Principles of Climatology

PRINCIPLES OF CLIMATOLOGY

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Preface

This book is a study of continuous changes in climatology. The use of the air mass concept in climatology, the principles of classifications and the evolution of the present day climates has been described briefly.

Climatology is the scientific study of climate. Climate is defined as weather patterns that have been averaged over a given period of time to obtain a consistent pattern of the expected atmospheric conditions. Weather is the atmospheric condition of a particular place over a short period of time, normally a day. Weather averaging for a long and indefinite period of time makes it possible to predict the climatic pattern of an area. Climatology is regarded as a subdivision of physical geography, atmospheric sciences, and earth sciences in general. Aspects of oceanography and biogeography have also been considered as part of climatology. Climatology focuses on aspects such as atmospheric boundary layer, circulation patterns, heat transfer in the globe, ocean interaction with the atmosphere and land surface, land use and topography.

Climatology has evolved from a simple theoretical bookkeeping activity to the current complex scientific and a practical field. Science is defined as the truths and facts that have been obtained through constant research, systematic methods, evaluation of phenomena, and observation. Climatology is therefore scientific method that involves all the aspects that define science. Other than the above stated aspects, to obtain climatic patterns, several scales and gauges are employed in the climatic research. Climatology is not only concerned with the climate of a place but it also establishes the reason for the fluctuation of climate in the area, how human activities lead to climatic variations, effects of the climate on human activities, and the characteristics of the climate. Climate also depends on the layers of the earth and atmosphere, a further manifestation of its scientific nature.

Climatology is important in determining the climatic patterns of a particular region. Establishing the climatic pattern is significant in deciding the economic activities that would thrive in that particular region. If the climate of a region is established to be cool and wet, it would be safe to conclude that agriculture might thrive in the region. Having a clear climate pattern makes it easier for people to understand the seasons of engaging in particular tasks. This is especially most important to tourists and farmers. Infrastructure development, especially buildings are dependent on climate. After a climatic pattern has been established, the engineers recommend the use of materials that would not only withstand the conditions but also protect the dwellers from any harsh climatic condition. Furthermore, climatology seeks to establish why climate varies from place to another.

The present book addresses itself to readers who wish to become familiar with the atmosphere, its structure, weather, climates and their special distribution as well as the influence of various geoclimatic regions.

—*Dr. Udhav Eknath Chavan*

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1

Nature and Scope of Climatology

INTRODUCTION

Climatology is the study of climate, scientifically defined as weather conditions averaged over a period of time, and is a branch of the atmospheric sciences. Basic knowledge of climate can be used within shorter term weather forecasting using analog techniques such as the El Niño – Southern Oscillation (ENSO), the Madden-Julian Oscillation (MJO), the North Atlantic Oscillation (NAO), the Northern Annualar Mode (NAM), the Arctic oscillation (AO), the Northern Pacific (NP) Index, the Pacific Decadal Oscillation PDO), and the Interdecadal Pacific Oscillation (IPO). Climate models are used for a variety of purposes from study of the dynamics of the weather and climate system to projections of future climate.

History

The earliest person to hypothesize the concept of climate change may have been the medieval Chinese scientist Shen Kuo (1031–95). Shen Kuo theorized that climates naturally shifted over an enormous span of time, after observing petrified bamboos found underground near Yanzhou (modern day Yan'an, Shaanxi province), a dry climate area unsuitable for the growth of bamboo.

Early climate researchers include Edmund Halley, who published a map of the trade winds in 1686, after a voyage to the southern hemisphere. Benjamin Franklin, in the 18th century, was the first to map the course of the Gulf Stream for use in sending mail overseas from the United States to Europe. Francis Galton invented the term *anticyclone*. Helmut Landsberg led to statistical analysis being used in climatology, which led to its evolution into a physical sciences.

PALEOCLIMATOLOGY

Paleoclimatology (also palaeoclimatology) is the study of changes in climate taken on the scale of the entire history of Earth. It uses a variety of proxy methods from the Earth and life sciences to obtain data previously preserved within (e.g.) rocks, sediments, ice sheets, tree rings, corals, shells and microfossils; it then uses these records to determine the past states of the Earth's various climate regions and its atmospheric system.

