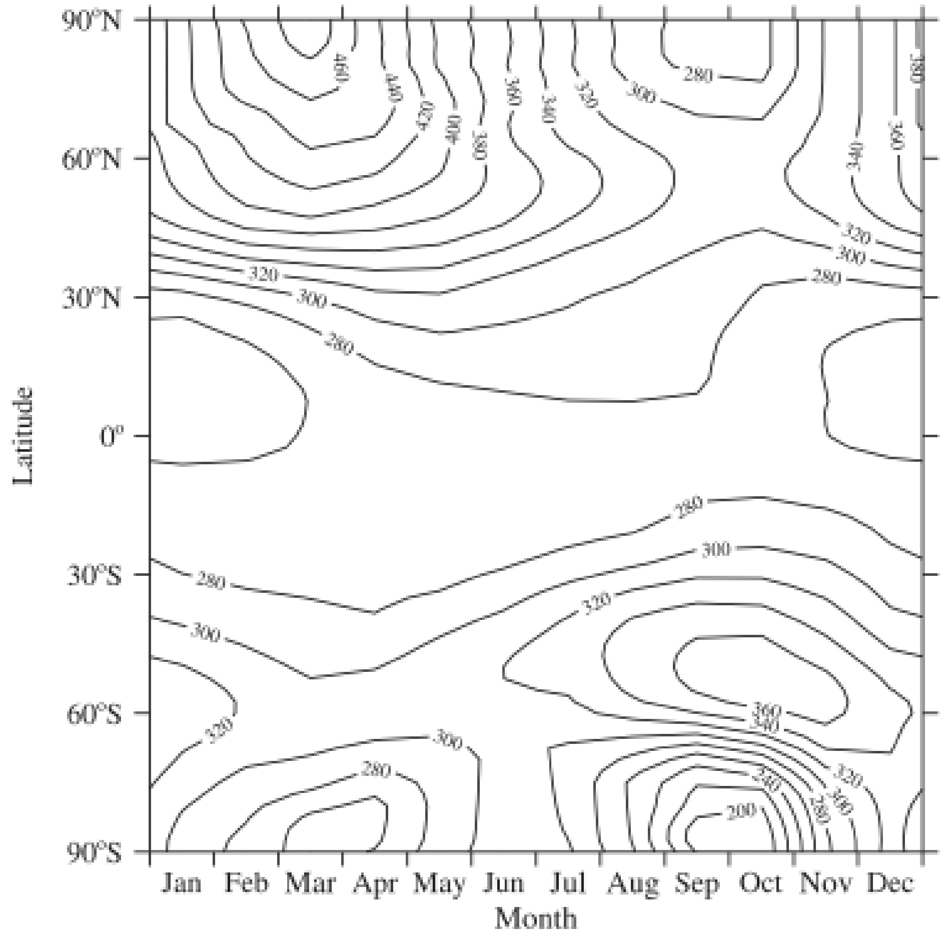
Rossby waves

Ozone

Zonal-mean volume mixing ratio of ozone (parts per million by volume), as a function of latitude and height, for January, based on the 5-year climatology of Li and Shine (1995). Data provided by Dr D. Li.



The observed annual cycle in column ozone, based on the 5-year climatology of Li and Shine (1995). The units are Dobson Units: see Section 6.7.DataprovidedbyDrD.Li.



Weather and climate

Content

سرفصل یا رئوس مطالب:

کلیاتی در مورد ساختار و خواص کلی جو،
نگاهی به مسأله تابش در رژیم گرمایی،
اثرات میدان مغناطیسی زمین و تغییرات آن،
درخشندگی ترکیبات استراتوسفر، ازن جوی،
ترکیبات و ساختار مزوسفر و ترموسفر،
میدان مغناطیسی زمین و تغییرات آن،
درخشندگی جو و نورهای قطبی

هدف درس:

آشنایی با علوم جو با دیدگاه فیزیکی

رئوس مطالب:

- کلیاتی درباره ساختار و خواص کلی جو،
- ترکیبات مشاهدات و اندازه گیری جوی،
- تغییرات زمانی پارامترهای جوی در سطح زمین،
- مبانی تابش،
- تابش خورشیدی و زمینی،
- نمودارهای ترمودینامیکی،
- جذب و پراکندگی تابش در جو،
- یونیزاسیون و یونسفر،
- تولید الکترون در اثر اشعه خورشید،
- تئوری تشکیل لایه های یونسفر،
- لایه های D، E و F،
- انتشار امواج الکترومغناطیس در جو

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اهداف کلی درس: آشنایی با مبانی فیزیک جو شامل حرکت، تولید، انتقال و اتلاف انرژی و فرایندهای تابشی

اهداف رفتاری:

سرفصل درس:

نظری:

فصل اول – مقدمه

سامانه فیزیکی جو، مدل های جوی یک مدل تابشی ساده، مدل ساده ای برای اثر گلخانه ای، گرمایش جهانی، برخی مشاهدات جوی شامل میدان های جهانی دما و باد؛ امواج گرانی، راسبی؛ ازن؛ توزیع افقی انتقال تابش.

