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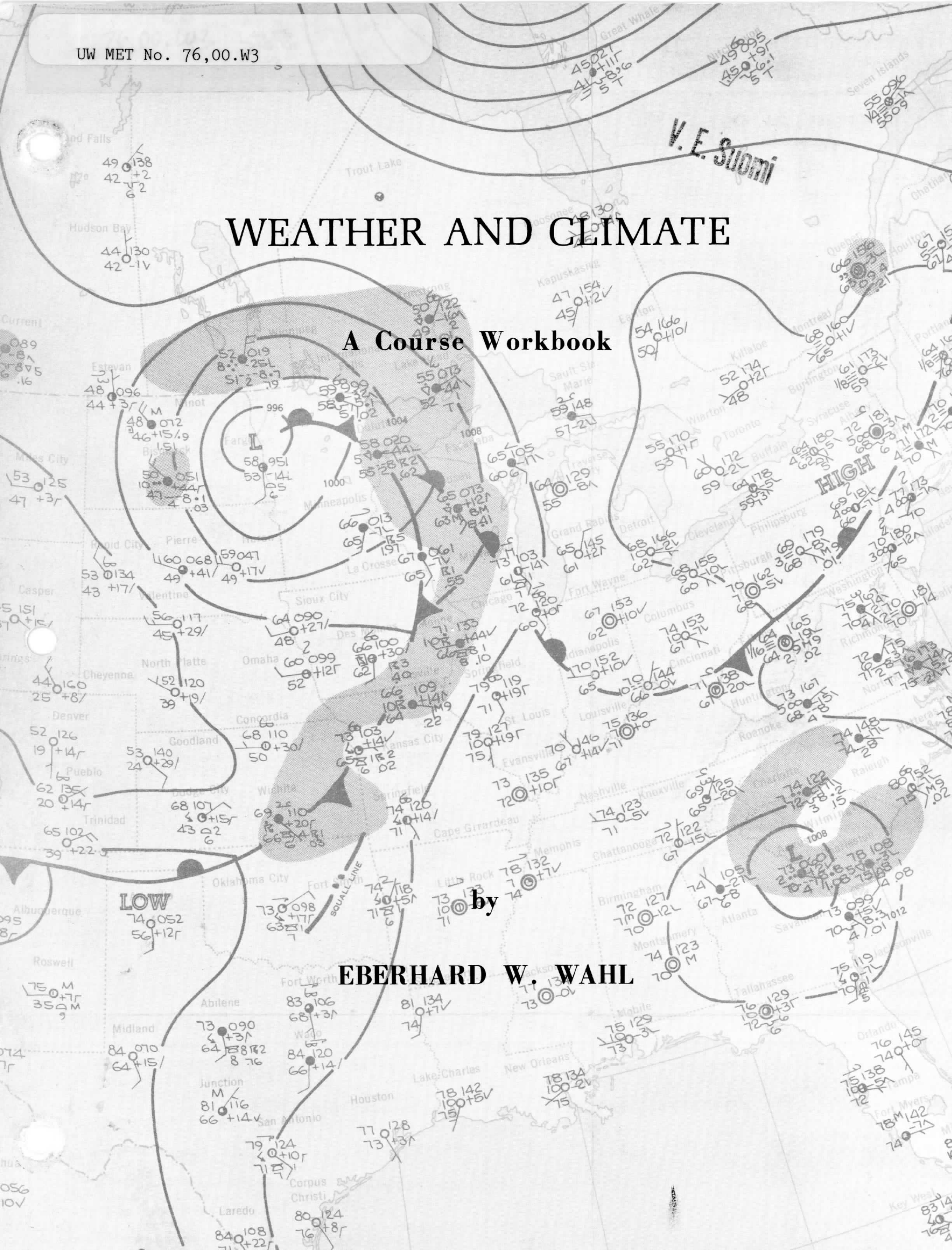
WEATHER AND CLIMATE

A Course Workbook

by

EBERHARD W. WAHL



WEATHER AND CLIMATE

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Eberhard W. Wahl

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INTRODUCTION

This workbook has been prepared as a classroom aid for use in a Weather and Climate course and was first used at the University of Wisconsin. It does not substitute for the textbook to be used in such a course (the author's choice was: The Atmosphere. - R.A. Anthes, H.A. Panofsky, J.J. Cahir, A. Rango, Charles E. Merrill, 1975. However, the workbook will give the student background material in key phrases, tabular and pictorial form to augment the lecture content for his own studies.

The course outline presented below is one possible arrangement of the material usually covered in such a course; however, other ways of presenting the subject matter may be as effective. The numbering of the pages of this workbook by chapter and page within the chapter will allow the use of the material in any desired order and should thus simplify its use even if the course outline is not adopted. Equally easy would be a replacement of the reading assignments in the textbook as listed at the beginning of each chapter by the corresponding pages of another textbook.

Students must realize that such a survey course is essentially a lecture course; the content of the lectures are only augmented by the material in this workbook and by the reading assignments in the required textbook. Nothing can substitute for conscientious attendance at the lectures and for careful note taking. The latter, however, is substantially simplified by the workbook.

The course as taught at Wisconsin consists of three lecture hours and one discussion hour per week for one semester. For use in the discussion sections which are held in small groups, and for home study of the individual student, a number of problem sets related to the content of each chapter have been added at the end of the chapters. It has been found most advantageous to solve these problems in order to clarify the concepts and thus deepen the understanding of various new facts which all too often overwhelm a student in such a course.

COURSE OUTLINE AND READING ASSIGNMENTS

Chapter		Approximate No. of Lecture Hours	Reading Assignment (text book)
I	Introduction	2	pp. 1-18
II	General characteristics of the Atmosphere	4	pp. 35-36, 61-67, Appendix A.1
	A. Composition and Extent		
	B. Vertical Structure		
	C. Quantities describing the state of the atmosphere and their observation		
III	Heating of the atmosphere	5	pp. 20-35, 58-61, 72-73
	A. Solar energy driving the atmosphere		
	B. Seasons and daily changes		
	C. Radiation processes		
	D. The heat budget		
IV	Thermodynamics of the atmosphere	6	pp. 35-41, 168-173
	A. Adiabatic processes		
	B. The thermodynamic diagram		
	C. Stability		
V	Clouds and Precipitation	7	pp. 94-103, 209-251
	A. The condensation process		
	B. The precipitation process		
	C. Clouds—Classification, appearance		
	D. Forms of precipitation		
	E. Optical phenomena		

VI	Atmospheric Motions	5	pp. 41-51, 105-116
	A. Definitions		
	B. Equations of motion		
	C. Approximation of the actual wind		
	D. Variation of wind with height		
VII	The general circulation	3	pp. 67-72, 78-79
	A. General arrangement		
	B. The three circulation zones		
	C. Upper level circulation		
	D. Special large-scale wind systems		
VIII	Second-order circulations—weather systems	6	pp. 42-58, 103-107, 116-128
	A. Cyclones and anticyclones		
	B. Airmasses and fronts		
	C. The weathermap		
IX	Third-order circulation—mesoscale systems	3	pp. 168-187
	A. Thunderstorms and tornados		
	B. Mountain wave		
	C. Local wind systems		
X	Climates of the earth	3	pp. 55-58, 80-92
	A. Climatic elements		
	B. Climate classification		
	C. Climatic zones		
	D. . . . and of other planets?		

CHAPTER I. INTRODUCTION

(Reading: pp. 1-18)

What is METEOROLOGY?

Role of atmosphere in life's existence.

Matter in the Universe—cosmic importance of atmospheres.

History of planetary atmospheres—how earth acquired its current gaseous envelope and what it means to life.

Primary vs. secondary atmosphere. Importance of ozone.

Brief history of man's interest in weather. Folklore to empirical rules to natural science.

Relation of meteorology to other natural sciences.

Some historical highlights in the development of meteorological knowledge. (Important facts underlined)

First recorded speculations on meteorological phenomena occur early in Greek history, due to dependence of Greeks on wind and weather. Hippocrates ($\sim$500 B.C.) mentions the effect of climate (Greek word!) on human health. <u>Aristotle</u> ($\sim$350 B.C.) compiled first "compendium" (4 books) called <u>Meteorologica</u> (i.e. pertaining to things "above earth", it also contains information on meteors which we now classify as belonging to astronomy). Around 100 B.C., both Greeks and Romans are known to have kept weather records. The first actual measurements (rainfall amount, wind direction) were recorded. There exists fragmentary evidence that about that time similar observations were made in the East Asiatic cultures (China, Korea). Weather lore was quite extensive and probably weather forecasts ("signs") were made for many practical uses.

As in all other natural sciences, stagnation set in after the fall of the classical civilizations and no real progress is apparent until well into the Medieval times. The renaissance of mathematics in Arabic cultures (Near East, North Africa, Spain) did not advance meteorology; no real interest in weather existed for these people.

New impetus to our science came in monasteries and early universities in Middle Ages, often related to astrological concepts. The first systematically continued series of weather observations (for 7 years) was made by W. Merle in Oxford, England (1337 - 1344 A.D.), still largely subjective (i.e. describing the weather without actual measurements). This could change only to more scientific methods after instruments for measuring properties of the atmosphere were invented: the thermometer by Galileo (1592/1612 A.D.), the barometer by Torricelli (1643 A.D.)

With beginning of renaissance the exact sciences of physics and chemistry emerged from pre-science concepts of astrology and alchemy. Air became a subject of investigation. Its composition was investigated. J. Mayow (1655) first realized that air must consist of different parts, one of which sustains the burning ("fire air"). Boyle ($\sim$1660) formulated the first gas laws. Chemically, the first actual constituent of air was isolated by J. Black (1750), namely CO_2. Around 1780, all major constituents had been isolated (O_2, N_2) by Rutherford, Priestley and Scheele. Only the noble gas Argon was not found at that time; it was confirmed only 1896 by the British Chemist Ramsey.

Thermometer scales and means of accurately calibrating thermometers were devised by Fahrenheit (1710) and Celsius (1742); the "absolute" temperature scale was introduced 1850 by Lord Kelvin (now called ^{o}K).

Modern developments: First reasonably regular meteorological observations since mid-seventeenth century in various European countries; in U.S. first records were compiled by J. Campanius (1664) in Delaware. All these records were taken at or near the earth's surface. The first attempts to observe parameters above the surface were made by Wilson (1749) by means of large kites. Remember Ben Franklin's experiments with lightening electricity. The hot air balloon was invented 1783 by Montgolfier, the hydrogen balloon by Charles (1784) - who also formulated the gas laws.

The concept of the "synoptic weathermap" developed in the early 19th century in both U.S. and Europe, e.g. Loomis (U.S.) between 1820 and 1840 and at the same time Leverrier (France) and Brandes (Germany). To use these maps in a timely fashion, one had to receive the observations from large areas very speedily; the invention of the telegraph (by Morse, 1840) was essential. The first regular weather maps in the U.S. were issued by A. Abbe in Cincinnati in 1869.

1870, Congress established the U.S. Weather Bureau; before that, observations in this country were taken by Army personnel. Science of meteorology now became distinctly separate from its parent sciences physics and chemistry. Meteorologists studied behavior of atmosphere using mathematical and physical tools to develop theories and make observations. Technical progress afforded new methods: first use of airplanes to observe the weather goes back to World War I.

Invention of radio was used in the development of the first radio sondes by Moltchanov (1928). One important milestone in understanding weather phenomena on a larger scale came with formulation of Polar Front theory by Norwegian scientists, foremost V. Bjerknes, in 1917. Richardson (1922) made the first attempt to "compute" the weather, by desk calculator!

The modern picture of the general circulation is based to a large extent upon the work of C. G. Rossby who first described the "long waves" in the westerlies, now called Rossby-waves (1939); he also discussed in detail the jet stream found in the early 1940's. Development of radar (around 1942 both in U. S. and England) also lead to weather radars. After World War II, surplus German V-2's were used to probe the atmosphere to very large heights (beyond 100 miles). At the same time, the development of electronic computers led to the first successful "numerical" weather predictions and modelling of the atmosphere, spearheaded by von Neuman, Charney, Phillips and Thompson (around 1949). Finally, space age took over in meteorology, too, with the launch of the first satellite devoted to meteorological research, Tiros I, on 1 April 1960. The first synchronous satellite, televising nearly half of Earth's cloud systems was launched in 1966 (ATS 1); the next one gave a "color picture" (ATS 3) in 1967. The first Synchronous Meteorological Satellite (SMS) was launched in 1974.

During the 1970's, the international meteorological community launched their most ambitious endeavor, the Global Atmospheric Research Program (GARP) and the associated World Weather Watch (WWW). Several Subprograms already have been conducted, e.g. BOMEX (Barbados Oceanographic and Meteorological Experiment) and GATE (Garp Atlantic Tropical Experiment), and FGGE (First Garp Global Experiment) will be conducted in 1978. An extremely large number of countries are cooperating under the auspices of the U.N. and the WMO (World Meteorological Organization). In addition to research directed toward the improvement of weather forecasting and of our theoretical knowledge about the physical foundations of meteorology a significant effort is being directed toward the better understanding of climate and climatic change.

Cultural importance of meteorology as part of man's dependency upon his environment.

Current status of meteorology as science

CHAPTER II. GENERAL CHARACTERISTICS OF THE ATMOSPHERE

(Reading: pp. 35-36, 61-67, Appendix A.1)

A. Composition and Extent.

Total mass of planet Earth: $6.6 \cdot 10^{27}$ g

Total mass of atmosphere: $5.6 \cdot 10^{21}$ g

Air = Mixture of gases (not chemical compound!)

There is no "air molecule"!

Natural air also contains solid and liquid admixtures (aerosol).

Clean air—gases only.

Dry air—gaseous constituents except (very variable amount of) water vapor.

Composition of dry air very constant from place to place, also with height [up to 20 (90) km].

Contains (by volume)

			First identified:
N_2	Nitrogen	78.09%	) ~ 1780 (Rutherford, Priestley, Scheele)
O_2	Oxygen	20.95%	)
A	Argon	0.93%	1896 (Ramsey)
CO_2	Carbon dioxide	0.03%	1750 (Black)

Some traces of other gases (less than 3/1000 %)

He, Ne, Kr, Xe, Rn; H; O_3

Water vapor content of clean (natural) air can vary from near zero percent (very dry cold arctic air) to about four percent (very moist tropical air).

Ozone (O_3), in spite of very small concentration, important constituent of atmosphere. Is formed in level 30-50 km high and in forming absorbs lethal ultraviolet (uv) radiation from sun. Ozone layer about 20-50 km, concentration 0.1-0.2 parts per million.

Process: Formation 1) $O_2 + (uv) \rightarrow O + O$ [uv < 0.24μ]

2) $O + O_2 + (M) \rightarrow O_3 + (M)$

Decay 3) $O_3 + (light) \rightarrow O_2 + O$ [light < $1.1\,\mu$]

4) $O_3 + O \rightarrow O_2 + O_2$

Mass of Atmosphere—about 5,600 mill. mill. tons; still a very tiny part of total mass of earth itself (about one millionth - 10^{-6}).
Replace weight of air by water—then total atmosphere equivalent to a shell of water covering earth 10 m (34 ft) thick.

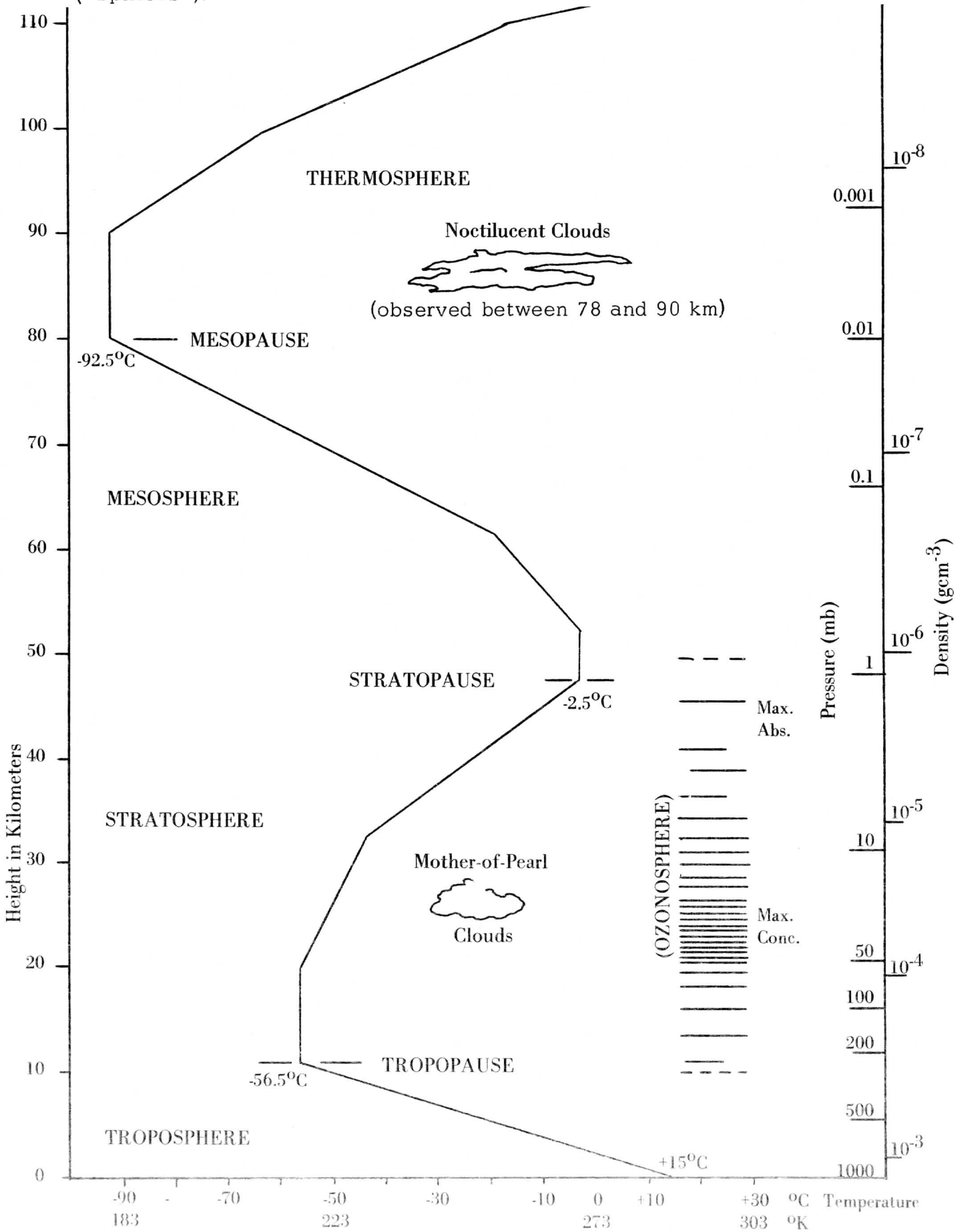
Extent of Atmosphere—while densest close to ground and getting thinner upward, does not have definite boundary toward space. When going up,

at 5.5 km height	—	1/2 of atmosphere below
at 11 km height	—	3/4 of atmosphere below
at 16.5 km height	—	9/10 of atmosphere below
at 32 km height	—	99/100 of atmosphere below

[Astronaut wings awarded when going above 50 mi = 80 km!]

B. Vertical Structure

Description by means of various shells with certain differences in behavior ("spheres").



Of chief interest in general meteorology: troposphere, stratosphere.
Observations up to 30-40 km by balloons + radiosondes
up to 80-150 km by meteor. rockets
above 150 km by earth satellites
Atmospheric composition up to ~ 90 km nearly constant (due to constant vertical + horizontal mixing). Above 90-100 km, composition changes—heavier gases decrease faster with height.

Eg. at 500 km height—atomic oxygen prevails (> 80%), some He
at 1000 km—He prevailing, very little O, some H
Further out, atmosphere merges with interplanetary gas (H, He)

Importance of very high layers (Thermosphere) due to electrification.
Ionosphere—effect on radiowaves (Electron—layers)
Exosphere—van Allen belts, magnetic field.

C. Quantities describing the state of atmosphere.

CGS—system: (used in science)	Length	1 cm	(1 inch = 2.54 cm)
	Mass	1 g	(1 pound = 454 g)
	Time	1 sec	(1 day = 86,400 sec)

Additional derived quantities of importance:

Velocity (a vector) $\vec{v}$ } change of distance in time [cm/sec]
Speed (a scalar) v

$$\vec{v} = \frac{\Delta d}{\Delta t}$$

Acceleration (vector) $\vec{a}$ change of velocity with time $\left[\frac{\text{cm/sec}}{\text{sec}} = \text{cm sec}^{-2}\right]$

$$\vec{a} = \frac{\Delta \vec{v}}{\Delta t}$$

Force (vector) $\vec{F}$ Mass times acceleration [g cm sec^{-2}]

$$\vec{F} = m \cdot \vec{a} \qquad \text{Unit 1 dyne}$$

Pressure (vector) $\vec{P}$ Force per area [g cm $\text{sec}^{-2}/\text{cm}^2$ = g cm^{-1} sec^{-2}]

$$\vec{P} = \frac{\vec{F}}{A} \qquad \text{Unit 1 dyne/cm}^2$$

This is the proper definition of P in mechanics. Pressure in a gas, however, acts in all directions at once. (Example: gas in interior of balloon expands the skin of balloon in all directions.) Thus, we will later consider gas pressure to be a scalar quantity!

Energy (scalar) E Force times distance $\left[\text{g cm sec}^{-2} \cdot \text{cm} = \text{g cm}^2\,\text{sec}^{-2} = \text{g}\left(\frac{\text{cm}}{\text{sec}}\right)^2\right]$

(also: mass times velocity squared)

Density (scalar) ρ mass per volume $[g\ cm^{-3}]$

$$\rho = \frac{M}{V}$$

Heat—Energy form = kinetic energy of molecules of body.

kinetic theory of gases.

Temperature—a measure of speed of molecules.

Absolute temperature scale (Lord Kelvin, ~ 1850) °K
Zero point—no remaining motion, molecules "frozen."
When heat added, molecules move faster, need more room—gas expands. For each °C (see below), increase in volume about 1 part in 273—

$$V_t = V_o\left(1 + \frac{1}{273}t\right) \qquad V_o = \text{volume at } 0°C$$
$$t = \text{degrees C}$$

Absolute zero point when $V_t = 0$, i.e. $t = -273°C$.

Other temperature scale 1) °C (Celsius or Centigrade) ~ 1742
zero point (0°C)—melting of ice/freezing of water
100°C—boiling pt. of water (at sealevel)
1°C = 1/100 of this difference.

2) °F (Fahrenheit) ~ 1710
zero (2) point (modern reference) 32°F same as 0°C
212°F same as 100°C
1°F = 1/180 of this difference

Conversion from any scale to any other:

1) °C = °K-273; °K = °C + 273

2) $°C = \frac{5}{9}(°F - 32)$; $°F = \frac{9}{5}°C + 32$

3) °F to °K: first go to °C, add 273.

Measurement of temperature—various forms of thermometers.

In the near future, all temperatures will be given in °C. Already now, all meteorologists report all (but local station temperatures in the U.S.) in °C. For scientific calculations one usually needs °K values.

Atmospheric Pressure

Physical unit 1 dyne/cm^2 very small, thus we use

1 millibar (mb) = 1,000 dynes/cm^2

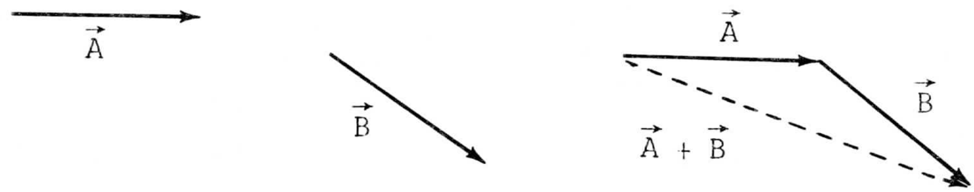
Pressure near sea level near 1,000 mb ± 20-40 mb

Pressure measurements by various forms of barometers. See text p.8.

Operations with Vectors (Addition, Subtraction)

To describe quantities which have both a size and direction (such as velocity or force) it is convenient to use vectors. A simple representation of a vector is an arrow, the length being proportional to the size of the quantity and the direction pointing in the direction of the quantity.

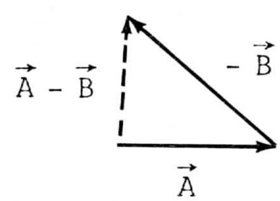
Vectors may be added and subtracted quite easily. To add two vectors: (1) place the tail of the second vector at the head of the first vector. (A vector may be moved about as long as its size and direction are kept the same), and (2) draw a new vector which has its tail at the tail of the first vector and its head at the head of the second vector. This new vector represents the sum of the two vectors. An example is shown below.



One vector may be subtracted from another by adding the negative of it to the first vector. The negative of a vector is a vector of the same size but of opposite direction. For example:

Thus to subtract Vector B from Vector A we merely add -B to A

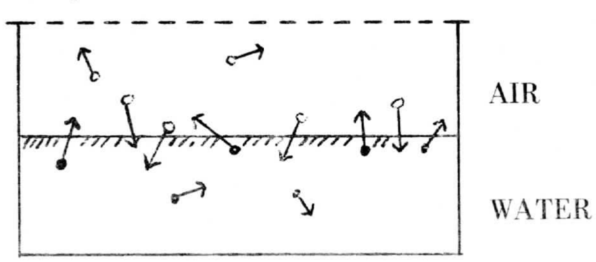
$$\vec{A} - \vec{B} = \vec{A} + (-\vec{B})$$



Above quantities valid for any gas. Specific quantities for measuring and defining water vapor content.

Concept of SATURATION

[Assumptions: Flat water surface, chemically clean water]



"As many molecules of H_2O return into liquid (condense) as leave liquid (evaporate) in given time interval."

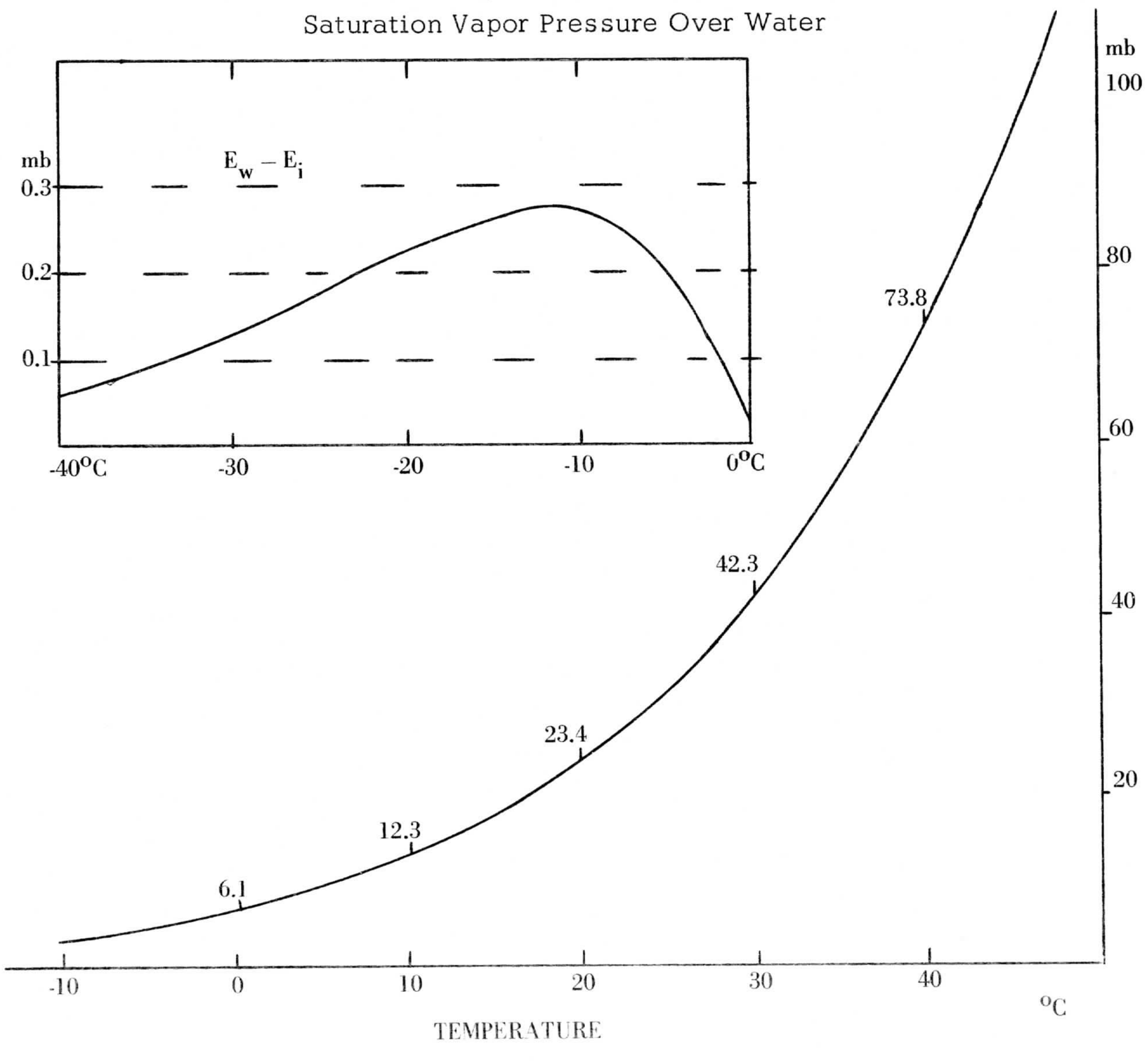
This requires, at a given temperature (i.e. molecular speed!) a definite number of H_2O-molecules in "Air."

Concept of PARTIAL PRESSURE

Total Pressure of gas mixture = sum of partial pressures of individual constituents.

If air in direct contact with open water surface, equilibrium when saturation reached. Partial saturation water vapor pressure then depends only on T. Called E_w (subscript w for (liquid) water)

For ice surface, E_i (i for ice) slightly lower than E_w at same temperature (below 0°C, i.e. supercooled water).



Quantities to specify H_2O-vapor content of air:

1) Partial water vapor pressure — e — (either e_w - over water or e_i - over ice)

2) Mixing ratio — w — no. of g H_2O vapor associated with 1 kg of dry air

3) Specific humidity — q — no. of g H_2O vapor contained in 1 kg of moist air

4) Absolute humidity — d — no. of g H_2O vapor in 1 cubic meter of moist air

5) Relative humidity — U — amount of H_2O vapor in percent of total amount possible at saturation (i.e. relative to saturation)

$$U = \frac{e}{E} \times 100 = \frac{w}{W} \times 100 \quad [W = \text{saturation mixing ratio}]$$

Measurement of moisture content

Dew point temperature ($U = 100\%$)
Dry bulb/Wet bulb temperature
Hygrometer, Psychrometer.

EXERCISES—Chapter II

1. Complete the following temperature conversions:

   a) -100° F = ________ °C = ________ °K

   b) -40° C = ________ °F

   c) +27° F = ________ °C = ________ °K

   d) 323° K = ________ °C = ________ ° F

   e) -10° K = ________ °F

2. Why can one not complete 1e)? Review definition of absolute or Kelvin temperature scale.

3. What are the units of measurement for the following quantities:

   a) velocity: ________

   b) density: ________

   c) energy: ________

   d) force: ________

   e) acceleration: ________

   f) pressure: ________

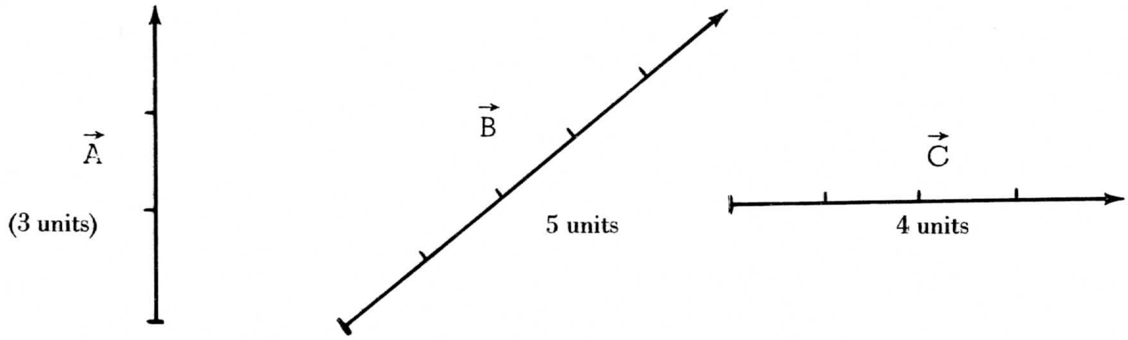
4. Compute the sea level pressure (using definition $p = F/A$) in $dynes/cm^2$ and mb. Given is:

   Density of mercury (Hg): $13.6\ g/cm^3$
   Acceleration of gravity (g): $980\ cm/sec^2$

   and atmosphere is supporting a column of 76 cm of Hg.

5. Why does the temperature increase upward in the stratosphere and why is the highest temperature found at the top of this layer in spite of the fact that most of the ozone is concentrated at levels around 20 km?

6. A group of vectors are given below. Do the requested operations, filling in the blanks with the letter of the appropriate vector



1. $\vec{A} - \vec{A} =$ ________
2. $\vec{A} + \vec{C} =$ ________
3. $\vec{B} - \vec{C} =$ ________
4. $\vec{A} + \vec{A} =$ ________
5. $\vec{B} - \vec{A} =$ ________
6. $\vec{A} + \vec{B} + \vec{C} =$ ________
7. $\vec{A} - \vec{B} =$ ________
8. $\vec{C} - \vec{B} =$ ________
9. $\vec{A} - \vec{B} + \vec{C} =$ ________
10. $\vec{C} - \vec{A} + \vec{B} =$ ________

7. Compute relative humidity U from:

a) $e = 5$ mb, $E = 25$ mb

b) $w = 3.5$ g/kg, $W = 14$ g/kg

c) $w = W = 10$ g/kg

CHAPTER III. HEATING OF THE ATMOSPHERE

(Reading: pp. 20-35, 58-61, 72-73)

A. Solar Energy driving the atmosphere.

HEAT = Energy contained in a body through motions of its molecules (or atoms)

Heat content depends on: Size (No. of molecules)
Temperature (their speed)
Composition (mass of molecules and phase state—solid, liquid, gas)

Specific heat: Amount of heat which has to be added to 1 g to increase its temperature by 1°C.

DEFINE: 1 calorie (cal): heat energy needed to heat 1 g of (pure) H_2O from 14.5°C to 15.5°C.

Forms of Heat
A) sensible heat
B) latent heat
C) heat equivalent of mechan. energy

Transfer of Heat (from warmer to colder body)

by
1) conduction
2) convection
3) radiation

Processes 1) and 2) require matter contact, 3) goes on also in vacuum.

Nearly all energy used on earth comes from sun!

Let solar energy received at top of atmosphere on an annual basis be $1 \cdot 10^9$ units. Comparable annual contributions from other natural sources are:

Heat from Earth interior (conduction and convection)	182,000
Heat (IR) radiation from full moon	28,100
Reflected visible radiation from full moon	11,200
Tidal power from kinetic and potential energy	17,000
All radiation from other stars	45
from man-made sources	
Fossil fuel burning (1970)	32,000
World electric power production	3,000
Hydroelectric power	1,300
Nuclear energy for power generation	20

Some facts about SUN.

A star (gas ball)

Temp. of visible sfc ~ 6,000°K
Temp. in center ~ 20-40 mill. °K

Mass of sun ~ 2×10^{33} g = ~ 300,000 x mass of earth

Diameter 1.4 mill. km = ~ 110 x earth diameter

Distance Sun - Earth (average) ~ 150 mill. km (93 mill. miles)

Light (radiation) from sun takes ~ 8 min to reach earth

(velocity of light 300,000 km/sec)

Energy of Sun generated in interior by Fusion Processes (as in Hydrogen bomb!).

Process: Fusion of 4 H-atoms into 1 He-atom + energy

$[4 \times 1.008 \rightarrow 1 \times 4.003 + \boxed{0.029}\,]$

Mass 0.029 converted into energy according to Einstein's Eq.

$$E = m \cdot c^2$$

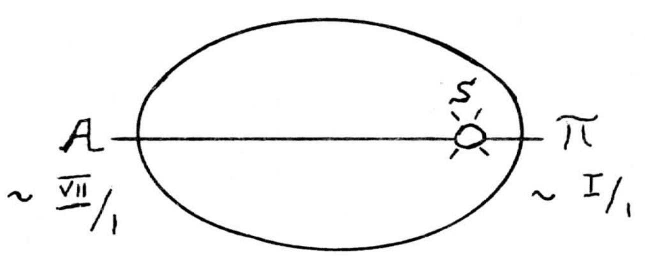
Sun converts each second about 4 mill. tons of H to Energy.
This energy filters outward and eventually is radiated into space. 150 mill. km away, Earth intercepts a tiny amount.

At earth distance, a disk of 1 cm^2 area (vertical to Sun's rays) intercepts amount of energy of

2 cal/min.
(outside of earth atmosphere)

$\boxed{\text{"SOLAR CONSTANT"} = 2\ cal/cm^2/min.}$

Orbit of Earth around the sun is an ellipse (Kepler). In one of the focal points is the sun. When earth closest to sun (perihelion) ~ 147 mill. km
farthest away from sun (aphelion) ~ 151 mill. km.



π PERIHELION

A APHELION

Earth thus receives a small amount more energy at π (~ 7%) than at A.
But this is not reason for our Seasons. (Note that π occurs in midwinter of the northern hemisphere!)

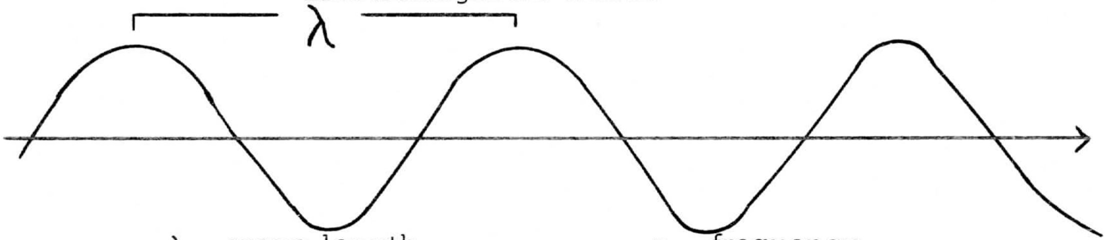
Explanation of seasons.

TILT of earth axis ~ 23 1/2 °

EFFECT of angle of incidence of sun's rays on sfc and of length of daylight.

LATITUDINAL Differences in received radiation energy.

RADIATION Energy Transfer by Electromagnetic Waves



λ = wave length ν = frequency

$\nu = \frac{c}{\lambda}$ or $\lambda = \frac{c}{\nu}$ (c = speed of light
$c = 3.10^5$ km sec.$^{-1}$ = 3.10^8 msec.$^{-1}$)

For heat and light radiation, λ is very small, measured in very small units.

Unit of Wavelength: MICRON $1\,\mu = \frac{1}{1{,}000}$ mm $= 10^{-6}$ m

(other units $1\,\mu\mu = \frac{1}{1{,}000}\mu$; $1\,\mathring{A} = \frac{1}{10}\,\mu\mu$)

Frequency (How many maxima pass by in 1 sec?)

Cycles per sec (cps); kcs = kilocps; mcs = mega-cps

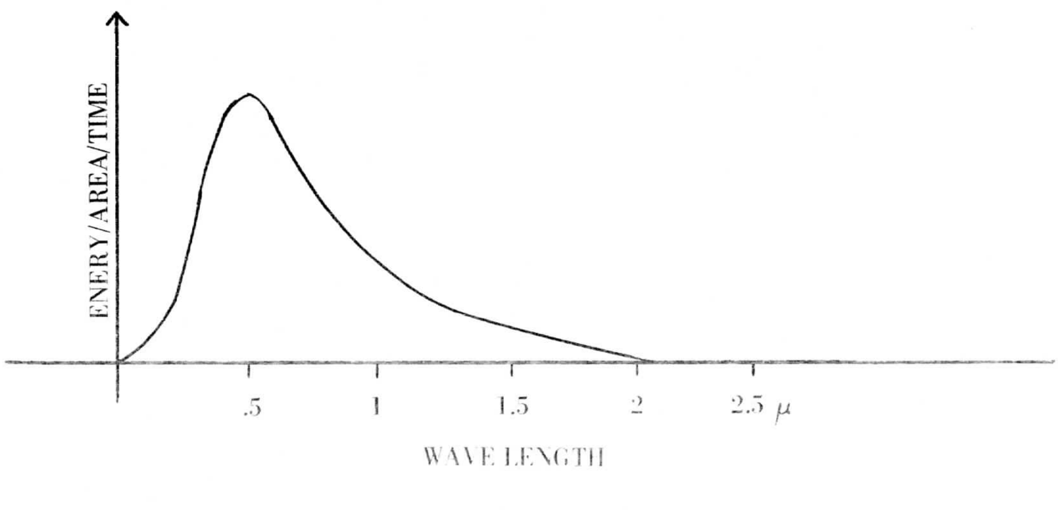
(now Hertz) hz kilohertz (khz) megahertz (mhz)

$= 10^3$ $= 10^6$ hertz

Example: FM-Radio 100 mhz

$$\lambda = \frac{c}{\nu} = \frac{3 \cdot 10^8}{100 \cdot 10^6} = 3 \text{ m wavelength}$$

SPECTRUM: Energy of radiation as function of λ or ν



We call:		Wavelength	
very long	-	cm to km	- Radio, TV
	-	mm	- Microwave
long	-	(1 mm) 1000 - 1 μ	- Infrared (IR)
visible	-	0.7 - 0.4 μ	- (Red - Yellow - Blue)
short	-	0.4 - 0.1 μ	- Ultraviolet (uv)
very short	-	0.01 - 0.1 μ	- EUV
---	-	100 - 1 $\mu\mu$	- X-ray
---		~ 0.001 $\mu\mu$	- γ-ray

LAWS OF RADIATION

1) Basic law (Planck) [1900, beginning of modern physics]
energy emitted at wavelength λ

$$W_\lambda = \frac{c_1 \lambda^{-5}}{(e^{c_2/\lambda T} - 1)}$$

c_2 related to "energy quantum" $h \cdot \nu$

total energy up to λ_0 $\quad W_{(0-\lambda_0)} = \int_0^{\lambda_0} W_\lambda \, d\lambda$

If one intergrates this, one obtains (for total energy, 0 - ∞)

2) Stefan-Boltzmann law:

$R = \sigma T^4$ (T in °K) R = Radiation energy

Example: two bodies $T_1 = 200°K$ $T_2 = 400°K$

$$R_1 = \sigma(200°K)^4 = \sigma \cdot 2^4 (10^2)^4 = \sigma \cdot 2^4 \cdot 10^8$$

$$R_2 = \sigma(400°K)^4 = \sigma \; 4^4 (10^2)^4 = \sigma \cdot 2^8 \cdot 10^8$$

$$\text{Ratio} \quad \frac{R_2}{R_1} = \frac{\sigma \cdot 2^8 \cdot 10^8}{\sigma \cdot 2^4 \cdot 10^8} = 2^4 = 16$$

"T doubles, R increases 16 (= 2^4) times."