Paleoclimatology has wider implications for climate change today. Scientists often consider past changes in environment and biodiversity to reflect on the current situation, and specifically the impact of climate on mass extinctions and biotic recovery.

Reconstructing Ancient Climates

Paleoclimatologists employ a wide variety of techniques to deduce ancient climates.

Ice

Mountain Glaciers and the polar ice caps/ice sheets are a widely employed source of data in paleoclimatology. Recent ice coring projects in the ice caps of Greenland and Antarctica have yielded data going back several hundred thousand years—over 800,000 years in the case of the EPICA project.

- Air trapped within fallen snow becomes encased in tiny bubbles as the snow is compressed into ice in the glacier under the weight of later years' snow. This trapped air has proven a tremendously valuable source for direct measurement of the composition of air from the time the ice was formed.

- Layering can be observed due to seasonal pauses in ice accumulation and can be used to establish chronology; associating specific depths of the core with ranges of time.
- Changes in the layering thickness can be used to determine changes in precipitation or temperature.
- Oxygen-18 quantity changes (ä^{18}O) in ice layers represent changes in average ocean surface temperature. Water molecules containing the heavier O-18 evaporate at a higher temperature than water molecules containing the normal Oxygen-16 isotope. The ratio of O-18 to O-16 will be higher as temperature increases and less as temperature decreases. Various cycles in those isotope ratios have been detected.
- Pollen has been observed in the ice cores and can be used to understand which plants were present as the layer formed. Pollen is produced in abundance and its distribution is typically well understood. A pollen count for a specific layer can be produced by observing the total amount of pollen categorized by type (shape) in a controlled sample of that layer. Changes in plant frequency over time can be plotted through statistical analysis of pollen counts in the core. Knowing which plants were present leads to an understanding of precipitation and temperature, and types of fauna present. Palynology includes the study of pollen for these purposes.
- Volcanic ash is contained in some layers, and can be used to establish the time of the layer's formation. Each volcanic event distributed ash with a unique set of properties (shape and color of particles, chemical signature). Establishing the ash's source will establish a range of time to associate with layer of ice.

Dendroclimatology

Climatic information can be obtained through an understanding of changes in tree growth. Generally, trees respond to changes in climatic variables by speeding up or slowing down growth, which in turn is generally reflected a greater or lesser thickness in growth rings. Different species, however, respond to changes in climatic variables in different ways. A tree-ring record is established by compiling information from many living trees in a specific area.

Older intact wood that has escaped decay can extend the time covered by the record by matching the ring depth changes to contemporary specimens. Using this method some areas have tree-ring records dating back a few thousand years. Older wood not connected to a contemporary record can be dated generally with radiocarbon techniques. A tree-ring record can be used to produce information regarding precipitation, temperature, hydrology, and fire corresponding to a particular area.

On a longer time scale, geologists must refer to the sedimentary record for data.

Sedimentary Content

- Sediments, sometimes lithified to form rock, may contain remnants of preserved vegetation, animals, plankton or pollen, which may be characteristic of certain climatic zones.
- Biomarker molecules such as the alkenones may yield information about their temperature of formation.
- Chemical signatures, particularly Mg/Ca ratio of calcite in Foraminifera tests, can be used to reconstruct past temperature.
- Isotopic ratios can provide further information. Specifically, the ä^{18}O record responds to changes in temperature and ice volume, and the ä^{13}C record reflects a range of factors, which are often difficult to disentangle.

Sedimentary Facies

On a longer time scale, the rock record may show signs of sea level rise and fall; further, features such as "fossilised" sand dunes can be identified. Scientists can get a grasp of long term climate by studying sedimentary rock going back billions of years. The division of earth history into separate periods is largely based on visible changes in sedimentary rock layers that demarcate major changes in conditions. Often these include major shifts in climate.

Corals

Coral "rings" are similar to tree rings, except they respond to different things, such as the water temperature and wave action. From this source, certain equipment can be used to derive the sea surface temperature and water salinity from the past few centuries. The ä^{18}O of coraline red algae provides a useful proxy of sea surface temperature at high latitudes, where many traditional techniques are limited.