فصل دوم – ترمودینامیک جو خشک

قانون گاز ایده آل، ترکیب جو، توازن هیدروستاتیک، آنتروپی و دمای پتانسیلی، مفهوم بسته و پایداری ایستایی، انرژی پتانسیلی در دسترس.

فصل سوم – ترمودینامیک جو نمناک

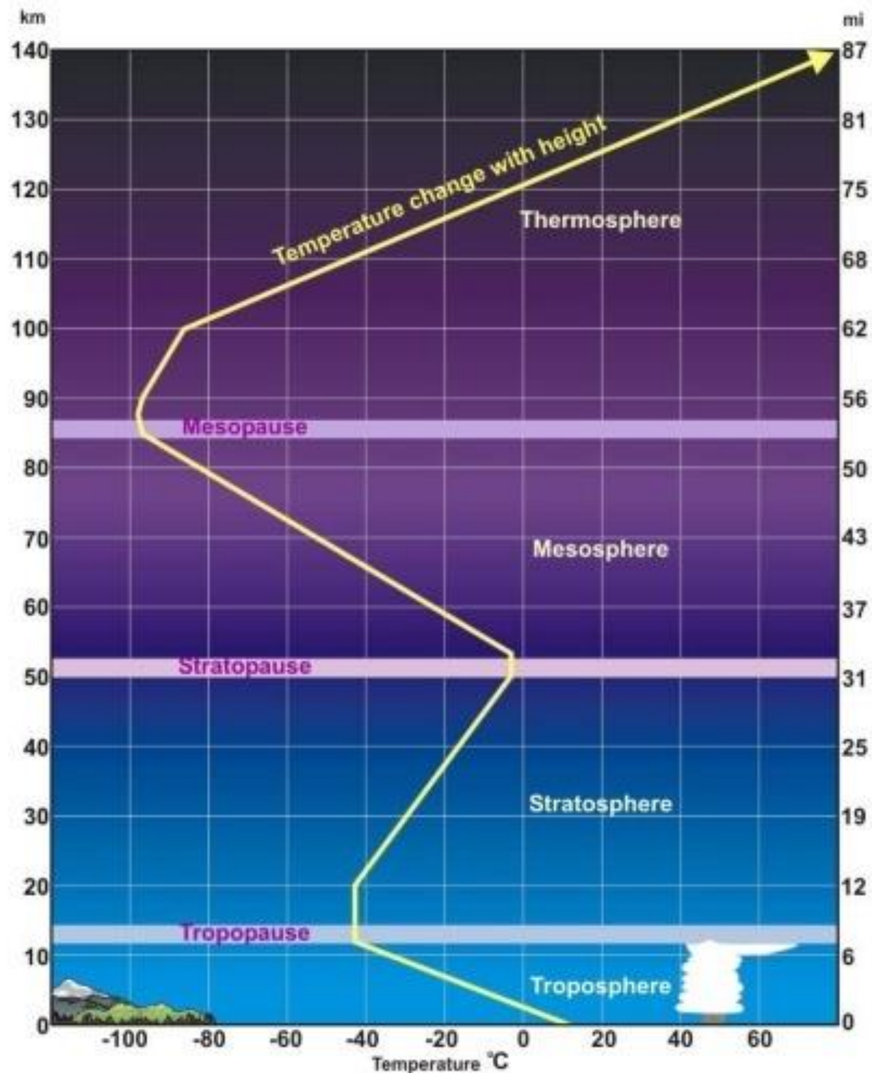
توصیف هوای نمناک، میعان و آزاد شدن گرمای نهان، آهنگ افت دمای بی درروی اشباع، اثر رطوبت بر پایداری ایستایی، نمودار skew-T.

فصل چهارم – ابر و بارش

هوامیزهای جوی، رشد قطرک به وسیله ی میعان، هسته سازی همگن، فرمول کلوین، هسته سازی ناهمگن، منحنی کوهلر، رشد قطرک به وسیله ی برخورد، رشد ذرات یخ، فرآیند برگرون.

فصل پنجم – تابش جوی

مفاهیم فیزیکی پایه، تابع پلانک، ترازمندی تابشی محلی، معادله ی انتقال تابش، طیف نگاری پایه، تراگسیلایی، جذب به وسیله



Atmosphere Layers and Their Mean Temperatures