3) Wien law (or displacement law)

$\lambda_{max} \cdot T$ = constant ($\cong$ 3,000, if T: °K, λ : μ)

λ_{max}: wavelength of maximum energy output.

Example: T = 300°K $\quad \lambda_{max} = 10\ \mu$ (i.e. IR)

T = 6,000°K $\quad \lambda_{max} = 0.5\ \mu$ (i.e. visible)

INTERACTION of Radiation and Matter.

1) From above laws evident that any matter with T > 0°k EMITS radiation.

2) All matter also ABSORBS radiation (may convert into other energy form, e.g. heat).
ALL bodies also will REFLECT radiation or SCATTER it.
Finally, a body may TRANSMIT (let through) some radiation.

ALBEDO: Ratio of $\frac{\text{REFLECTED RADIATION}}{\text{TOTAL RECEIVED}}$

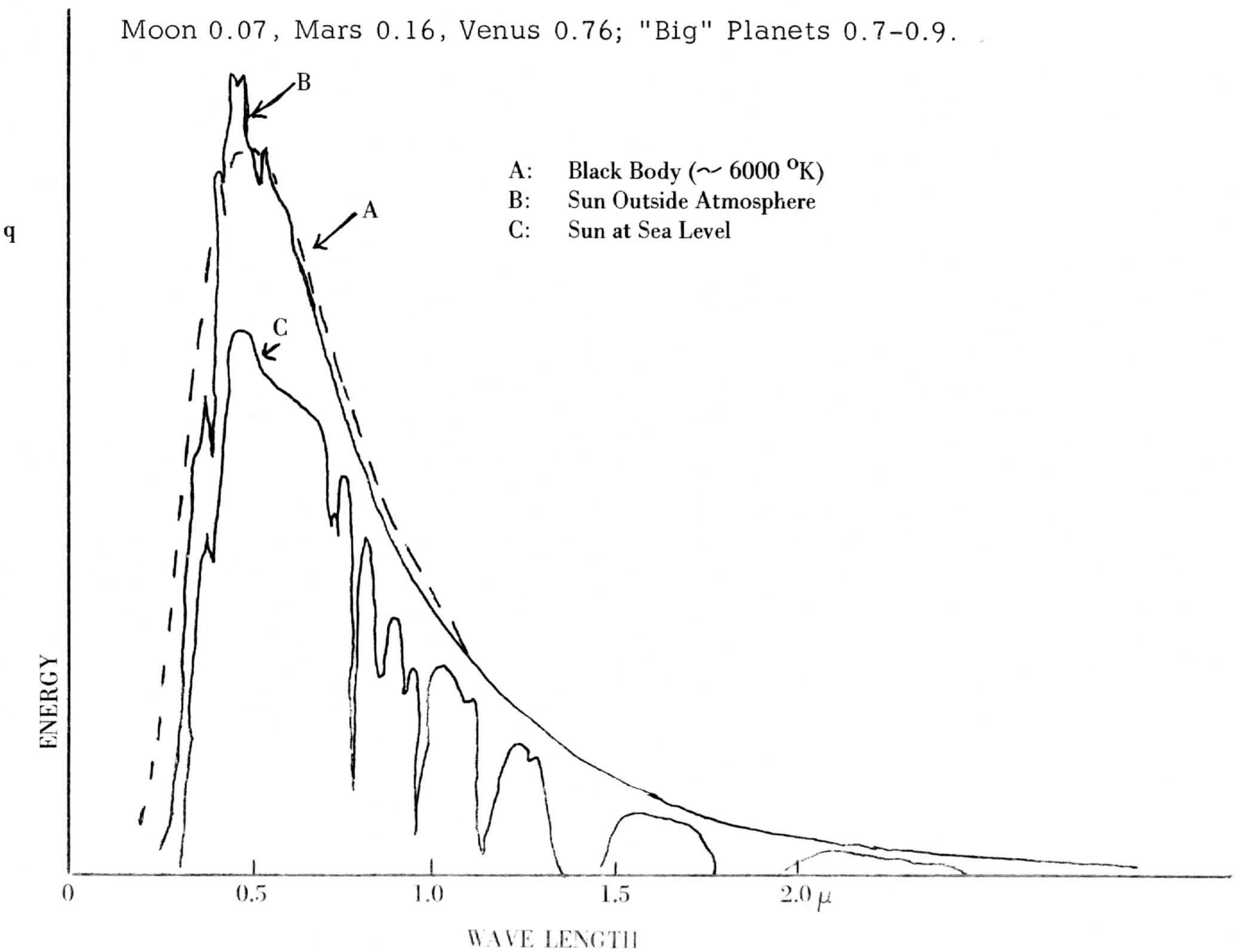
ALBEDO for natural bodies dependent on wave length.

E.g. for snow $A_V \sim 0.9$ $A_{IR} \sim 0.1$

Some typical albedo values: (for visible light, e.g. seen from space)

Oceans	about	0.06 [very black!]	
Sand, Soils		0.05 (lava) - 0.45 (desert sands)	
Forests		0.1 - 0.3	
Thin clouds		0.4 - 0.6	
Thick clouds		0.6 - 0.9	[very white]
Snow		0.8 - 0.95	
Planet Earth		0.30(±.02)	

Moon 0.07, Mars 0.16, Venus 0.76; "Big" Planets 0.7-0.9.



SOLAR ENERGY AVAILABLE

A) Outside atmosphere - SOLAR CONSTANT S

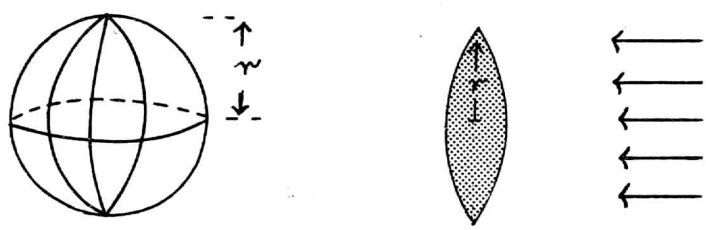
"Amount of energy intercepted by 1 cm^2, perpendicular to Sun's rays, at mean distance Sun/Earth, in 1 min."

$S = 2$ cal/cm^2/min.

Depends only on amount of energy radiated by sun and distance of the cm^2 (planet, space ship etc.) from sun. Assuming constant energy output from sun, S decreases proportional to the square of distance. (Inverse square Law). For example: Jupiter is approx. 5 times as far away from sun than Earth. The solar constant for Jupiter then will be

$$S\ (\text{Jup.}) = \frac{1}{5^2}\ S\ (\text{Earth}) = \frac{1}{25} \cdot 2 = 0.08\ \text{cal/cm}^2/\text{min.}$$

Result: The farther planet from sun, the colder it will be!
Average amount for 1 cm^2 of earth:



sfc area (sphere) $4\pi r^2$

Area (disc) πr^2

Total amount of solar radiation received by earth in 1 min

$$R_1 = \pi r^2 \cdot S \quad (r = \text{Earth radius in cm})$$

Distributed evenly over sphere, each cm^2 receives

$$R_2 = \frac{\pi r^2 S}{4\pi r^2} = \frac{S}{4} = 0.5\ \text{cal/cm}^2/\text{min.}$$

However, part of this amount is reflected back immediately. Planet Earth Albedo 0.30; i.e. 30% reflected, 70% retained. Actual available (useable energy thus

0.35 cal/cm^2/min.

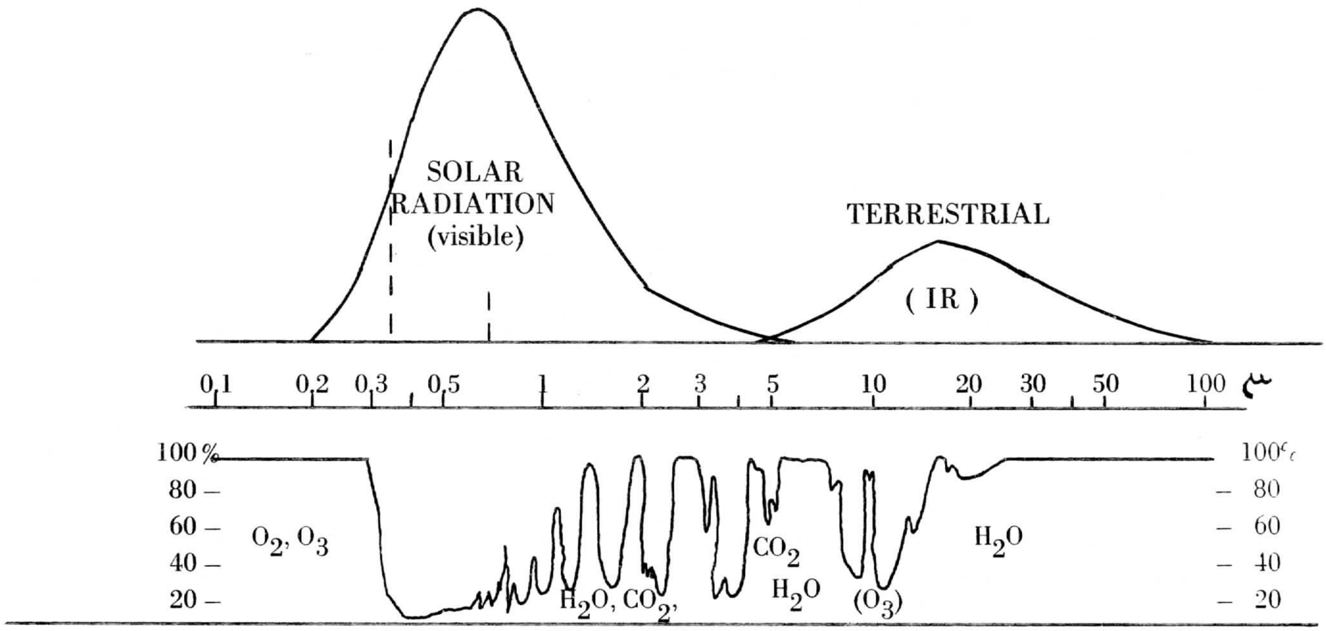
Total amount for earth very large (surface $\sim 5.1 \cdot 10^{18}$ cm^2)

Each day, Earth retains $\sim 3 \cdot 10^{21}$ cal; each year $\sim 10^{24}$ cal.

But: if one would use this energy to heat the water in the oceans ($\sim 10^{24}$ g), temperature of oceans would only increase by ~ 0.003°C/day.

B) USE OF ENERGY IN EARTH/ATMOSPHERE SYSTEM

Of energy not immediately rejected ("albedo"), part is absorbed in atmosphere, part will penetrate to surface.



ABSORPTION IN ATMOSPHERIC GASES

Most (~43% out of 70%) of incoming solar energy (sunlight, λ_{max} ~0.5μ) reaches ground, heats surface. However, much of outgoing energy (IR, λ_{max} ~12μ) from surface is absorbed in atmosphere then re-radiated from atmosphere to space. See Figure.

a) In visible light (0.3-0.65μ) atmosphere is quite transparent, lets light-energy in.

b) In IR, H_2O vapor is nearly opaque, especially in regions 3-7μ, 14-15μ and >22μ.

c) In IR, CO_2 absorbs strongly in wavelengths >1μ in various regions.

d) O_3 has a strong absorption band near 10μ, in addition to its absorption of UV below 0.3μ.

e) Clouds in atmosphere absorb and reflect IR from ground, prevent escape of IR to space.

Thus: Solar radiation can get through atmosphere but IR hardly can get out - keeps earth warm like a blanket or window panes of a greenhouse:

GREENHOUSE EFFECT

Major contributors to it: H_2O, CO_2.
CO_2 evenly mixed in atmosphere, its contribution the same everywhere. However, H_2O varies widely from place to place. Thus greenhouse effect changes chiefly due to changes in H_2O vapor (and cloudiness).

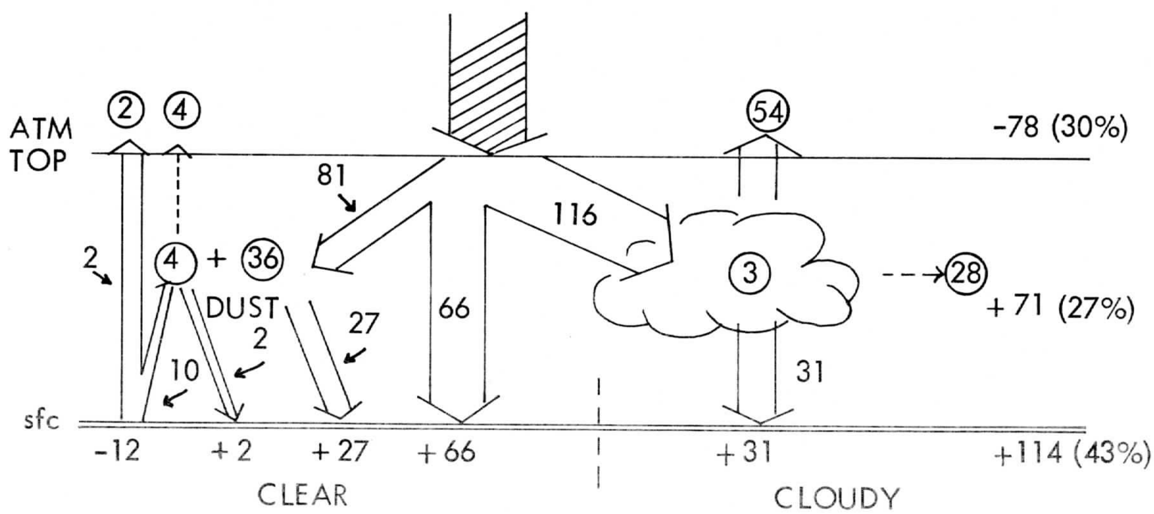
Result: Warmest where most water vapor in atmosphere (other conditions equal) - compare nights in Florida vs. Arizona. Cloudy nights are warmer than clear nights.

Initial conclusion: Most incoming energy goes to surface. Most outgoing energy goes to space from atmosphere.

C) GLOBAL HEAT BUDGET (PLANET EARTH)

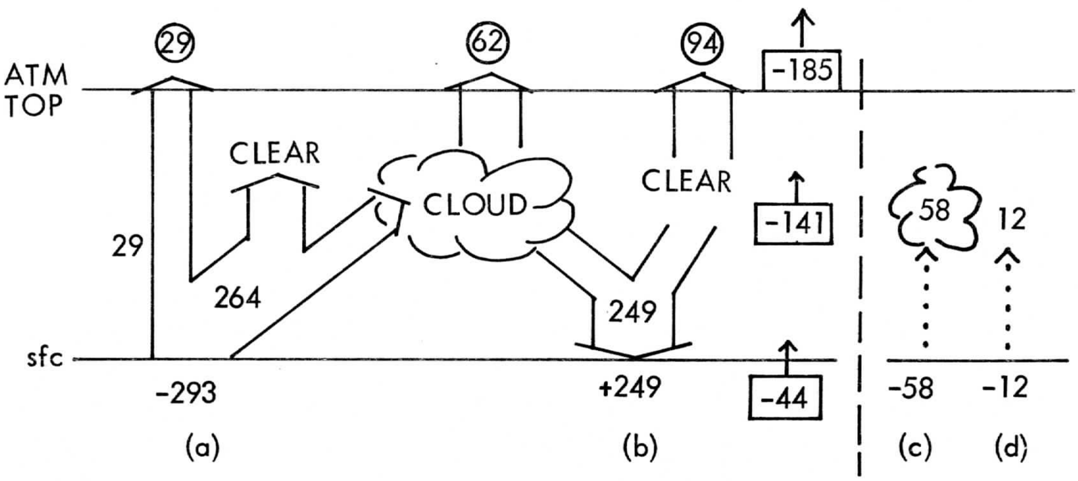
Basic fact: Incoming energy nearly perfectly balanced by outgoing energy (when averaging over, say, year or decade). Only minute amounts stored or released in chemical processes.
Examples: Plants use energy (photosynthesis) to grow. Coal, oil, peat are stored energy from earlier times. Now we burn this, release energy again.
Heat budget: Account of where energy goes, how various parts of system retain or transmit energy etc. Usually obtained over a year (recall elliptic earth orbit!)

1. Incoming (solar) radiation (in k cal/cm^2/year).
   Total amount outside atmosphere (per cm^2)
   0.5 cal/min·1440 min/day·365 days~263,000 cal/yr = 263 kcal/yr.



Retained in earth-atmosphere (263 - 78) = 185.
Of this useable energy, 71 (38% of 185) absorbed in atmosphere.
114 (62% or 185) absorbed at surface.

2. Outgoing (IR) radiation from planet (same units)



IR RADIATION

NON-RADIATIVE PROCESSES.

Radiation loss from surface 44 (24%) (a+b)
from atmosphere 141 (76%)

(a) IR radiation from surface to atmosphere and space (based on T_{sfc})
(b) Back radiation/reflection of IR from atmosphere to surface.

3. Total Heat Budget (1 and 2 together)
Net gain at surface from solar radiation 114 = 62% of energy retained.
Net loss from surface by IR, however, only 44 = 24%.
In order to achieve balance, 70 (= 38% of energy) has to be transported from surface to atmosphere upward by other than radiation processes.
These other processes are: (see Figure: Non-radiative processes) transport of latent heat and transport of sensible heat.

(c) Latent heat transport: evaporation of sfc, condensation in atm.: 58 units;

(d) Sensible heat transport: conduction, convection from sfc to atm.: 12 units.

c + d = 70 units; of this 58/70 ~ 83% latent heat (E)
12/70 ~ 17% sensible heat (H)

Ratio H/E = Bowen ratio = 0.21 (whole earth)

Earth-atmosphere works essentially like a steam engine: water evaporation in boiler (sfc), condensed in condenser (atmosphere) Work performed in this engine results in weather!

D. LATITUDINAL CHANGES OF GLOBAL HEAT BUDGET TERMS

Because the earth is a sphere, more incoming energy reaches low latitudes than region near poles. Amount of energy lost also decreases from equator to pole but more slowly (depends on sfc temperature).

Lat. Zone (degrees)	Area (% of earth)	Absorbed (solar, $cal/cm^2/min$)	Emitted (IR, $cal/cm^2/min$)	Difference
0 - 20	34	.45	.36	+ .09
20 - 40	30	.40	.37	+ .03
40 - 60	22	.26	.33	- .07
60 - 90	14	.15	.31	- .16

Equatorwards of 40°, surplus (more in, less out)
Polewards of 40°, deficit (less in, more out)
Therefore - transport of energy required from Eq. to Pole.

Across Lat. 20° -- $2.5 \cdot 10^{19}$ kcal/year

30° -- 3.6·

40° -- 3.9· ← Maximum!

50° -- 3.4·

60° -- 2.4·

70° -- 1.2·

This transport not exclusively in atmosphere but also in oceans. Current estimate (still relatively inaccurate!):

By a) Sensible heat exchange ~50%
(warm air poleward, cold air equatorward)

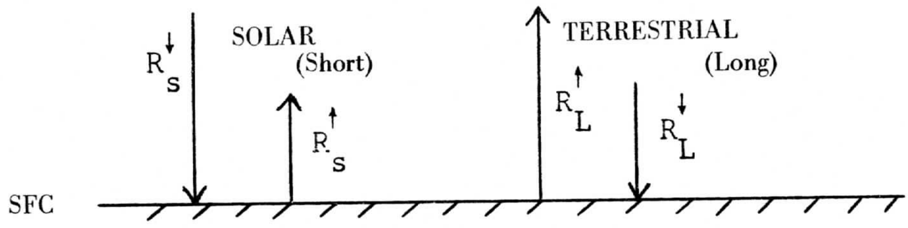
b) Latent heat ~25 - 30%
(water vapor moved by air motions as in a))

c) Ocean currents ~20 - 25%
(warm currents toward pole etc.)

(Example: Gulf stream warm, Labrador current cold)

E. LOCAL HEAT BUDGET

As for the earth as a whole one also can establish the heat budget for a particular piece of the earth surface. Can be either solid or liquid (land or water).



$$R_{net} = (R_s^{\downarrow} - R_s^{\uparrow}) + (R_L^{\downarrow} - R_L^{\uparrow})$$

$R_s^{\downarrow}$: Incoming solar radiation reaching surface.

$R_s^{\uparrow}$: Reflected solar radiation from surface.

(Solar = Shortwave)

$R_L^{\uparrow}$: Outgoing IR radiation from surface.

$R_L^{\downarrow}$: Back radiation from atmosphere to surface.

(IR = Longwave)

R_{net}: Net radiation (positive = gain, negative = loss)

Two extreme cases:

a) At night: $R_s^{\downarrow} = R_s^{\uparrow} \equiv 0$ (no solar radiation)

then $R_{net} = (R_L^{\downarrow} - R_L^{\uparrow}) < 0$ (net loss)

Surface loses energy by radiation.

b) Daytime, near noon (when $R_s^{\downarrow}$ at maximum):

then $(R_s^{\downarrow} + R_L^{\downarrow}) > (R_L^{\uparrow} + R_s^{\uparrow})$, i.e. $R_{net} > 0$ (net gain).

Surface gains energy from radiation.

When energy (heat) is gained from radiation at surface it must be transported away (sfc cannot store heat). Similarly, when heat is lost from sfc, it must be transported to sfc from somewhere.

In both cases, transports go to or come from:

soil or water below or air above.

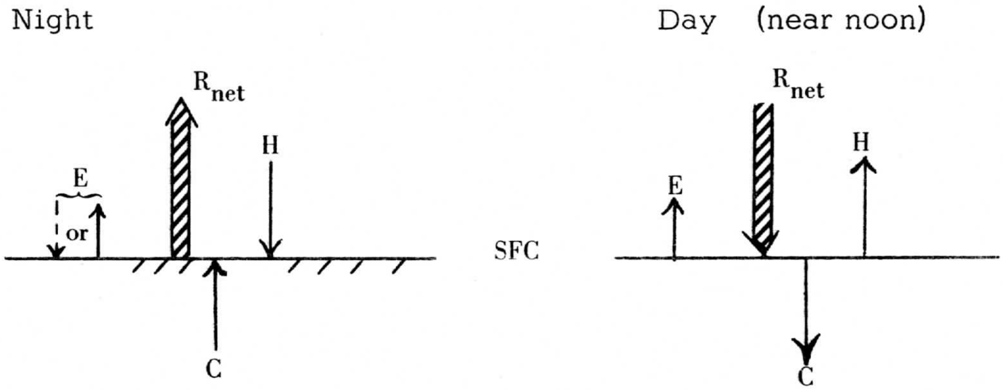
At any time, these transported amounts must equal the amount of heat required by radiation balance. R_{net}.
Balance commonly written as:

$$R_{net} = C + E + H$$

C: Conduction into/out of subsurface (and convection in water)

E: Latent heat (evaporation or condensation)

H: Heat transport into/out of air (conduction, convection)



Discussion of these 3 terms:

1) C: If C to soil, T_{soil} will rise.

<u>DIURNAL</u> VARIATION OF SOIL TEMPERATURE (from observations in a desert in Southern Russia) (Daily Extremes)

Desert Soil at	Time of Max. T	T_m(°C)	T_n(°C)	ΔT(°C)
sfc	1:15 p	32	5	27
depth of 5 cm	4:30 p	19	9	10
10 cm	5:30 p	18	10	8
20 cm	8:15 p	16	12	4
40 cm	3:55 a	14.7	14.1	0.6

<u>ANNUAL</u> VARIATION OF MEAN SOIL TEMPERATURE

at:	Coldest Month		Warmest Month		ΔT
air (+2 m)	- 0.4°C	(I)	24.4°C	(VIII)	24.8
at sfc	+ 0.6°C	(I)	33.1°C	(VIII)	32.5
depth of 0.2 m	+ 1.6	(I)	30.7	(VIII)	29.1
0.4	+ 2.9	(I)	28.9	(VIII)	26.0
0.84	+ 5.5	(II)	26.2	(VIII)	20.7
1.65	+ 8.3	(II)	22.2	(VIII)	13.9
3.26	+11.9	(IV)	17.7	(X)	5.8
3.99	+12.5	(IV)	16.5	(X)	4.0
6.47	+13.8	(VI/VII)	15.3	(XII)	1.5

below ~ 12-15 m $\Delta T < 0.1°C$.

In <u>Oceans</u>, C reaches deeper (convection added)

<u>Diurnal</u> variations ~ 1/2°C
<u>Annual</u> near Equator ~ 2°C
at ~ 50°N ~ 5°C

More in shallow waters near coasts.

2) E: Depends on moisture at sfc and in soil.
E large over oceans (<u>free</u> sfc), increased by wind.
E less over land (~ 1/3). Least in deserts.

3) H: usually larger over land.
Ratio H/E (Bowen ratio) over oceans ~ 0.1, land ~ 1
[over Australia (desert) ~ 2.1]

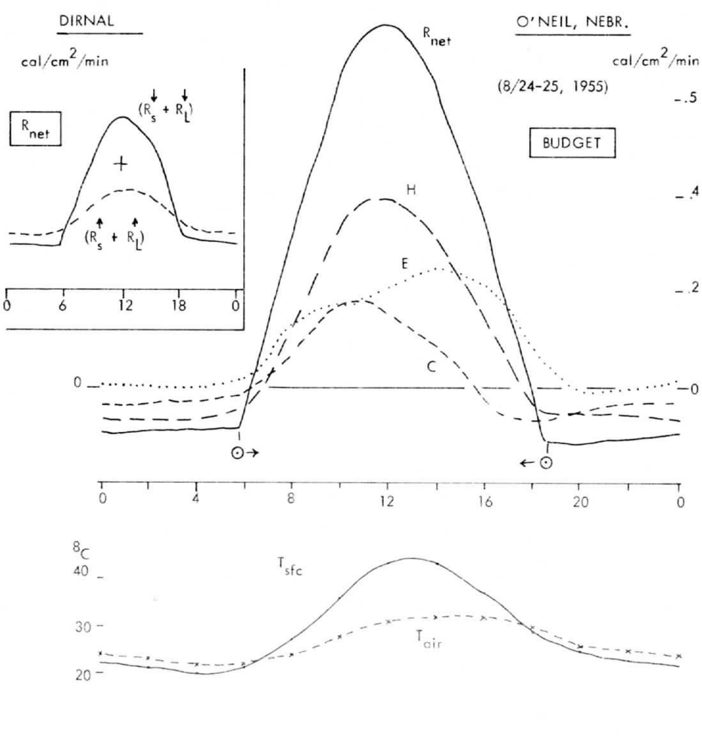
Effect of heat budget equation is diurnal/annual variation of temperature in air near sfc (i.e. what we normally experience and measure).

Result of R_{net} and its distribution by C, H and E is ultimately the diurnal and annual variation of temperature in the air near the ground (Shelter temperature), i.e., what we usually measure and experience.

As examples are given, in the next two figures, the diurnal and the annual heat budget as calculated from observations. Such calculations go far to quantitatively understand the physical mechanisms which determine the climate of a location and ultimately of Earth as a whole. Called Climat<u>onomy</u> (Lettau).

<u>a) Diurnal radiation and total heat budget</u>

(O'Neill, Nebr. - based on observations 8/24-25, 1953.)



Typical values (cal/cm^2/min)

1. Radiation balance terms.

At 0^h (midnight): $R_s^{\downarrow} = 0$ $\quad R_s^{\uparrow} = 0$ $\quad R_L^{\uparrow} = 0.591$; $R_L^{\downarrow} = 0.495$

12^h (noon) : $R_s^{\downarrow} = 1.280$; $R_s^{\uparrow} = 0.256$; $R_L^{\uparrow} = 0.778$; $R_L^{\downarrow} = 0.509$

2. R_{net} and non-radiative heat balance terms.

At 0^h : $R_{net} = -0.096$; $C = -0.036$; $E = +0.008$; $H = -0.068$

12^h : $R_{net} = +0.766$; $C = +0.153$; $E = +0.211$; $H = +0.391$

Since this calculation was using August data from Nebraska, there is a considerable surplus in R_{net} (Midsummer, sun very high in sky). More than half of the available energy goes, during the day, into the air (H), while C becomes negative already in midafternoon (heat from soil to surface). Evaporation carries heat also away from surface during most of the day except for short time in late evening where dew was forming (E<O)

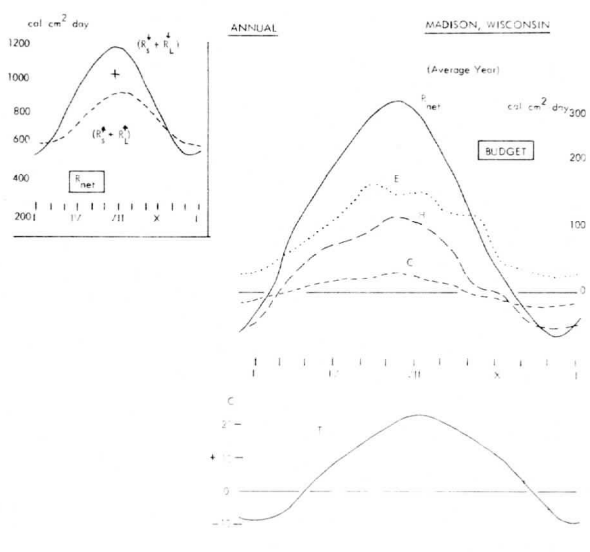
T_{sfc} is temperature <u>at</u> soil surface (radiative interface).

T_{air} is measured at screen height (air temperature at 2 m).

T range is considerably higher <u>at</u> sfc (23°C) than in air 2 m above sfc (10°C). Maximum: at sfc $12\text{-}14^h$, in air $14\text{-}16^h$ (~2 hrs later).

b) Annual radiation and total heat budget

(Madison, Wisc. - Average Year)



Typical values (cal/cm^2/day!)
[To convert to cal/cm^2/min; divide by 1,440 min/day; e.g. 720 cal/cm^2/day $\equiv$ 0.5 cal/cm^2/min]

1. Radiation balance terms

In January: $R_s^{\downarrow} = 156$; $R_s^{\uparrow} = 60$; $R_L^{\uparrow} = 545$; $R_L^{\downarrow} = 405$

July : $R_s^{\downarrow} = 520$; $R_s^{\uparrow} = 83$; $R_L^{\uparrow} = 838$; $R_L^{\downarrow} = 669$

2. R_{net} and non-radiative heat balance terms.

In January: $R_{net} = -44$; $C = -19$; $E = +25$; $H = -50$

July : $R_{net} = +268$; $C = +19$; $E = +149$; $H = +100$

At Madison, E is the largest contributor for transporting heat away from sfc throughout the year. In winter (XI - II) both air (H) and soil (C) supply heat to sfc. Plotted temperature is screen T, very small difference to T_{sfc}.

Note that both $R_L^{\uparrow}$ and $R_L^{\downarrow}$ increase toward summer due to increase of T_{sfc} (= radiating surface)

General Conclusions:
Both diurnal and annual heat budgets depend on location and physical factors such as surface albedo and availability of water for evaporation.

Over land, less evaporation than over oceans - thus more of R_{net} goes to H, C.

Over deserts, E very small, thus high H, C - results are large diurnal and annual variations.

Over tropical oceans, E very large, rest chiefly into C (mixing in deep water), results are very small diurnal and annual variations.

Cloudiness and/or high water vapor in atmosphere increases $R_L^{\downarrow}$, decreases thus R_{net} and moderates diurnal and annual variations. Dry and cold air (little H_20 vapor!) acts in opposite way.

Additionally (especially on annual budget) one has to take account of horizontal transport ("advection") by the atmosphere and oceans.

Near oceans, climate is moderated, becomes more oceanic; far away from oceans, climate is more continental. In Mid-latitudes (with prevailing west winds) West coasts are warmer than East coasts (Seattle vs. Boston).

Ocean current also very important: London nearly 10° farther north than Boston but much milder.

Far inside continents, no moderating effects of oceans left - both diurnal and annual variations increase.

Largest annual variation inside (in Eastern part) of largest continental landmass on Earth - Eurasia.

In Northern Siberia: Verkhoyansk ($67\frac{1}{2}$°N, $133\frac{1}{2}$°E)

January $T = -46.8°C$, July $T = +15.7°C$;

Annual variation, $T = 62.5°C = 112.5°F$!

Even at South pole, ΔT is not as large.

December (summer!) $T = -25°C$, July (winter) $\Delta T = -61°$, T

$\Delta T = 36°C \sim 65°F$.

At Madison, $\Delta T = 30.6°C = 55.1°F$, less than half of that in Northern Siberia; it's not so bad here, after all!

EXERCISES—Chapter III

1. How many calories are required to bring 50 g of water, initially at 50° F, to the boiling point (i.e. 100°C)?

2. List the planets of the solar system, in order of increasing distance from the sun.

3. The solar constant (at the mean distance of earth from the sun) is given as 2 cal/cm^2/min. What is the value of this constant for Mars which is approximately 1.5 times as far away from the sun than the earth? What is it for a point only half as far away as earth?

4. Latent heat is either stored or released when the chemical compound H_2O undergoes a phase change. When (i.e. during which transition) is latent heat being stored and when is it released?

5. The distance of earth from the sun is 150 mill. km. Convert this distance into miles (1 mile = 1.6 km).

6. Using the proper radiation law, compute:
   a) The wavelength of maximum energy for a body which has a temperature of 6,000° k.
   b) The λ_m for T = 12,000° k.

7. A spaceship arrives (maybe some centuries from now!) at a star and takes up an orbit around it which has the same distance from that star as has earth from the sun. The astronauts measure the "solar constant" and find it to be 32 cal/cm^2/min.
   a) Can you compute the surface temperature of this star?
   b) How far outward would they have to move their ship to find the same solar constant as we have for earth?
   c) What would be the wavelength of maximum energy output for this star? In what region of the spectrum is this wavelength situated?

8. A rotating spherical artificial satellite (diameter 1 meter) has an albedo of 0.5.
   a) How many cal. would be absorbed by one cm^2 of its surface on the average.
   b) If I increase the diameter of the satellite to 10 meters, what is the change in the amount of energy received per cm^2?

9. Assume that the earth axis was exactly vertical to the center orbit. What would be the result for the climate of, say, Madison? When would it be winter, when summer—and would these differences be larger or smaller than they are now? Would you be able to see the sun when standing at the North Pole?

10. Plot, on graph paper, the ΔT-values of the table on p. III-10 as a function of height or depth. Why is the maximum where it is? Also plot the average temperature (coldest plus warmest month/2). How does this quantity behave, as compared to the individual temperature values?

CHAPTER IV. THERMODYNAMICS OF THE ATMOSPHERE

(Reading: pp. 35-41, 168-173)

A. Adiabatic Processes

1) Variables for describing state of a gas

p	pressure	[mb]
T	temperature	[°K]
v	volume	[cm^3]
or ρ	density	[= mass/volume, g cm^{-3}]

2) Gas Laws (Equation of State)

a) Boyle (~ 1660) pv = constant, if T held constant.

b) Charles (~ 1780) T proportional to v , if p = constant.

3) Together—Eq. of State: a) $pv = MRT$ (Gas constant R = 2870)

if p in mb, v in cm^3, T in °K
M in g

b) $p = \rho RT$

c) $p\alpha = RT$ $\alpha = \frac{1}{\rho}$, specific volume

(Main use of 3) Substitution of variables for others which may be easier to measure.

4) With previous definition of 1 cal (see Chapter III A)

(General) Specific heat [No. of cal required to heat 1 g of substance by 1°C]

Defined for 1 g: $C = (\frac{\Delta h}{\Delta T})$

$C_{(water)} = 1$

$C_{(silver)} \sim \frac{1}{10}$ [most substances, C < 1]

Dealing with gases, it is more practical to use volume.

$\Delta h = M \cdot C \cdot \Delta T$ (for M g of matter)

With $\rho = \frac{M}{v}$: $\Delta h = \rho v C \Delta T$

for $\alpha = \frac{1}{\rho}$ (i.e. v = 1, unit volume)

Heat capacity $\frac{\Delta h}{\Delta T} = \rho C$ [for gases]

5) Specific heat at constant volume: $c_v = [\frac{\Delta h}{\Delta T}]_{v = \text{const.}}$

at constant pressure: $c_p = [\frac{\Delta h}{\Delta T}]_{p = \text{const.}}$

$c_p = c_v + R$

Experiment of R. Mayer (1842)

a) Piston fixed - V = const.
For $\Delta T = 1°C$, $\Delta h = 0.17$ cal/g of air

b) Piston loose - p = const. [equal to $p_{outside}$]

For $\Delta T = 1°C$. $\Delta h = 0.24$ cal/g of air.

Difference is equal to energy required to move piston upward against outside pressure. (.07 cal)

6) First Law of Thermodynamics

(General form of equation)

a) $\Delta h = C_v \Delta T + p \Delta \alpha$

b) $\Delta h = C_p \Delta T - \alpha \Delta p$ ← often used.

7) Hydrostatic Equation [for atmosphere]

$\Delta p = - g \rho \Delta z$ (Δz: change of height)

8) Adiabatic Process $\underline{\Delta h = 0}$

i.e. no change in heat energy during process.
(specifically by adding heat <u>from the outside</u> to gas volume under study)

Opposite process called <u>diabatic</u>
(e.g. gas volume in contact with hotter/colder surface; then transfer of heat energy by conduction into/out of volume. Also possible heating by absorption of radiation or cooling by radiation loss.)

From Eq. 6 b) $0 = c_p \Delta T - \alpha \Delta p$

follows $\underline{c_p \Delta T = \alpha \Delta p}$

This means: if $\Delta p < 0$ (pressure decreases) then $T < 0$ also!

9. In atmosphere, "gas" means the usual gases O_2, N_2 etc.

<u>and</u> some water vapor.

a) As long as water vapor does not condense ($U < 100\%$!) no latent heat is released and energy content of air volume does not change:

<u>Dry-adiabatic</u> process.

b) When water vapor starts condensing ($U = 100\%$), then latent heat is released and energy content of air volume (in the form of sensible heat) increases - but <u>was not added from the outside!</u> Then we call this:

<u>Moist-adiabatic</u> process.

10) Lapse Rate.

Defined as <u>decrease</u> of temperature with height increase, usually given in °C per 100 m. Symbol γ

$$\gamma = -\frac{\Delta T}{\Delta z} \quad \left(\frac{°C}{100m}\right)$$

a) Actual Lapse Rate.

As observed in the atmosphere, averaged over a specific height interval = layer.

1. Example: At height of 1,000m, T = +10°C

At height of 2,000m, T = + 2°C

Decrease is +8°C over 1,000m height difference, i.e. lapse rate $\gamma = -\left(\frac{-8}{1,000}\right) = +0.8°C/100m$

Note that γ is positive when T decreases with height. This is the normal state of affairs in the troposphere (see page II-3).

2. Example: At 1,000m, T = +10°C
at 2,000m, T = +12°C

Decrease is −2°C per 1,000m, $\gamma = -0.2°C/100m$. This behavior is opposite to the normal way, called Inversion.

b) Theoretical for adiabatic processes.
From First law of thermodynamics (6b) one can calculate by how much T would change in an adiabatic process if an air parcel would move upward or downward in the atmosphere - which is equivalent to a change in pressure.