Limitations

A consortium called the European Project for Ice Coring in Antarctica (EPICA) has drilled at Dome C in the East Antarctic ice sheet, and retrieved an ice core which dates to roughly 740,000 years old. The project has been working in the same location to core a final section that could reach back to 900,000 years or even further. The international ice core community has, under the auspices of International Partnerships in Ice Core Sciences (IPICS), defined a priority project to obtain the oldest possible ice core record from Antarctica, an ice core record reaching back to or towards 1.5 million years ago. The deep marine record, the source of most isotopic data, only exists on oceanic plates, which are eventually subducted — the oldest remaining material is 200 million years old. Older sediments are also more prone to corruption by diagenesis. Resolution and confidence in the data decrease over time.

Notable Climate Events in Earth History

Knowledge of precise climatic events decreases as the record goes further back in time. Some notable climate events:

- Faint young Sun paradox (start)
- Huronian glaciation (~2400Mya Earth completely covered in ice probably due to Great Oxygenation Event)
- Later Neoproterozoic Snowball Earth (~600Mya, Precursor to the Cambrian Explosion)
- Andean-Saharan glaciation (~450Mya)
- Carboniferous Rainforest Collapse (~300Mya)

- Permian-Triassic extinction event (251.4Mya)
- Paleocene-Eocene Thermal Maximum (Paleocene-Eocene, 55Mya)
- Younger Dryas/The Big Freeze (~11Kya)
- Holocene climatic optimum (~7-3Kya)
- Climate changes of 535-536 (535-536 AD)
- Medieval warm period (900-1300)
- Little ice age (1300-1800)
- Year Without a Summer (1816).

ICE CORE

An ice core is a core sample that is typically removed from an ice sheet, most commonly from the polar ice caps of Antarctica, Greenland or from high mountain glaciers elsewhere. As the ice forms from the incremental build up of annual layers of snow, lower layers are older than upper, and an ice core contains ice formed over a range of years. The properties of the ice and the recrystallized inclusions within the ice can then be used to reconstruct a climatic record over the age range of the core, normally through isotopic analysis. This enables the reconstruction of local temperature records and the history of atmospheric composition.

Ice cores contain an abundance of climate information. Inclusions in the snow of each year remain in the ice, such as wind-blown dust, ash, bubbles of atmospheric gas and radioactive substances. The variety of climatic proxies is greater than in any other natural recorder of climate, such as tree rings or sediment layers. These include (proxies for) temperature, ocean volume, precipitation, chemistry and gas composition of the lower atmosphere, volcanic eruptions, solar variability, sea-surface productivity, desert extent and forest fires.

The length of the record depends on the depth of the ice core and varies from a few years up to 800 kyr for the EPICA core. The time resolution (i.e. the shortest time period which can be accurately distinguished) depends on the amount of annual snowfall, and reduces with depth as the ice compacts under the weight of layers accumulating on top of it. Upper layers of ice in a core correspond to a single year or sometimes a single season. Deeper into the ice the layers thin and annual layers become indistinguishable.

An ice core from the right site can be used to reconstruct an uninterrupted and detailed climate record extending over hundreds of thousands of years, providing information on a wide variety of aspects of climate at each point in time. It is the simultaneity of these properties recorded in the ice that makes ice cores such a powerful tool in paleoclimate research.

Structure of ice Sheets and Cores

Ice sheets are formed from snow. Because an ice sheet survives summer, the temperature in that location usually does not warm much above freezing. In many locations in Antarctica the air temperature is always well below the freezing point of water. If the summer temperatures do get above freezing, any ice core record will be severely degraded or completely useless, since meltwater will percolate into the snow.

The surface layer is snow in various forms, with air gaps between snowflakes. As snow continues to accumulate, the buried snow is compressed and forms *firn*, a grainy material with a texture similar to granulated sugar. Air gaps remain, and some circulation of air continues.

As snow accumulates above, the firn continues to densify, and at some point the pores close off and the air is trapped. Because the air continues to circulate until then, the ice age and the age of the gas enclosed are not the same, and may differ by hundreds of years. The gas age–ice age difference is as great as 7 kyr in glacial ice from Vostok.

Under increasing pressure, at some depth the firn is compressed into ice. This depth may range between a few to several tens of meters to typically 100 m for Antarctic cores. Below this level material is frozen in the ice. Ice may appear clear or blue.

Layers can be visually distinguished in firn and in ice to significant depths. In a location on the summit of an ice sheet where there is little flow, accumulation tends to move down and away, creating layers with minimal disturbance. In a location where underlying ice is flowing, deeper layers may have increasingly different characteristics and distortion. Drill cores near bedrock often are challenging to analyse due to distorted flow patterns and composition likely to include materials from the underlying surface.

Characteristics of Firn

The layer of porous firn on Antarctic ice sheets is 50–150 m deep. It is much less deep on glaciers. Air in the atmosphere and firn are slowly exchanged by molecular diffusion through pore spaces, because gases move toward regions of lower concentration. Thermal diffusion causes isotope fractionation in firn when there is rapid temperature variation, creating isotope differences which are captured in bubbles when ice is created at the base of firn. There is gas movement due to diffusion in firn, but not convection except very near the surface.

Below the firn is a zone in which seasonal layers alternately have open and closed porosity. These layers are sealed with respect to diffusion. Gas ages increase rapidly with depth in these layers. Various gases are fractionated while bubbles are trapped where firn is converted to ice.

Coring

A core is collected by separating it from the surrounding material. For material which is sufficiently soft, coring may be done with a hollow tube. Deep core drilling into hard ice, and perhaps underlying bedrock, involves using a hollow drill which actively cuts a cylindrical pathway downward around the core.

When a drill is used, the cutting apparatus is on the bottom end of a drill barrel, the tube which surrounds the core as the drill cuts downward around the edge of the cylindrical core. The length of the drill barrel determines the maximum length of a core sample (6 m at GISP2 and Vostok). Collection of a long core record thus requires many cycles of lowering a drill/barrel assembly, cutting a core 4–6 m in length, raising the assembly to the surface, emptying the core barrel, and preparing a drill/barrel for drilling.

Because deep ice is under pressure and can deform, for cores deeper than about 300 m the hole will tend to close if there is nothing to supply back pressure. The hole is filled with a fluid to keep the hole from closing. The fluid, or mixture of fluids, must simultaneously satisfy criteria for density, low viscosity, frost resistance, as well as workplace safety and environmental compliance. The fluid must also satisfy other criteria, for example those stemming from the analytical methods employed on the ice core. A number of different fluids and fluid combinations have been tried in the past. Since GISP2 (1990–1993) the US Polar Program has utilized a single-component fluid system, n-butyl acetate, but the toxicology, flammability, aggressive solvent nature, and longterm liabilities of n-butyl acetate raises serious questions about its continued application.

The European community, including the Russian program, has concentrated on the use of two-component drilling fluid consisting of low-density hydrocarbon base (brown kerosene was used at Vostok) boosted to the density of ice by addition of halogenated-hydrocarbon densifier. Many of the proven densifier products are now considered too toxic, or are no longer available due to efforts to enforce the Montreal Protocol on ozone-depleting substances. In April 1998 on the Devon Ice Cap filtered lamp oil was used as a drilling fluid. In the Devon core it was observed that below about 150 m the stratigraphy was obscured by microfractures.

Core Processing

Modern practice is to ensure that cores remain uncontaminated, since they are analysed for trace quantities of chemicals and isotopes. They are sealed in plastic bags after drilling and analysed in clean rooms.

The core is carefully extruded from the barrel; often facilities are designed to accommodate the entire length of the core on a horizontal surface. Drilling fluid will be cleaned off before the core is cut into 1-2 meter sections. Various measurements may be taken during preliminary core processing.

Current practices to avoid contamination of ice include:

- Keeping ice well below the freezing point.
  - At Greenland and Antarctic sites, temperature is maintained by having storage and work areas under the snow/ice surface.
  - At GISP2, cores were never allowed to rise above-15 °C, partly to prevent microcracks from forming and allowing present-day air to contaminate the fossil air trapped in the ice fabric, and partly to inhibit recrystallization of the ice structure.
- Wearing special clean suits over cold weather clothing.
- Mittens or gloves.
- Filtered respirators.
- Plastic bags, often polyethylene, around ice cores. Some drill barrels include a liner.
- Proper cleaning of tools and laboratory equipment.
- Use of laminar-flow bench to isolate core from room particulates.