1. For a dry-adiabatic process:

We had $C_p \Delta T = \alpha \Delta p$

Using definition of $\alpha = \frac{1}{\rho}$ and hydrostatic equation (7.)

$$\Delta p = -g \cdot \rho \cdot \Delta z$$

we can obtain

$$C_p \Delta T = \frac{1}{\rho} -g \cdot \rho \cdot \Delta z$$

which leads to

$$\Delta T = -\frac{g}{C_p} \Delta z ;$$

$$\gamma_d = -\left(\frac{\Delta T}{\Delta z}\right) = +\frac{g}{C_p} \cong 1°C/100\text{ m}$$

2. Moist-adiabatic process:
Here we have to use the full equation 6b)

$$C_p \Delta T = \alpha \Delta p + \Delta h_\ell$$

where Δh_ℓ now stands for the internally released latent heat of condensation. As above we can arrive at

$$\Delta T = -\frac{g}{C_p} \Delta z + \Delta h_\ell$$; allows to obtain γ_s

This says that the change in temperature ("decrease") is smaller than in the dry adiabatic case. The difference is due to the release of the latent heat which is used to heat the parcel of air (add energy). Δh_ℓ depends on amount of water vapor condensed which in turn depends itself on the temperature of the air. Warm air can (and usually does) contain more water vapor, thus more can be condensed during lifting of parcel. The difference $\gamma_d - \gamma_s$ will be larger for warm air than for cold air. Values of γ_s can be calculated knowing T, p of air (which determine H_2O saturation water vapor pressure or any other measure of H_2O content). In lower part of troposphere and at "usual" temperature ($\sim$+10°C) a reasonable value for lapse rate is

$$\underline{\gamma_s \sim 0.6 \text{ to } 0.7°\text{C}/100 \text{ m}}\ .$$

11) Vertical Motions in Atmosphere

Two kinds: a) Relatively fast motions of individual "air parcels" upward—turbulence, convection—imbedded in actual atmosphere (speeds in m/sec); b) Large widespread up—or down motion of large layers of atmosphere as a whole, which are generally very slow (speeds in cm/sec)

Example for a) Development of "vertically growing" clouds (cumulus) and of "thermals"; air bubbles over locally heated surfaces (sand heated by sun)

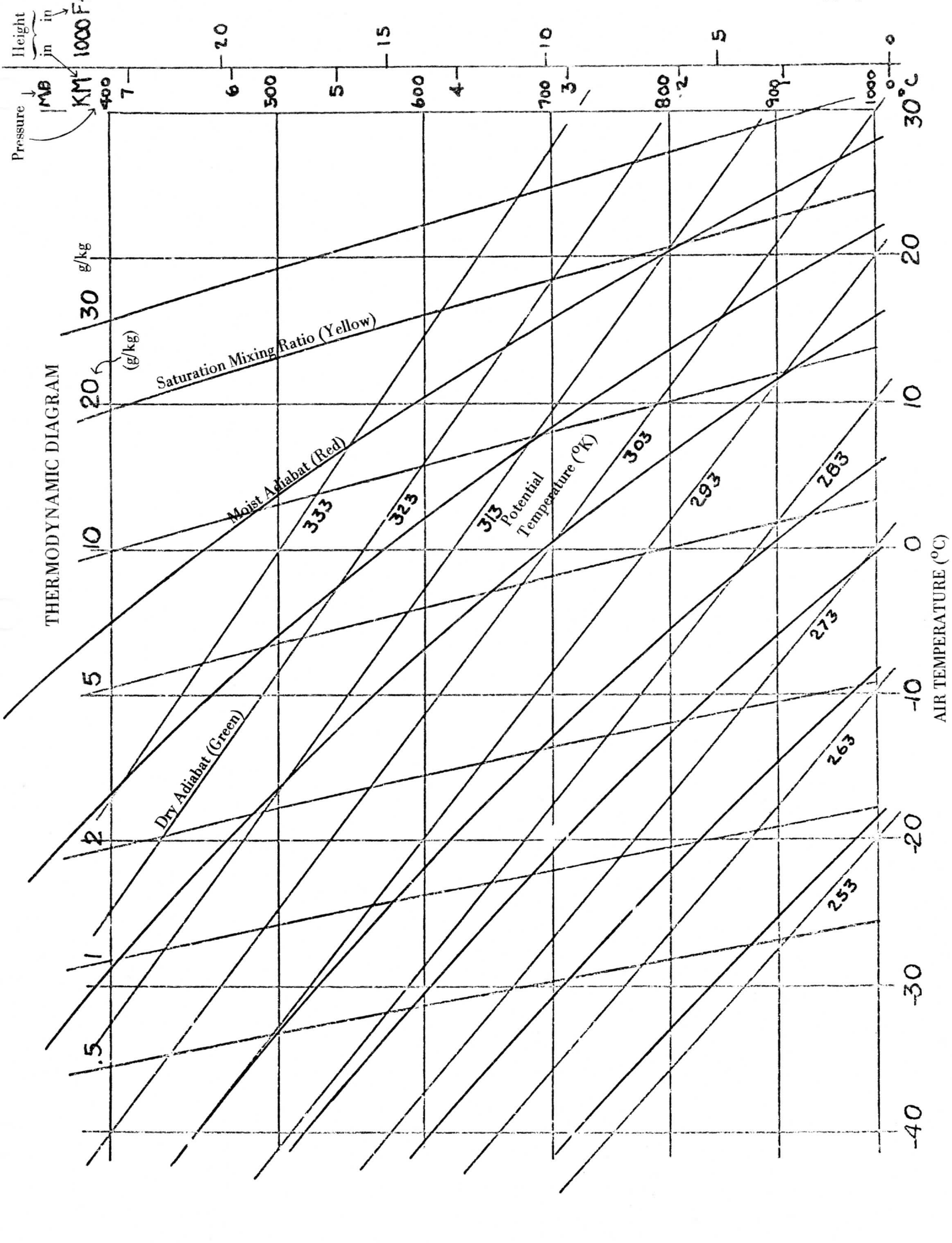
for b) Movement of warmer air mass (lighter air) over colder (heavier) air near surface; dynamic downward sinking of large upper air masses.

Especially case a (where motions are fast and usually rather localized) results in nearly adiabatic processes (motion is too fast to allow exchange of heat between moving parcel of air and surrounding air encountered during rise or fall).

In case b one can often also assume near-adiabatic behavior because all air is moving together up or down—heat cannot be dispersed. Thus—adiabatic processes important in understanding thermodynamic behavior of air in motion.

With equations above, one can investigate this behavior. However, rather than calculate every time the necessary values, meteorologists have developed various methods of solving them graphically by means of different kinds of charts and nomograms.

The various diagrams have been developed chiefly to allow solutions for a wide variety of problems. Here the simplest form of diagram is being used relating T (°C) to p (mb) and using as measure of H_2O vapor content the mixing ratio w (g/kg). The diagrams on the following pages are a simplified version of the diagram in use at weather stations in the U.S.


THERMODYNAMIC DIAGRAM
Pressure
MB
Height
in KM
in 1000 Ft.
g/kg
(g/kg)
Saturation Mixing Ratio (Yellow)
Moist Adiabat (Red)
Dry Adiabat (Green)
Potential Temperature (°K)
333
323
313
303
293
283
273
263
253
AIR TEMPERATURE (°C)

USE OF THIS DIAGRAM

I. Graphically investigate adiabatic processes.

II. Graphically represent measured actual conditions in the atmosphere (p, T, U at various levels) at given time at certain location.

III. Draw conclusions on future behavior of atmosphere (= prediction) combining I and II.

I. Graphical investigation of adiabatic processes in atmosphere. (see p. IV-8)

1) Dry adiabatic process.

From A (p = 900 mb, T = -10°C) dry air is lifted to A_1 (p = 700 mb). Graph shows that T = -29°C (along dry adiabat)

Note that air cannot contain more than 0.5 g/kg of water vapor since W at A_1 is 0.5 and we specified dry adiabatic process (i.e. air cannot reach saturation)

2) Moist adiabatic process.

From B_1 (p = 880 mb, T = -1°C) saturated air is lifted to B_3 (p = 590 mb, T = -22°C) along moist adiabat.

[Compare: If air had been dry, process would have followed dry adiabat, then B_3 would be at (590 mb, -30°C)]. Warming of 8°C is due to release of latent heat.

Note: Moist adiabatic process only for saturated air.

3) Combination of both processes.

Air initially at B (1000 mb, +10°C) but not saturated. (here U = 50%) Therefore, when forceably lifted, follows dry adiabat.

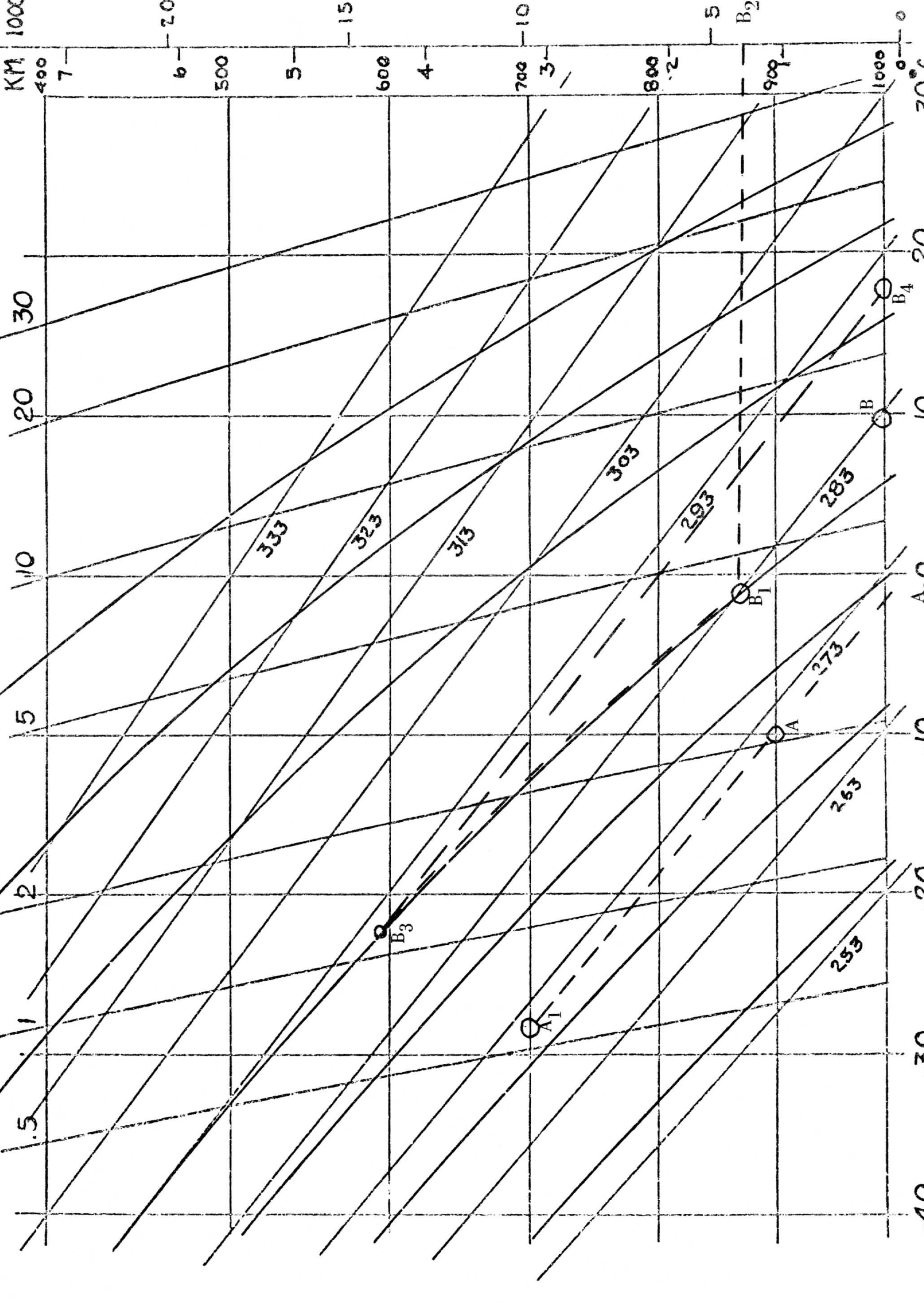
At B, we read off W = 8 g/kg. With U = 50%, actual mixing ratio

$$w = 8 \cdot \frac{50}{100} = 4 \text{ g/kg}.$$

Dry adiabat is followed as long as air not saturated, i.e. $w < W$. With decreasing T, W also decreases. (E.g. at p = 900 mb, T ~ +1°C and W = 5 g/kg; w = 4 g, i.e. $w < W$)

ADIABATIC PROCESSES

THERMODYNAMIC DIAGRAM



When air has been forced upward to point B_1, we find that now $W = 4$ g/kg. We also know that the actual mixing ratio was (and has remained to be) $w = 4$ g/kg. At B, therefore, we now have

$$w = W \text{ , i.e. saturation } (U = 100\%!)$$

Any air parcel forced upward further must follow the moist adiabat (i.e. undergo a moist-adiabatic process). During this process, some of the water vapor will have to condense in order to keep the actual mixing ratio equal to the saturation mixing ratio required by the lower temperature which ensues in the lifting process.

Level at which saturation is reached when air is forced upward from the initial level (at which process started, i.e. where T, w etc. were given) is called

LCL = Lifting Condensation Level.

Air now forced upward to B_3 along moist adiabat. At B_3, air is still saturated but $W \cong 1$ g/kg. Thus—from B_1 to B_3 air must condense out surplus of water vapor (4 g - 1 g = 3 g/kg). Water vapor goes into liquid state (cloud droplets and finally rain which falls out!).

Assume that we now move the air from level B_3 downward (fast). Immediately, T will increase and also W, but $w = 1$ g/kg stays the same (assuming no evaporation of cloud droplets). Result: $w < W$, the condition for a dry-adiabatic process. Therefore—when air, initially saturated, is moved down it will undergo a dry-adiabatic process; on graph, it will follow the dry adiabat.

Since initially non-saturated air certainly will move downward dry-adiabatically, we find the important general rules:

UP—non-saturated air—dry adiabatic
saturated air—moist adiabatic

but

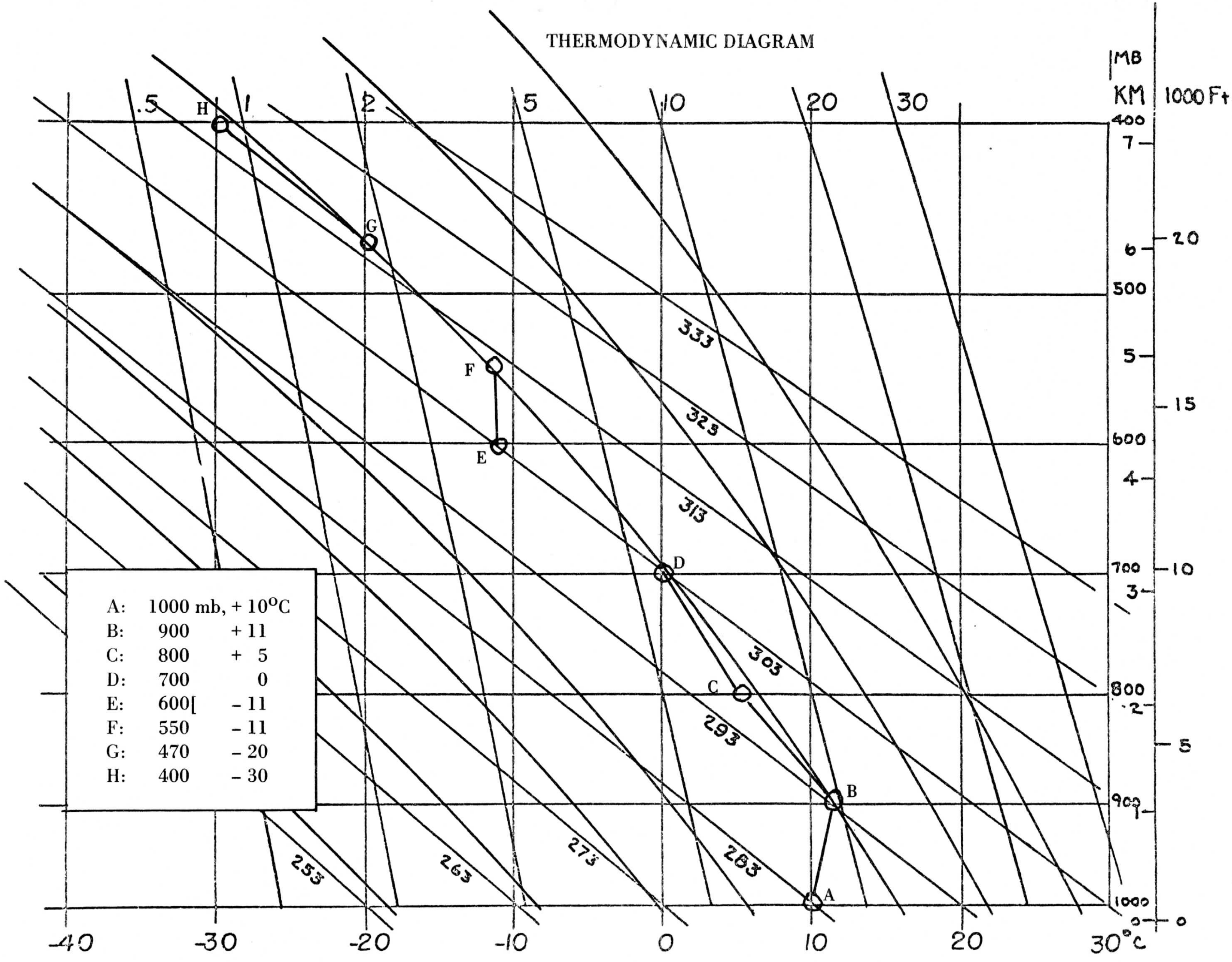
DOWN—always dry adiabatic.

Thus, moving air down from B_3 brings it to B_4 along dry adiabat to (1000 mb, +19°C)

Definition: Potential Temperature T_p (see text, p. 111).

Temperature of an air parcel which is moved dry adiabatically to the level of 1000 mb. (given in °K)

THERMODYNAMIC DIAGRAM



In diagram: Any point (T, p) situated on a particular dry adiabat has the same T_p. Thus we can label the dry adiabats with T_p.

The one going through (1000 mb, 0°C) is $\underline{T_p = 273°K}$

Another quantity (will not be used in this course) is Equivalent potential temperature.

Defined by: Temperature which air would achieve if

a) all water vapor would be condensed (i.e. all latent heat released and used to heat air)

b) then brought dry adiabatically to 1000 mb.

(see text, page 112)

II. Plotting of actual state of atmosphere (see p. IV-10)

= Sounding. Radiosonde reports (p, T, U)

Height given frequently by radar distance + angle. (Also, from drift of balloon, one obtains wind velocities.)

Plot of sounding idealized. Points can be at any value of p. "SIGNIFICANT POINTS"

Definition of Isothermy [T stays same]
Inversion [T increases with height]

III. Combining I and II for PROCESSES in ACTUAL atmosphere. [Sounding on page IV-13 is the same as the one on page IV-10, with U = 100% added at Point A]

1. Determine the LCL.
Since **U** = 100% at Pt. A, any forcing of air at A upward would immediately result in condensation, i.e.

LCL is at level of Pt. A = 1000 mb.

(Probably, one has already fog at the surface)

2. However, the sounding as given could have been an early morning sounding. When sun comes up, sfc layers will be heated from sfc by conduction, i.e. temperature will rise at level of Pt. A (i.e. sfc). When this happens, U at level A will decrease. E.G. T rises to +13°C, then W = 10 g/kg; w still is the same as before, i.e. w = 8 g/kg, thus U = 80%. Since air at this temperature is warmer than the air in vicinity, it can rise upward; with $w < W$, it rises dry-adiabatically until its T is equal to surrounding air in free atmosphere (as given by sounding).

- - - Broken line upward from T = 13°C, p = 1000 mb. Note that, where this line meets sounding, W still larger than actual w—no condensation yet. With further heating, same process continues. At A_0 (T = 20°C) air can rise up to Pt. B.

Why does air stop at sounding?—If it should move higher up, it becomes colder than surrounding air (the "environment"), thus also more dense (i.e. heavier) and will fall back to a point where it is of equal density—i.e. at the same T as sounding.

When warming at sfc has reached A_1 (~ 22°C), air will rise to sounding at height of Pt. B_1. Arriving there, its w (= 8 g/kg) now is equal to W = 8 g/kg at this point. Thus U = 100% —air is saturated and, from this level on, has to follow moist adiabat. At B_1, therefore, condensation will occur. Process to lift air to this level is CONVECTION (air "bubbling" up from sfc due to sfc heating). This condensation level is therefore called

CCL = CONVECTIVE CONDENSATION LEVEL.

Thus CCL at Pt. B_1 (~ 820 mb).
This is cloud base of convective clouds (CUMULUS)

3. Cloud top height

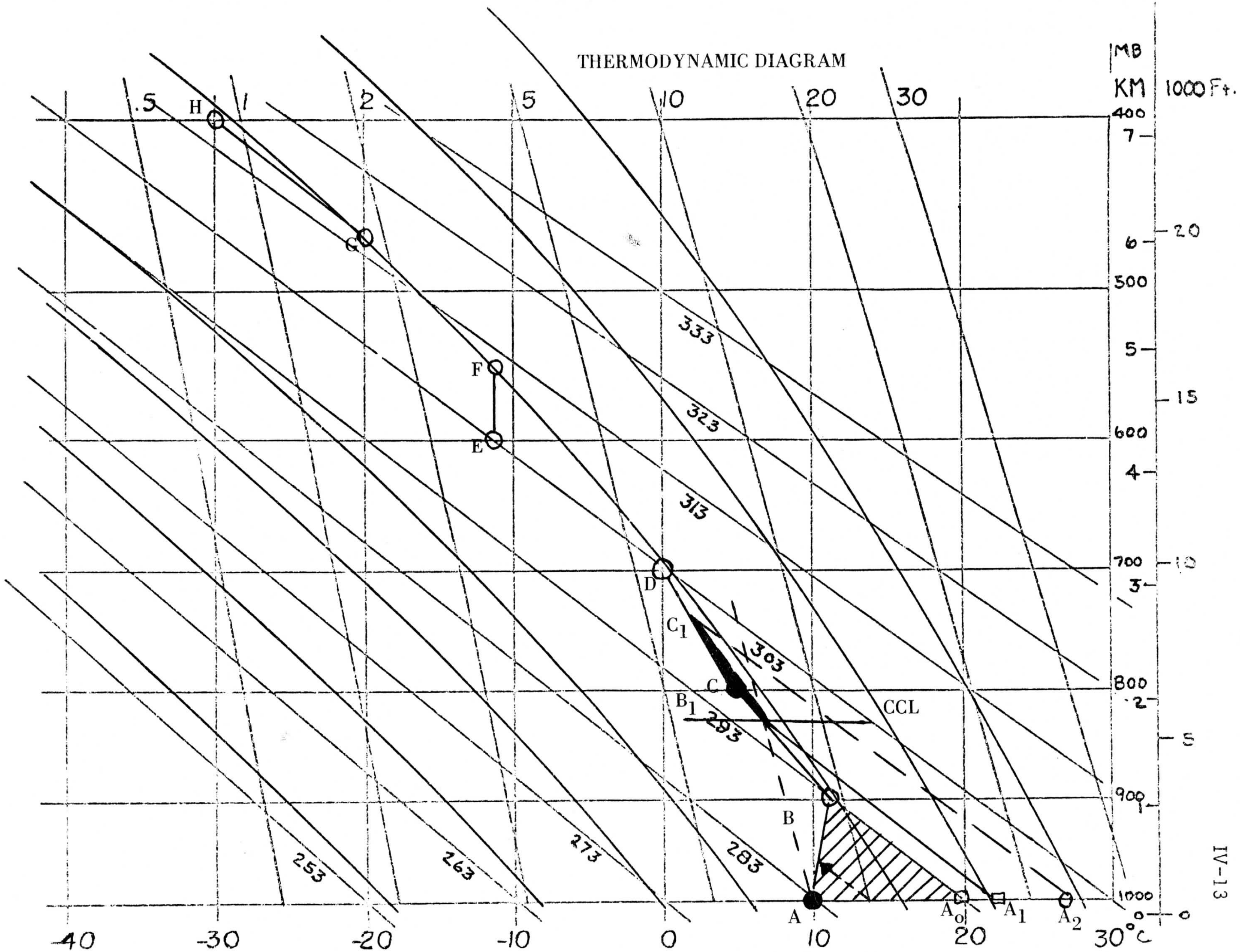
   Air reaching B_1 from below now rises along moist adiabat. Since its T during rise is still higher than environment (i.e. sounding between B–C–D) it can continue upward—until its T again becomes equal to that on sounding which happens at Pt. C_1 (~ 730 mb). During ascent, air remains saturated (U = 100%), but some of water vapor is condensed into cloud droplets. At C_1, we can read off W = 6.5 g/kg; thus, air reaching C_1, also has w = 6.5 g/kg. In cloud formed between the CCL and C_1 (the cloud top level), each kg of ascending air thus has lost (8 - 6.5) = 1.5 g/kg of water vapor which now is in cloud as liquid water droplets.

4. When above process is going on, the environment (i.e. the original sounding) obviously cannot remain unchanged. By sfc heating, the sounding is being modified—until finally (when clouds have formed) the sounding will be given by

   A_1 — B_1 — C_1 — D etc.

   Originally we had a surface inversion (A - B) which, during the day, is being destroyed through convection from the surface. This is a very frequent and usual change in soundings which goes on without any change in the air mass. Similarly, when at night the sfc cools, a new sfc inversion can be established—again on purely local grounds. Note that this heating at sfc is adding external heat energy—a diabatic process!

THERMODYNAMIC DIAGRAM



At A: U = 100%

C. <u>Stability</u> (see sounding on page IV-15)

Air parcel, imbedded in actual atmosphere (environment), has same (p, T).

Assume that it is suddenly disturbed by vertical gust, i.e. forced up— or downward. When gust acceleration stops, either one of two things can happen:

> If (small) motion induced by gust continues (in spite of friction in surrounding air which normally would stop motion) and in fact motion accelerates <u>without</u> further pushing, the situation is
>
> UNSTABLE.
>
> If, on the other hand, motion is stopped and, in fact, parcel attempts to return to original level, we have a situation which resists the change, i.e. is
>
> STABLE.

Let T_1 be temperature of parcel after being pushed to level at pressure p and

T_2 be temperature of environment at this level.

If $T_1 > T_2$, then parcel is warmer and thus less dense than air of the surrounding atmosphere. It is therefore lighter and will have "lift" like a "hot-air balloon." This lift will act as a force, causing an acceleration and parcel will go on rising, may even accelerate to higher speed, after gust has stopped.

This situation ($T_1 > T_2$) clearly is UNSTABLE.

If $T_1 < T_2$, the parcel is more dense (heavier) than environment, has "negative lift" like a "lead balloon." When lifting force of gust stops, parcel will fall back to previous conditions. This then is a STABLE situation.

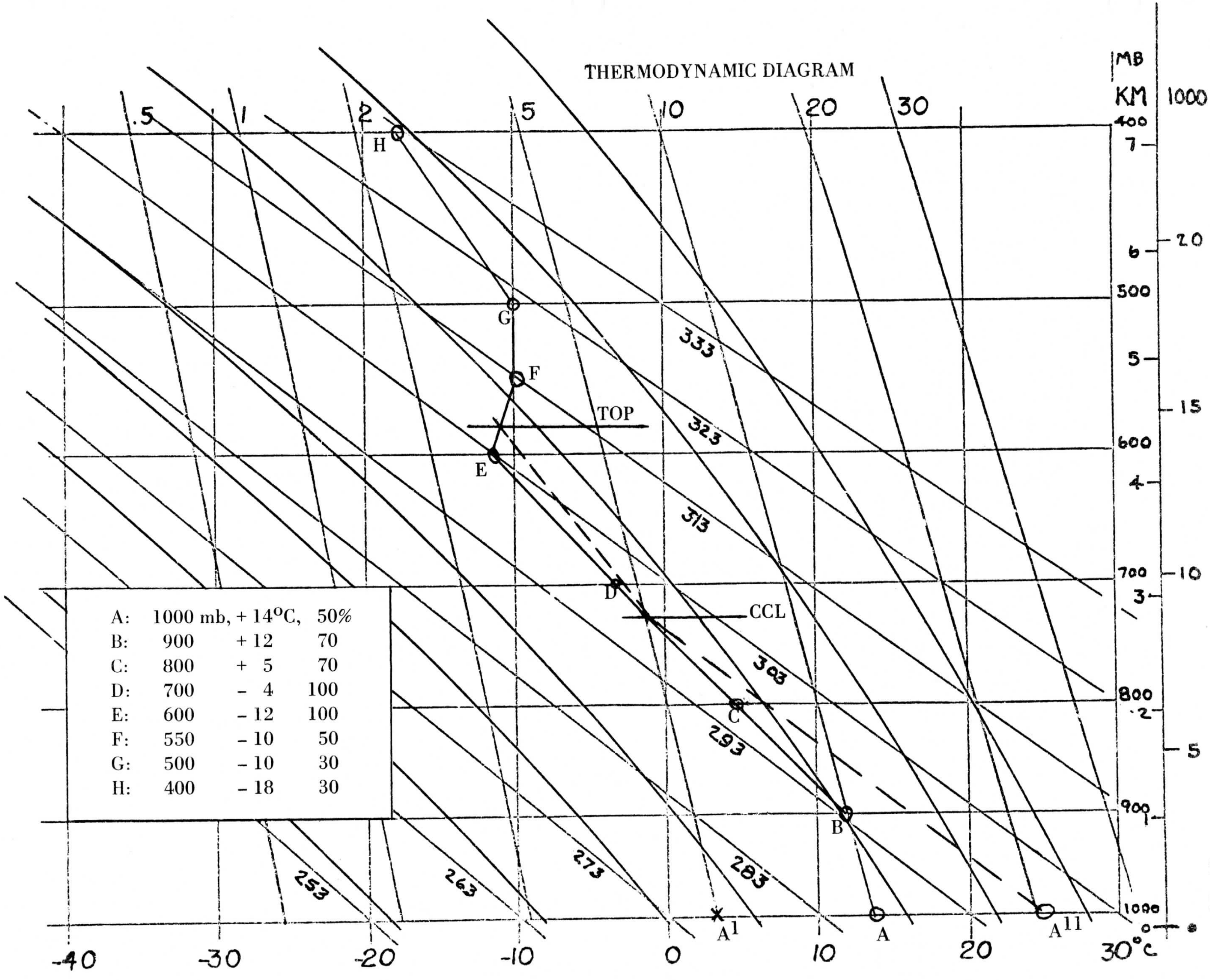
If $T_1 = T_2$, parcel will be as dense as environment. According to Newton laws, no force acts when gust force stops. Parcel will continue moving through its inertia until motion is stopped by friction. Since this happens rather fast, this case usually is also a STABLE situation since motions don't continue on their own accord. Therefore, we usually call all situations STABLE, when we find

$$T_1 \leqq T_2$$

Now let's look at a layer of a sounding, e.g. layer BC in sounding on page IV-15.

Temperature in environment changes (nearly) linearly between B and C. If I disturb a parcel at B, I can judge its stability by comparing the

THERMODYNAMIC DIAGRAM



A:	1000 mb,	+ 14°C,	50%
B:	900	+ 12	70
C:	800	+ 5	70
D:	700	– 4	100
E:	600	– 12	100
F:	550	– 10	50
G:	500	– 10	30
H:	400	– 18	30

sounding with the adiabatic (either dry or moist) lapse rate as given by the dry or moist adiabats through pt. B. Similarly, I can do it for any point between B and C and will find the same result. Thus, to judge the stability of air within the layer BC, we only have to apply the criterion to the lower edge of layer, i.e. at Pt. B.

Stability therefore applies to layer—not just a single point.

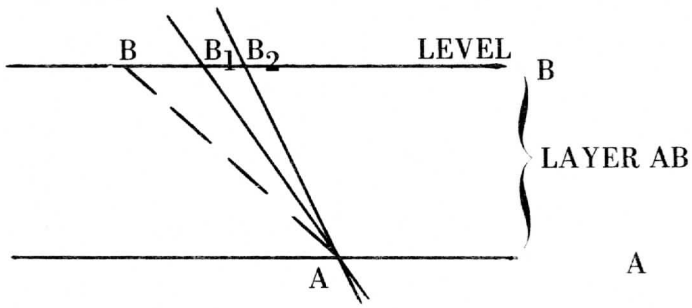
Easiest way to find stability:

Compare actual lapse rate γ in layer (as given by sounding between B and C, for example)

with dry-adiabatic lapse rate γ_d
and moist-adiabatic lapse rate γ_s

existing at bottom of layer under consideration. The latter are given on thermodynamics diagram by dry and moist adiabat drawn through bottom point.

Case I.



$\boxed{\gamma > \gamma_d > \gamma_s}$

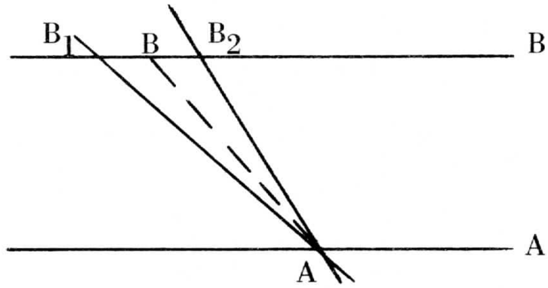
AB: Environment = Sounding
AB_1: Dry adiabat; AB_2: Moist adiabat.

B to left of both B_1 and B_2 (i.e. T at B lower)

Parcel of air at A pushed up: if not saturated, follows dry adiabat AB_1; always is warmer than environment, will continue upward—unstable. If saturated, follows moist adiabat AB_2, again always warmer = unstable. Thus: Conditions in layer (A-B) always unstable—called ABSOLUTELY UNSTABLE.

This condition normally does not occur in free atmosphere. Only possible under very special circumstances (e.g. near the ground when continuous heating occurs = diabatic effects). When found in sounding, usually one can assume error in sounding transmission!

Case II.



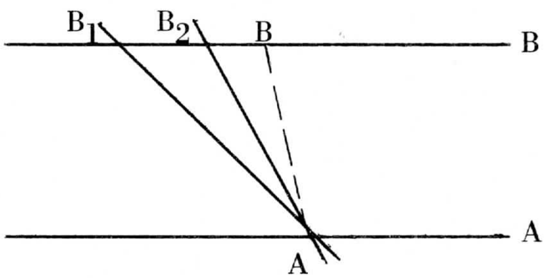
$\gamma_d \geq \gamma \geq \gamma_s$

B to right of B_1 <u>but</u> to left of B_2 ("in-between")

For "dry" parcel of air—STABLE (at B_1 air cooler than environment at B)
<u>BUT</u> for "moist" parcel—UNSTABLE. Called <u>CONDITIONALLY STABLE</u>

[If B coincides with either B_1 or B_2, case is included here.]

Case III.



$\gamma_d > \gamma_s > \gamma$

B to right of both B_1, B_2 (i.e. always air parcel cooler than environment)—always stable. Called <u>ABSOLUTELY STABLE</u>.

Note that any isothermy or inversion layer always is absolutely stable. ($\gamma \leq 0$!)

<u>Apply these criteria to sounding</u>, p. IV-15.

As plotted, no Case I = absolutely unstable layers
<u>Verify</u>.
Case III—absolutely stable: A-B, E-F, F-G, G-H
(all are to right of moist adiabat through bottom point)
Therefore, Case II must be remaining layers:
conditionally stable—B-C, C-D, D-E.

CHANGE OF STABILITY

a) By heating or cooling top or bottom of layer (e.g. by radiation = diabatic process)
b) By moving whole layer up—or downward (e.g. vertical motions on large scale)

Case a.

Heating bottom or cooling top
decreases stability of layer

Cooling bottom or heating top
increases stability of layer

Case b.

Whole layer being raised
decreases stability

Whole layer being lowered
increases stability

[Verify this on diagram!]

Consequences:

a) Cloud tops are radiating, i.e., are cooling. If this goes on for some time, layer containing cloud will become more unstable and may allow further growth or more condensation.
b) Air moving downward (such motion on large scale called "SUBSIDENCE") will become more stable, may even develop from normal lapse rate to isothermy or inversion.

In fact, many inversions observed in free atmosphere are due to subsiding motions (case b).
Exception usually only in surface inversions, which develop through cooling at surface (case a).

EXERCISES—Chapter IV

1. Study and verify the Home Study Assignment—Part I.

2. Complete the questions, etc. on the Home Study Assignment—Part II, following the same procedures as in Part I.

   In the subsequent exercises, make use of the various thermodynamic diagrams and/or formulas and concepts discussed in this chapter.

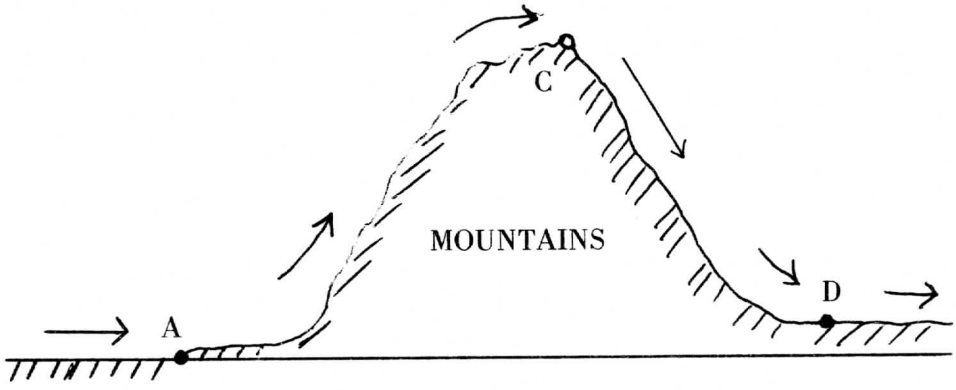
3. An air parcel is being moved dry-adiabatically from the sea level to a height of 3 km. At sea level, its temperature was known to be +10°C. Determine its temperature at the top of its motion.

4. Another air parcel starts out at sea level with the same temperature of +10°C. It is saturated at this level.
   a) Determine its mixing ratio at sea level.
   b) Determine its temperature at a height of 3 km.
   c) How much water will be condensed during the ascent per kg of dry air?
   d) If the temperature of the surrounding air (environment) at 3 km is -3°C, will the parcel continue moving on its own accord or will it be stopped?

5. Assume a parcel of air in the atmosphere at a pressure of 800 mb and a temperature of -12°C.
   a) How many g of water vapor could be associated at this point with one kg of dry air?
   b) If one measures a relative humidity of 75%, how many g of water vapor are associated with 1 kg of air?
   c) If one heats the air parcel (without changing its pressure) to +10°C, what will be the relative humidity if, originally it was 100%?
   d) To which level do you have to lift the air parcel (having a relative humidity of 50% at the original pressure and temperature) to reach saturation?

6. Determine the potential temperature of air at
   a) p = 880 mb, T = +10°C
   b) p = 795 mb, T = +20°C
   c) p = 600 mb, T = -20°C
   d) p = 1000 mb, T = +27°C

7. The "FOEHN" or "CHINOOK" problem. Reading: text pp. 281-282, 290.



At A, air arrives (at the coast) at the windward side of a mountain barrier. After being forced upward across the mountains it will descend again into the plains and finally arrive at D.

Given is: a) Properties of air at A: p = 1000 mb, T = +10°C, U ≈ 60% (w = 5 g/kg)
b) Air at D, before air crossing the mountains arrives, has following properties: p = 930 mb, T = -10°C, U = 100%.

a) Determine the LCL at windward side of mountains. (p, T, U, w)
b) Determine the properties of the air at C on top of the mountains (height 12,000 ft = 640 mb) (T, w, U).
c) Determine the amount of condensed water vapor per kg of dry air during the ascent.
d) Determine the properties of the air arriving at D—assuming D at 2,500 ft = 930 mb. (T, w, U)
e) What is the change in T at station D? in U? in w?
f) What is the potential temperature at A, C and D?
g) What happens to a snow cover existing at D before the chinook arrives and what will happen after its arrival?
h) Will any snow existing on the leeward side of the mountains (i.e. above D) melt with the arrival of the chinook? Below which height will this melting occur, if at all?

Home Study Assignment—Part 1

In order to deepen your understanding of the Thermodynamic Diagram and its use, you are urged to plot the sounding given below and verify the statements accompanying it.

PLOT the following sounding which could represent the state of the atmosphere over Madison some early summer morning: Pt A at 1000 mb, +13°C, Rel. Hum. U = 100%. Subsequent points have pressure and temperature as follows: B = 900 / + 15; C = 800 / + 5; D = 700 / 0; E = 600 /-8; F = 550 / -8; G = 500 / -15; H = 400 / -20.

VERIFY the following statements:

1. Mixing ratio at A equals 10 g/kg.
2. A - B indicates an inversion, E - F an isothermy.
3. Absolutely stable layers are A - B, C - D, E - F, G - H.
4. Conditionally stable layers are B - C (= dry adiabatic), D - E (moist adiabatic)
5. As given above, F - G is absolutely unstable (Lapse rate is larger than dry adiabatic)—probably this layer has been recorded or transmitted with an error.
6. Assuming that no change in the atmospheric structure occurs during the day above B, one can estimate the development of this sounding during this day due to the diurnal heating of the surface layer.
7. During the morning, sfc will heat up. Until a sfc temp of +23°C is reached, no ascending air can rise beyond Pt. B (The sfc inversion has to be cleared away.)
8. When sfc. temp. +23°C, Rel. Hum. has dropped to about 50%.
9. When sfc temp. $\geq$ +23°C, air can rise beyond Pt. B; the CCL is then reached at about 870 mb (= 1,400 m)
10. Air ascending to CCL is saturated there. It will now follow the moist adiabat and will keep on going up until being stopped in layer E - F at ~ 570 mb (4,600 m)
11. In each kg of air rising to the top, 6 g of H_2O vapor will condense to water droplets, i.e. a cloud will form.
12. This will be a vertically developing cloud, i.e. CUMULUS.
13. Therefore, we can expect (a) a rather fast rise of the temperature near the surface, without cloud formation, by about 10°C (from 13 to 23°C), then a rapid buildup of Cu clouds which are fast growing up to roughly 4 1/2 km (about 15,000 ft). These clouds naturally will decrease the direct solar radiation and thus the further rise of the temperature at the surface is slowed down.