For shipping, cores are packed in Styrofoam boxes protected by shock absorbing bubble-wrap. Due to the many types of analysis done on core samples, sections of the core are scheduled for specific uses. After the core is ready for further analysis, each section is cut as required for tests. Some testing is done on site, other study will be done later, and a significant fraction of each core segment is reserved for archival storage for future needs. Projects have used different core-processing strategies. Some projects have only done studies of physical properties in the field, while others have done significantly more study in the field. These differences are reflected in the core processing facilities.

Ice Relaxation

Deep ice is under great pressure. When brought to the surface, there is a drastic change in pressure. Due to the internal pressure and varying composition, particularly bubbles, sometimes cores are very brittle and can break or shatter during handling. At Dome

C, the first 1000 m were brittle ice. Siple dome encountered it from 400 to 1000 m. It has been found that allowing ice cores to rest for some time (sometimes for a year) makes them become much less brittle.

Decompression causes significant volume expansion (called relaxation) due to microcracking and the exsolving of enclathratized gases. Relaxation may last for months. During this time, ice cores are stored below-10 °C to prevent cracking due to expansion at higher temperatures. At drilling sites, a relaxation area is often built within existing ice at a depth which allows ice core storage at temperatures below-20 °C.

It has been observed that the internal structure of ice undergoes distinct changes during relaxation. Changes include much more pronounced cloudy bands and much higher density of "white patches" and bubbles. Several techniques have been examined. Cores obtained by hot water drilling at Siple Dome in 1997–1998 underwent appreciably more relaxation than cores obtained with the PICO electro-mechanical drill. In addition, the fact that cores were allowed to remain at the surface at elevated temperature for several days likely promoted the onset of rapid relaxation.

Ice Core Data

Many materials can appear in an ice core. Layers can be measured in several ways to identify changes in composition. Small meteorites may be embedded in the ice. Volcanic eruptions leave identifiable ash layers. Dust in the core can be linked to increased desert area or wind speed.

Isotopic analysis of the ice in the core can be linked to temperature and global sea level variations. Analysis of the air contained in bubbles in the ice can reveal the palaeocomposition of the atmosphere, in particular CO_2 variations. There are great problems relating the dating of the included bubbles to the dating of the ice, since the bubbles only slowly "close off" after the ice has been deposited. Nonetheless, recent work has tended to show that during deglaciations CO_2 increases lags temperature increases by 600 +/-400 years. Beryllium-10 concentrations are linked to cosmic ray intensity which can be a proxy for solar strength.

There may be an association between atmospheric nitrates in ice and solar activity. However, recently it was discovered that sunlight triggers chemical changes within top levels of firn which significantly alter the pore air composition. This raises levels of formaldehyde and NOx. Although the remaining levels of nitrates may indeed be indicators of solar activity, there is ongoing investigation of resulting and related effects of effects upon ice core data.

Core Contamination

Some contamination has been detected in ice cores. The levels of lead on the outside of ice cores is much higher than on the inside. In ice from the Vostok core (Antarctica), the outer portion of the cores have up to 3 and 2 orders of magnitude higher bacterial density and dissolved organic carbon than the inner portion of the cores, respectively, as a result of drilling and handling.

Paleoatmospheric Sampling

As porous snow consolidates into ice, the air within it is trapped in bubbles in the ice. This process continuously preserves samples of the atmosphere. In order to retrieve these natural samples the ice is ground at low temperatures, allowing the trapped air to escape. It is then condensed for analysis by gas chromatography or mass spectrometry, revealing gas concentrations and their isotopic composition respectively. Apart from the intrinsic importance of knowing relative gas concentrations (e.g. to estimate the extent of greenhouse warming), their isotopic composition can provide information on the sources of gases. For example CO_2 from fossil-fuel or biomass burning is relatively depleted in ^{13}C.

Dating the air with respect to the ice it is trapped in is problematic. The consolidation of snow to ice necessary to trap the air takes place at depth (the 'trapping depth') once the pressure of overlying snow is great enough. Since air can freely diffuse from the overlying atmosphere throughout the upper unconsolidated layer (the 'firn'), trapped air is younger than the ice surrounding it.

Trapping depth varies with climatic conditions, so the air-ice age difference could vary between 2500 and 6000 years (Barnola et al., 1991). However, air from the overlying atmosphere may not mix uniformly throughout the firn (Battle et al., 1986) as earlier assumed, meaning estimates of the air-ice age difference could be less than imagined. Either way, this age difference is a critical uncertainty in dating ice-core air samples. In addition, gas movement would be different for various gases; for example, larger molecules would be unable to move at a different depth than smaller molecules so the ages of gases at a certain depth may be different. Some gases also have characteristics which affect their inclusion, such as helium not being trapped because it is soluble in ice.

In Law Dome ice cores, the trapping depth at DE08 was found to be 72 m where the age of the ice is 40±1 years; at DE08-2 to be 72 m depth and 40 years; and at DSS to be 66 m depth and 68 years.

Paleoatmospheric Firn Studies

At the South Pole, the firn-ice transition depth is at 122 m, with a CO_2 age of about 100 years. Gases involved in ozone depletion, CFCs, chlorocarbons, and bromocarbons, were measured in firn and levels were almost zero at around 1880 except for CH_3Br, which is known to have natural sources. Similar study of Greenland firn found that CFCs vanished at a depth of 69 m (CO_2 age of 1929).

Analysis of the Upper Fremont Glacier ice core showed large levels of chlorine-36 that definitely correspond to the production of that isotope during atmospheric testing of nuclear weapons. This result is interesting because the signal exists despite being on a glacier and undergoing the effects of thawing, refreezing, and associated meltwater percolation. ^{36}Cl has also been detected in the Dye-3 ice core (Greenland), and in firn at Vostok.

Studies of gases in firn often involve estimates of changes in gases due to physical processes such as diffusion. However, it has been noted that there also are populations of bacteria in surface snow and firn at the South Pole, although this study has been challenged. It had previously been pointed out that anomalies in some trace gases may be explained as due to accumulation of in-situ metabolic trace gas byproducts.

Dating Cores

Shallow cores, or the upper parts of cores in high-accumulation areas, can be dated exactly by counting individual layers, each representing a year. These layers may be visible, related to the nature of the ice; or they may be chemical, related to differential transport in different seasons; or they may be isotopic, reflecting the annual temperature signal (for example, snow from colder periods has less of the heavier isotopes of *H* and *O*). Deeper into the core the layers thin out due to ice flow and high pressure and eventually individual years cannot be distinguished. It may be possible to identify events such as nuclear bomb atmospheric testing's radioisotope layers in the upper levels, and ash layers corresponding to known volcanic eruptions. Volcanic eruptions may be detected by visible ash layers, acidic chemistry, or electrical resistance change. Some composition changes are detected by high-resolution scans of electrical resistance. Lower down the ages are reconstructed by modelling accumulation rate variations and ice flow.

Dating is a difficult task. Five different dating methods have been used for Vostok cores, with differences such as 300 years at 100 m depth, 600yr at 200 m, 7000yr at 400 m, 5000yr at 800 m, 6000yr at 1600 m, and 5000yr at 1934 m. Different dating methods makes comparison and interpretation difficult. Matching peaks by visual examination of Moulton and Vostok ice cores suggests a time difference of about 10,000 years but proper interpretation requires knowing the reasons for the differences.

Ice Core Storage and Transport

Ice cores are typically stored and transported in refrigerated ISO container systems. Due to the high value and the temperature-sensitive nature of the ice core samples, container systems with primary and back-up refrigeration units and generator sets are often used. Known as a Redundant Container System in the industry, the refrigeration unit and generator set automatically switches to its back-up in the case of a loss of performance or power to provide the ultimate peace of mind when shipping this valuable cargo.

Ice Core Sites

Ice cores have been taken from many locations around the world. Major efforts have taken place on Greenland and Antarctica. Sites on Greenland are more susceptible to snow melt than those in Antarctica. In the Antarctic, areas around the Antarctic Peninsula and seas to the west have been found to be affected by ENSO effects. Both of these characteristics have been used to study such variations over long spans of time.