14. Depending on time of year and location, there may exist a possibility that sfc temp still <u>may</u> rise to about +30°C. (This may be unlikely at Madison but could occur, say, in Texas or Oklahoma).

15. In this case, air starting from sfc would form clouds at ~ 790 mb (i.e. CCL will rise). Such air could then pass the isothermy beyond Pt. F and clouds could build up even higher to about 430 mb (6 1/2 km). In this case, we have high Cu congestus which easily could develop into Cb (temperature on top is about -18°C).

16. Thus there could be thunderstorms or showers in the afternoon—<u>if</u> sfc temp > 30°C (= 86°F).

<u>Home Study Assignment</u>—Part 2

Plot the following ascent for a late winter morning at Madison.

PTA: 1000 mb, -8°C, 100% rel. hum.; B: 900 mb, -5°C;
C: 800 mb, -10°C; D: 700 mb, -20°C; E: 650 mb, -10°C;
F: 500 mb, -25°C; G: 400 mb, -30°C.

(Be careful to plot this sounding accurately; use straight edge!)

Answer the following questions, either in numbers or in as few words as possible. The first <u>13</u> questions are closely related to those in Part I of this assignment; the others are additional new questions.

A. <u>First 13 Questions</u>:

1. Mixing ratio at pt. A: ______
2. Inversion: Layer __________ Isothermy: __________
3. Absol. stable layers: ______________________
4. Condition. stable layers: __________________
5. Absol. unstable layers: ____________________
6. No answer required.
7. To form clouds, surface (sfc) to be heated to: ______ °C
8. Relative humidity (RH) at sfc when heated to this point: _______%
9. CCL at ______ mb, equal to ______ m or _______ ft.
10. Clouds so formed will develop up to ______ mb = _______ m thickness of clouds thus ______ m or _______ ft.
11. Kind of clouds formed is called ______

12. Describe day (sounding made at sunrise) assuming nothing else changes in atmosphere: ______________________________
____
____

13. Could a Cb form: ________
What temp at sfc needed? ________
Is this likely to happen? ________

B. The additional 15 questions

1. Convert all temperatures (Pts. A - H) into °F.
A: ______ B: ______ C: ______ D: ______ E: ______
F: ______ G: ______

2. List the potential temperature of all points:
A: ______ B: ______ C: ______ D: ______ E: ______
F: ______ G: ______

3. An airplane flying at the 400 mb level pressurizes its cabin to an interior pressure of 800 mb. Assume that at 400 mb, the air is saturated (i.e. plane in clouds).
What is temperature and rel. humidity of air coming from compressor?
T = __________ R.H. = __________

4. If I want, in cabin, a temp. of 20°C and R.H. of 50%, how many grams of water per kg of air do I have to add to the compressed air?

____

5. Assume, in addition to the values stated at the beginning, that, at Pt. E we have measured a R.H. of 100%.
How many grams of water vapor are associated with each kg of dry air at this height? ________

6. A inversion, as found above Pt. D, is often (but not always) the result of warm air aloft, having been moved into the area from southern regions and being lifted in this process slowly over the lower, colder air. The result is then a cloud layer above Pt. E, extending in this case upward to Pt. F.
What is the R.H. in this layer (E - F)? __________

7. What is the mixing ratio at Pt. F in this case? ____________

8. Since the cloud results from air being moved upward through this layer, I can compute the amount of liquid (or solid) water in each kg of cloud air. It is ________ grams.

9. What kind of cloud would you think is this?

____

10. What is the thickness of the cloud layer above?
    __________ m or __________ ft.

11. Under such circumstances, is a development such as described in questions A 1-13 very likely? __________

12. If precipitation would fall out of these clouds, what kind would you expect? _________

13. In questions A 1-13, a R.H. at Pt. A of 100% was assumed. If we change this assumption to 50%, where (at which height) would you then find the CCL? _____________

14. What would this mean with regard to the cloud formation by convection? ______________________________

15. Why is this case a rather common one in winter?
    (Remember that cold air in Madison usually comes from Canada!)

    ____

    ____

CHAPTER V. CLOUDS AND PRECIPITATION

(Reading: pp. 94-103, 209-251)

A. Condensation Process

Concept of "supersaturation" (U > 100%) in clean air.

Condensation Nuclei

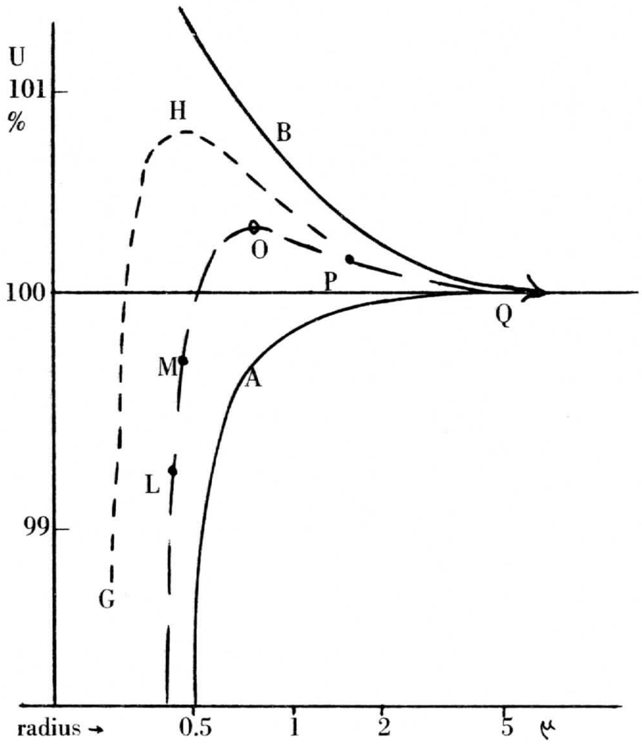
a) Salt crystals (NaCl) size ~ 1/10 μ
b) Combustion products (forest fires, volcanoes, man)

Hygroscopic. Number: 10-1000 per cm^3.

Solute effect. Lowering of saturation vapor pressure in close vicinity of nucleus "helps" (i.e. condensation starts at U < 100%)

Curvature effect. Surface tension over strongly curved surface is increased, "hinders" (adding H_2O molecules to droplet requires outside U > 100%)

Both effects acting together in initial stages of droplet formation.



Curve A: Growth of droplet from nucleus if no curvature

Curve B: Growth of droplet if no solute effect;

Curve L M O P Q—Example of growth under influence of both effects: below L, strong solute effect
L → M: Curvature very strong, continuously higher U required
M → O: at that size, actual supersaturation necessary
O → P → Q: Curvature decreases (size increases), but also NaCl concentration decreases.
Beyond Q - Growth at ~ 100%.

Curve G → H → for smaller droplet. Usually such droplets will not grow due to vicinity to drops already larger (which can grow at lower U and thus will "steal" H_2O molecules from smaller droplets).

Result—usually only "a few" drops (as compared to available nuclei) will grow beyond O → P.

Time of growth: from nucleus to droplet of 10 μ ~ 1 sec. Increases in size to 100 μ (i.e. volume 1000 times) ~ 1-5 min, to 1000 μ (= 1 mm) ~ 3 hours, to 3 mm ~ 24 hours. Raindrop size 0.5-7 mm. But: rain often falls shortly after clouds have built up. How can that be? Need additional mechanism.

Falling of drops through air—terminal velocity.

Terminal velocity of raindrops and cloud droplets in still air

Diameter (micron)	Rate of fall m/sec.	Rate of fall in 1 min.	Type of drop
5,000	9.1	546 m	Large rain drop
2,000	6.5	390 m	Medium rain drop
1,000	4.0	240 m	Small rain drop
500	2.1	126 m	Large drizzle
200	0.7	42 m	Drizzle
100	0.3	18 m	Large cloud droplet
50	0.076	5 m	Ordinary cloud droplet
10	0.003	18 cm	Incipient cloud droplet
2	0.00012	7 cm	Nuclei
1	0.00004	2 mm	Nuclei

Drops larger than ~ 7 mm will disintegrate when approaching velocities of ~ 10 m/sec (which is below their terminal velocity). Thus: If time too short, droplets are too small; then they fall very slowly—this obviously cannot explain rain. (Also—drops falling through air with U < 100% will start to evaporate)

ICE NUCLEATION

Condensation Process can (and does) occur both at T > 0°C and T < 0°C (supercooled). Efficiency of nuclei, however, decreases with low T and at very low T (-25°/-30°C) usually H_2O vapor will not condense but sublimate (form ice crystals).

Again nuclei required, but they don't have to be hygroscopic. Any kind of particle will do (dust, mineral crystals, etc.) (ICE NUCLEI).

In fact, with dust available, ice crystals can and often do form already at T near -8 to -10°C.

Basic form of crystal: hexagonal plates or needles. Further growth to snow crystals depends on T and supersaturation in vicinity of crystal. Infinite variety.

Such low T occur in atmosphere at heights of 5-10 km (depending on season, latitude, etc.)—thus in those heights usually ice crystal clouds if conditions favorable (i.e. saturation).
Ice clouds called "CIRRUS" (various types)

Sizes of Atmospheric Impurities and Hydrometeors

Raindrops	500 - 7,000 μ (0.5-7 mm)
Drizzle	100 - 500 μ (0.1-0.5 mm)
Mist, fog; Cloud droplets	5 - 100 μ
Salt nuclei	0.1 - 5 μ "giant" up to 15μ
Combustion nuclei	0.01 - 0.3 μ
Dust (mineral)	0.1 - 100 μ
Dust motes (visible)	> 10 μ
Heavy industrial dust	> 100 μ

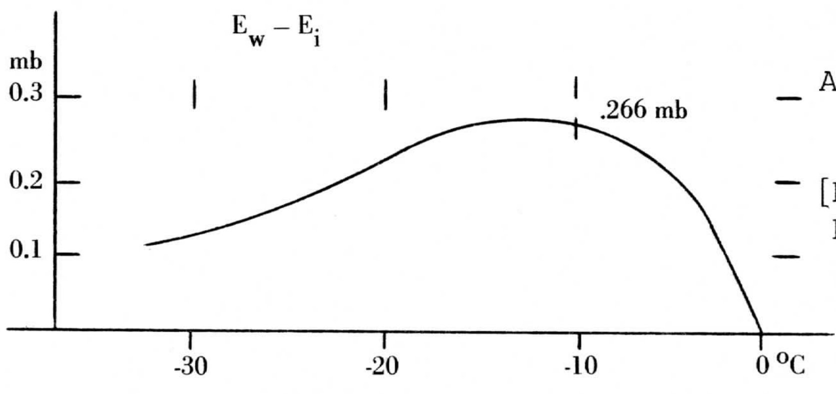
Hayfever pollen 15-60 μ
Tobacco smoke ~ 0.3 μ
Industrial smoke 0.1 μ - 10 μ, with admixture of larger particles
Wavelength of yellow light 0.5 μ

B. Precipitation Process

I. Bergeron—Findeisen Process (1928)

Growth of cloud droplets to rain drops with the active help of ice crystals injected into cloud.

Difference in saturation water vapor pressure between liquid water and ice ($E_w - E_i$)



At -10°C, ΔE = 0.266 mb

[E_w = 2.863 mb
E_i = 2.597 mb]

EXAMPLE:

Assume a cloud droplet (~ 10 μ) and an ice crystal in close vicinity. Air around them has actual e = 2.8 mb. 2.8 < 2.863, i.e. U_w (rel. hum. with respect to water droplet) is therefore less than 100%; cloud droplet will slowly evaporate. But: 2.8 > 2.597, i.e. $U_i > 100\%$ (supersaturated); thus H_2O molecules will be deposited (sublimate) on ice crystal.

Result: Ice crystal grows at expense of cloud droplet; eventually will use up the molecules.

When ice crystal grows, its velocity downward grows: fall through cloud and can find more cloud droplets. It also will encounter other ice crystals—but while droplets may "bounce" off each other, ice crystals stick (snowball!) and form larger snowflakes. Thus: growth of ice crystals using up cloud droplets. Ice crystals finally come down to earth (snow fall) or reach layers where T > 0°C, will melt into large drops (rain fall).

This process (described here in a very simplified way) is assumed to account for the majority of cases where raindrops develop out of cloud droplets.

Availability of ice crystals is usually assured because they form by sublimation in the cirrus-levels and slowly drift downward.

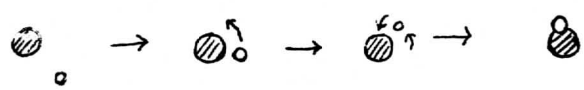
II. Other Processes

1) Sometimes (especially in tropics) rain forms in clouds which are totally below the freezing level (warm clouds). Thus no ice crystals can be involved. Cloud droplets here combine by capturing neighboring droplets and thus grow also.

Two ways of capture:

a) Direct -

b) Wake -



Process not very efficient since two drops hitting each other can also result in a number of small drops rather than one big one. (splattering)

2) Today suspected that static electricity in falling drops has rather important role in capture process. (We know that electricity exists in clouds—lightning!)

3) Finally, raindrops can occur as result of melting hail stones, especially in thunderstorms. These probably account for largest type of drops in showers, often starting off the rain fall.

. Clouds

Two ways of classification:
a) Water vs. ice cloud (physical)
b) Level and/or development of cloud (geometric)

a) Distinction between ice clouds (Cirrus) and water clouds (all others, except top of cumulo nimbus, anvil and virga)

b) Two basic groups: <u>stratiform</u> (horizontal)
<u>cumulo form</u> (vertical)

subdivided into 4 cloud families
1) High - base above 20,000 ft (or cirrus)
2) Middle - base between 8,000 and 20,000 ft (Alto....)
3) Low - base below 8,000 ft
[Families 1-3 basically "layers"]
4) Vertically developed (cumulo....)

Distinction between <u>high</u> and <u>middle</u> essentially by makeup—

Temperate Latitudes. | High (cirrus) <u>always</u> ice cloud
Middle <u>always</u> water cloud, at least partially.

Cloud families again subdivided into several cloud types.

To enter cloud type into Wx map, we use symbols.

Cloud Families and Types

Family	Type	Abbreviation	Symbol
I High	(base > 20,000 ft) (ice)		
	Cirrus	Ci	
	Cirrostratus	Cs	
	Cirrocumulus	Cc	
II Middle	(base < 20,000; > 8,000 ft) (water)		
	Alto stratus	As	
	Thick As or Nimbo-stratus	Ns	
	Alto cumulus	Ac	
	Alto stratus + Alto cumulus	As + Ac	
III Low	(base < 8,000 ft) (water)		
	Stratus	St	—
	Stratocumulus	Sc	
	Fractostratus (Scud)	Fs	- - -
IV Vertical developing	(base <u>usually</u> < 8,000 ft)		
	Cumulus	Cu	
	Cumulus congestus	Cu(cong)	
	Cumulonimbus	Cb	

Lecture on clouds with pictures. [see also pictures in textbook]

Special cloud types: Ac castellanus
Ac lenticularis
Mammatus forms.

Note special difference between Cu congestus and Cb. (The latter requires beginning of conversion to ice cloud.) Anvil, virga.

In addition, (near) surface phenomena:

Fog (thick, visibility below 1 km = 5/8 mi)	≡	water droplets
Fog (thin), also called Mist (1-2 km)	=	
Haze (visibility > 2 km = 1 1/4 mi)	∞	(often dust)
Smoke (industry, forest fires, etc.)		
Sandstorm, Duststorm		
Blowing Snow		

D. Forms of Precipitation and Associated Phenomena

	Symbol		Symbol
Rain	●	Thunderstorm	
Drizzle	,	(with hail)	
Snow	*	Lightning	
Hail	△	Funnel cloud	)(
Sleet		(tornado)	
Freezing rain		Dust Devil	
Shower	▽	Waterspout	
e.g. rain shower			
snow shower			

[Definitions, see p.V-8]

Used singly or in combination. Heavier amounts often indicated by multiple symbols.

For example:

Heavy, continuous rain

Snow/rain mixed

Thunderstorm with rain

If occurrence during last hour, but not at time of observation, a "]" added, e.g. ·] = rain in past hour.

Brief discussion of thunderstorm.
(Read description in textbook, pp. 163-180)
Cells with system of up—and down drafts
3 stages—growing (mostly up)
mature (up and down)
dissipating (mostly down)

Transition from Cu cong. to Cb—ice transformation in top parts. Formation of ANVIL (ice—cirrus). Hail formation in shower cloud. Multiple up-down. Electric phenomena. Lightning: within cloud (cell to cell), between clouds, cloud-ground.

Energy in ☈ largely from latent heat release. Thus: most in tropics, least (or none) in Arctic.

World pattern of occurrence: rarely poleward of latitudes 60-65° Heaviest in tropics over land: Central Africa, Indonesia. Also highest Cb—due to height of tropopause at Equator.

In U.S.: Maximum in Florida, about 60 days per summer; more than 50% in June, July, August. Secondary maximum in Colorado/New Mexico with about 40 days per summer.
Rather rare in colder season; then exclusively airmass or coldfront thunderstorms, require advection of warm moist air.

Related to ☈ also TORNADOES.
(Read text, pp. 180-186)
Very localized but extremely destructive. Windspread near center estimated > 400 mph; wind pressure over 700 to 1000 lb/ft^2.
Funnel cloud, developing from above, touching ground. Path on ground fortunately short (rarely over a few miles).
Forecast from SEVERE STORM CENTER (Norman, Oklahoma)
Tornado Watch. Good progress in last decade; decrease of loss of life and injuries in spite of denser population.

Occurrence: 80% between noon and 9 p.m.

Most in broad area from Okalhoma → NE (Tornado Alley)—indicates requirement for warm moist Gulf air to supply moisture for latent heat release. Often occur in groups—i.e. if one was reported in vicinity, probability of another nearby is rather high. Take cover (SW corner of basement—i.e. corner toward approaching funnel).

Occur practically always together with widespread ☈ and/or hailstorms. Speed of forward motion widely variable (on average 10-30 mph).

DEFINITION OF PRECIPITATION PHENOMENA

Type	Symbol	Description/Size
Rain	•	liquid drops 0.5–7 mm
Drizzle	,	liquid drops 0.1–0.5 mm (seem to float, not fall)
Snow	*	solid flakes, from ice crystals variety of forms—see textbook pictures
Granular Snow		very small pellets, like frozen drizzle
Sleet	$\triangle$	layer ice pellets, frozen rain [in England, sleet denotes snow & rain mixed!]
Glaze or Freezing Rain		supercooled raindrops (will freeze when hitting surface)
Ice needles	$\leftrightarrow$	especially in Arctic, Antarctica, at very low temperatures single ice columns or hexagonal plates often seem to fall out of blue sky.
Hail	$\triangle$	soft—snowlike center with thin ice shell, soft, will bounce very little when hitting ground. small—heavier ice shell, will bounce; diameter 2–5 mm ordinary to large—hard throughout, often in distinct layers, "lumps" of ice, > 5 mm.

Above fall out of clouds to the ground. Some forms of condensation also occur directly at or near the ground, as follows:

Dew	liquid droplets form at sfc (grass, leaves, etc.) ["dewpoint": $U_W = 100\%$]
Frost	(solid) sublimation on surface = ice ["frostpoint": $U_i = 100\%$]
Rime	solid deposit on surface or exposed objects due to supercooled drops freezing upon contact. Often found on mountains when clouds drift by. Buildup especially strong on windward side. Ice storms in the NE, related to "icing" of aircraft wings.

FOG

If condensation process (described above for formation of clouds) occurs near the ground, we have FOG. Different kinds due to difference of formation:

I. Radiation Fog:
Cooling at sfc—ground fog.
Cooling at airmass boundary—inversion fog.
Latter grows downward from inversion.

II. Advection Fog:
Warm air moves over cold surface (i.e. grows upward)
over land—e.g. warm air moves over snow
over ocean—moves over cold ocean current
(most famous: Labrador, Newfoundland)

III. Evaporation Fog:
Cold air moving over warm surface (usually water)
Steaming of lake in late fall; fog over leads in Arctic Ocean.
Combination of evaporation and mixing of air close to water surface.
on small scale, bathtub in cold room; breathing out into cold air (air from lungs very moist).

IV. Upslope Fog:
Air being forced upward in valleys or along slope reaches LCL, forms cloud near ground.

V. Smog:
Fog formation aided by large amounts of combustion particles. Irritation of throat and nose by admixture of various pollutants. Dissipation frequently prevented by low-level inversion acting as lid on convection; concentration of irritants increases continuously.

All the above processes require the existence of sufficient water vapor in the atmosphere. Comes into atmosphere from surface.

EVAPORATION

I. From open water surfaces (oceans, lakes, streams, etc.):
depends on 1) temperature of water
2) water vapor pressure of air above
3) air circulation (wind)
4) salinity of water

II. From soil:
depends on 1) texture of soil
2) water content
3) permeability of soil

III. From plants:
depends on 1) species
2) size of leaf surface
3) growing conditions
4) availability of water in plant or soil

Evaporation is not easy to measure in many cases.
Evaporimeters (water or soil)
Lysimeters (test beds with plants)

Important quantity in agriculture. Major topic in field of Micrometeorology and Agricultural Meteorology.

Artificial Stimulation of Precipitation ("Rainmaking") [see text, pp. 214-219]

Some success in limited way when conditions are favorable.
While experimentally established, most claims of commercially feasible operations highly exaggerated. Proof of positive success extremely difficult—usually not known whether it would not have rained anyway!

3 Methods:

1) Dispersing of Dry Ice (= solid CO_2 snow) into cloud by airplane. Process: Air being cooled locally so that ice crystals can form; CO_2 snow is very cold! Then Bergeron Process can take over.

2) Dispersing of Silver Iodide into cloud (usually by burning of fuel containing small amounts of AgI which are carried into smoke) from airplanes or from ground burners.
Process: Silver iodide crystals have structure very similar to ice crystals. Substitute for ice crystals in Bergeron process and may initiate it.

3) Sprinkling of finely dispersed water drops into cloud form. Introduction of these "large drops" may lead to additional growth of others by coagulation (capture).

Rainmaking only one aspect of larger problem of WEATHER and CLIMATE MODIFICATION.

A large number of legal, ethical and cost-efficiency problems have to be solved. Problem is becoming tremendously important—could mean life or death for millions of humans.

In 1966, American Academy of Science issued a substantial report on problem which goes into many of the scientific and other problem areas.

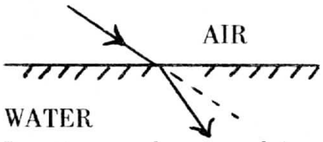
Problem certainly not yet even faintly close to solution—estimates of time for reaching substantial progress range from decades to centuries!

E. Optical Phenomena in Atmosphere (See text: Chapter 9, pp. 233-251)

I. In clear air

a) Refraction.

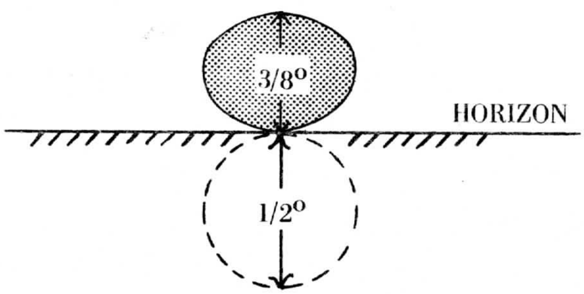
Light ray is bent when passing from less dense to more dense medium (for example, at surface of water)



In atmosphere, objects appear lifted near horizon. Sun may still appear to be above horizon when geometrically it already has set (i.e. is below horizon).

Refraction depends very strongly on height angle above horizon. At horizon lifting about 1/2°.

Sun in horizon appears oval—rays from lower rim lifted more than those from upper rim.



Result: daylight lasts a few minutes longer than it should without atmosphere.

b) Scattering.

Light waves are scattered by encounter with small objects. Amount of energy scattered in various directions depends on size of particles in air, in relation to wave length.

If particles are larger than wave length, then

MIE scattering

Energy is proportional to $1/\lambda^2$
[4 times as much scattered if wave length decreases to one half]

but

if particles much smaller than wave length ($< 1/10\ \lambda$), then

RAYLEIGH scattering

Energy proportional to $1/\lambda^4$
[16 times as much scattered if $\lambda_2 = 1/2\ \lambda_1$]

Molecules in air (O_2, N_2, etc.) are that small. Thus violet and blue light ($\lambda \sim 0.3/0.4$) much stronger scattered than red light ($\lambda \sim 0.7$). Result: Sunlight, on way through atmosphere, loses much more blue than red light—and the longer path of sun's rays through atmosphere (evening or morning), the redder the sun.

Scattered light, eventually reaching eye from "somewhere in sky," on other hand will be blue! "Blue Sky."

When air contains dust (particles larger than λ (vis.)) effect is much less pronounced (only 4 times as much blue than red, MIE), thus whitish tinge of sky.

Very fine dust in high atmosphere occasionally from volcanic eruptions (stays in atmosphere often for months to years). Brilliant colors at sunrise or sunset (very long paths of light through air and dust).

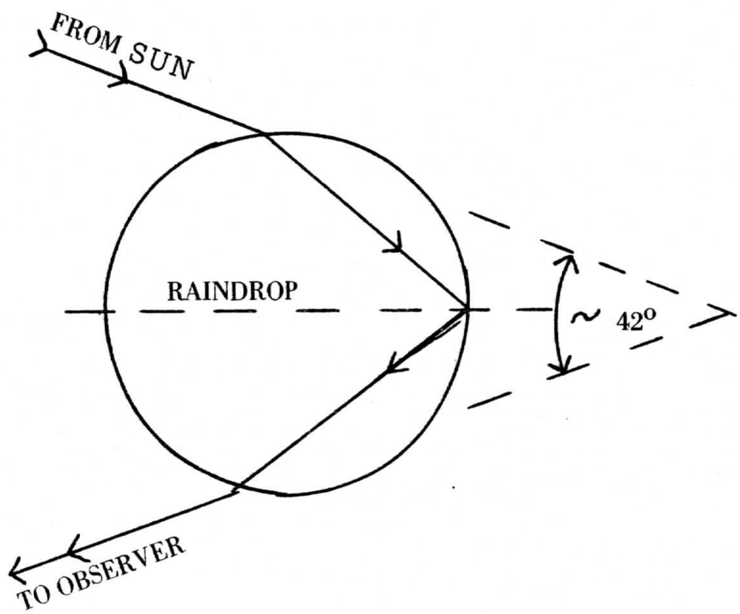
II. In air containing water droplets

a) Rainbow produced by combined action of refraction (air/water boundary) and reflection.

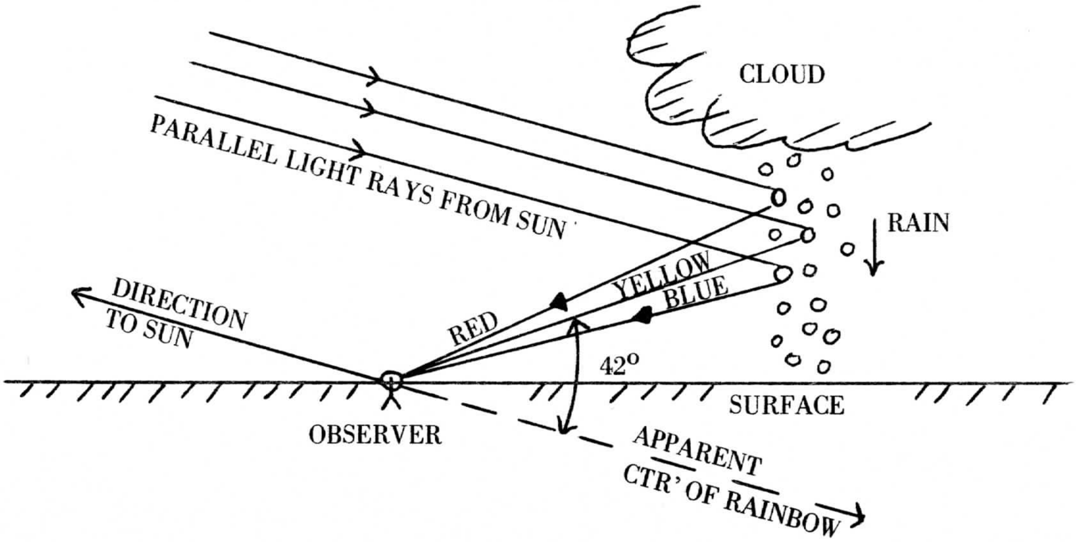
Primary rainbow—radius 42°
around a point opposite to sun as seen by observer.

Schematic

1) In a single drop:



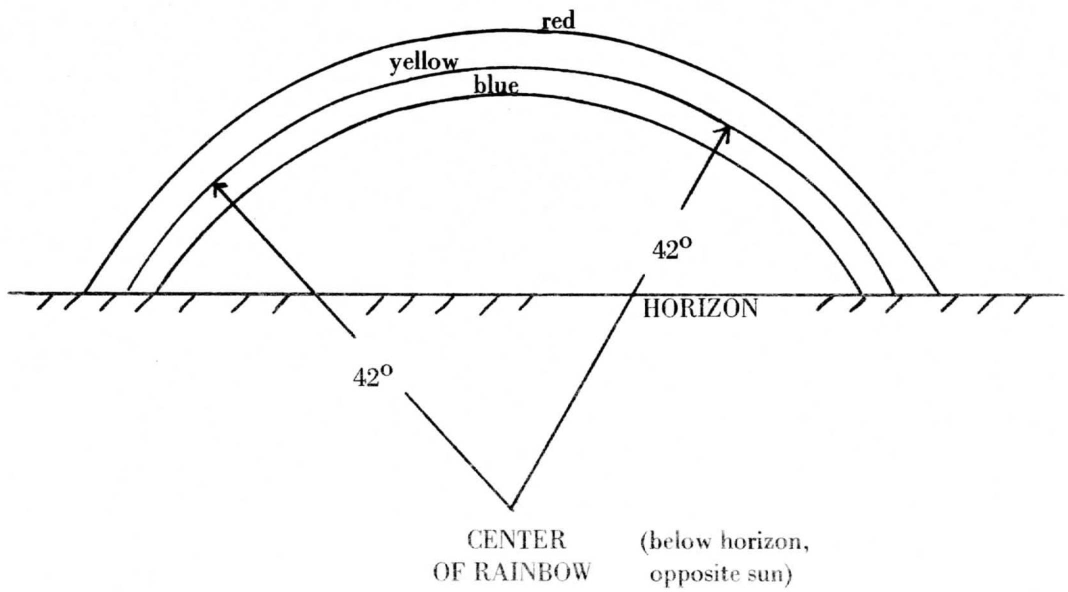
2) Many falling drops combine to form rainbow. Refraction is dependent on wave length of light: red refracted less than blue light, but when combined with reflection, the red rays appear above (outside of) the blue rays.



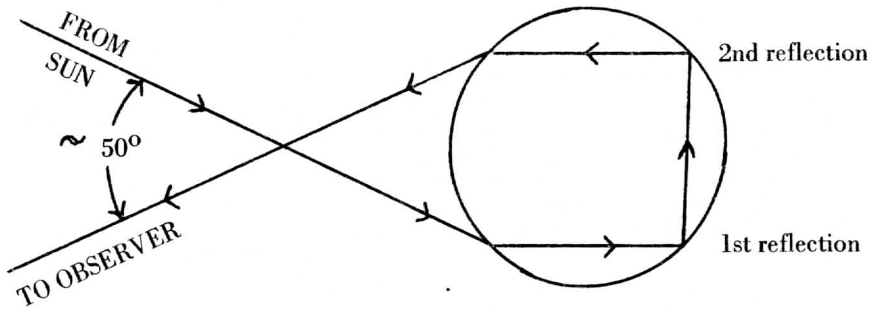
Color sequence in primary rainbow thus:

Red on outside

Blue on inside



Secondary rainbow (usually much fainter) has double reflection inside drop which reverses color sequence to outside blue, inside red. Seen outside primary at radius of 50°.



The smaller drops, the more washed-out are the colors. In drizzle or fog, we have occasionally a "fog-bow" looking only faintly colored, more whitish.

b) Corona
While rainbow appears opposite to sun, one sometimes can see refraction of light from sun in cloud droplets (water). Color separation very weak so that often only whitish glow around sun (or moon) is seen. If colors, then red appears on outside (as in rainbow). Intensity decreases very fast away from sun or moon; size of corona inversely related to drop size in cloud: the smaller drops, the larger corona. In general, size rarely more than a few degrees (or a few diameters of sun/moon).
Note that corona is seen only in clouds containing water droplets!

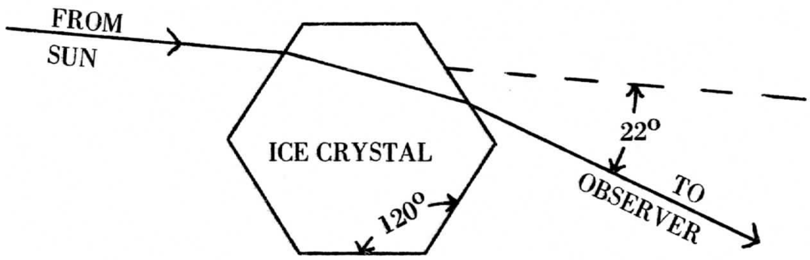
c) Glory
When on top of clouds (mountain, airplane), often brilliant colored phenomenon around shadow of object (observers head or airplane, observer is in). Due to backward scattered light from cloud droplets, colors due to interference of many such beams. Each observer sees glory only around his own head! Outlines shadow of person on top of mountain—immensely magnified—against clouds below. "Brocken Ghost."

III. In air containing ice crystals

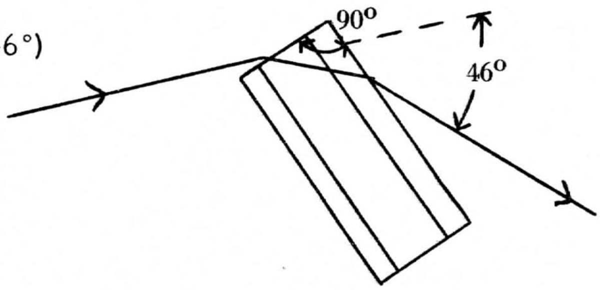
a) Halo
A distinct (usually whitish) ring around sun or moon with a radius of 22°(regular halo) or 46° (large ring). Due to refraction of light in crystals.

Schematic:

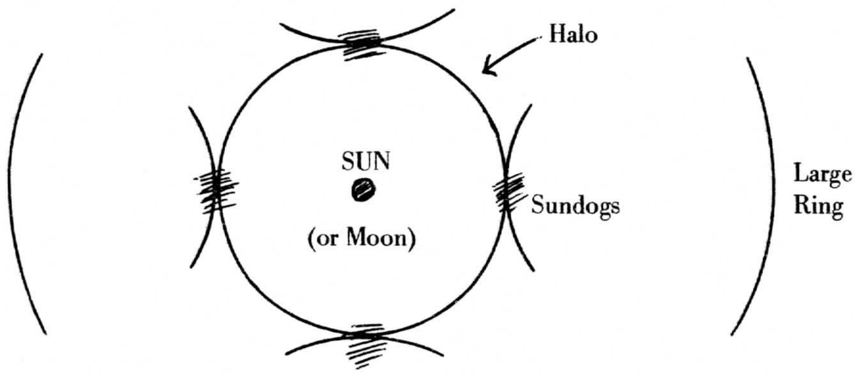
a) Ordinary or regular halo (22°)



b) Large Ring (46°)



b) Together with halo sometimes occur SUN DOGS, MOCK SUNS, PILLARS, etc. Also additional arcs tangential to halo at top, sides and bottom. All essentially due to same basic phenomena: refraction in ice. E.g. If ice needles fall downward and most are falling parallel to each other (small end forward) then spots at same height as sun/moon appear brighter, etc.



Note: Halo requires ice cloud between observer and sun or moon; occurs in cirrus = ice cloud. Corona requires water cloud; occurs in Alto stratus = water droplets. Easiest way to distinguish between Cs and As!

Since frequently Ci or Cs indicates approach of warm front (bad weather system), a halo is a fair predictor for rain or snow in the next 24-36 hours. One of the better means of predicting approach of bad weather if one does not have access to other Wx information!

EXERCISES—Chapter V

1. What is the difference between a condensation and a precipitation process?
2. What is the essential difference between condensation and sublimation nuclei?
3. Discuss the specific role of ice crystals in the formation of raindrops.
4. How do you distinguish between Cs and As?
   What optical phenomena can occur in Ci or Cs and in As or Ac?
5. How does one distinguish between "rain" and "rain shower"?
6. What is the definite difference between Cu congestus and Cb?
7. Explain formation of larger-sized hail stones.
8. What is the basic principle and schematic construction of a lysimeter? Why is it used on agricultural experimental stations?
9. Why is the secondary rainbow so much weaker than the primary?
10. Why has the sun a reddish color at sunset or sunrise? Why does it also appear to be flattened?
11. Why does the sky appear blue?
12. Difference between corona and halo?

CHAPTER VI. ATMOSPHERIC MOTIONS

(Reading: pp. 41-51, 105-116)

A. Definitions

Atmosphere moves (wind!)—transports energy, water vapor, etc.
Laws of Motion (early formulation, Newton)

3 Laws:

1) Object at rest remains at rest unless acted upon by an unbalanced force.
   Object in motion will continue in straight line with constant speed (i.e. with same velocity) unless acted upon by an unbalanced force.

2) [For constant mass:] The acceleration of an object is in the direction of the (unbalanced) force acting on the object

$$\vec{F} = m \cdot \vec{a} \quad (\vec{F}, \vec{a}, \text{same direction})$$

3) To every action there is an equal and opposite reaction.

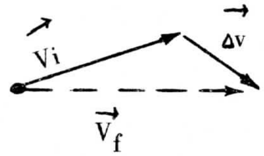
Any motion (velocity) has direction and magnitude, i.e. is a vector (see p. II-6).

Consequences of 3 laws:

1) If no unbalanced force acts, velocity of body remains the same (at rest—velocity = 0!)

2) If a body shows an acceleration, an unbalanced force must act. Direction of force given by direction of acceleration.

$$\text{Acceleration} = \text{change of velocity with time } \frac{\Delta \vec{v}}{\Delta t}$$



$\left.\begin{array}{ll}\vec{v}_i & \text{initial} \\ \vec{v}_f & \text{final}\end{array}\right\}$ velocity

Force must act in direction of $\Delta\vec{v}$.

From $\vec{F} = m \cdot \vec{a}$ we can compute $\vec{a}$ if m and $\vec{F}$ are known:

$$\vec{a} = \frac{\vec{F}}{m}$$

Example: $\vec{F}$ can be force of gravity.
$\vec{a}$ then acceleration of gravity
(often also called $\vec{g}$ = 980 cm/sec^2 at sea level)

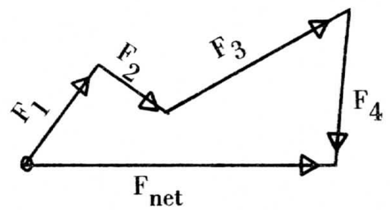
As long as $\vec{g}$ not balanced, body will accelerate. Terminal velocity (i.e. when acceleration = 0) reached when $\vec{g}$ is balanced (counter-acted) by frictional force of air on body (acting opposite to $\vec{v}$). Then no unbalanced force left, thus no acceleration = constant velocity.

EQUATION OF MOTION (General Case)

Usually more than one force can act on body. Total effect is equal to sum of all acting forces =

Resultant force ($\vec{F}_{net}$)

Since $\vec{F}$ are vectors, must be added as such.



(sum of)

$$\vec{F}_{net} = \sum_i \vec{F}_i$$

EXAMPLE: Car on road

$$\vec{F}_{net} = \vec{F}_1 + \vec{F}_2 + \vec{F}_3 + \vec{F}_4$$

$\vec{F}_1$: Driving force from motor applied to wheels

$\vec{F}_2$: Friction on road

$\vec{F}_3$: Air resistance

$\vec{F}_4$: Gravity

$\vec{F}_1$ directed in direction of travel

$\vec{F}_2$, $\vec{F}_3$ opposite to travel

$\vec{F}_4$ can be positive (downhill) or negative (uphill)

If car has constant velocity, $\vec{F}_{net} = 0$!

Then $\vec{F}_1$ must be balanced exactly by the sum of the other three forces.

B. Equation of Motion for Atmosphere

Similar form, but component forces given specific names and letter symbols:

$\vec{F}_{net} = \vec{G} + \vec{P} + \vec{M} + \vec{C}$ (Important! Remember them!)

$\vec{G}$: gravitational force; $\vec{P}$: pressure gradient force

$\vec{M}$: frictional force; $\vec{C}$: coriolis force .

Consider each separately before putting them together:

I. $\vec{G}$ = gravitational force. Acts downward.

Actually, we normally combine pure gravitational force and centrifugal force (due to earth rotation) to $\vec{G}$. [at Equator: 978 cm sec^{-2}, at Poles: 983 cm sec^{-2}]
It is chiefly important for non-horizontal motions (up/down)—should be balanced by some other force because motions in atmosphere are predominantly horizontal (velocities horizontal of order of m/sec, in vertical of order of cm/sec —except cumulus=small-scale)

[If $\vec{G}$ would be unbalanced, e.g. in body falling free, then large vertical motions would result]

II. Pressure Gradient Force $\vec{P}$

Concept of Gradient:
"Field" of values (in space) of a quantity. Pick a specific point, determine change of value of quantity over a (small) specified distance in all possible directions. Except in very special cases, one can find two directions (opposite to each other) in which this change is largest (either towards smaller or larger values).

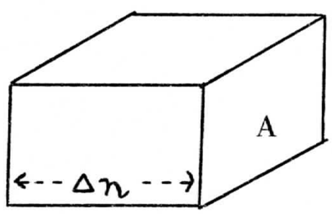
Gradient is defined by
a) one of these directions—customarily toward smaller value
b) the magnitude of the change along this direction over a specified distance

Gradient thus is a vector!

For quantity: Atmospheric pressure

Pressure Gradient $\vec{PG} = (\overrightarrow{\frac{\Delta p}{\Delta n}})$, direction to lower pressure.

Pressure gradient $\vec{PG}$ exerts a force $\vec{P}$ upon a "parcel of air" and thus will lead to an acceleration, producing wind!
Parcel of air enclosed in box with long dimension in direction of Δn, cross section area of A.



Pressure acting on "left" side of box p_1
on "right" side of box p_2
Force exerted from the left $F_1 = p_1 \cdot A$
from the right $-F_2 = p_2 \cdot A$
(opposite!)

Pressure gradient force = (net force) $= \vec{P} = \vec{F_1} + \vec{F_2} = A(p_1 - p_2) = A \cdot \Delta p$

Pressure gradient itself was $\vec{PG} = (\frac{\vec{\Delta p}}{\Delta n})$

thus $\vec{\Delta p} = \Delta n \cdot \vec{PG}$

or

$\vec{P} = \vec{PG} \cdot \Delta n \cdot A$, but $\Delta n \cdot A$ = volume V.

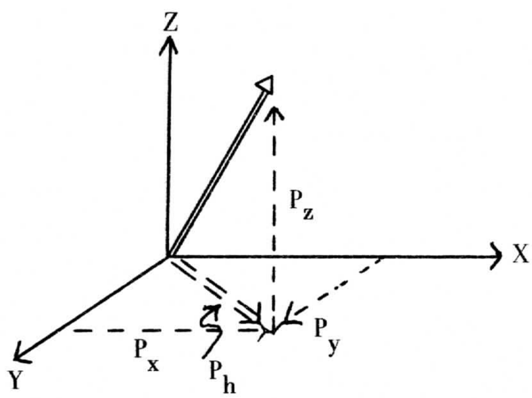
For unit mass (e.g. 1 g) V is "specific volume" $\alpha = \frac{1}{\rho}$

then

Pressure gradient force $\vec{P} = -\alpha \cdot (\frac{\vec{\Delta p}}{\Delta n})$ or $= -\frac{1}{\rho}(\frac{\vec{\Delta p}}{\Delta n})$. Minus because

$\vec{P}$ is directed along direction of Δn, opposite to Δp (i.e. from larger to smaller pressure values). In 3-dimensional space, this can be any direction.

Concept of "components," in three directions x, y, z.



$$P_x = -\frac{1}{\rho}\frac{\Delta p}{\Delta x}; \quad P_y = -\frac{1}{\rho}\frac{\Delta p}{\Delta y}; \quad P_z = -\frac{1}{\rho}\frac{\Delta p}{\Delta z}$$

[also $P^2 = P_x^2 + P_y^2 + P_z^2$]

Previously (p. IV-2) we had used <u>hydrostatic</u> <u>equation</u>
[Change of pressure with height]

$$\Delta p = -g\rho\,\Delta z$$

From this follows $\underline{g = -\frac{1}{\rho}\frac{\Delta p}{\Delta z} = P_z}$

Since P_z is directed upward (toward low pressure), but g (acceleration of gravity) downward, both values are equal in magnitude but opposite in direction, thus they will balance each other.

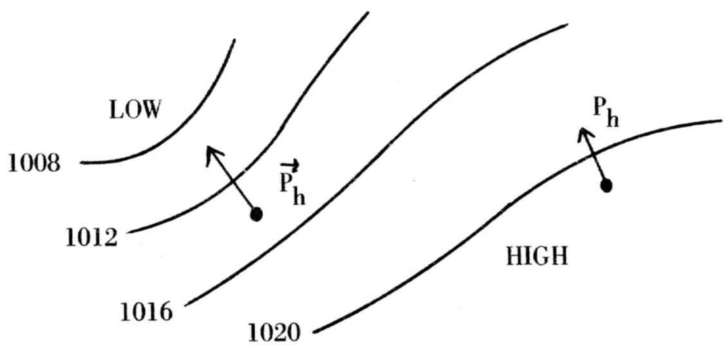
"Vertical component of pressure gradient force is balanced by gravitational force ($\vec{P}_z$ upward, $\vec{G}$ downward)"

Balanced force = no resulting acceleration (Newton)

However, the horizontal pressure gradient forces (P_x, $P_y \rightarrow \vec{P}_h$) are not balanced, thus can cause acceleration.

Horizontal pressure gradients $\vec{P}_h$ from:

a) Map of constant level pressures—isobars (often, constant level is "sea level")



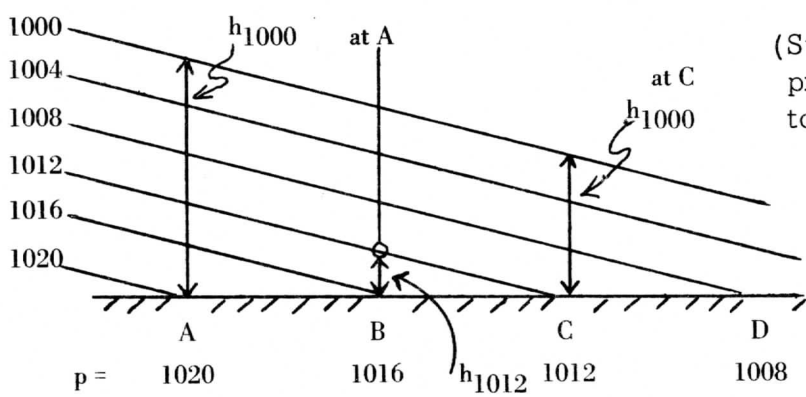
$$\vec{P}_h = -\frac{1}{\rho}\frac{\vec{\Delta p}}{\Delta n}$$

if Δp = 4 mb (constant) then:

the smaller Δn (distance, the larger $\frac{\Delta p}{\Delta n}$

"Closely spaced isobars, strong pressure gradient."

b) Map of "height of constant pressure level"
= "Pressure contour maps"
(usually used for upper layers of atmosphere)



(Surfaces of constant pressure slope upward toward high pressure)

One draws a map of the heights (at various stations) of a specific "constant pressure" surface (usually specified levels as 1000 mb, 850 mb, 700 mb, 500 mb, 300 mb, etc.)

Slopes of these constant pressure surfaces are very gentle (of order of 10 m per 100 km or 1:10,000) so that one can derive $\vec{\Delta p}/\Delta n$ from spacing of contours.

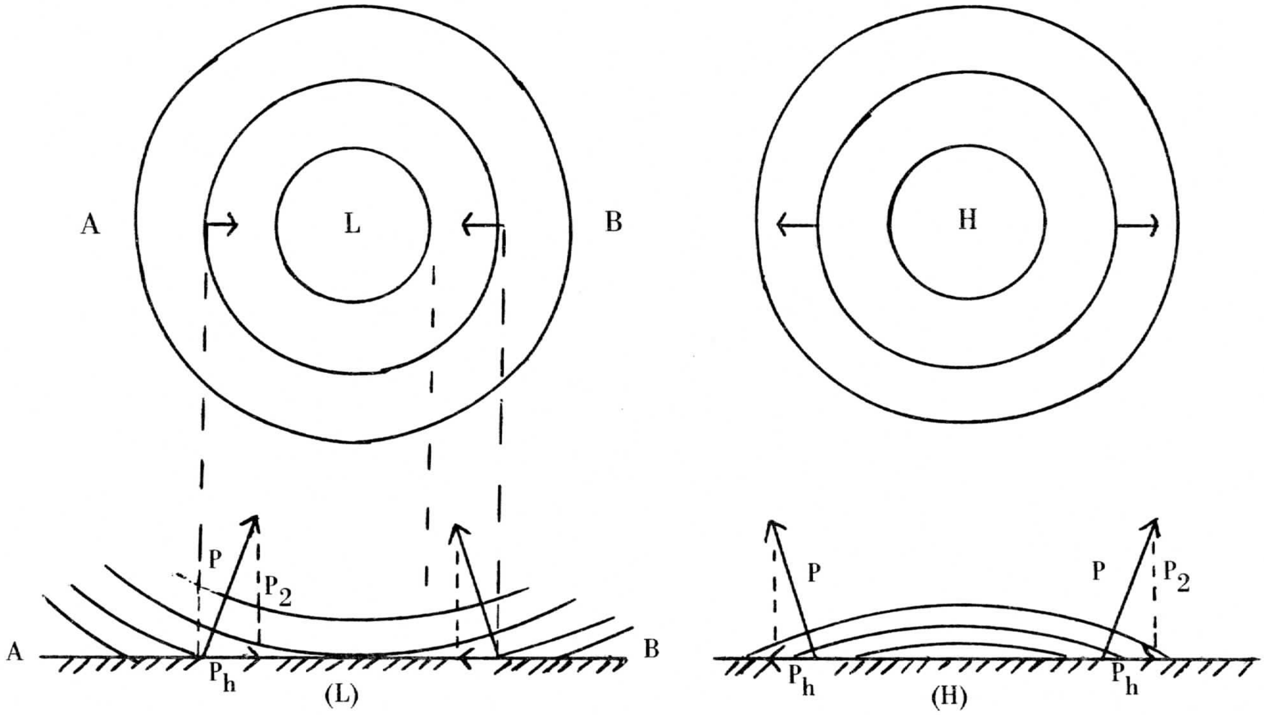
(Replace Δp, using hydrostatic equation, $\Delta p = -g\rho\Delta z$ i.e.)

$$\vec{P}_h = -\frac{1}{\rho}\frac{\Delta p}{\Delta n} = -\frac{1}{\rho}\cdot -g\rho\cdot\frac{\Delta z}{\Delta n} = g\frac{\Delta z}{\Delta n}$$

Note that factor $1/\rho$ disappears, g is constant, so that

$g\dfrac{\Delta z}{\Delta h} = \vec{P}_h$ is the same for all levels.

Relation between $\vec{P}$ and $\vec{P}_h$ in Low and High:



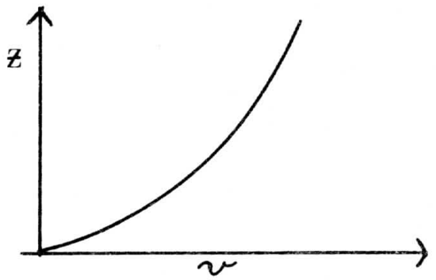
While $\vec{P}$ is directed chiefly upward (toward lower pressure), $\vec{P}_h$ is directed toward inside of low
outside of high, because

$\vec{P}_h$ always directed toward low pressure across isobars (or contour lines), perpendicular to the local isobar at the given point of interest.

III. Frictional Force $\vec{M}$

Main friction occurs by air flowing over surface terrain. "Wind resistance!"

Wind<u>speed</u> change with height



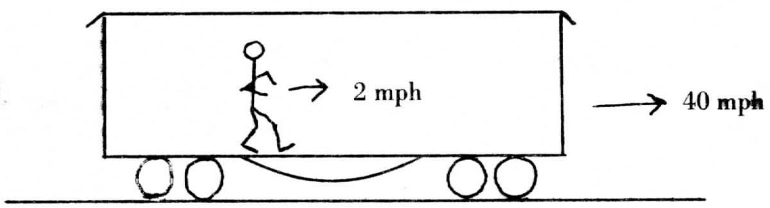
Complicated problem if one wants to formulate it exactly. However, common sense allows to draw certain conclusions:

a) The rougher the surface, the more friction
b) The farther away from surface, the less effect on flow of wind. Experience (and theory) has shown that effect of frictional forces becomes negligible at height levels of above 1,000 m or 3,000 ft <u>above ground</u>.
c) If air flow is more turbulent, then "air bubbles" rising will have to be accelerated by flow at greater height—this requires energy which is supplied by velocity in upper layer; consequently wind will be affected more (and to greater height) if more turbulence. Result: in <u>daytime</u> (more ascending motions due to less stable lapse rates) frictional effect will reach <u>higher up</u>, in the <u>night</u> (less turbulence due to stable lapse rates) frictional effect is more <u>confined to lower layers</u>.
d) Frictional effects on oceans (smooth sfc) usually much less than over land. Thus, over oceans close to surface stronger winds than over land—other conditions being equal (i.e. driving forces, etc.).

IV. Coriolis Force $\vec{C}$ (G. C. Coriolis, ~ 1884)

Concept of "relative motion." Newton laws refer to a specific reference system. E.g. "straight" line—has to be defined. Until now, we have not yet defined our system of reference.

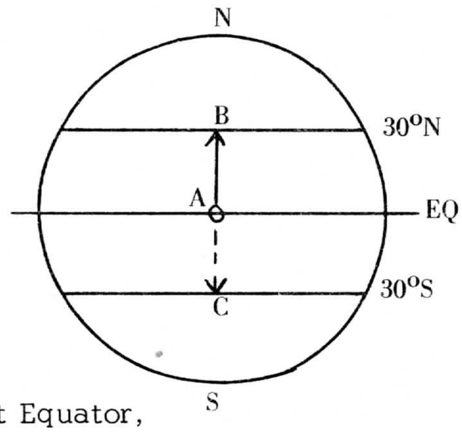
Example: Man moving (walking) inside railroad car which itself moves along.



Velocity of man in car? With respect to what?
Relative to car 2 mph, but relative to earth sfc 42 mph

Since we live on a rotating earth, our "reference system" usually is given by longitude/latitude on earth, i.e. fixed to earth. However, different points on earth also move due to rotation of earth, and more specifically, move with different speed!

Thought experiment: Launch missile at equator toward a point exactly north (i.e. same longitude) at 30°N



Assume speed of missile such that flight time is 10 min = 600 sec. (~ 5 1/2 km/sec)

When launched at A, missile has velocity v_m toward N from launch, but also, on rotating earth, has the velocity of point A due to earth rotational speed, v_E .

At Equator,

$$v_E = \frac{40{,}000 \text{ km}}{24^h} \sim 463 \text{ m/sec.}$$

When arriving at 30°N, it has moved 30 latitude degrees (~ 3300 km) to north but also 600 sec x 463 m/sec ≅ 280 km toward the East. Where is now Pt. B after these 10 min? At latitude 30°N, tangential speed of Earth is slightly smaller than at Equator (~ 401 m/sec). Accordingly, Pt. B also has moved East, but only by ~ 240 km. Result: Missile, aimed originally at B, now hits a point (280-240) = 40 km to the East of B.

An observer at A, looking along flight path, thus sees the missile not going straight toward target, but deviating to the right (curving off target).

Obviously—the longer flight, the larger deviation. This effect of earth rotation thus important for large-scale motions over long distances, rather unimportant for short distances.

Verify the next two statements:

1) If missile launched from B toward A, deviation as seen from B is also toward the right!
2) If missile had been launched from A toward the south (i.e. into the southern hemisphere toward Pt. C) deviation, looking in direction of flight, would have been to the left!

Discussion of the Coriolis force and its effect.

The total Coriolis force vector is a three dimensional vector which has, at any point of the Earth surface, a specific horizontal and vertical component - somewhat similar to $\vec{P}$. Since we will be talking chiefly about horizontal (or very near to horizontal) motions, only the horizontal component $\vec{C}_n$ will be needed - corresponds to P_n discussed above. It should be understood that our thought experiment showing the deflection of N/S or S/N motion is incomplete. For example, it is very difficult to show in this way that E/W or W/E motion also will be deflected. The actual theory in three dimensions, however, can prove this but requires mathematics beyond the scope of this course.

When looking in the direction of (horizontal) motion, the deflection was shown to "force" the object to the right on the Northern and to the left on the Southern hemisphere. Also we see that this deflection only occurs in a reference system which is fixed to the rotating earth. From space the motion appears to be straight without this effect. This says that a) on a non-rotating planet $\vec{C}$ does not exist and b) that this force depends on the motion of the object - obviously a object at rest does not experience this force. The Coriolis force is an apparent force. Change of direction of a motion (mathematically expressed by a velocity vector) means that the velocity vector changes. In order to explain this change of the $\vec{v}$ vector, we must assume that an additional force acts to imbalance the force leading to the motion itself. This "apparent force" identified above as $\vec{C}$ leads to an apparent acceleration (due to the rotation of the Earth) called the Coriolis acceleration $\vec{a}_C$ [recall $\vec{F} = m \cdot \vec{a}$]. Its horizontal component can be described by the two values: magnitude and direction.

Magnitude: $a_c = v \cdot 2\Omega \cdot \sin\phi$

v = speed of object, ϕ latitude of point where a_c acts

(+ on N.H., - on S.H.), Ω = rate of angular speed of Earth

(equal to $\frac{360°}{24h} = 0.000073$ radian/sec).

Quantity: $2\Omega\sin\phi$ called Coriolis parameter f.

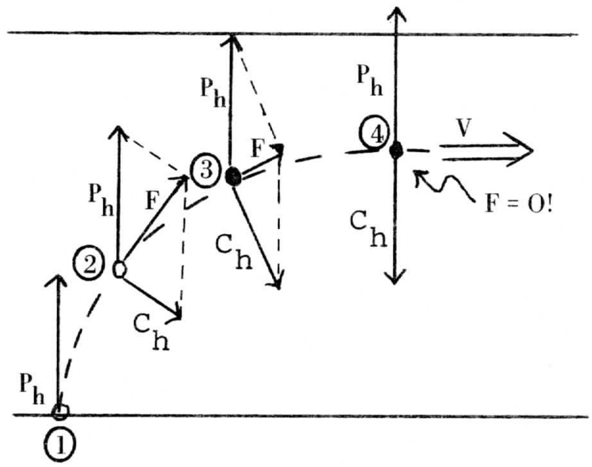
Direction: perpendicular to direction of motion (i.e. to $\vec{v}$) to the right on N.H., to the left on S.H.

From this formula also follows that $a_c = 0$ for $\phi = 0$. (Coriolis force, horizontal component is zero at Equator) Conversely, a_c is largest at the two poles ($\phi = \pm 90°$)

[For completeness: the magnitude of the vertical component of the Coriolis acceleration is $v \cdot 2\Omega\cos\phi$]

C. Application to Atmospheric Motions

(Northern Hemisphere)   Low



High

1 $\vec{C}_h = 0$ because $v = 0$; $\vec{F} = \vec{P}_h$

2, 3 $\vec{C}_h$ partially compensates $\vec{P}_h$; $\vec{F} = \vec{P}_h + \vec{C}_h$

4 $\vec{C}_h = -\vec{P}$ (opposite)
$\vec{F} = 0$--no acceleration
i.e. $\underline{\vec{v} = \text{constant}}$.

Parcel at 1 starts motion, following direction of $\vec{P}_h$; however, coriolis acceleration turns motion to right until finally, at 4 , equilibrium is reached, constant velocity resulting from disappearing of net force.

[Note that this is a very "approximate" treatment of a more complicated problem; in nature nearly complete balance will usually exist!]

From above, we see that, at final stage (pt. 4) a very simple state of affairs exists:

In equilibrium, $\vec{P}_h$ and $\vec{C}_h$ are equal but opposite and the (constant) wind velocity is a vector whose direction is parallel to isobars. In Northern Hemisphere, low pressure to left, in Southern Hemisphere to right when looking into direction to which wind is blowing.

Note that we have not included $\vec{M}$ (friction), i.e. we should be talking about wind above, say, 1,000 m height above ground. We know direction—what about speed?

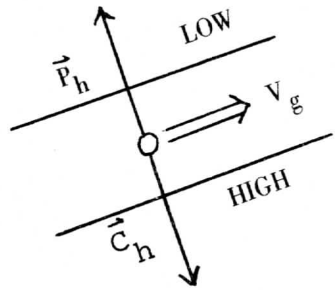
D. Geostrophic Wind

We had before $\vec{F}_{net} = \vec{G} + \vec{P} + \vec{M} + \vec{C}$

We already know (1) that $\vec{G}$ is balanced by $\vec{P}_z$
(2) that $\vec{M}$ is nearly zero for heights of more than 3,000 ft (1,000 m) above ground.

If we now look at the case of a "steady" wind (pt. 4 in diagram) then $\vec{F}_{net} = 0$; using these simplifications we can write the general equation for $\vec{F}_{net}$ as

$$\vec{F}_{net} = 0 = \vec{P}_h + \vec{C}_h$$



From this equation we can compute $\vec{v}_g$ (where subscript g denotes "geostrophic"); actually we only need v (speed = magnitude of velocity; we already know direction!)

We had $P = -\frac{1}{\rho}\frac{\Delta p}{\Delta n}$ and also $C_h = V \cdot 2\Omega \sin\phi$

the two together should be zero, i.e. $-\frac{1}{\rho}\frac{\Delta p}{\Delta n} + v_g \cdot 2\Omega \sin\phi = 0$

Solved for v_g:

$$v_g = \frac{1}{\rho \cdot 2\Omega \sin\phi} \cdot \frac{\Delta p}{\Delta n}$$

Thus, if I know a) spacing of isobars or contours
b) latitude

then I can directly compute speed of geostrophic wind!

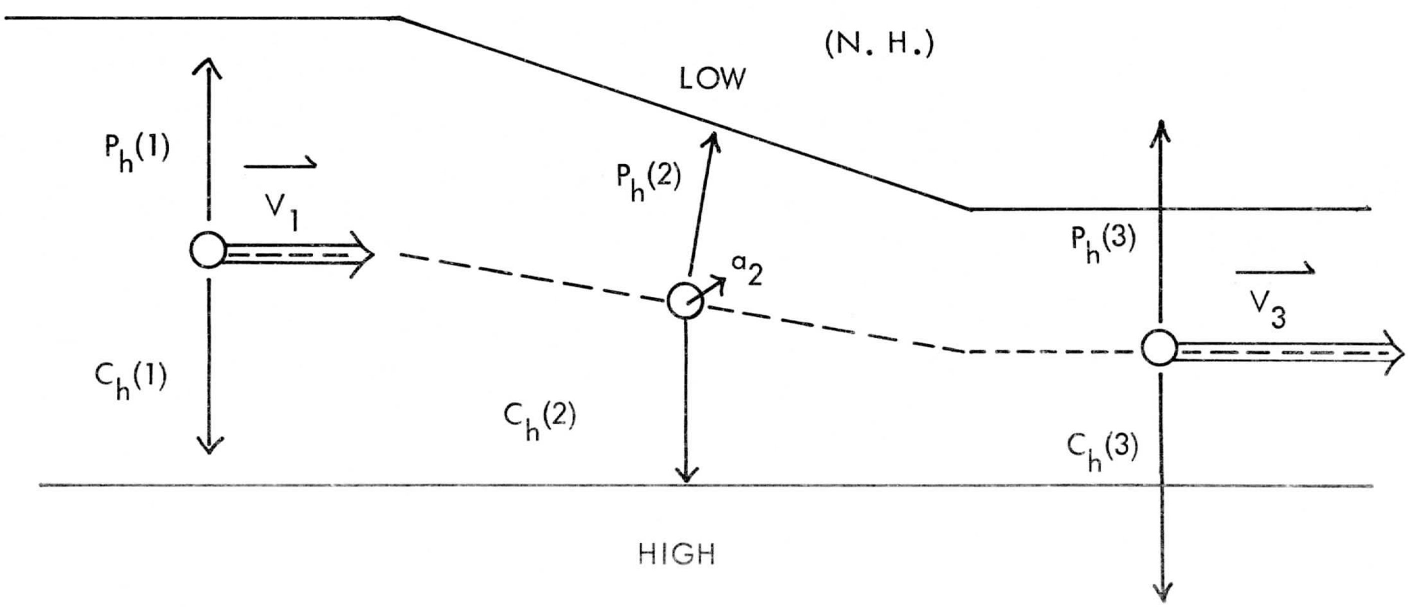
Note, that for $\phi = 0°$ (i.e. at equator) our formula is

$$v_g = \frac{1}{\rho \cdot 2\Omega \sin 0°} \cdot \frac{\Delta p}{\Delta n}$$

and $\sin 0° = 0$. Consequently, it is not possible to evaluate v_g at equator since denominator of ratio is zero!

Result: The geostrophic assumption is poor in regions close to the equator (e.g. ± 5-10° latitude)

E. Convergence—Divergence



At Pt. 1: $P_h(1) + C_h(1) = 0$; same at Pt. 3.

Steady geostrophic wind, parallel to isobars

At Pt. 2: $P_h(2) + C_h(2) \neq 0$

Wind still has retained prior velocity vector from Pt. 1 (approximately, inertia) (thus $C_h(2) = C_h(1)$), but $P_h(2) > P_h(1)$ since isobars have converged and pressure gradient force has thus increased. Also, the direction of isobars through Pt. 2 is different from that at Pt. 1. Result: The addition of the two vectors leads now to an additional acceleration vector a_2 which will accelerate air parcel:

a) points "forward", will thus increase velocity
b) points toward low pressure, will also turn motion slightly toward low pressure, allow cross-isobaric flow of air from higher to lower pressure. This in turn can change the pressure field itself.

(If air only would move always parallel to once existing isobars, no air could be transported out of highs or into lows, i.e. pressure could not change!)

The same analysis can be made for <u>diverging</u> isobars. The result is a deceleration vector, pointing "backward" and slightly toward high pressure, resulting in cross-isobaric flow toward the high.

Thus:

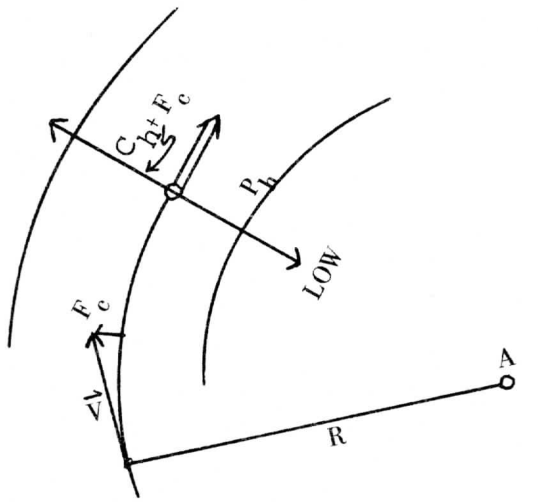
CONVERGENCE - Increase in speed, cross-isobaric flow toward low pressure

DIVERGENCE - Decrease in speed, cross-isobaric flow toward high pressure.

These effects are small locally but can act over long distances and in the whole atmosphere. They are extremely important for the understanding of changes in the pressure and wind fields in the atmosphere.

F. Gradient Wind

If isobars are not straight lines but curved, wind is forced to move (parallel to isobars) on a curved path.



$\vec{F}_C$

Centrifugal force results, directed toward the outside.
Since P_h is given, now the balance is achieved by combined action of $\vec{C}_h$ and $\vec{F}_C$. $\vec{C}_h$ thus can be smaller, resulting in $\vec{V}$ also being smaller.

If motion curves around Pt. A with radius R then $\vec{F}_C = V^2/R$, directed outward.

Result: for cyclonic curvature (around Low)

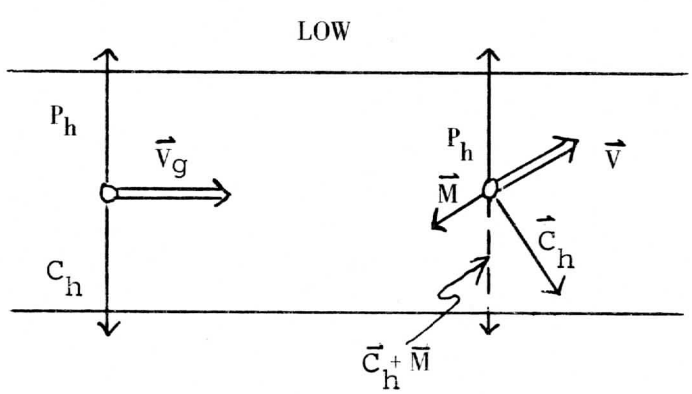
Gradient wind slightly smaller than Geostrophic.

This is example for Southern Hemisphere. Note wind is clockwise around low.

Effect normally very small (curvature usually very slight because R is very large—e.g. 500 miles or so). Only time meteorologists use gradient wind is in computing windspeed in hurricanes/typhoons (high V and small R).

Difference in this case can reach values of up to 20%.
No need to consider the anticyclonic case.

G. Effect of Friction ($\vec{M}$)



without friction:

1 $\vec{P}_h + (\vec{C}_h) = 0$ Geostrophic

with friction:

2 $\vec{P}_h + (\vec{C}_h + \vec{M}) = 0$

or

$-\vec{P}_h = (\vec{C}_h + \vec{M})$

$\vec{C}_h \perp$ to $\vec{V}$

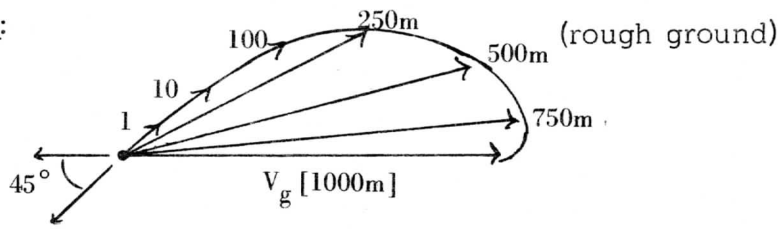
$\vec{M}$ opposite to $\vec{V}$

$\vec{P}_h$ now balanced by sum $(\vec{C}_h + \vec{M})$

Result: Wind now deviates toward low pressure

Wind speed less (since $\vec{C}_2 < \vec{C}_1$)

Ekman spiral:



Lowest speed and largest deviation from direction of geostrophic wind in the layers close to the ground. When ascending, largest change first in wind speed, in higher layers then stronger change in wind direction (turning toward the geostrophic wind). On top of friction layer, finally, wind becomes equal to geostrophic wind ($\vec{M}$ goes to zero).

Deviation in lowest layers depends on strength of frictional effect—
smooth surface—angle ~ 20-25°
rough surface—angle up to 45°

Over oceans (very smooth as compared to nearly all land) frictional effect much smaller—thus usually stronger wind near surface than over land when same pressure gradient exists.

Results of frictional effects:

a) Without $\vec{M}$

Wind geostrophic
i.e. parallel to isobars
thus <u>no</u> cross-isobaric flow

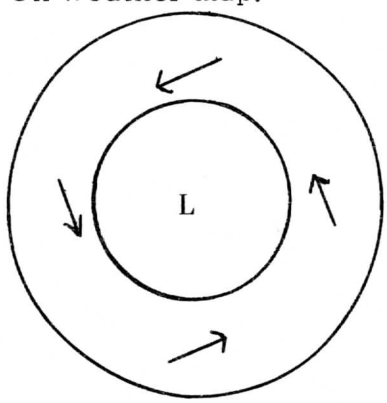
b) With $\vec{M}$ (near surface)

Wind deviating from geostrophic
i.e. across isobars (at angle)
thus cross-isobaric flow from higher to lower pressure

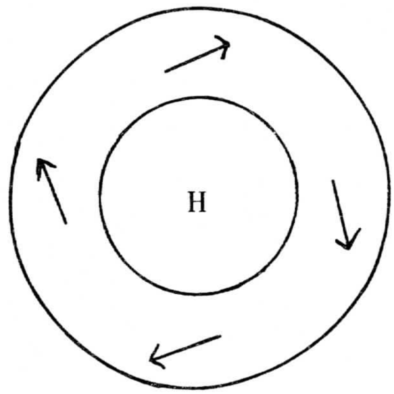
In a Low, air (near sfc) is transported into system—CONVERGENCE

In a High, air (near sfc) is transported out of system—DIVERGENCE

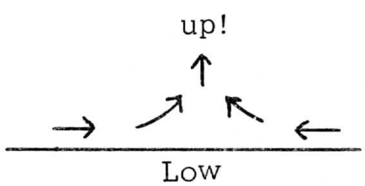
On weather map:

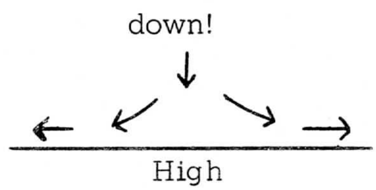


(N. HEM.)



In cross section:





CONVERGENCE
thus means upward motions

Air moved up: follows dry-adiabat until saturation, then moist adiabat
Result: condensation, clouds, possibly precipitation

"Bad" Weather

DIVERGENCE
downward motions

Air moved down: follows always dry adiabat
Result: Drying out, dissolving of clouds, clearing

"Fair" Weather

H. Variation of Wind with Height

Wind in any given level is determined by pressure field. Thus, wind will change with height if pressure field changes with height. The latter can (and does) occur due to horizontal temperature differences.

From hydrostatic equation $\Delta p = -\rho \cdot g \cdot \Delta z$

and equation of state $p = R\rho T$ or $\rho = \frac{P}{RT}$

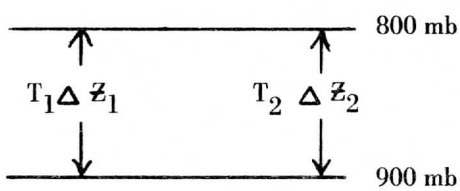
we obtain

$$\Delta z = -\frac{RT}{g}\left(\frac{\Delta p}{p}\right)$$

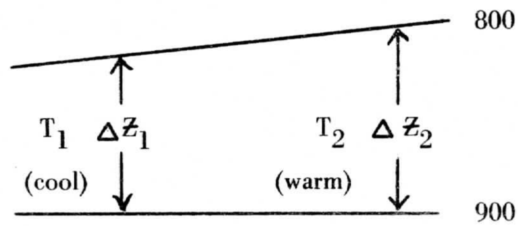
For a given (vertical) pressure difference Δp, the height difference of the two isobaric surfaces is directly related to the temperature of the air in-between. Thus, if this temperature changes from one place to another, the height difference Δz also will change, resulting in a change of the contour field with height. Consequently the wind field (which can be derived from the contour field) also will change with height.

Example:

a)



b)



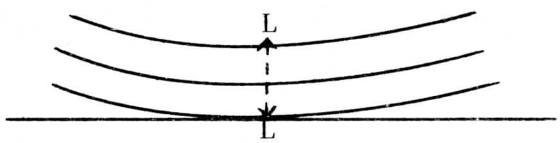
if $T_1 = T_2$, then $\Delta z_1 = \Delta z_2$

if $T_1 < T_2$, then $\Delta z_1 < \Delta z_2$

Now consider cross section through a Low:

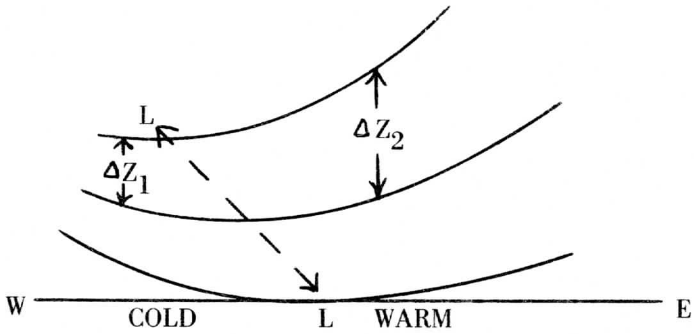
(Contour surfaces)

a) Temperature throughout the same:



Since T is the same in all regions of Low, Δz is the same; all contours are parallel, axis of low is vertical, shape of contour sfc's stays the same.

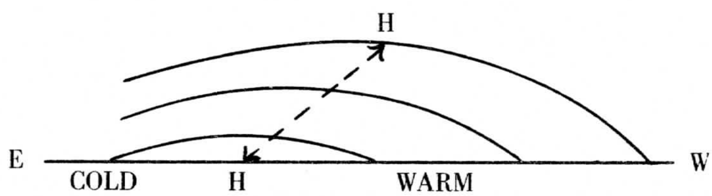
b) Temperature on "left" side is lower than on right side:



Now $\Delta z_1 < \Delta z_2$, so that contours no longer parallel.

Axis of Low tilts toward cooler air!

Similarly, in a HIGH, axis also will lean, but toward warmer air.



Remembering the flow around a low (N. Hem.: counterclockwise) or a high (clockwise), we can also see that:

Usually cold air (with N. wind) behind low/in front of high.

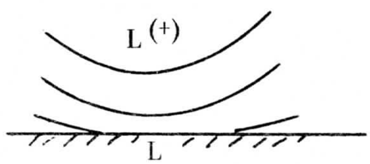
Thus, the upper low will be usually behind (west of) sfc low; the upper high also behind (west of) sfc high.

Finally, situation of a low or high containing in their centers either warm or cold air.

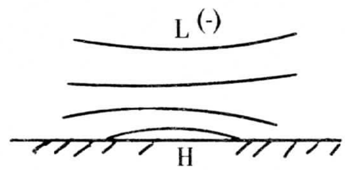
Schematic:

Low | High

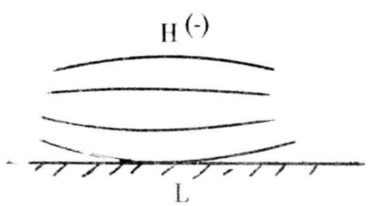
Cold core 1)



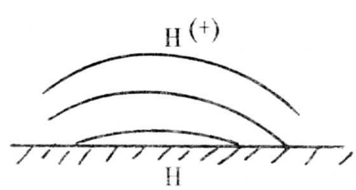
2)



Warm core 3)



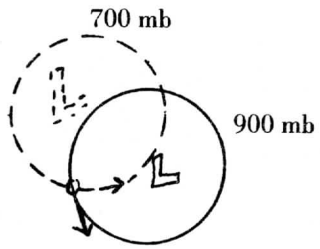
4)



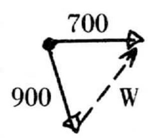
Cases 1)	Cold core Low:	Intensifies with height
4)	Warm core High:	
Cases 2)	Cold core High	Intensity decreases with height and may change over to opposite system at high level (i.e. sfc L → upper H or sfc H → upper L)
3)	Warm core Low	

From change of pressure field, one then can also obtain wind changes.

For example: A sfc Low with cold air in WNW



At Station A: 900 mb wind NW
700 mb wind W



Thus, we have a wind SHEAR vector
(difference vector $\vec{V}_{700} - \vec{V}_{900}$)
This vector points in such a way that <u>cold</u> air is to left.

In usual (geostrophic) wind <u>low pressure</u> to left (N. Hem.)
In shear vector ("thermal wind") <u>low temperature</u> to left.

<u>Note</u> that this vector is called "<u>thermal wind</u>"—but it does <u>not</u> really represent a true wind vector, but a vector describing the wind <u>change</u> with height, due to the horizontal temperature difference.

In general, wind changes both direction <u>and</u> speed with height—one can always express this change by means of a thermal wind vector. Knowing the temperature field one can construct the thermal vector field resulting from the temperatures—then by adding this to the wind at a lower (or higher) level, one can construct the wind field at the other level.

EXERCISES—Chapter VI

1. Convert mph to m/sec: wind speed of 10 and 30 mph.

2. A car drives along a straight highway with a speed of 30 mph (at time $t = 0$)
   a) One minute later ($t = 1$ min) the car has been accelerated to a speed of 60 mph. Calculate the acceleration (in units miles per minute squared).
   b) The car is brought to a complete stop from 30 mph within 2 seconds. Calculate the (negative) acceleration in the same units as above.
   c) Convert the two accelerations obtained in (a) and (b) to units cm/sec^2.
   d) Compare (as ratios) these accelerations to the unit of "1 g" (= 980 cm sec^{-2}, acceleration of gravity) i.e. how many g do you experience in the two cases above—and in which direction do these g-forces act?

3. The earth revolves around the sun in an ellipse. Obviously, then, it does not follow a straight line path. What do you have to assume to reconcile this fact with Newton's laws?

4. a) The earth is not a perfect sphere but slightly oblate. Why?
   b) The oblateness of the earth is only 1 part in about 297; for Jupiter, we observe an oblateness of about 1 part in 10. Why?

5. Compute the magnitude of the (total) pressure gradient force P, given the three components

$$P_x = +40 \text{ dynes}, \quad P_y = -30 \text{ dynes}, \quad P_z = 0 \text{ dynes}.$$

(Only for students who had high school physics and math! [Nos. 6-9])

6. Knowing the fact that the upward pressure change between 0 and 1 km is about equal to 100 mb, calculate the mean air density in this layer. Assume $g = 10^3$ cm sec^{-2} and calculate in powers of 10 to simplifiy matters.