Greenland

The first to winter on the inland ice was J.P. Koch and Alfred Wegener in a hut they built on the ice in Northeast Greenland. Inside the hut they drilled to a depth of 25 m with an auger similar to an oversized corkscrew.

Station Eismitte

Eismitte means Ice-Centre in German, and the campsite was located 402 kilometers (250 miles) from the coast at an estimated altitude of 3,000 meters (9,843 feet). As a member of the Alfred Wegener Expedition to Eismitte in central Greenland from July 1930 to August 1931, Ernst Sorge hand-dug a 15 m deep pit adjacent to his beneath-the-surface snow cave. Sorge was the first to systematically and quantitatively study the near-surface snow/firn strata from inside his pit. His research validated the feasibility of measuring the preserved annual snow accumulation cycles, like measuring frozen precipitation in a rain gauge.

Camp VI

During 1950-1951 members of Expeditions Polaires Francaises (EPF) led by Paul Emile Victor reported boring two holes to depths of 126 and 150 m on the central Greenland inland ice at Camp VI and Station Central (Centrale). Camp VI is in the western part of Greenland on the EPF-EGIG line at an elevation of 1598 masl.

Station Centrale

The Station Centrale was not far from station Eismitte. Centrale is on a line between Milcent (70°18′N 45°35′W, 2410 masl) and Crête (71°7′N 37°19′W), at about (70°43′N 41°26′W), whereas Eismitte is at (71°10′N 39°56′W, ~3000 masl).

Site 2

In 1956, pre-International Geophysical Year (IGY) of 1957-58, a 10 cm diameter core using a rotary mechanical drill (US) to 305 m was recovered. A second 10 cm diameter core was recovered in 1957 by the same drill rig to 411 m. A commercially modified, mechanical-rotary Failing-1500 rock-coring rig was used, fitted with special ice cutting bits.

Camp Century

Three cores were attempted at Camp Century in 1961, 1962, and again in 1963. The third hole was started in 1963 and reached 264 m. The 1963 hole was re-entered using the thermal drill (US) in 1964 and extended to 535 m. In mid-1965 the thermal drill was replaced with an electro-mechanical drill, 9.1 cm diameter, that reached the base of the ice sheet in July 1966 at 1387 m. The Camp Century, Greenland, (77°10′N 61°08′W, 1885 masl) ice core (cored from 1963–1966) is 1390 m deep and contains climatic oscillations with periods of 120, 940, and 13,000 years. Another core in 1977 was drilled at Camp Century using a Shallow (Dane) drill type, 7.6 cm diameter, to 100 m.

North Site

At the North Site (75°46′N 42°27′W, 2870 masl) drilling began in 1972 using a SIPRE (US) drill type, 7.6 cm diameter to 25 m. The North Site was 500 km north of the EGIG line. At a depth of 6–7 m diffusion had obliterated some of the seasonal cycles.

North Central

The first core at North Central (74°37′N 39°36′W) was drilled in 1972 using a Shallow (Dane) drill type, 7.6 cm diameter to 100 m.

Crête

At Crête in central Greenland (71°7′N 37°19′W) drilling began in 1972 on the first core using a SIPRE (US) drill type, 7.6 cm diameter to 15 m.

The Crête core was drilled in central Greenland (1974) and reached a depth of 404.64 meters, extending back only about fifteen centuries. Annual cycle counting showed that the oldest layer was deposited in 534 AD. The Crête 1984 ice cores consist of 8 short cores drilled in the 1984-85 field season as part of the post-GISP campaigns. Glaciological investigations were carried out in the field at eight core sites (A-H).

Milcent

"The first core drilled at Station Milcent in central Greenland covers the past 780 years." Milcent core was drilled at 70.3°N, 44.6°W, 2410 masl. The Milcent core (398 m) was 12.4 cm in diameter, using a Thermal (US) drill type, in 1973.

Dye 2

Drilling with a Shallow (Swiss) drill type at Dye 2 (66°23′N 46°11′W, 2338 masl) began in 1973. The core was 7.6 cm in diameter to a depth of 50 m. A second core to 101 m was 10.2 cm in diameter was drilled in 1974. An additional core at Dye 2 was drilled in 1977 using a Shallow (US) drill type, 7.6 cm diameter, to 84 m.