7. Assume a horizontal pressure gradient of 10 mb per 100 km; also assume that a vertical pressure change of 10 mb corresponds to a $\Delta z = 100$ m. Calculate the slope of the pressure surface (1:x) and the angle it makes with the horizontal.

8. Compute the Coriolis acceleration (cm sec^{-2}) for an air parcel moving with a speed of 10 m/sec
   a) at latitude 30°N ($\sin 30° = 0.5$)
   b) at the equator
   c) at the North Pole ($\sin 90° = 1$)

9. Calculate the geostrophic windspeed for a pressure gradient (at sea level) of 7.3 mb per 1,000 km (at latitude 30° N)
   a) in units m/sec
   b) in units mph

10. The wind at the 900 mb level is observed to come from the West. At 500 mb, the wind is observed to come from the North with the same speed. In which direction from the station do you find the cold air in the layer 900-500 mb?

CHAPTER VII. THE GENERAL CIRCULATION

(Reading: pp. 67-72, 78-79)

1. Scales of Motion in Atmosphere

Earth—a planet, energy received from sun ultimate cause of all processes on earth and in atmosphere. But: Most of radiation energy first goes to surface (solid or liquid) and then is redistributed into atmosphere or soil/water by very small-scale processes (conduction, convection at boundary layer). Result is that there are a large variety of scales of motions and processes, from very small ones to largest possible, i.e. of global scale. Similarly, certain processes may go on in extremely short time spans while others take very long times to get going or to achieve changes from one state to another. This infinite variety of scales (both in dimension and time) has been subdivided for convenience into four scale sizes. This is convenient for a number of reasons. For example, if I talk about global processes, I usually do not have to take into account small-scale local influences or short-term variations; on the other hand, when investigating short-term, small-scale processes, I usually do not have to concern myself with such effects as year-to-year changes in albedo, etc.

One can arrange these scales of processes in the following way:

Order	Usual name	Scale/length	Scale/time	Examples
1st	General circulation	global 10,000-1,000 km	years-months-weeks	Westerlies in temperate latitudes
2nd	Synoptic Scale Systems	continental Wx-systems (low, high) 1,000-100 km	weeks-days	Individual Low or High
3rd	Meso - Scale Systems	local 100-0.1 km	days-hours-minutes	Thunderstorm
4th	Micro - Scale Systems	small 100m-less 1 cm	minutes-seconds and less	Turbulence, "air pocket"

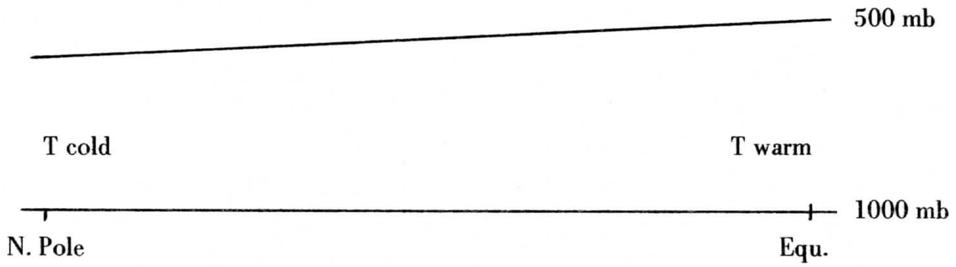
In the following paragraphs, these different scales and examples of the processes will be discussed in detail.

2. General Circulation - Basic arrangement

a) The Westerlies

Basic feature of physics of earth-atmosphere system is the fact that maximum energy is received near equator in Tropics and least energy near Poles. To achieve equilibrium, energy must be transported: exported from Tropics, imported to Polar regions.

To simplify, let's assume that sea level pressure over whole earth is the same. Then 1000 mb pressure surface (contour) is horizontal and thus parallel to sea level. Result would be no geostrophic wind (no pressure gradient). However, Tropics are warm, Poles are cold—thus the horizontal change (N. Hem. from south to north) in temperature will introduce a change in Δz (distance of pressure surfaces [contours] at higher levels).



E.g. 500 mb contour surface slopes toward the pole.
Result is: Lower pressure to the North, therefore wind from the West (due to Corioli's force)
i.e., we have Westerly winds in temperate latitudes.

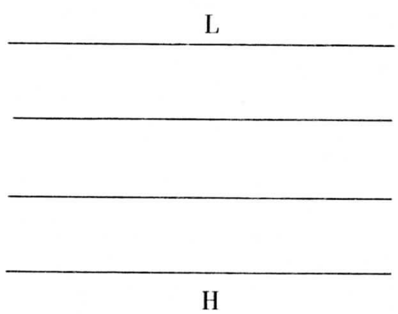
On Southern hemisphere, same thing happens; now low pressure to the South, but Coriolis force acts also in opposite direction; therefore again

Westerly winds in temperate latitudes.

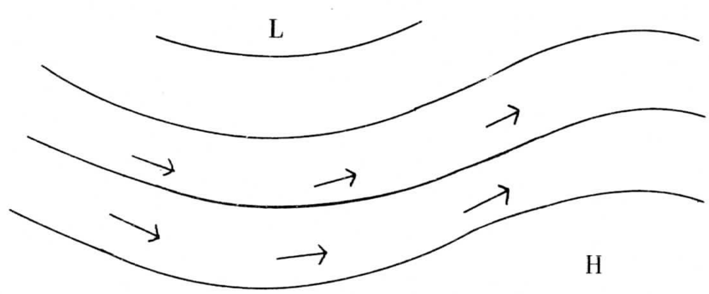
This is true even if there were no winds at all at lower levels. If one had already W-winds in some regions, they would increase in strength considerably with height—if we had E-winds near the ground, they would probably decrease and, in upper levels, change over to W-winds.

The "Westerlies" in the temperate latitudes of both hemispheres, however, could not transport energy from equator toward the poles if they would blow, in all places, strictly from West to East. Remember, that air above the friction layer obeys very nearly geostrophic conditions.

In order to transport heat = energy, north or south wind components have to exist; in this case, Northerly flow will bring cold air towards warmer regions and Southerly flow will bring warmer air into cooler areas. Averaged around the whole hemisphere (e.g. along one latitude circle), the sum total of these windspeed components can be zero, however a "net transport" of energy northward still will result. The flow in the westerlies then is no longer straight W-E but becomes "distorted" into a long-wave pattern.



Straight W-E from No transport N/S



Distorted "wave" flow transport N/S

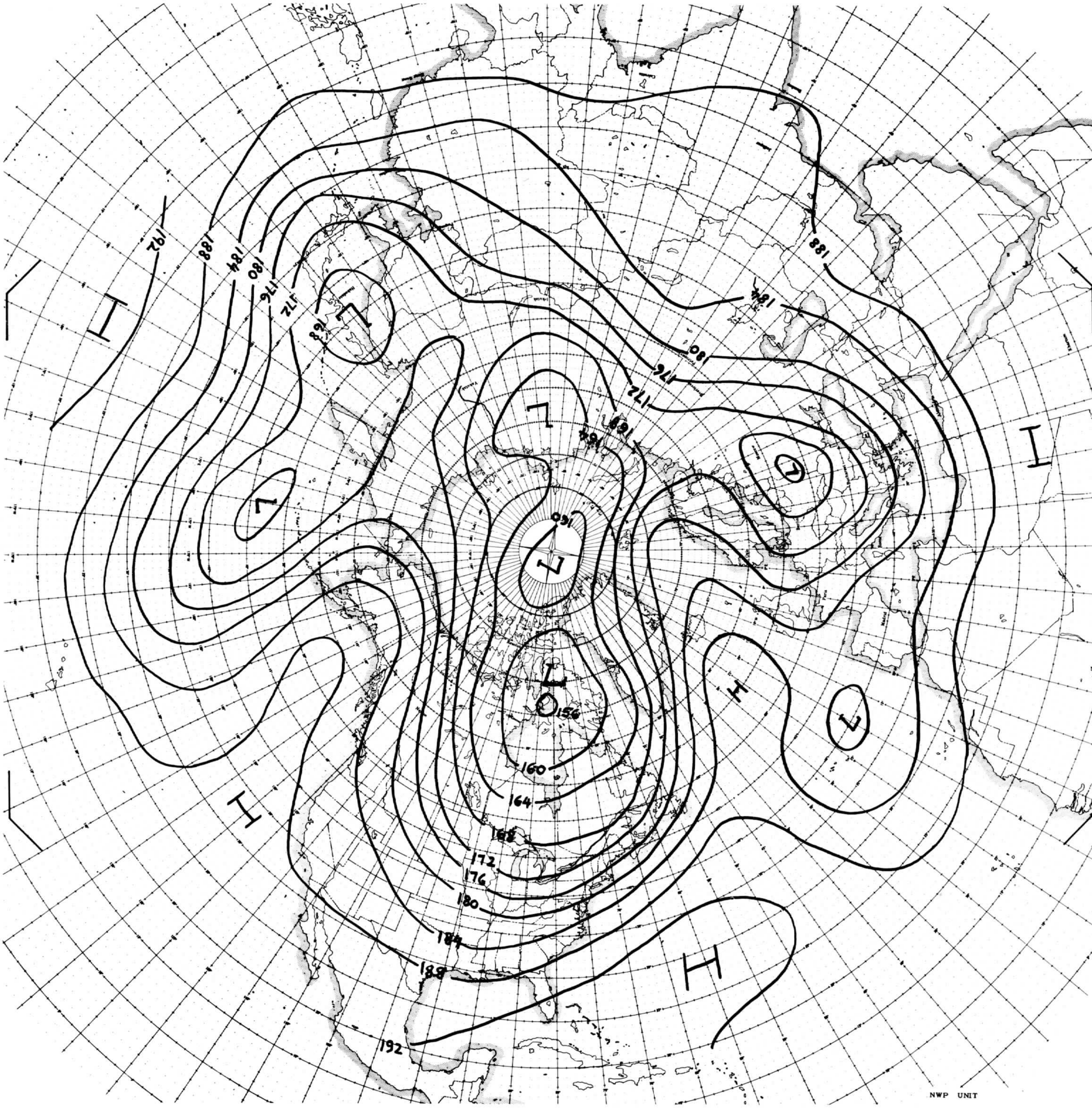
These long waves in the westerlies are called ROSSBY waves after C. G. Rossby who first explained the dynamic reasons for their existence in 1940. The Rossby waves constitute one of the most important features of the general circulation. Their behavior is closely related to the energetics of the circulation, their motion regulated by the speed of the westerly flow and their composite action leads to the essential physical properties of the large-scale motions in our atmosphere.
(See the maps of the 500 mb level on pages VII-5, VII-6,7)

To sum up: The main motion of the atmosphere is essentially from W to E (except in the lowest layers, say between sea level and 700 mb where easterly motions can exist). This means that the bulk of atmosphere moves bodily from West to East—carrying along the small disturbances (Lows, Highs). Called "STEERING" of low level systems by upper level flow (like eddies in stream carried along by current).

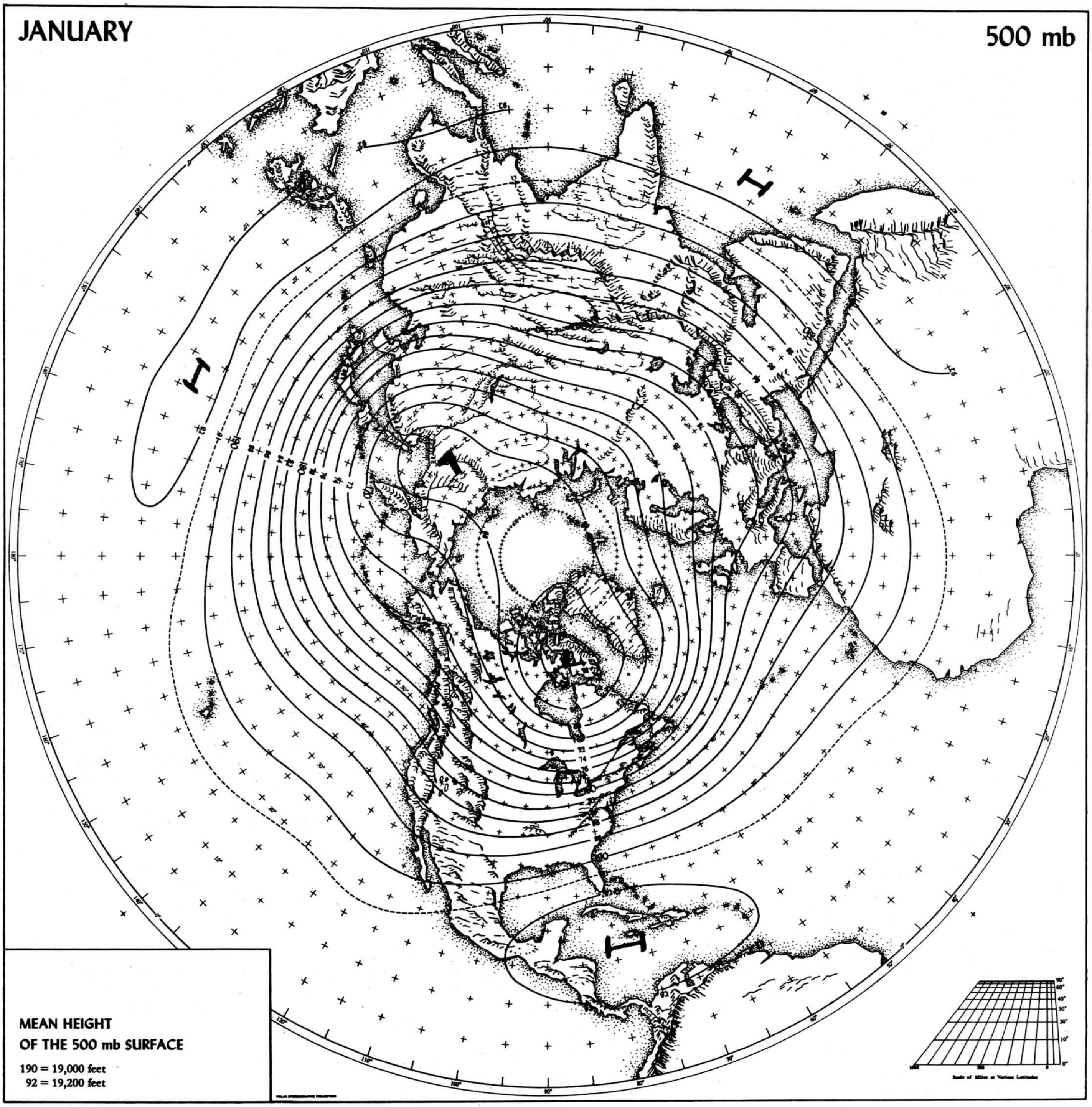
Westerlies—in temperate latitudes—are thus the most important wind system on earth.

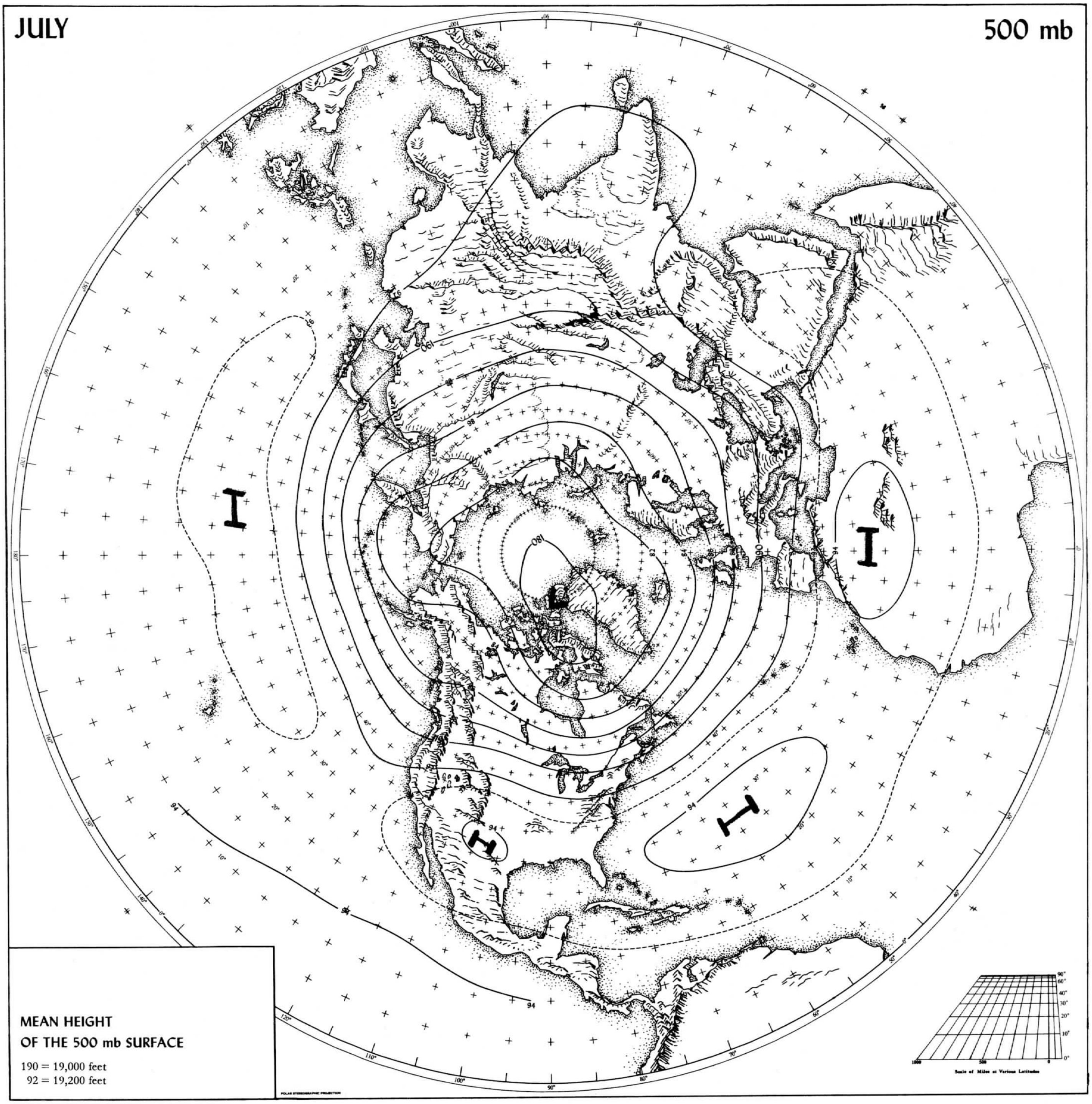
Strength of Westerlies varies with seasons—because temp. difference Eq.-Pole does [Eq = Tropics nearly no annual variations but Poles colder in Winter than in Summer]. Strongest Westerlies thus in Winter, least in Summer. (See 500 mb maps for pressure gradient differences between summer and winter.)

When comparing Northern and Southern Hemisphere, recall that N. Pole in Winter is quite a bit warmer than S. Pole in its winter (where sometimes, met. obs. have shown $T < -100°F$). Even in summer, N. Pole is somewhat warmer ("ocean") than ice-covered Antarctica ("continent"). Result is that Westerlies in Southern Hemisphere are somewhat stronger than those in Northern hemisphere at the corresponding season. In addition, near sea level, temperate latitudes in N. Hem. are to large extent over continental areas (increased sfc friction) while, in S. Hem., they are mostly over ocean. "Roaring Forties"—Gales in vicinity of Cape of Good Hope and Cape Horn. In N. Hem. 2 centers of sub-arctic low pressure (Iceland, Aleutians), in S. Hem. a near-continuous belt of low pressure at latitude 50°S.



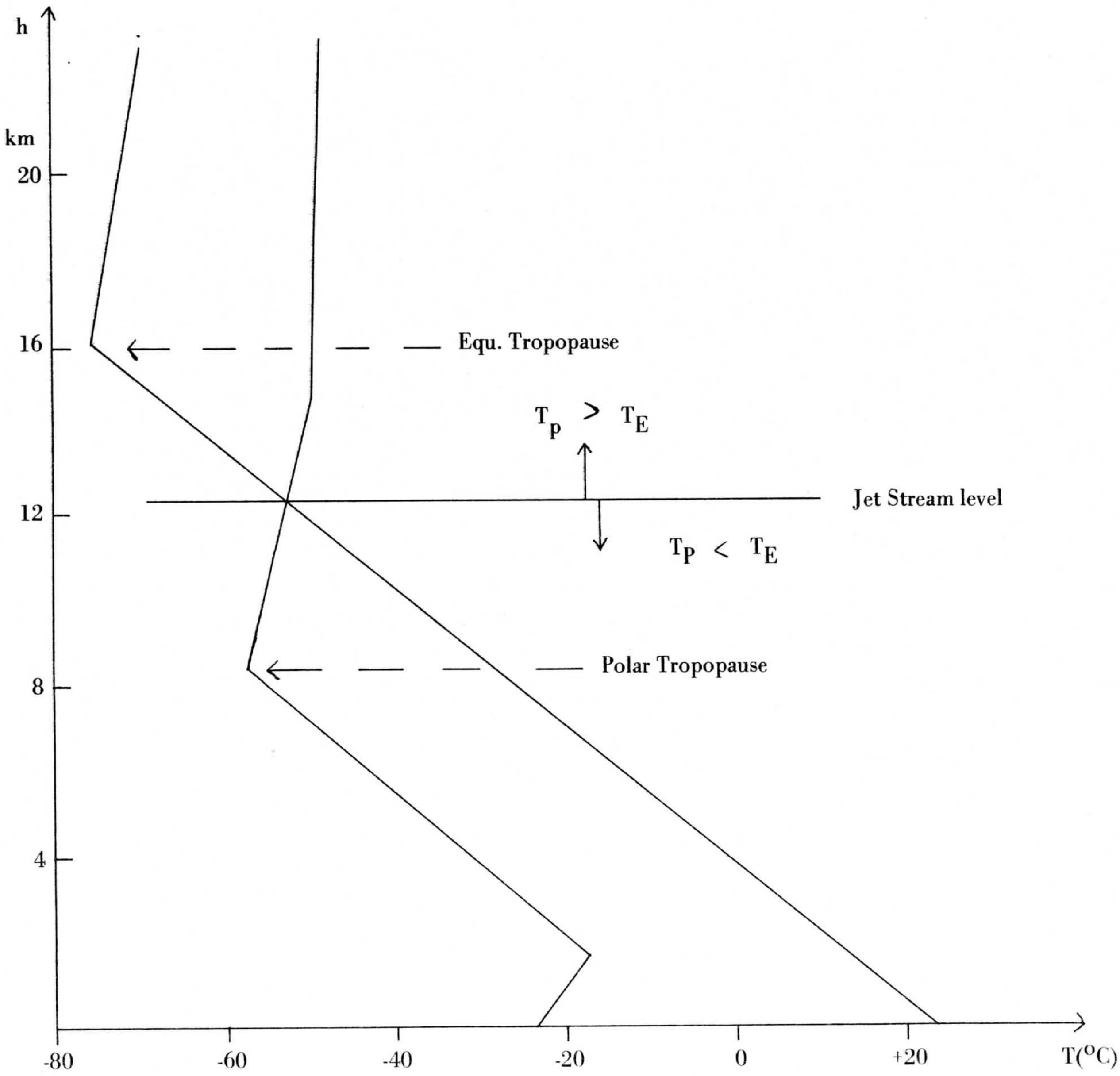
500 mb
1963 Jan. 16
00 Z


JANUARY
500 mb
MEAN HEIGHT
OF THE 500 mb SURFACE
190 = 19,000 feet
92 = 19,200 feet


JULY
500 mb
MEAN HEIGHT
OF THE 500 mb SURFACE
190 = 19,000 feet
92 = 19,200 feet
180
190
94
94
94
94
L
H
H
H
H
H

b. Extent of Westerlies in the Vertical.
The Jetstream

Recap: Westerly flow due to temperature difference Equator/Poles. (Thermal wind effect)



As long as T_P (near poles) is less than T_E (near Equator) windspeed with height will increase.

Thus Westerlies will intensify upward to level where temp. difference disappears (~ 12 km on average).

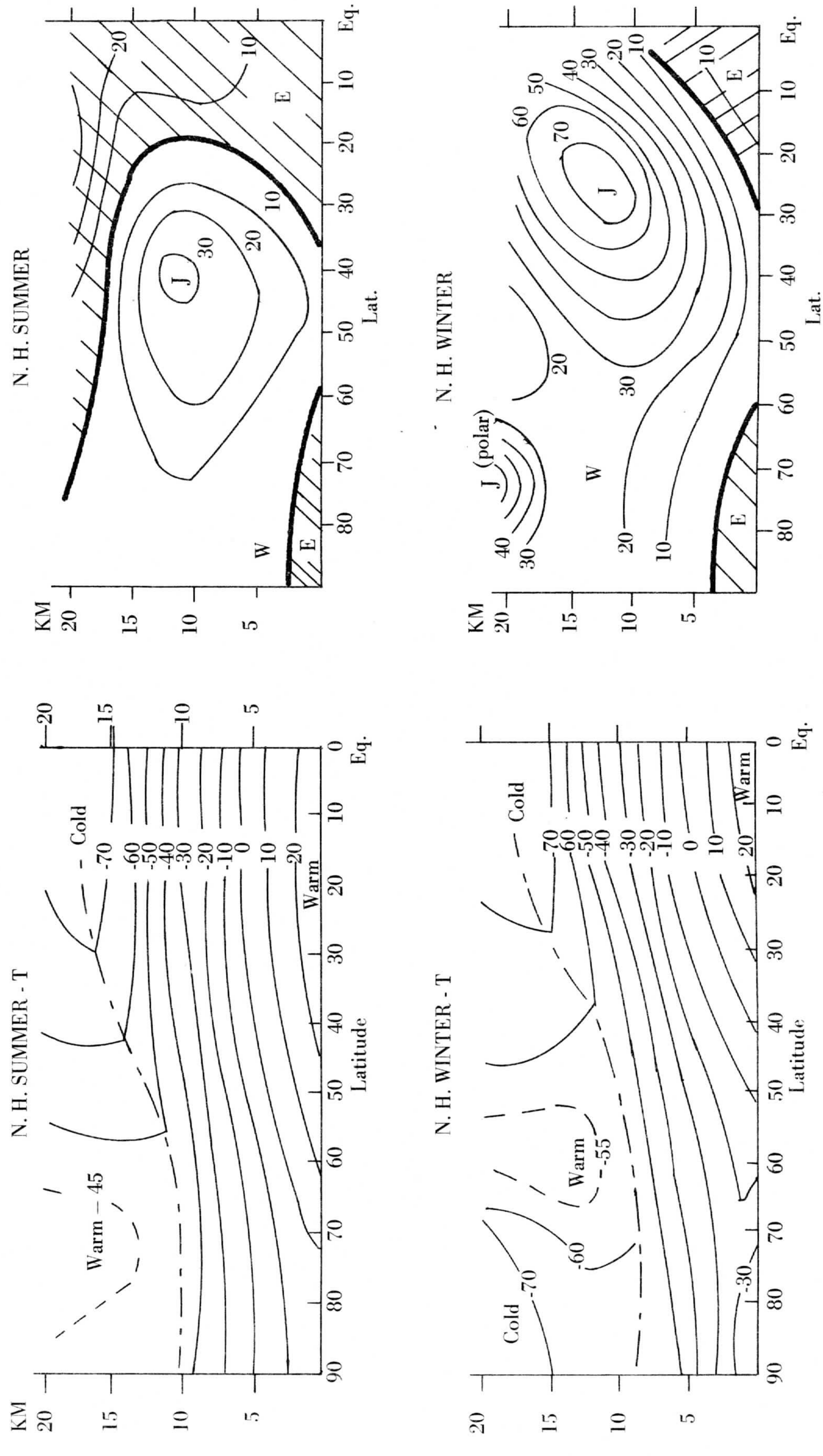
Above this, air over Eq. is colder than air over Poles; thus Westerlies will decrease in intensity.

Result: Windspeed maximum at ~ 12 km height
"Jetstream level."

If one goes even higher up in the atmosphere the westerlies will further decrease and, in fact, change to Easterlies. This happens at and above appr. 30 km. In these levels we find usually Westerlies in Winter, transitions (light winds) in Spring and Fall and Easterlies in Summer.

High altitude research balloons flying at heights between 30 and 40 km often are launched in Spring and Fall to facilitate tracking and recovery of payload. They may stay over launch area or nearby for several days. In 1975 one was seen near Madison for more than 2 days and payload landed on third day near Fond du Lac.

Diagrams on next page show Pole-Equator cross section for summer and winter conditions of temperature and wind speeds (in latter diagrams, hatched area on graphs indicate easterly winds).



N. H. SUMMER - T
N. H. WINTER - T
N. H. SUMMER
N. H. WINTER
KM
Latitude
Lat.
Eq.
Cold
Warm
J (polar)
W
E
J

Strongest contour (pressure) gradients in temperate latitudes, thus strongest W-winds in these regions. Position of Jet stream:

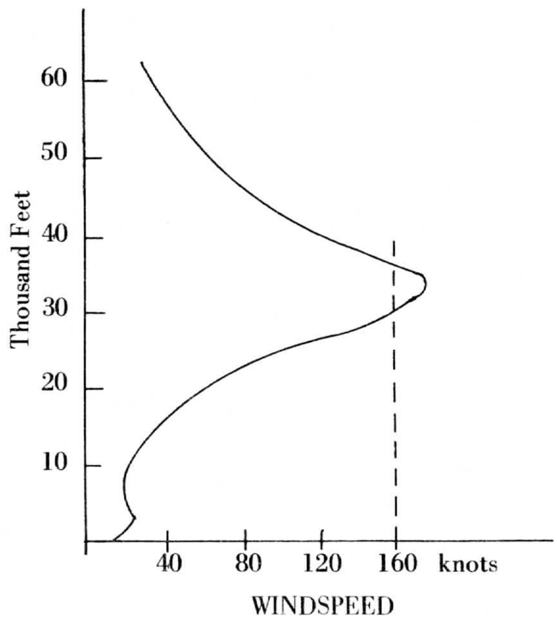
Winter 35°/40° N, Summer 50°/55° N
(Shifts along with sun)

Shows similar Rossby waves as all upper level flow—"meanders" all around hemisphere. Regions of strongest winds frequently over East coasts of large continents—NE—U.S. and Japan.

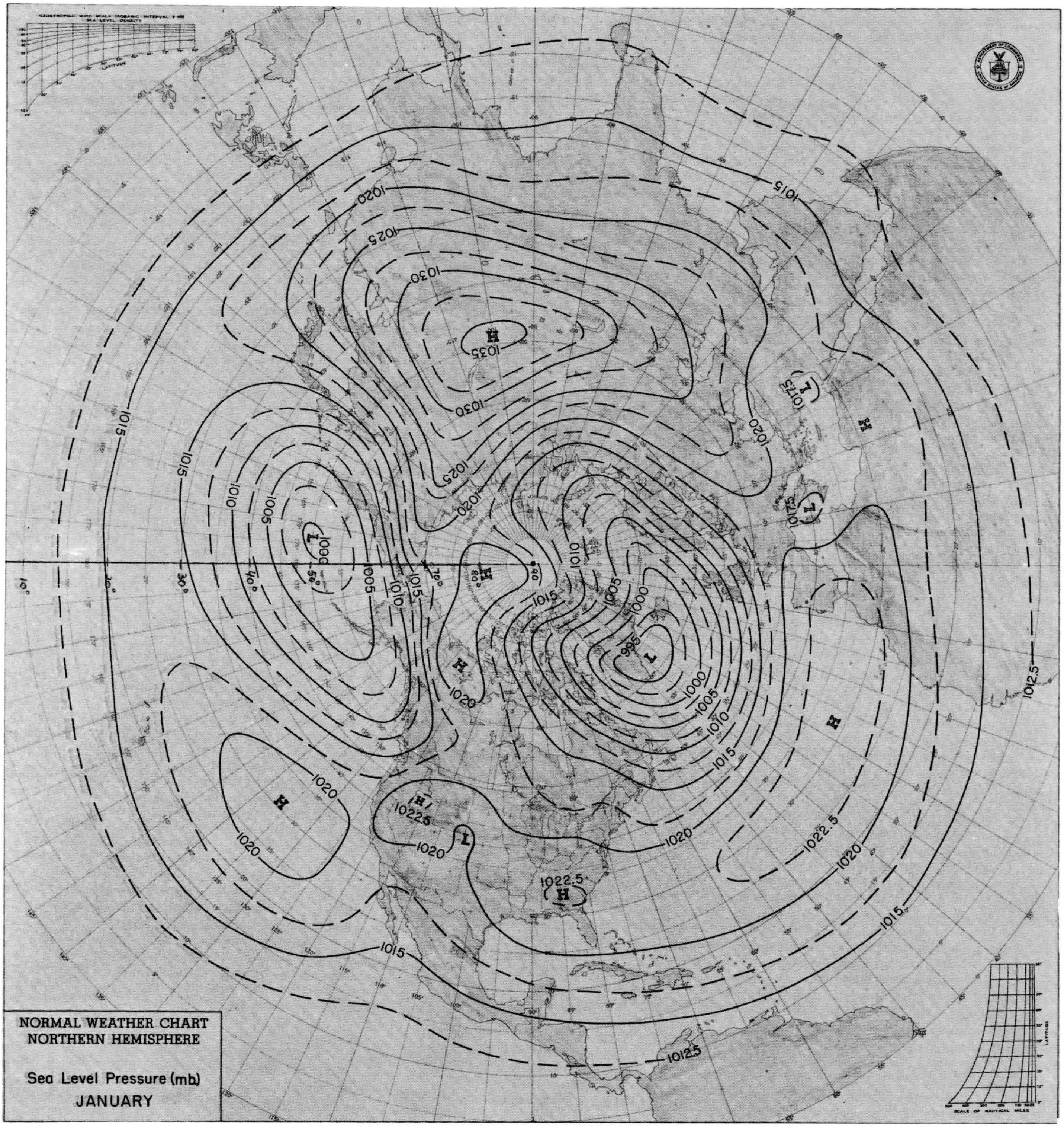
Windspeeds (average) Winter 75 mph
Summer 30 mph

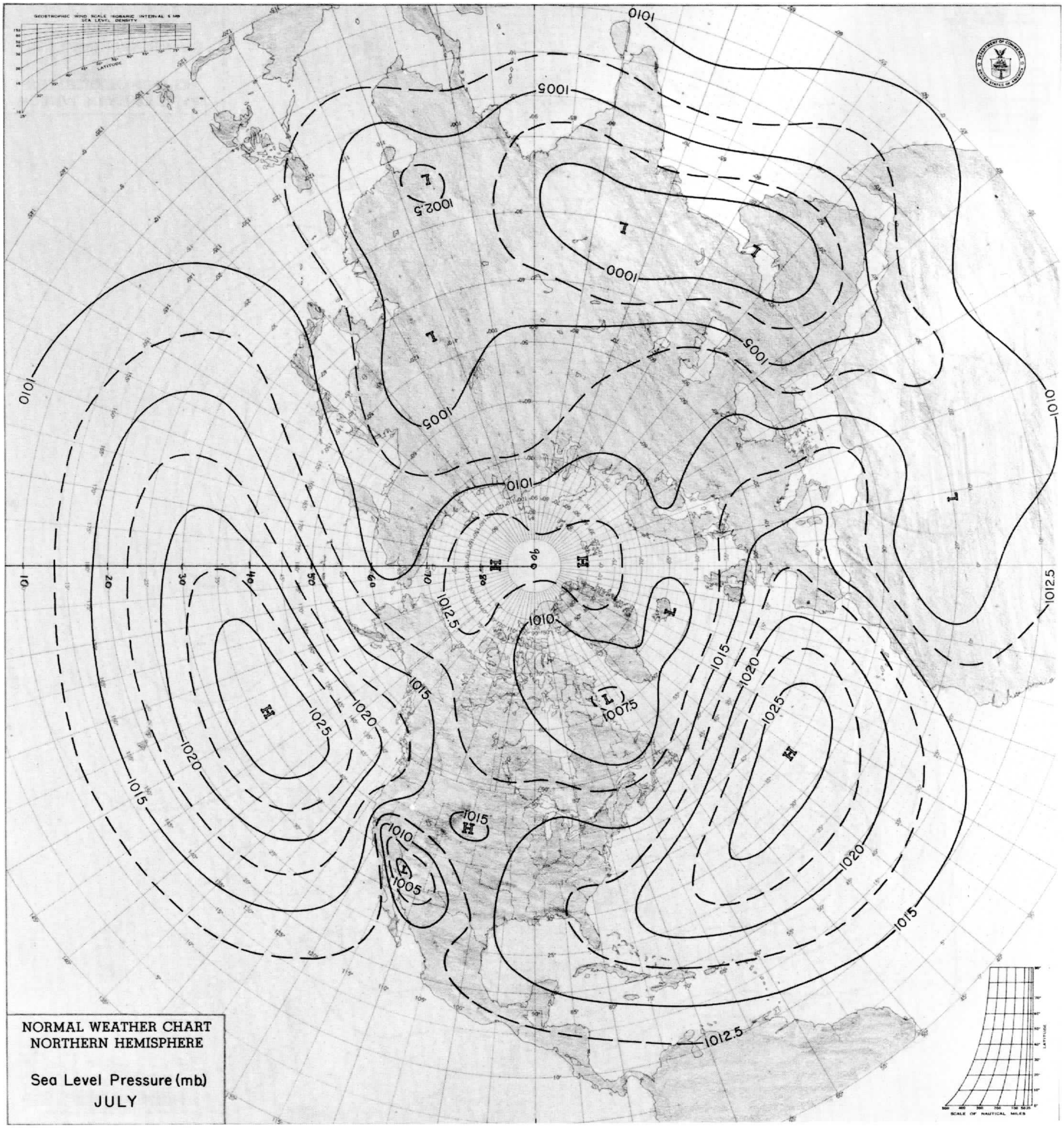
Frequently very high (above 100-150 mph)

Note effect on air travel, especially long-distance. U.S.-Europe tailwinds, Europe-U.S. headwinds.



Jetstream, observed at
Seattle – 1958 April 21


NORMAL WEATHER CHART
NORTHERN HEMISPHERE
Sea Level Pressure (mb)
JANUARY
1035
1030
1025
1020
1015
1012.5
1010
1005
1000
995
1017.5
1022.5
H
L


NORMAL WEATHER CHART
NORTHERN HEMISPHERE
Sea Level Pressure (mb)
JULY
1000
1002.5
1005
1007.5
1010
1012.5
1015
1020
1025
H
L
SCALE OF NAUTICAL MILES

c. Polar Easterlies.

From maps of sea level pressures, areas of lowest pressure not at pole but at ~ 60-65°N (Aleutian Low, Icelandic Low).
Over Arctic Ocean all year round cold air forms Cold Polar High.
Low level winds consequently Easterlies

But:

Cold high decreases in strength with height and, already at 600-700 mb level has changed into large low. Thus, polar Easterlies are only low level wind system; above them, in upper levels Westerlies reach northward into close vicinity of pole.

d. Subtropical Highs

A belt of generally high pressure is found at latitudes between 25-20°N (in Winter) and 35°-40°N (in Summer). Contain tropical (i.e. warm) air. Since warm highs intensify with height, these highs become more intense upward. Warmest air situated toward Equator—thus axis tilts toward this region.

Subtrop. highs—separate in the two hemispheres in low levels become one big high pressure system circling the earth over the Equator in upper levels. Called: semi-permanent—are noticeable in all seasons (permanent) but shift slightly with sun's position w.r.t. equator (semi . . .).

Subtropical (warm) highs are most pronounced over oceans (Atlantic, Pacific, Indian Ocean), somewhat less distinct over continents.

In Highs, usually divergence and thus downward motions (subsidence). Results in dissolving of clouds (dry-adiabatic descent of air), decrease of rainfall—climate becomes one of rain deficiency, i.e. steppe or desert. In fact, all big deserts are essentially due to influence of subtropical highs.

Note latitudes of great deserts in relation to subtropical highs:

N. Hem.: Sahara, Arabia, Iran & NW India, SW N. America
S. Hem.: Kalahari, Ctr. Australia, Pampas in S. America.

Over oceans, subsidenic forms distinct inversion which effectively puts lid on development of Cu (from moist warm sfc air)—TRADE INVERSION. Cu all seem to have about same level of tops—only occasionally one may break through.

e. Trades

Equatorwards from subtropical highs we must have Easterly flow (pressure in vicinity of Equator slightly lower). Air flowing out of S.T. Highs (on N. Hem.) from NE—and since highs are extremely stable, wind will be extremely steady. These very large-scale winds (chiefly over oceans) called TRADE WINDS. (Name from old sailing ship days, very faborable). The most steady winds on earth (Effect on trees, buildings, etc.).

In tropics therefore, weather moves from East to West—in contrast to our temperate latitudes where weather comes from the West. Chief difference in atmospheric motions in lower levels.

Another climatic difference: In temperate latitudes, the West coast usually are more moist (compare W Coast (Washington, Oregon) with New England) but: In tropics, East Coasts are moist/humid (Florida vs. S. Calif.; Caribbean Islands). Also, on mountaineous islands, East slopes are wet.

One can see already that General Circulation is a major factor in determining climate of a location.

f. Inter-Tropical Convergence Zone (ITC)

Trades approach Equator both from Northern and Southern Hemispheres —they converge.

Air thus has to rise near Equator to allow new air to move in along trades. A zone of irregular winds and large-scale upward motions results—characterized by heavy clouds and high rainfall amounts. Trade inversion disappears in this region and thus high vertical development (transport of water vapor and thus latent heat) is possible in this ITC. Its development not equally strong in all parts of equatorial regions—over continents usually less pronounced than over oceans. Problem is that much of tropical regions are over oceans where very few observations available. In fact, tropical meteorology currently undergoes violent re-appraisal—due largely to new tool: Met Satellite. Latest in series of TIROS, NIMBUS, TOS is ATS (Advanced Technology Satellite).

A "synchronous satellite"—remains over fixed point of Equator (first about 151°W—over Pacific) and can photograph all of Pacific (from US West Coast to Japan–New Zealand) once every 20 min. Launched Dec, 1966, on station Jan. 67. (U. Wis.—Suomi)

The most recent meteorological satellites launched are: ATS-6 (May 74), SMS-1 (May 74) and SMS-2 (Feb. 75) [Synchronous Met. Sat.] and GOES-1 (Oct. 75) (all synchronous) and NOAA-4 (Nov. 74) and LANDSAT-2 (Jan. 75).

Area of trades and ITC a special area of interest because, in these regions, much of solar energy is introduced into atmosphere and thus has to be transported from tropics to poles.

Atmosphere is essentially a heat engine:

Tropics = boiler,
Poles = condensors.

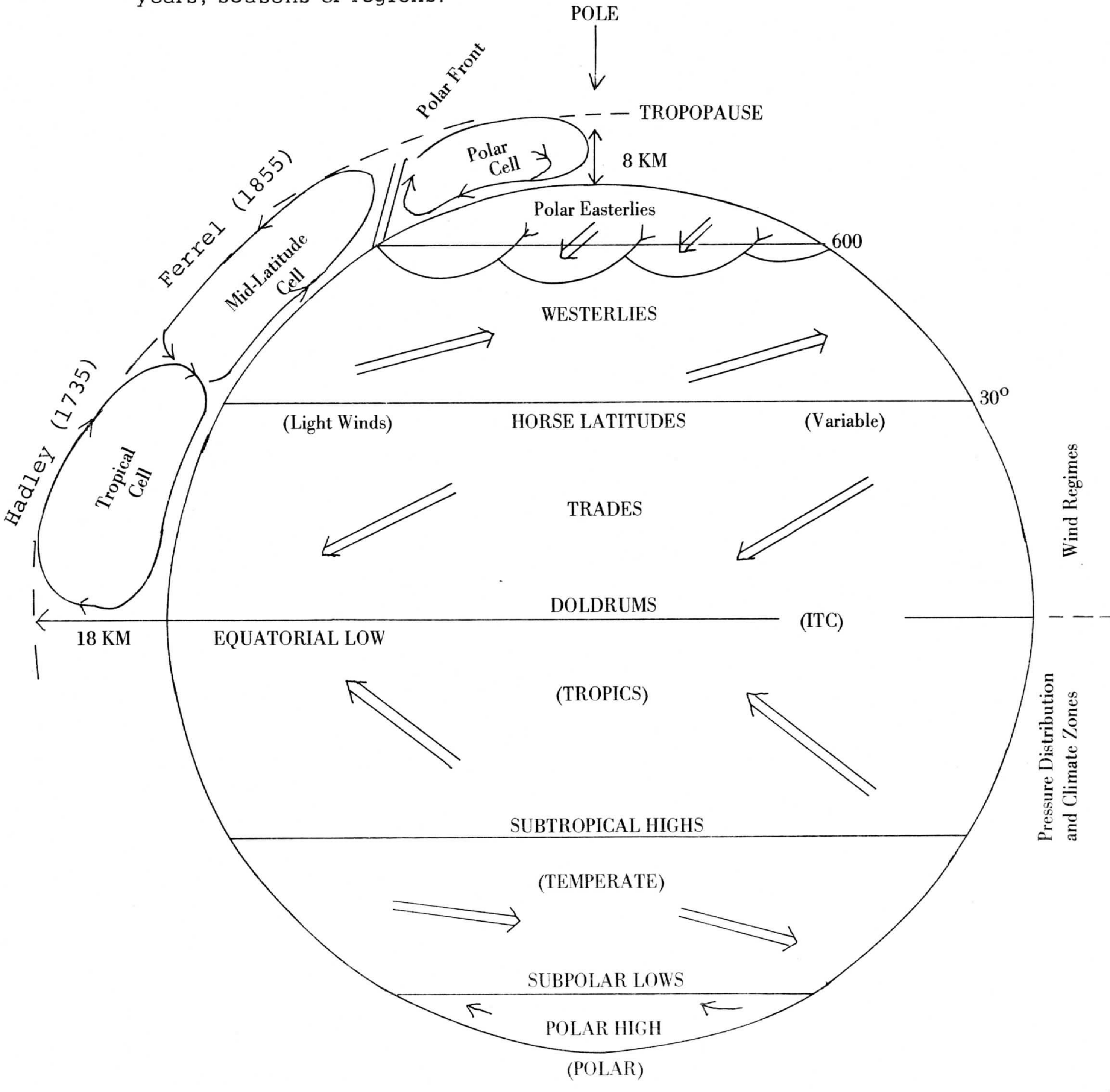
Transport <u>from</u> <u>sfc</u> <u>to</u> <u>atmosphere</u> (vertical-up) by

a) latent heat (evaporation, condensation) ~ 80%
b) sensible heat (conduction, convection) ~ 20%

Transport from Equator to Poles (horizontal-N/S) by

a) sensible heat ~ 50%
b) latent heat ~25 - 30%
c) ocean currents ~20 - 25%

These figures only averages—can vary considerably in different years, seasons or regions.

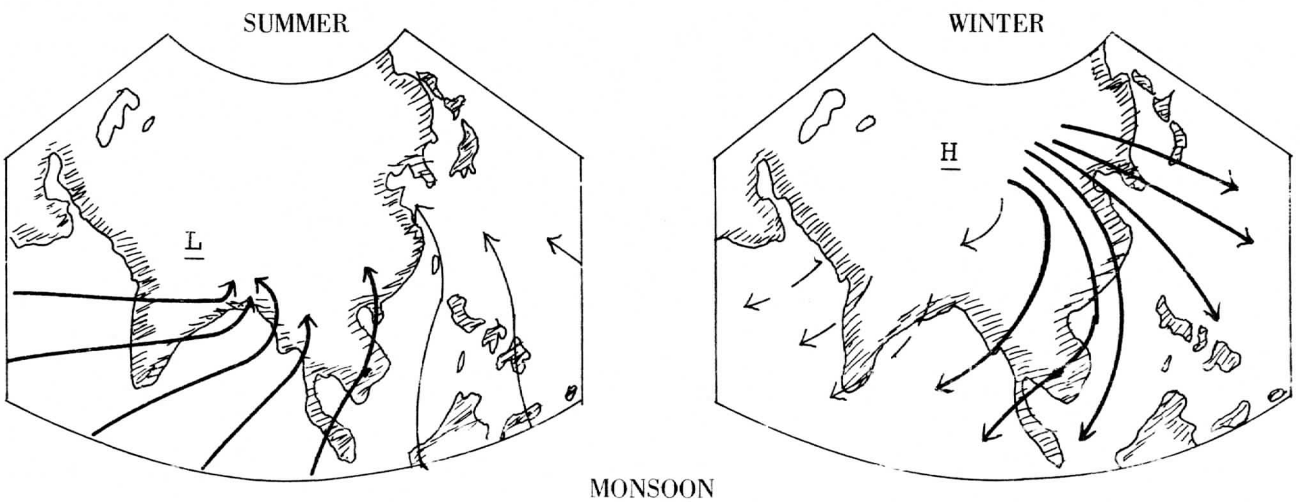


3. The Monsoon

Above parts of General Circulation are world-wide (global) and essentially due to geometric effects of earth's shape upon solar radiation received in various latitudes, coupled with secondary variations through year due to inclination of earth axis (seasons). Another, at least "near-global" physical process was not yet taken into account, namely distribution of oceans and continents on earth surface. In heat budget, difference chiefly is in different heat capacity and albedo of two kinds of surfaces, to small extent also difference in elevation.

In summer (high solar input) continent will heat up more and faster than ocean where energy goes into water and, in fact, is even distributed further downward by convection. Result is decrease in density of air over continents relative to oceans, i.e. a semi-permanent (seasonal) summer Low.

In winter (low solar input) continent will cool more, especially when snow covered (IR albedo of snow is very high, thus large radiation losses). Result is strong increase in air density—seasonal Winter High. Chief development of this kind of seasonal systems occurs over biggest continent—Asia.



MONSOON

Strength of monsoon given by thickness of arrows.

Over N. America, such system can only be seen in rudimentary form on mean maps, but no large-scale consistent wind systems develop. Over Asia, however, we find a consistent system of seasonally changing winds—MONSOONS.

Summer monsoon: Low over northwestern India. Influx of air, especially from Indian Ocean. Warm, moist air—when flowing north will be forced to rise on southern slopes. Soon LCL reached, clouds & precipitation—monsoon rains. Since tropical air contains large amounts of moisture, very heavy summer rains. Reach all the way to Himalayas where (~ Early June) heavy snow falls. After crossing these mountains, air has lost most of its moisture—Tibet is desert country.

Winter monsoon: High (cold air) over Siberia. Outflow toward oceans. But cold air near ground, cannot cross Himalayas (dammed up)—will thus flow out mainly over Manchuria, China into Pacific. Dry cold winds, carrying frequently large amounts of dust from deserts and steppes of Mongolia far out into Pacific Ocean.

Monsoons are not really "global" circulation systems but of such dimensions (affect nearly 1/2 of N. Hem.) that customarily included in General Circulation discussion.

Somewhat similar monsoon-type circulations also occur in other regions, e.g. Africa, South of Sahara and even U.S. Southwest ("rainy" season in W. Texas in Mid-Summer).

4. Causes of General Circulation

(1) Geometry of earth (a sphere), modifying solar input
(2) Tilt of earth axis (seasons)
(3) Rotation of earth (Coriolis force)
(4) Heat capacity differences between continents and oceans and albedo changes at sfc.

Theory for explaining details not yet completely perfected—largely due to lack of sufficient data (2/3 of earth sfc is ocean where not many data are available). We can explain by theory the gross features but still lack understanding of many details. Applying theory recently attempts to also predict general circulation on other planets (especially Venus and Mars). Such attempts help to understand effects of various assumptions (e.g. composition of atmosphere affects radiational heat balance, etc.) and thus allows to improve theory.

Concept of various (vertical & horizontal) "cells" in circulation systems. Hadley cell regime vs. Rossby wave regime.

Attempts to model circulation by computers and by analog experiments (dishpan experiments). New breakthrough now possible with tremendous increase of data collected by satellites and steadily increasing power of scientific and operational computing systems.

EXERCISES—Chapter VII

1. List, for each of the 4 orders of circulation, at least three examples from your own knowledge of weather phenomena.
2. Verify that effect of temperature gradient and Coriolis force in the two hemispheres results in both cases in Westerlies.
3. Assume an upper level high and superimpose its circulation upon a general westerly flow. What kind of upper level feature will result?
4. Calculate the (average) wave length of Rossby waves in the case of 3 and 5 waves around 45° N, both in degrees longitude and in km.
5. Describe the most important seasonal differences occurring at the 500 mb level and at sea level, on basis of maps given in Chapter VII.
6. Discuss briefly the most obvious changes in the General Circulation if we would assume that the Earth axis were exactly perpendicular to the plane of the Earth orbit.
   What would happen to the weather at Madison?

CHAPTER VIII. SECOND-ORDER CIRCULATIONS—WEATHER SYSTEMS

(Reading: pp. 42-58, 103-107, 116-128)

A. Cyclones and Anticyclones in temperate latitudes

1. Basic pattern

From previous chapter obvious that weather in tropics is distinctly different from that of extra-tropical areas. (Extra-tropical: between subtrop. H and poles.)

Main difference:

Within tropics, horizontal temperature differences very small, thus little horizontal but large vertical transport of energy, etc.
Outside tropics, strong temperature gradients and thus large horizontal transports (largest in Westerlies ~ 40° N/S!)

Also: more important for us—circulation within the belt of Westerlies. Thus begin with these—will mention tropical weather and circulation later in chapter.

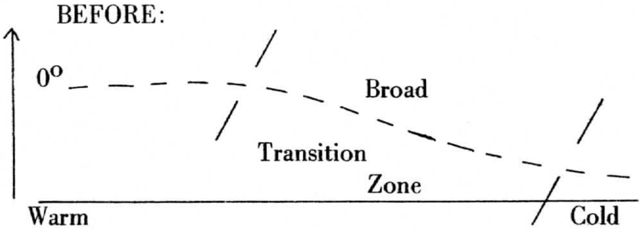
Within Westerlies (part of General Circulation) we have imbedded second-order systems—the familiar LOWS and HIGHS of temperate latitudes.
LOW also called Cyclone (with cyclonic flow)
HIGH also called Anticyclone (with anticyclonic flow)
Size of these systems between 100 and 1,000 km. They are most pronounced in our daily weather maps and thus are also called
SYNOPTIC SCALE FEATURES.

Recall Rossby waves in upper levels of troposphere [Any one day, one may find between 3 and 6-7 such waves circling the globe—with wave length thus of 50-100° longitude]. In lower layers (say near sea level), we find then closed lows or highs—related to upper flow through the thermal wind, caused by motions of cold and warm air. (Tilt in Low or High axis with height.)

2. Frontal zone

Without Rossby waves (i.e. straight W-E flow), no transport of air in north/south direction. But with them, cold air (in back of trough) moves toward Equator, warm air (in front of trough) toward poles. This increases strength of horizontal temperature gradient

BEFORE:



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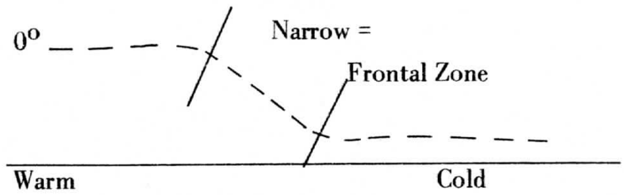


First proposed as major mechanism of forming Lows and Highs, by J. Bjerknes (1919)—(Bergen school).

<u>POLAR FRONT THEORY</u>

Since air in tropics/subtropics is warm and air in polar regions is cold, the (often strong) transition zone between subtropical and polar air was called the polar front.

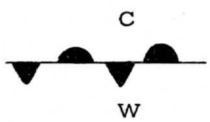
Definition of "a front":

A zone of pronounced horizontal temperature gradient separating two masses of air with different T and thus density. Frequently, other met. elements also change considerably across this zone.

Drawn on Wx-maps as Cold—or Warm front (depending on which air mass is actively moving with respect to the other.)

<u>Symbols</u>:  cold front

 warm front

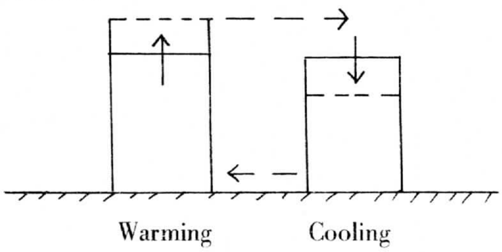
If front is <u>stationary</u>, 

If front <u>occluded</u> (explained later!)  [combination of CF and WF Symbols]

3. Cyclone development

Once front has reached a certain sharpness, CYCLOGENESIS (i.e. forming of low) can occur.

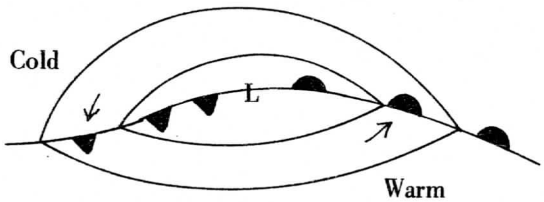
Very schematic:



Vertical cell forms; twisted by Coriolis force into horizontal—producing cyclonic motion around vertical axis.

(Similar thing can occur by disturbing flow near sfc by friction or local irregularities—see "Storm," pp. 7-8!)

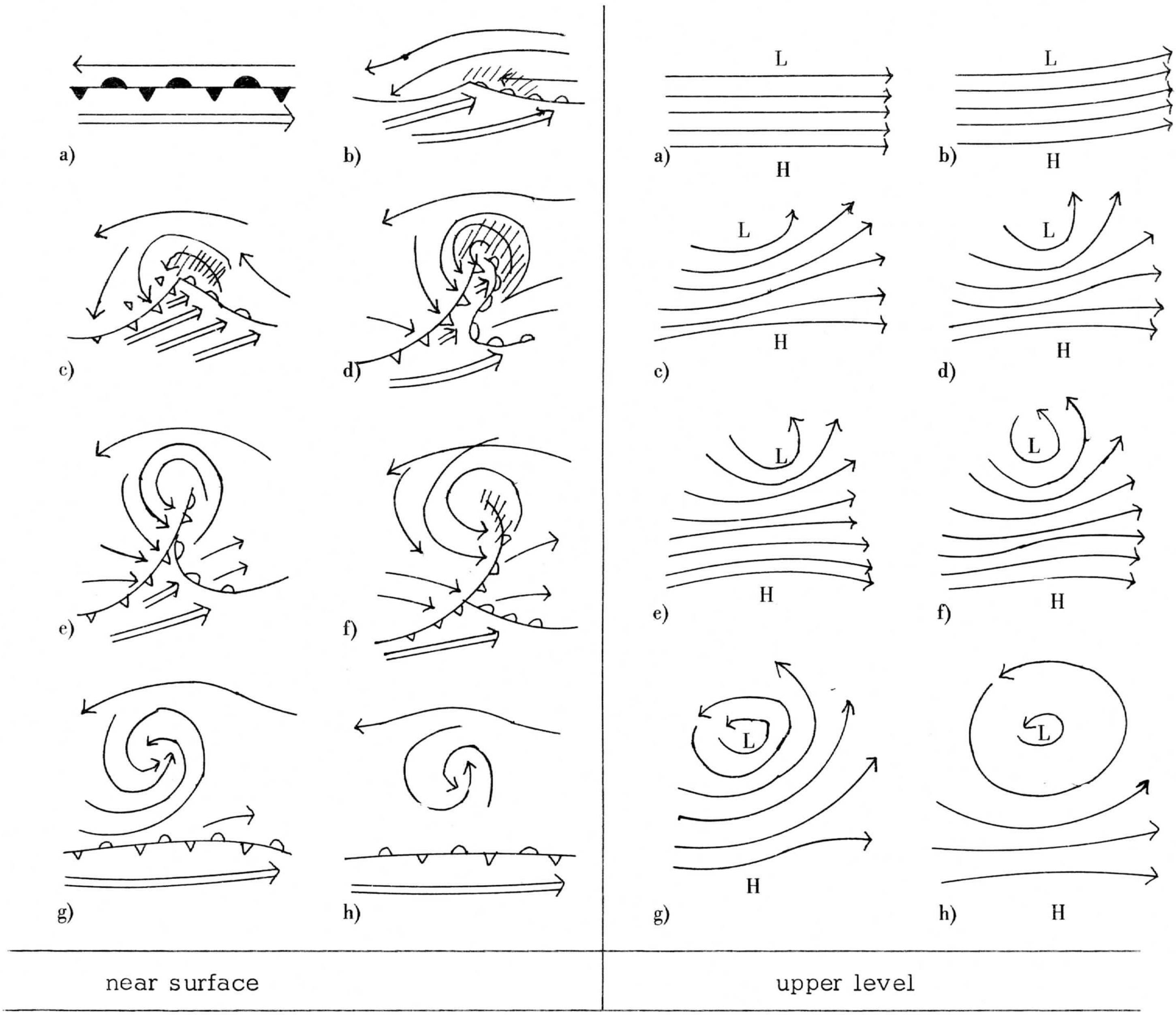
This initial cyclonic flow will, on Wx-maps, appear as a minor "wave" on polar front (cool air starts moving south (N. Hem.) behind cyclonic eddy (west), warm air moves north in front of eddy (east).



Energy of cyclone development comes from sinking of cold (heavier) air and rising of warm (lighter) air: Conversion of potential to kinetic energy.

(Example in mechanics: Weights on old-fashioned clock.)

Further "life cycle" in development:



LIFE HISTORY OF AN IDEALIZED CYCLONE

a) original polar (near-stationary) front
b) beginning cyclogenesis (wave cyclone)
c) cyclone strengthens (well-developed wave)
d) warm sector narrows (cyclone matures)
e) occlusion near center (see below)
f) occlusion nearly complete (cold center, dying cyclone)
g) Polar front reforms south (cyclone nearly dead!)
h) final stage, cyclone disappears.

Motion of such cyclone "steered" by upper level (westerly flow). One has to remember, however, that cyclone near sfc also will modify to some extent the upper flow (trough/ridge pattern).

4. Anticyclones

Previously (in Chapter VII) already discussed 2 possible types of anticyclones (HIGHS) in temperate latitudes:

Type a) cold core high;
b) warm core high.

Type a) is the usual high pressure area resulting from advance of cold air behind a low southward (in N. Hem.). This high has characteristic flow pattern of all highs, i.e. divergence near sfc and subsidence, often with formation of inversion aloft. Result is frequently dissolution of clouds, thus increase in solar radiation penetrating to sfc and subsequent convective activity during day time. Often Cu are restricted to lower levels by upper inversion. Sometimes also, inversion may lead to stratiform cloud shields. These highs are directly part of the cyclone/anticyclone system; they move along in Westerlies behind lows. Since this high is containing cold air, its intensity decreases with height. In upper layers it is usually recognizable as the ridge in the westerly wave pattern.

Type b) is occurring less frequently. It contains warm air, thus intensifies with height and often is then the reason for a disruption of the usual westerlies (because it forms a strong upper level high with closed contour lines)—it "blocks" the normal flow—therefore called "blocking high."

Its formation not too well understood in detail. Usually develops in one of two ways:

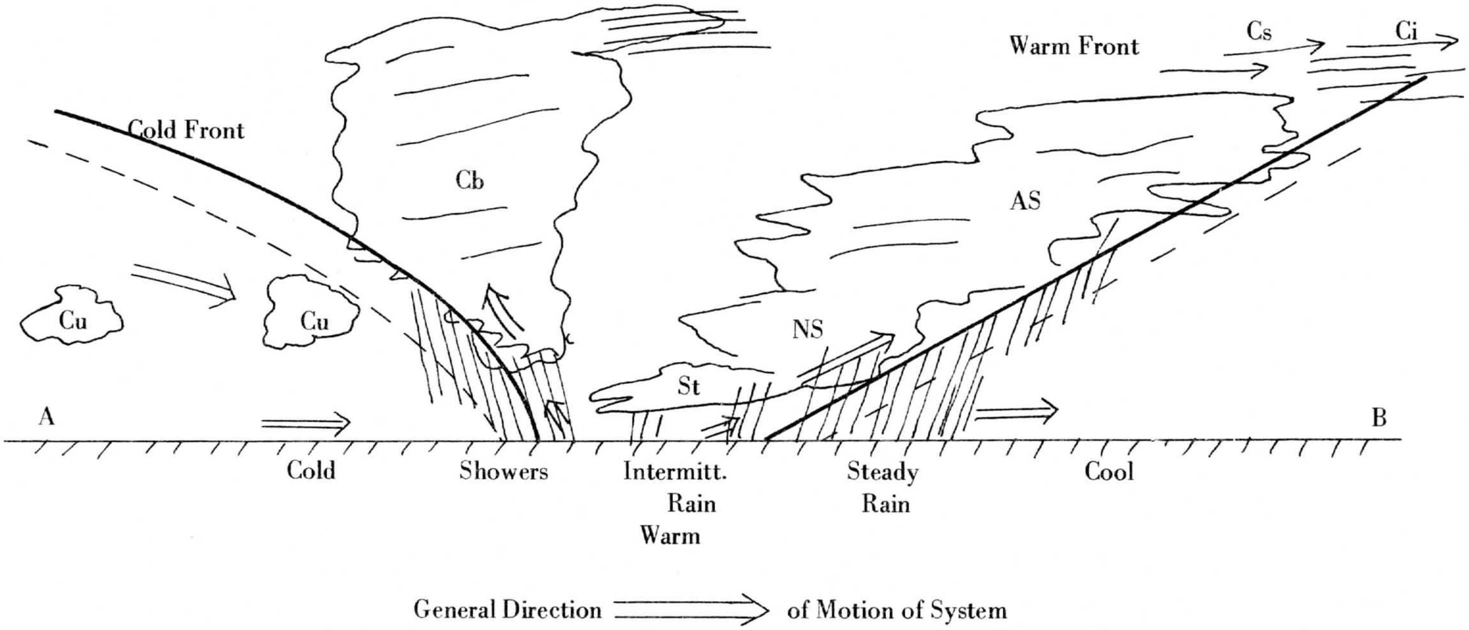
1) Subtropical (semi-permanent) high forms an extension northward, pushes polar front far north, then establishes this extension as a separate warm high;
2) Original cold high (large outbreak of polar air) becomes stationary, warms up due to strong insolation and mixing in of warm air aloft—slowly converts into (relatively) warm high which may develop further into true blocking high.

Obvious that, due to divergence in highs, no fronts can develop (frontogenesis requires convergence, i.e. moving warm and cold air together). Anticyclone thus a more homogeneous system, usually containing only one kind of air (either cold or warm). Cyclones are the more important, active weathermakers, while anticyclones tend to preserve weather as it was.

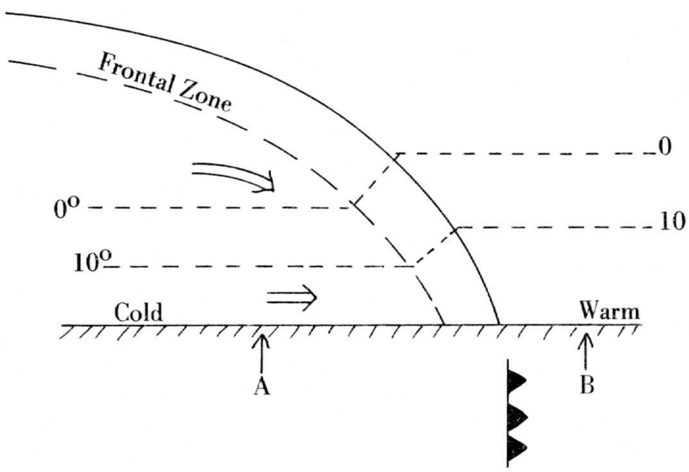
Concept of persistence in Wx forecasting—most valuable when one is located in anticyclone.

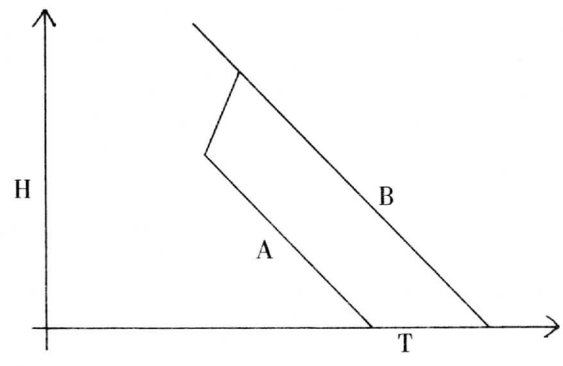
B. Fronts

Schematic description of cyclone at height of development (stage c or d) by means of cross section A-B (south of Low Center thru warm sector):



1. Cold front



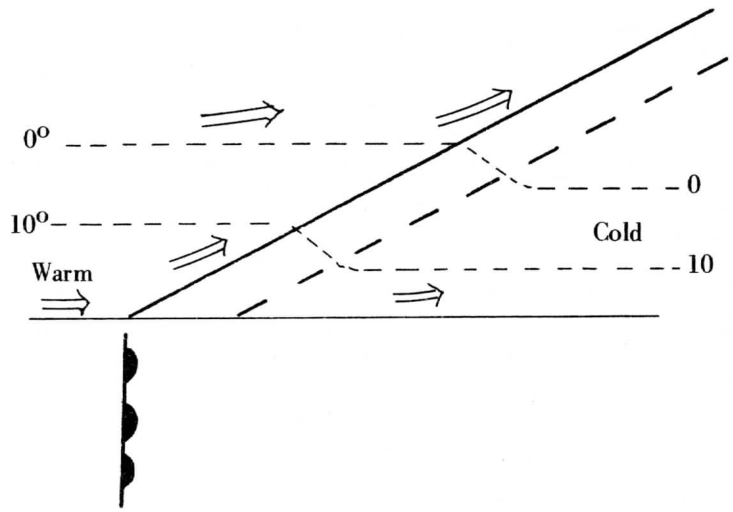


Cold air advances faster than warm air retreats. At front, thus warm air forced upward. Behind CF, cold air will show sinking motions. If warm air is moist, forced upward motion will result in violent vertical motions and buildup of Cb (showers) Slope: 1:30 to 1:200. Steep near sfc due to friction.

Wx phenomena with passage of CF at a station:

(1) Wind shift clockwise (i.e. SW → NW); called "veers" in contrast to "backs." [see Wx-map.]
(2) pressure rises strongly (cold air heavier)
(3) temperature falls after frontal passage
(4) showers may occur, or even thunderstorms
(5) noticeable change in air mass (e.g. warm to cold, humid to dry)

2. Warm front



Warm air moves faster than cold air in front of it. Warm air lighter, will ride up over cold air (large-scale slow upward motion, formation of layer clouds—beginning from top/right side: Ci + Cs, As + Ac, Ns - St, with steady rain fall, increasing toward sfc front.

Slope: 1:100 to 1:300 (gentler than CF)

Wx phenomena with advance of WF (and passage) at a station:

(1) Wind shift again clockwise (e.g. SE → SW),
(2) pressure falls with approaching front—after passage, falls less or remains same with irregular fluctuations,
(3) temperature rises when front close to station, after passage usually stays same,
(4) layer clouds thicken with approach of front and often steady precipitation falls. After passage, rain usually becomes intermittent or may stop.
(5) noticeable change in air mass (e.g. cold to warm).

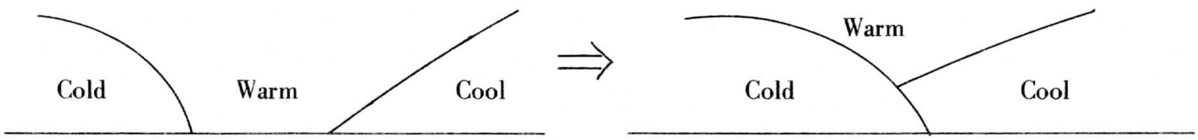
3. Warm sector

Area between CF and WF in (open) cyclone. Usually the moist area between two drier masses. Contains the warmest (subtropical) air of system. Often low clouds linger on after passage of WF. Rain becomes intermittent or may stop altogether. In moist air often very low clouds (Fs) may occur. Temperature high and rather stable. This warm air is very necessary ingredient in energy driving cyclone. When it gets cut off (i.e. CF starts to overtake WF), beginning of decay of cyclone.

4. Occlusions

Develop due to different speed of air in front and back of cyclone.
Two types: a) Cold front type b) Warm front type.

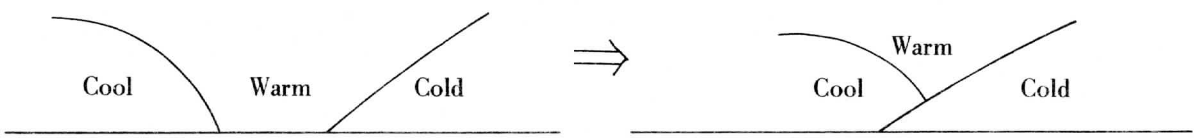
Type a) T of air behind CF cold, lower than of air ahead WF (Cool)



a) COLDFRONT TYPE

Wx behavior during passage similar to CF passage.

Type b) T behind CF (cool) higher than ahead of WF (cold)



b) WARMFRONT TYPE

Wx behavior during passage similar to WF passage.

The older cyclone, the more occluded. Potential energy is used up more and more, kinetic energy generation slows down, cyclone decays and finally disappears when cold air has filled up the whole system. Thus original CYCLOGENESIS changes to CYCLOLYSIS .

5. Areas of cyclogenesis

In general along polar front in areas where strong horizontal T-gradients occur or are generated.

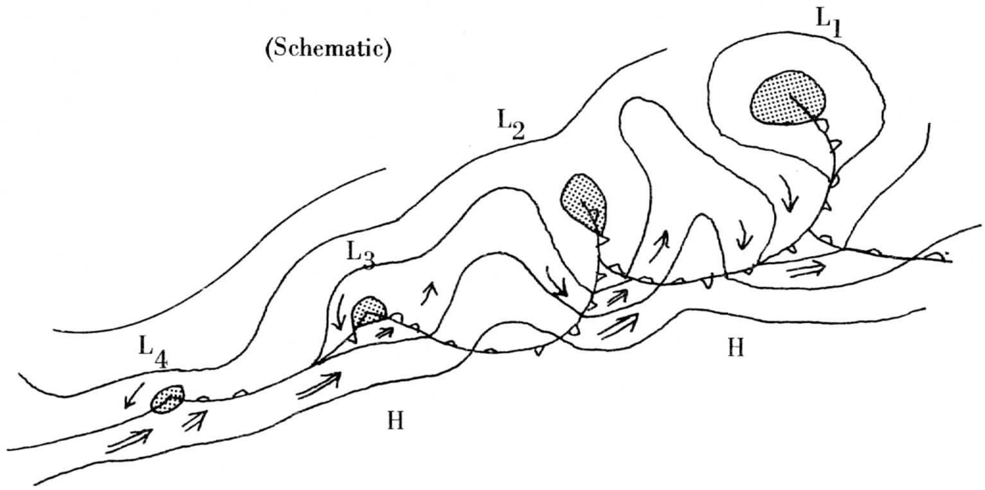
One region specifically favorable: where westerlies carry cold dry continental air out over warmer moist oceans, i.e.

East Coast of continents (Asia-near Japan)
(US East Coast-Hatteras to New England)

Another area where, on Eastside of meridional mountain chains, cold air from North funneled south to meet subtropical air, e.g. East of Rocky Mountains (Colorado, Alberta), East of Alps, etc.

All such areas are usually active year round but intensity of cyclogenesis much more pronounced in cold seasons. Then all circulation speeds up, in response to requirement of much stronger energy transport from Eq. to Poles. Cyclogenesis related to upper levels—Jet stream maxima occur also over East coasts. LOWS and HIGHS are the essential mechanism to transport this energy, together with upper level Rossby waves.

6. Cyclone Families



After one cyclone has formed, cold front behind it is essentially another very active (concentrated) part of original polar front. Thus it is possible that, on this front, another wave cyclone may form. Process can go on for some time; we may find a sort of chain of lows strung out—No. 1 possibly in last decay stage (heavily occluded), No. 2 beginning to occlude (mature), No. 3 a vigorous wave cyclone and No. 4 just getting started as initial wave. Whole "family" usually moves along in similar path, the later ones often a bit more southerly. Result in weather at a given station is then a succession of WF/CF passages with rain—warm sector—cool backside high—next WF rain, etc.—sequence of rainy and fair days.

This system normally ends with one of the "cold air outbreaks" behind CF establishing a cold large high which cuts off flow of subtropical air to the North and thus establishes a broad frontal zone (inactive due to relatively wide separation of true cold and warm air).

C. Air Masses

Already used concept (subtropical vs. polar air) without really defining it. Various ways of defining possible. Here: Airmass—the direct opposite of Front. If you have a front ("fast transition zone"), you have to know what fronts separate!

Definition: AIR MASS = a large mass of air which is relatively homogeneous throughout its horizontal extent. At a given level (height or pressure), temperature, water vapor content, lapse rate are not varying much through its region of existence.
Then front is the (narrow) zone of transition between two such airmasses—usually differing (due to ΔT) also in density.

Airmass characteristics are very much determined by the surface over which they are existing—because most of energy received from sun first goes to sfc and then is transported up into atmosphere.

Review Radiation budget (p. III-9)

$$R_{net} = C + E + H$$

To allow a large volume of air to acquire homogeneous character, the above 3 quantities should be reasonably constant in area; then this air can become a "meteorological air mass." This requires:

(a) Homogeneous surface (then all three are constant!)
e.g. ocean area, snow-covered continent, desert or steppe
(b) Time—air must remain over area for sufficient time (i.e. several days or weeks) to achieve equilibrium between R_{net} and air characteristics (lapse rate, etc.). This, in turn, means low windspeeds (to keep air in area) i.e. either semi-permanent highs or lows. (warm: subtropical highs, cold: polar or winter Siberian highs, Aleutian or Icelandic lows)

We usually define 4 major air masses—based on warm vs cold and moist vs dry:

	Continental	Maritime
Polar	cP	mP
Tropical	cT	mT

cP (continental polar) and mT (maritime tropical) are considered primary airmasses—originating in continental polar highs [Arctic ocean ice-covered, acts just like solid continent] or in subtropical (permanent) oceanic highs.

mP and cT usually are secondary airmasses—originally may have come from cP or mT but have been modified by travelling for long time over (near) homogeneous ocean areas (mP) or continental desert areas (cT).

1. cP: Source regions (N. Hem.) Siberia, Alaska + NW Canada, NE Canada/Labrador—in Winter also Arctic Ocean; (S. Hem.) Antarctic Continent. Over U.S. we sometimes use a sub-airmass class cA (continental Arctic) which is very cold cP. In summer, cP is rare on N. Hem. (even shores of Arctic Ocean are rather warm, Arctic Ocean has many open areas). True cP in summer actually only found over Antarctica, due to ice cap. The only icecap on N. Hem. (Greenland) is too small—also does not have a Polar High associated with it. When cP still at source, air mass forms due to long nights, small (or no) solar radiation, no heat supplied from ground below—snow/ice is good insulator, (Eskimo igloo!). Formation of strong sfc inversion; this increases vertical stability. No moisture source, turns very dry ($w < 2$ g/kg). Tropopause level usually very low (~ 8 km).

2. mT: Source regions (both hemispheres) subtropical high cells over oceans, throughout year available. At source, air mass develops over warm water sfc (T ~ 70-90° F); consequently large evaporation and very high humidity (w often 20 to 40 g/kg). Lapse rate usually conditionally stable/near moist adiabatic. Often contains haze and/or mist (fog droplets, salt particles) "turbid" air. Tropopause over it usually very high (~ 16-18 km).

3. mP: Source regions—high latitude (polar + subpolar) oceans. Usually develops from cP (or cA) which has been moved out over these oceans into semi-permanent lows (Aleutian, Icelandic lows). Since oceans much warmer than continents, air near sfc becomes rather unstable—humidity goes up due to evaporation. This speeds up modification from cP to mP by vertical mixing-overturning. Actually becomes airmass only because area over which it moves is very large and very uniform. Near (secondary) source (i.e. oceans): Moderately cool-subarctic water temp. 35-45° F winter, 45-55° F summer. Condit. stable—near moist adiabatic. Much clouds, showers. Total moisture content less than mT but more than cP.

4. cT: Source regions—large deserts and steppes in subtropics (Sahara, Arabia-Iran, S. China, Australia, Central S. Africa). In source regions, cT has very high temperatures—heated from hot surface

(large insolation, no clouds, no water vapor). Very low relative humidity ("desert air") but not necessarily low mixing ratios (w can be as high as 10-20 g/kg). Lapse rate highly unstable (near dry-adiabatic), in lowest layers, due to intense heating near ground often even absolutely unstable. This leads to anomalous refraction of light rays—Fata morgaria, double images, etc. Due to large lapse rate, often very strong turbulence, but not much cloudiness (CCL very high, due to low relative humidity). cT, when moving out of desert usually is modified very quickly due to large turbulence and thus fast efficient mixing of lower and upper air.

Modification of airmasses—occurs when airmass leaves source region and thus moves over sfc with other chacteristics.

Example: Fresh cP moving south from source region (N. Canada) to U.S. usually is colder than sfc in U.S.; heat will be added on sfc—airmass becomes less stable, will warm up and become more cloudy. Fresh mT from Gulf moving north into U.S.—in winter usually is warmer than sfc, will be cooled from below (fog, low clouds)—in summer can be still cooler than surface and thus may become more unstable (showers, thunderstorms, tornadoes).

Airmasses over the U.S.—Source regions:

cA from Arctic Ocean/N. coast of Canada/Alaska;
cP from N. Central Canada/Alaska;
mP (due to predominant westerly flow) mainly from Pacific; near East Coast occasionally from Atlantic;
mT from Gulf, sometimes in Eastern states from Atlantic near Bermuda;
cT (rare) from SW-states (Arizona/New Mexico desert).

Effect on our local weather (Madison):

cA cold, clear winter days, N-winds (usually light) strong outgoing (IR) radiation at night—brings coldest days of winter with $T_n < 0°$ F. (Absolute Minimum -37° F., Jan. 30, 1951)
cP similar, but not quite as cold. Both low humidity (relative and total).
mP the usual cold air behind a Cold front. Not as cold as cA or cP, but more cloudiness. In summer and fall, often showery weather due to higher humidity and less stable lapse rate. W-winds.
mT warm sector weather—humid, rainy. Southerly winds, often rather unstable, but occasionally clearing.
cT very rare here; if occurring, very high temperatures (Record breakers!) with rather low relative humidity. Not too many clouds; if some, then usually at high levels—CCL is very high.

D. Tropical Weather Systems

Brief look only. Distinctly different from extra-tropical systems discussed before. Several reasons for this:

a) Tropics are rather uniform—large parts are oceans, insolation in region between two bands of subtropical highs also rather uniform, modified essentially only by variation in cloudiness—Trade Cu vs. ITC cloud systems.
b) Horizontal Corioli's force disappears at equator—thus geostrophic wind assumption not valid there. Result is that pressure gradients are very small (uniformity) and that winds have little relation to gradients.
c) No airmass contrasts—air all over is essentially mT—has no fronts of any consequence. Even if cP or mP should travel far enough south to reach subtropics, they will be modified so much that they no longer can be classified "polar."
d) General motion of air in Tropics from East to West.

Until recently, tropical meteorology was more conjecture than facts, due to lack of data. Only recently we learn more due to satellites. Much still is rather problematic—new discoveries occur constantly. Still, major features at least known even if not completely understood yet.

Normally, weather in tropics very uniform and regular—much regulated by diurnal rather than seasonal changes. Vertical motions rather than horizontal air movements often the most important.

However, regular patterns sometimes disrupted by perturbations in steady flow of air from subtropical highs to ITC. Such perturbation classified into several types. (Relate to regions north of ITC.)

a) Easterly waves—form south of subtropical highs.
Very minor disturbance in pressure field or even wind field, but noticeably by line of showers and/or thunderstorms traveling toward West, disrupt usual tradewind pattern. When hitting islands, bring occasionally heavy squalls and rainfall.
b) Shear lines—often remnants of cold fronts sweeping down into subtropics. Air north of these cold fronts already very much modified, only very slightly cooler than air south of near stationary "front." Hardly airmass boundary, no directional convergence but air north of shear line slightly faster moving. Results in mixing of air along shear line leading to turbulence, showers and generally unsettled weather.
c) Weak tropical depressions—actually low pressure areas, however very hard to find in pressure field due to small pressure gradients and sparsity of stations. Large moisture content, coupled with convergence due to friction near sfc leads to precipitation on a large scale even with slightest provocation. Note that most of

precipitation in tropics is occurring by widespread rain, not by random showers occurring within airmasses. Such tropical weak cyclones may travel for large distances, continuously renewing their moisture supply from evaporation and carrying rain area along.

d) Tropical storms—sometimes (why?) develop from easterly waves or weak cyclones. Probably combination of various favorable factors. Once started, the latent heat liberated by condensation aloft keeps them going, they may grow to considerable size. Windsystem gets organized to true cyclonic vortex, clouds build up to great heights and finally they may develop into the next class—

e) Hurricanes (in Atlantic), Typhoons (East Asia) or Cyclones (Indian Ocean).—Names given to tropical storms once max. windspeed exceeds 75 knots (1 knot = 1 nautical mile/hour). When tropical cyclone develops into this hurricane, it often appears to be more concentrated than the previous less intense and less well-organized cyclone. [See text, pp.149-168,289-290]

Energy to build up and maintain hurricane comes also chiefly from vertical transport of latent heat. When supply of evaporation is cut off—i.e. hurricane moves inland, it will die down rapidly. Most, however, stay over tropical oceans—less than half ever influence inhabited areas. If they do, extensive damage, not only from winds (often > 120 knots) but also from rain, storm seas, high tide effects on coastal areas, etc. Pressure in center ("eye" of hurricane) often very low—measured up to 80 mb lower than outside. Air seems to have been pumped out like in centrifugal vacuum pump. Move originally like easterly waves from East to West south of subtropical highs but may recurve around western edge (near continents) toward N and finally move into Westerlies north of highs. Motions rather slow (10-15 mph), often erratic and difficult to forecast. Reason is that this very intense high-reaching system actively distorts upper flow which normally would steer it. Most impressive (apart from high wind and mountainous seas) is eye of hurricane, a nearly clear and calm area in center, surrounded by the wall of clouds churning around it. During last decade, much research has been done by means of airplane flights through storms at various levels (hurricane hunters); recently also satellites allow to pinpoint positions and intensity on basis of typical cloud patterns.

Season of occurrence—usually late summer and fall in Atlantic, but exceptions do occur (one recently developed in January). Over Western Pacific they can occur (as typhoons) nearly any time. Most (~ 20-25/year) in that area; they also are the most intense and destructive (often windspeeds up to and over 150 knots). Over Atlantic about 5-15/year. Usually affect either coasts of Carribbean Gulf (on their East-West course) or move toward Florida/Georgia coast and subsequently up north (steered around Bermuda High) as far as New England. Normal decay either over land (cutting off moisture supply and increased friction = convergence) or over colder water which limits evaporation also. Decay into intense cyclones of temperate latitudes, causing strong storm lows which frequently can be followed on through westerlies all the way towards and into Europe across the North Atlantic.

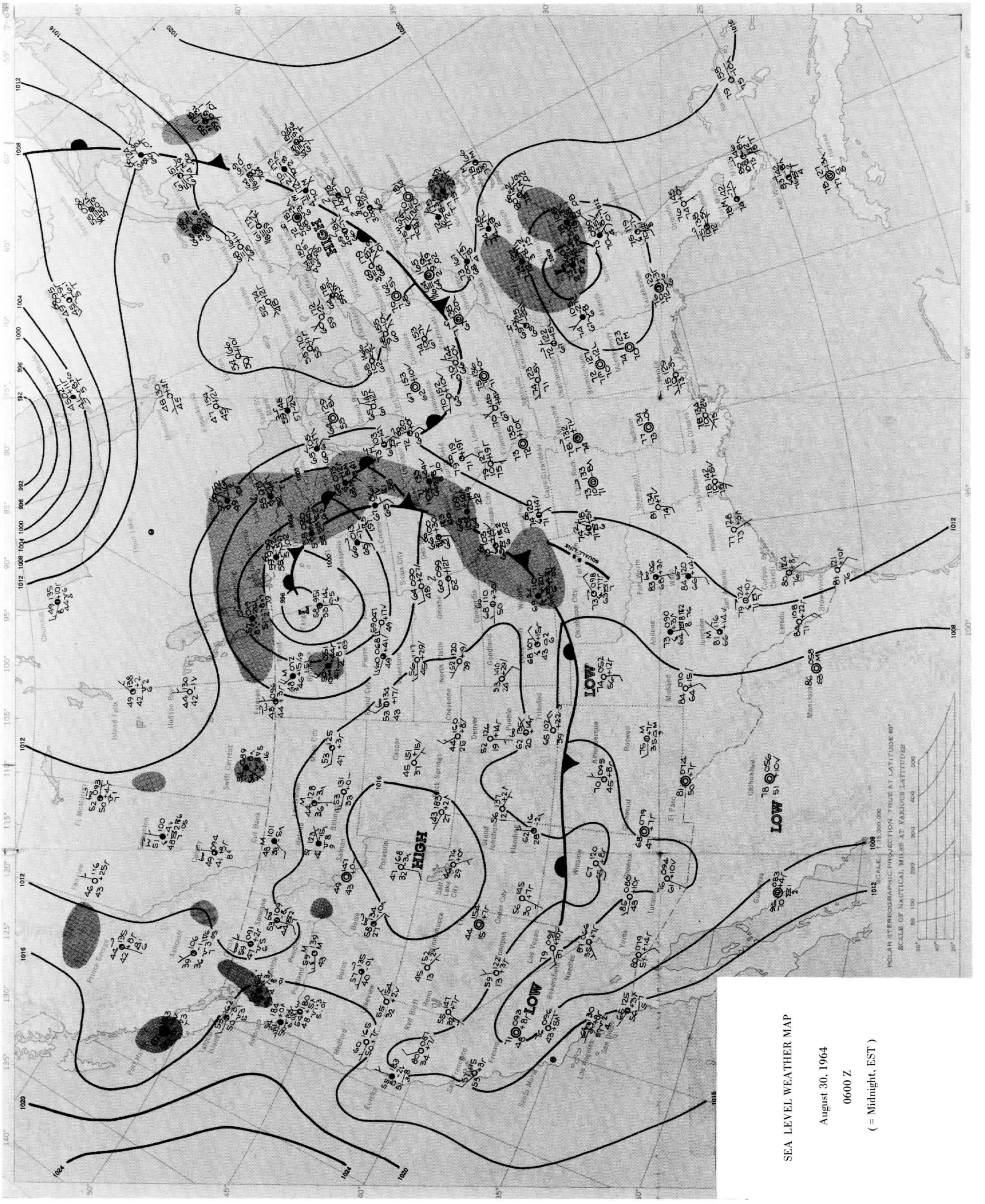
Project: Storm Fury—an attempt to closely investigate and possibly attempt to modify hurricanes. Control obviously extremely difficult—energy in hurricane equal to that of an atom bomb let off every few seconds. Lots of unsolved problems—any control only possible by trigger action—possibly by preventing formation in early stages.

But: if we could kill all hurricanes, how would this affect the rainfall distribution in, say, Florida? And would energy now used up and/or transported in such systems appear somewhere else in atmosphere?

E. The Weather Map

Having discussed the various "synoptic scale systems" in detail, we now can put things together—as meteorologists do daily several times in many hundreds of places all over the world—in their weather maps. They are a rather unique thing—no matter what language you speak or what script you use, you can "read" Wx maps of any country no matter where! Due to international agreement on symbols and the way maps are plotted, any U.S. meteorologist can not only read a Chinese or Russian Wx map but make his forecast from it, plot it up and the Chinese or Russian will again understand what the American has plotted. International agency (attached to U.N.) is WMO—World Meteorological Organization.

Concept of "synoptic" = "seen simultaneously."
Plot of weather elements at large number of stations, observed all at same instant [time in Universal time (UT) also called Z (Zebra time) or GT (Greenwich time)]. Standard times 00Z, 06Z, 12Z, 18Z. At each station, one enters observations according to "station model"—every symbol or figure, located in a specific place, has a very specific meaning.



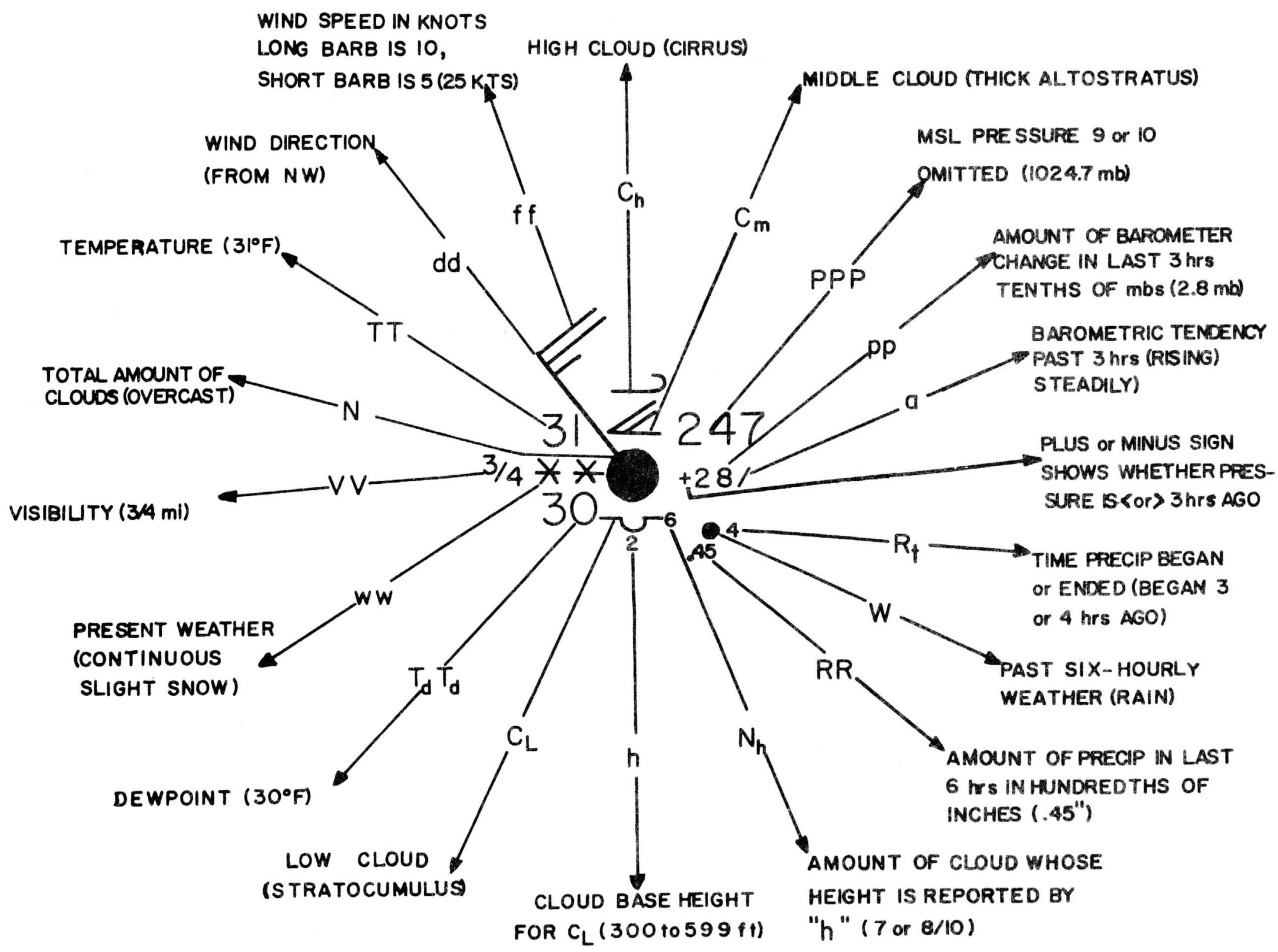
SEA LEVEL WEATHER MAP

August 30, 1964

0600 Z

(= Midnight, EST)

STATION MODEL



From data plotted on such maps at stations, meteorologists analyze map—draw isobars, determine location of fronts, indicate areas of rain/snow or, if so desired, of low clouds, clear weather and many other things. Above weather map is considerably simplified—normally maps are drawn to much larger scale and much more detailed. Using this map, its predecessors (to find out development of systems) and adding to sea level maps also upper level maps—to obtain information of upper flow to estimate steering of lower systems—one attempts to forecast development and motion, in effect tries to plot (construct) a forecast map some time (i.e. 6 hours or 24 hours) in the future. Then, local weatherman tries to determine effect of now predicted features on his own station weather.

Much of work nowadays done in large central forecast centers (in U.S. Washington, D.C.). Motions and development now mostly not "estimated" by subjective means but "computed" by numerical methods—using the basic Equations of Motion and adding mathematically effect of diabatic processes (latent heat, frictional effects, etc.) Need for very large computers, running 24 hrs/day. Started with advent of earliest computers (late 1940's—John v. Neuman/Charney at Princeton) after pioneering first attempt (by hand calculators!) by Richardson in 1922.

Maps transmitted to individual stations (e.g. Truax Field or our Dept.) by Facsimile printer circuits. Other Wx information also transmitted by teletypewriter circuits. Complete international exchange. For reasonable forecasts beyond, say, 2-3 days, one needs data covering essentially the whole hemisphere! Fantastic amount of work needed to allow TV-forecaster to give you "next day's weather"!

EXERCISES—Chapter VIII

1. On graph paper, draw the expected barograph and thermograph curves for the case of
   a) a warm front, warm sector and cold front passage at a station.
   b) passage of a warm- and of a cold front occlusion.

   In addition to these curves, also indicate when and what kind of precipitation will fall.
2. Discuss briefly the reasons for the known fact that all synoptic-scale weather systems are more intense in Winter than in Summer.
3. What are the essential differences between a high behind a cold front and a high in the Subtropics?
4. Discuss the reasons for the difference in slopes (near the surface) of cold and warm fronts.
5. Plot a schematic picture of a low pressure system (with a partially occluded frontal system) and indicate the areas of different kind of clouds and the general area of steady and of showery precipitation.
6. Which airmasses dominate the region of
   a) Seattle,
   b) New Orleans,
   c) Boston,
   d) Miami

   in summer and in winter?
7. A hurricane is located 100 miles east of Miami. What kind of weather do you expect at Miami (wind, temperature, clouds, precipitation, effect on high tide water level, etc.)?
8. Can you describe the weather you would experience on a cruise through the Trades of the North Atlantic and then approaching and crossing through the ITC?
9. Why can a mountain close to or even at the equator still have glaciers and permanent snow fields on top (as has the Kilimandjaro and peaks in the Andes)?
10. On the top of Himalaya peaks (e.g. Mt. Everest) extremely severe weather conditions have been reported. No attempt to scale these mountains has been made other than in the months of April through early June.
    a) Why no ascents at other seasons?
    b) What kind of wind and temperature are found at these heights of over 25,000 ft?
    c) Why are the winds and temperatures so extreme?
11. You see a shield of cirrostratus approaching Madison from the West. If this is related to a warm front, can you estimate how far away the intersection of the front with the surface is from Madison? If the low moves with 20 mph, when would the front reach the Madison area? What kind of clouds would you expect to develop during this approach?

CHAPTER IX. THIRD-ORDER CIRCULATIONS—MESOSCALE SYSTEMS

(Reading: pp. 168-187)

Next lower scale (between 100 km and 100 meters) is called "Mesoscale." Time scales of processes going on at this scale tyically are of order of hours rather than days. Typical processes or phenomena going on at this type of scale are: a) growth and decay of individual Cu; behavior of a whole thunderstorm, development of a tornado. [These are essentially all processes related to vertical stability in conjunction with water vapor/ liquid water phase change.] b) Lee waves behind mountain ranges, Ac lenticularis, local cloud systems over isolated islands [meso-scale processes due to local disturbances in the horizontal wind field] and c) local wind systems like sea breeze, mountain/valley winds, diurnal wind variations, katabatic winds [processes due to topographic features or differential heating of varied surfaces].

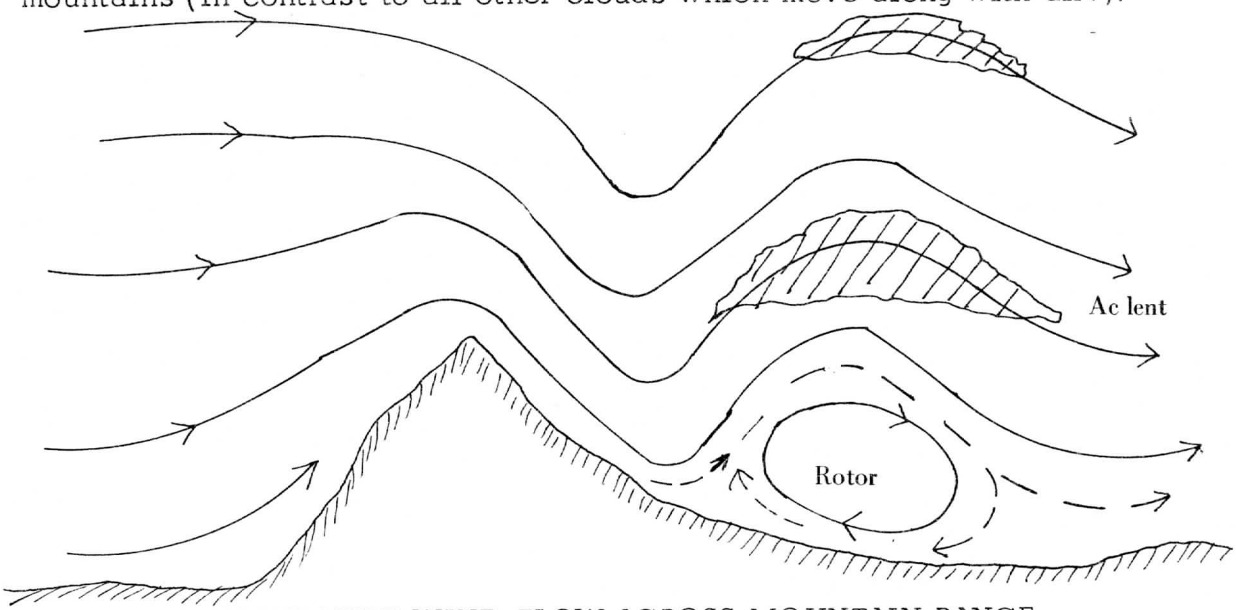
Treatment of these mesoscale processes in this course will be rather brief. Heavy reliance on textbook, in Chapters (pages) included in assignments for Chapter VIII.

A. <u>Thunderstorms and Tornadoes</u>

See text reading assignments in Chapter V, p. V-7. Extensive treatment of topic. Read it!

B. <u>Mountain Lee Waves—Ac lent.</u>

Resulting from (usually) Westerlies encountering a N-S running mountain range (e.g. Sierra Nevada, Rockies). Air is being forced upward; when crossing mountain chain, oscillations are set up in flow leading to <u>stationary</u> wave formation in lee (behind) mountain range. Where air again rises, adiabatic cooling will result in clouds <u>if</u> moisture sufficient. These clouds form over top part of wave, will remain in place with regard to mountains (in contrast to all other clouds which move along with air!).



SCHEMATIC WIND FLOW ACROSS MOUNTAIN RANGE

Called Ac lenticularis (= lens-shaped). Air moves through cloud—thus cloud indicates a process, is made up at any time of different H_2O-droplets. When air again descends to trough of (vertical) wave, droplets evaporate—cloud remains behind and clear area develops in trough. In certain areas one can see a whole series of such clouds—spaced 1 wavelength apart downstream from mountains.

An occasional special feature is the "rotor" indicated in figure. It can, under favorable circumstances, lead to an also stationary rotor cloud (low level) which modifies weather close to mountain range.

Such waves also occur on other wind systems, e.g. those related to cyclones, and mountains don't have to be very high. Well known also in area of Alleghenies where successive series of ridges reinforce wave pattern.

Somewhat related to it are clouds forming over islands, especially in tropical ocean. Air flowing for hundreds of miles undisturbed over ocean suddenly encounters a small island—this a) forces air slightly upwards and b) changes suddenly the frictional effect. Also, island may be heated by sun and thus is warmer. Result is usually development of a Cu-type cloud, which will constantly reform over the island. On flights or sailing in tropical seas, such clouds often can be seen long before one sees the actual island.

C. Local Wind Systems

1. Land-Sea breeze (text, p. 58)

Due to differential heating of land during day and cooling at night. Result of this is a slight change in air density (and thus pressure). Over land daytime local low—nighttime local high. Air will move toward low pressure—but since distances and pressure gradients are so small (order of a few miles or a few tenths of mb only), Coriolis force has no real chance to act. Thus flow of cool air (to replace air ascending due to heating over land) from sea to land in daytime. Arrives over land usually in early afternoon only since time is required for heating. Then, when air cools at night, circulation reverses—air moves from land to sea. Can occur at ocean coasts but also at shores of any reasonably large body of water, e.g. Great Lakes.

One special requirement for having this effect active is that the overall general flow is light (i.e. usually in a High)—otherwise the normal wind system, e.g. associated with a low or the Westerlies will totally obliterate this wind system. Explains, in summer at certain days why, for example, afternoon temperature at Chicago's waterfront may be as much as 5-10°F lower than at O'Hare field.

2. Mountain-Valley winds

Similar effect of heating of valley floor and slopes in daytime and cooling at night also causes flow up or down valleys and slopes (up during day, down at night). Added to down flow of cool air is cooling by outgoing radiation which offsets any (usually only slight) dry-adiabatic warming.

Connected to this is the nighttime flow of cool air into depressions. Cool air is heavier—flows like water down and collects at low spots in topography. Especially in calm weather, these cold pools can become quite distinct. If, in addition no clouds in sky, then strong cooling by radiation and, even in summer, chance of local frost—danger to agriculture (e.g. cranberry bogs). In 1965, such damage occurred in Wisconsin on July 6th!

3. Katabatic winds (text p. 290)

Above example is a small-scale type of Katabatic wind = flow of air due to action of gravity. On larger scale, they are also called fall-winds or drainage winds. Typical examples: Bora (N. coast of Adriatic Sea): cold air originates on plateau inside Yugoslavia, rushes down toward coast as a cold wind. Mistral: N-wind down the Rhonevalley (France) into Mediterranean, cold air forming inside Central France or having arrived there as cold front from the North. Santa Ana—coast of California. Especially violent due to "ducting" in canyons of Santa Ana Mountains.

All wind systems discussed above are usually of rather small horizontal dimensions and also occur only in a rather shallow layer near the surface.

4. Foehn (Chinook) (see pp. IV-20) —Discussed there in detail.

5. Sirocco/Harmattan

Sirocco: Actually not a separate wind system but a local name given to very hot southerly winds blowing dry desert air from Sahara into Mediterranean Sea. In desert itself, it is called Khamsin—usually connected with tremendous amounts of desert sand/dust = sandstorm.

Harmattan: The opposite kind of medium-scale wind—outflow of relatively cool air in winter (radiation cooling during winter nights is largely due to lack of water vapor) toward the South (Gulf of Guince) replacing there the moist mT which dominates this area in other seasons. Brings welcome relief from usual heat but also carries quite a bit of dust along.

6. Blizzard

This again is not a "special wind system" but rather a name given a particular combination of wind and low temperature, occurring in winter in Central plains (and also in other areas favorable—e.g. Antarctica, Siberia (there called buran—or, if connected with driving snow, purga). Essential is a) high windspeeds, usually connected with outbreak of cold arctic air and b) very low temperature. This combination has a very high chilling effect on man and animals, is thus extremely dangerous and thus has been given a special name.

7. Small-scale systems;see text, pp. 51-52, 80-88.

Below Mesoscale we still have MICROSCALE. We include here all processes and phenomena which have (length) scales below, say 10-100 meters—down to cm and even mm, time scales of minutes or seconds. Often motions are collectively called turbulence—individual "air bubbles"—air parcels, air pockets, etc. While we have not said so, much of our discussion in thermodynamics was concerned with this scale—individual air parcel imbedded in environment. This scale very important—because exchange of heat, energy, etc. at earth sfc goes on essentially in this scale, within "boundary layer." Only after energy is available in atmosphere it is then converted to the larger-scale motions—energy transfer in the scale spectrum. However, this microscale is extremely difficult to handle both theoretically and by measurement. Totally new observation methods have to be used—one needs extremely accurate and fast responding instruments—located in close vicinity to each other. Our previous discussion of frictional effects already has indicated some of the difficulties.

Importance in practical problems also very great—e.g. for meteorological application in field of agriculture, city planning, pollution and so on. On this scale also most of direct effects on individual man—biometeorology. Effect of heat, radiation, wind, etc. is microscale (man ~ 6 feet high!).

EXERCISES—Chapter IX

1. Where are the regions a) in the U.S., b) in the world with the highest numbers of thunderstorms during the year?
2. Assuming that the really big thunderstorms develop up to and slightly above the tropopause—what vertical heights would you expect for such a storm a) over Indonesia, b) over Madison, c) over Oslo?
3. Why do hailstones frequently show a layered cross section?
4. Is it advisable to fly through a thunderstorm? Why or why not?
5. Why does one see houses literally explode when hit by the funnel of a tornado?
6. When is the preferred time and place of occurrence of tornadoes (time of year as well as time of day).
7. Besides the examples given in Ch. 9, can you give a few more regions on earth where you would expect to see Ac lent—and under what conditions?
8. When the sea breeze reaches the beach you are swimming on, what would you expect to observe?
9. Write down at least 3 examples each of weather systems or processes belonging to each of the four scales discussed in Chapters 7-9.

CHAPTER X. CLIMATES OF THE EARTH

(Reading: pp. 55-58, 80-92)

A. Climatic Elements

Definition: Climate is the collective state of the earth Atmosphere for a given place within a specified interval of time. (Landsberg)

a) Collective state—contrasted to "individual state" = weather
b) given place—has to be specified in each case. A broad definition, can range all the way from a single field or anthill or even plant ("micro . . .") to larger places (city, county, state, country) ("local") to zones (tropics, temperature, arctic) ["macro . . ."] to the whole earth itself.
c) interval of time—always reasonably long (several years) but has to be defined also—"of our lifetime" (about 30 years—this time span is the most commonly used one to define "climatic normals"—now: 1931-1960), of our century—of historical times—of a geological era. (Ice ages vs. now, etc. climate of the Mesozoic).

Climatology can be approached in various different ways.

a) Descriptive—usually for a given region; since it is highly related to landforms and important for cultural and population studies, it is usually taken as part of studies in geographical sciences.
b) Physical—attempt to understand the particular climatic character of region, zone or whole earth in terms of average energy input and energy conversions. This requires a large amount of meteorological knowledge, applied to various aspects of climate—is generally thought to be intimately linked to study of meteorology.
c) Dynamic—goes even further, attempts to relate climatic character and its changes in time to the general circulation and the other scales of motion to understand interaction; definitely a part of meteorological climatology.

There are also other, more specialized branches of climatology.

a) Micro-climatology—related to climate of small scale features
b) Bio-climatology—related to effect of meteorological and climatic variables upon living beings—plants, animals, men.
c) Upper air climatology—climate of "free atmosphere" under influence of various physical processes such as radiation, solar changes, etc.
d) Airmass climatology—effect of airmass motions and character on area under study, a part of dynamic climatology.

e) Paleo-climatology—study of the climate of the past attempting to determine and to understand it.

Climatic elements are those which can be obtained by combining the "individual" to the "collective state" of the atmosphere—i.e. they are essentially the same elements which the meteorologist measures daily. However, treatment is different and thus resulting conclusions are also distinctly different. Using daily observed values, climatologist also derives a number of additional variables meaningful only for describing collective state, e.g. No. of days with rain/month, average sunshine duration on a specific date, average date of first frost, etc. etc.

Climatic character of a given place determined by

a) its location with respect to the energy input from sun—i.e. its latitude. Was already discussed in Gen. Circ. chapter
b) its location with respect to distribution of land and oceans—effect of heat capacity (and heat admittance) land vs. water, friction, etc.
c) its location with respect to semi-permanent centers of General Circulation.
d) its height above sea level—effect of decrease of pressure and temperature with height.

Recognizing these 4 important factors, one can usually predict quite well the basic climatic pattern in a specific area—a good alien climatologist landing with a spaceship on earth should be able to do this!

Analysis of climatic data—
Since we want the "collective state," large amounts of "individual" data have to be combined in various ways. Today done exclusively by computer data analysis. Most important mathematical tool is statistics in all its varied modern forms. In modern climatology we not only want to know the "mean" value of a certain climatic element but usually also the likelihood of having specified deviations from this mean value—and climatic elements can be extremely complicated and complex.

Examples:

1) Likelihood of wind gust exceeding 100 mph at a specific mountain top, in course of year.—Required for estimating safety factor to build a radar dome at this location.
2) Average length of growing season at a location and various other agriculturally important quantities such as total rainfall during this time, insolation, evaporation, water losses, etc.—to estimate commercial value of planting a specific new (or newly developed) crop.

Representation of climatic data—
By tabulations, maps, (statistical) distribution plots, empirical numerical expressions, etc.—adapted usually to specific use to be made. General

collections of data either in map (Atlas) or tabular form [Climatic Summary of U.S.—from U.S. Weather Bureau].

Climatic changes—

All know that climate cannot possibly stay the same for eternity—we had ice ages. Paleoclimatology interested in this. Since accurate meteorological records usually don't reach back more than at most 150 years, we have to evaluate climate in past eras from indirect evidence. Climatology (especially in this aspect, but also due to its many relations to man's current endeavors) thus becomes a truly "interdisciplinary science." Work together with historians, archaeologists, anthropologists, geologists, physicists, etc.

Example: Climate may change—this has effect on kind of forest growing in area, also may change water temperature of lake in vicinity. Now: trees throw off seed pollen; if forest kind changes, composition of pollens, buried in soil also will change. At a given level, archaeologists find Indian village remnants; refuse heap shows shells of various mussels which the Indians ate. By dating time of habitation (radio carbon dates) one knows time; from pollen profile, what kind of forest, from mussel shells what kind of water temperature—result: a certain temperature value for that time. [In reality, things are usually much more complicated.]

From this kind (and other) of evidence, we now know that climate changes occur within centuries, millenia and over even longer (geological) time periods.

Theory of causes for climatic changes are still far from complete and by no means generally accepted. Can list a few basic reasons why climate could change, either in a particular region or all over earth. (This list by no means complete; just a sample of a few more obvious causes.)

a) Changes in solar radiation (solar constant)—Physical changes in sun itself. Changes in space between sun and earth, e.g. interplanetary dust. Changes in the orbit of earth (perihelion/aphelion distance).
b) Changes in distribution of solar radiation, transmission of radiation through atmosphere.—Changes in atmospheric composition (amount of water vapor, CO_2, dust, etc.) by volcanoes, man's activity, plant vs animal balance. Also secondary changes of surface or atmospheric albedo (large ice and snow shields, increased cloudiness in various zones, etc).
c) Change in location of a region with respect to earth axis—polar wandering, continental drift, mountain building, changes in shore lines and surface geography (lakes, inland seas, salt- vs. fresh water).

All these will, to a larger or lesser degree, determine changes in general circulation (which provides the balance of energy distribution of earth) or at least the effect of such changes upon the climate of a specific region. Man until now has lived within climate—his activities were only passively affected. In recent decades, it appears as if man—now being a major force in shaping earth—also starts actively influencing climate in a big way (has done it for millenia on a smaller scale). It becomes imperative that we understand what makes climate—to prevent our inadvertently changing climage in undesirable ways.

B. Climate Classification

Men like to catalogue things—climate of a region no exception. Has many practical advantages. Basic climatic classification is related to the large climate zones on earth—Tropics, Subtropics, temperate and polar zones. These can be defined geometrically. But, due to secondary causes (ocean/continent, etc.) a somewhat better classification (even on large scale) should be based on actual observations of climatic quantities, the most important for man and his environment are: Temperature, rainfall and some minor ones such as sunshine duration, cloudiness, humidity.

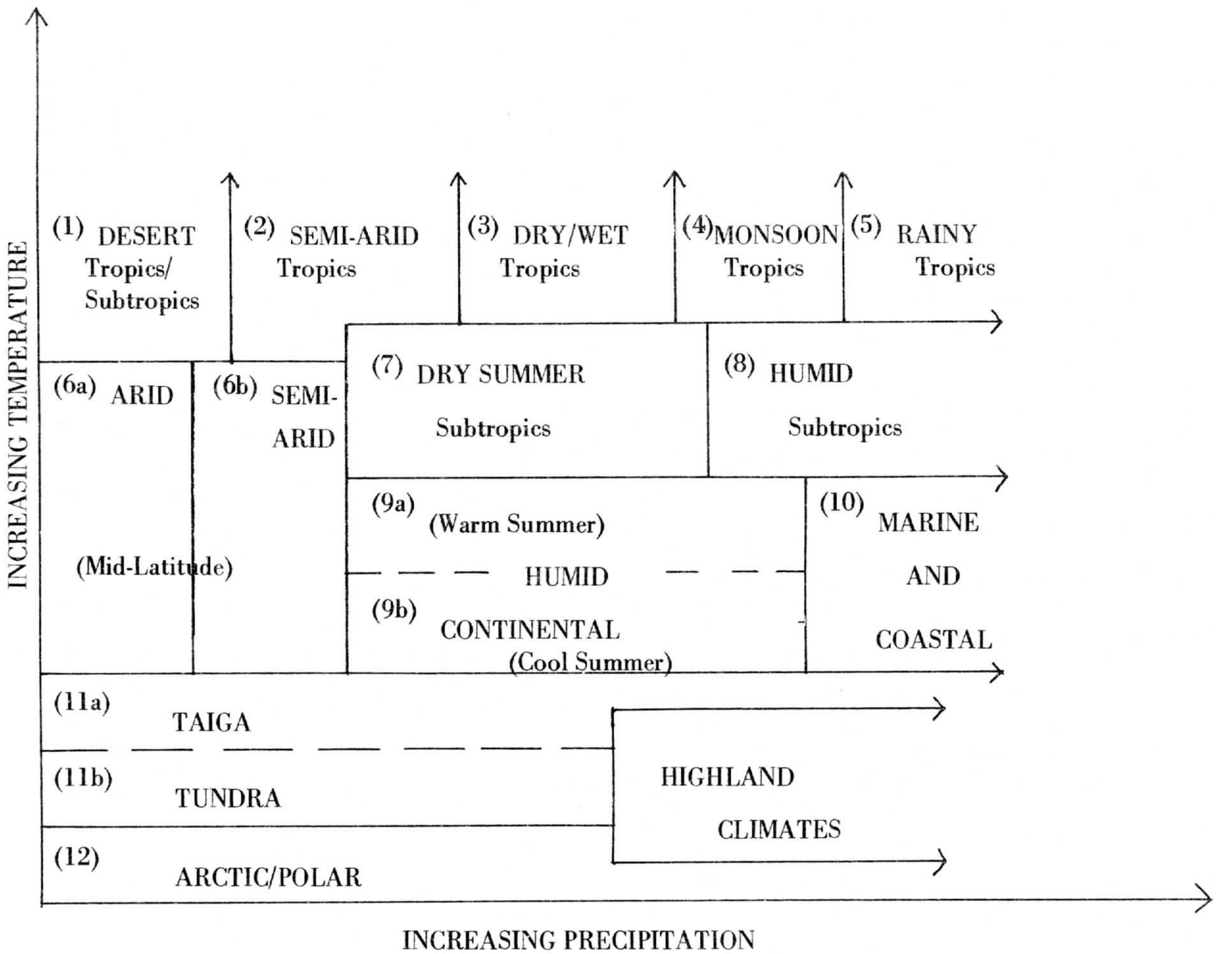
All such classifications, however, are at best abstractions; if one really wants to understand climate, one has to do detailed studies for that particular region.

Most accepted Climate Classification according to W. Koeppen (1900, 1918 revised and extended).

Later further expanded by Geiger and Trewartha (Prof. Geography, U. Wisconsin, now retired).

Other classification based upon slightly different principles by C. W. Thornthwaite (1931) and W. G. Kendrew (1961).

One can schematically locate the various more commonly used and generally known climatic regions in a table combining temperature and precipitation character.



Examples:
(1) Central Sahara, Arabia, Peruvian desert, Ctr. Australia
(2) Sudan, Mato Grosso
(3) Kenya
(4) parts of India, Burma, etc.
(5) Central Africa, Indonesia
(6a) Parts of Jordan, Syria
(6b) Southwestern U.S.
(7) Mediterranean, S. California
(8) Caribbean, parts of Brazil
(9a) Southern U.S., Southern Russia
(9b) Europe, eastern U.S., Southern Argentina
(10) Ireland, Oregon, Japan + many islands
(11a) Siberia
(11b) N. Canada
(12) Antarctica, Shores of Arctic Ocean
(13) Tibet, locations on mountains (Andes, Himalayas)

More detailed climatic classification maps are found in a number of climatology textbooks, atlases, etc.

C. Climatic Zones

Koeppen Basic Classes:

First (capital) letter: large-scale climatic zone
Second letter (capital or lower case): next order subdivision.
Additional classification by added letter symbols (not given here).

A	Tropical Rain Climate	Af	tropical rain forest (all year round rain)
		Aw	Savanna (winters dry)
B	Dry Climates	BW	desert
		BS	steppe
C	Moderate Climates	Cf	oceanic climate (all year rain)
		Cw	winter dry
		Cs	summer dry
D	Moderate Continental Climates	Df	continental, year-round precipitation
		Dw	continental, dry winter
E	Polar Climates	ET	Tundra, Taiga
		EF	glaciated, ice caps, ice covered ocean

Wisconsin, according to this classification, is denoted Df as is most of the Midwest up to Canadian border.

D. Climate of Other Planets

With our new space age, man starts looking outward. How can we estimate climate of another planet?

a) Amount of solar radiation (solar constant for other planets) depends on distance from sun. Zone of life in a solar system.
b) Composition of atmosphere determines distribution of incoming energy—greenhouse effect on earth—H_2O vapor, CO_2, Ozone, etc.
c) Distribution of land vs. water bodies—if there are any. Depends also on total pressure.
d) Rotational period of planet (Coriolis force influence) will determine motions, i.e. general circulation.

When all these things are known, one can make attempt to estimate climate of a planet.

Example: Mars—Know that a) less solar energy, b) atmosphere contains very little oxygen but mostly CO_2, c) no open water, in fact very little water overall, d) very low pressure. Based on all this, one can estimate that Mars is extremely unfavorable for life (as we know it). In 1976, two U.S. Mars landers (Viking 1, 2) landed safely on Mars and obtained valuable scientific data, including confirmation of our theories on Mars' atmosphere. No clear signs of life were found, however - as expected!

Venus another problem—sfc totally obscured by very high dense clouds. In the 1970's, Venus probes have explored planet (mostly by remote sensing from space but also by landings of space capsules from Russian Venera probes). Very high atmospheric pressure (CO_2) (appr. 90-100 times of our own), very strong winds in upper levels, temperatures on Venus' surface of the order of 600 to 750°K.

Nevertheless—when man will eventually reach planets, our theories regarding climate will have to be tested and, in fact, many problems may become easier by not being bound to one celestial body only. In fact, climatological theories may allow us to estimate whether life can or cannot exist. From all our knowledge today we can at least abstract one thing: Earth is a pretty nice place to live on—if only mankind would be a bit more sensible!

EXERCISES—Chapter X

1. What kind of problems would you investigate (give a specific example) if you were working in
   a) Microclimatology
   b) Biometeorology
   c) Paleoclimatology?

2. From archaeological evidence one has found that in prehistoric times Indian tribes in E. Colorado were cultivating corn (maize). Can you infer from this the change in the climate of this region from then to today?