Summit Camp

The camp is located approximately 360 km from the east coast and 500 km from the west coast of Greenland at (Saattut, Uummannaq), and 200 km NNE of the historical ice sheet camp Eismitte. The closest town is Ittoqqortoormiit, 460 km ESE of the station. The station however is not part of Sermersooq municipality, but falls within the bounds of the Northeast Greenland National Park.

An initial core at Summit (71°17′N 37°56′W, 3212 masl) using a Shallow (Swiss) drill type was 7.6 cm in diameter for 31 m in 1974. Summit Camp, also Summit Station, is a year-round research station on the apex of the Greenland Ice Sheet. Its coordinates are variable, since the ice is moving. The coordinates provided here (72°34′45"N 38°27′26"W, 3212 masl) are as of 2006.

South Dome

The first core at South Dome (63°33′N 44°36′W, 2850 masl) used a Shallow (Swiss) drill type for a 7.6 cm diameter core to 80 m in 1975.

Hans Tausen (or Hans Tavsen)

The first GISP core drilled at Hans Tausen (82°30′N 38°20′W, 1270 masl) was in 1975 using a Shallow (Swiss) drill type, 7.6 cm diameter core to 60 m. The second core at Hans Tausen was drilled in 1976 using a Shallow (Dane) drill type, 7.6 cm diameter to 50 m. The drilling team reported that the drill was stuck in the drill hole and lost.

The Hans Tausen ice cap in Peary Land was drilled again with a new deep drill to 325 m. The ice core contained distinct melt layers all the way to bedrock indicating that Hans Tausen contains no ice from the glaciation; i.e., the world's northernmost ice cap melted away during the post-glacial climatic optimum and was rebuilt when the climate got colder some 4000 years ago.

Camp III

The first core at Camp III (69°43′N 50°8′W) was drilled in 1977 using a Shallow (Swiss) drill type, 7.6 cm, to 49 m. The last core at Camp III was drilled in 1978 using a Shallow (Swiss) drill type, 7.6 cm diameter, 80 m depth.

Dye 3

The Greenland Ice Sheet Project (GISP) including Dye 3 was a decade-long project to drill 20 ice cores in Greenland.

Renland

The Renland ice core from East Greenland apparently covers a full glacial cycle from the Holocene into the previous Eemian interglacial. It was drilled in 1985 to a length of 325 m. From the delta-profile, the Renland ice cap in the Scoresbysund Fiord has always been separated from the inland ice, yet all the delta-leaps revealed in the Camp Century 1963 core recurred in the Renland ice core.

GRIP/GISP

The GRIP and GISP cores, each about 3000 m long, were drilled by European and US teams respectively on the summit of Greenland. Their usable record stretches back more than 100,000 years into the last interglacial. They agree (in the climatic history recovered) to a few metres above bedrock. However, the lowest portion of these cores cannot be interpreted, probably due to disturbed flow close to the bedrock. There is evidence the GISP2 cores contain an increasing structural disturbance which casts suspicion on features lasting centuries or more in the bottom 10% of the ice sheet. The more recent NorthGRIP ice core provides a undisturbed record to approx. 123,000 years before present. The results indicate that Holocene climate has been remarkably stable and have confirmed the occurrence of rapid climatic variation during the last ice age.

NGRIP

The NGRIP site was chosen to extract a long and undisturbed record stretching into the last glacial. NGRIP covers 5 kyr of the Eemian, and shows that temperatures then were roughly as stable as the pre-industrial Holocene temperatures were.

NEEM

The North Greenland Eemian Ice Drilling (NEEM) site is located at 77°27′N 51°3.6′W, masl. Drilling started in June 2009. The ice at NEEM was expected to be 2545 m thick. On July 26, 2010, drilling reached bedrock at 2537.36 m.

Plateau Station

Plateau Station is an inactive American research and Queen Maud Land traverse support base on the central Antarctic Plateau. The base was in continuous use until January 29, 1969. Ice core samples were made, but with mixed success.

Byrd Station

Marie Byrd Land formerly hosted the Operation Deep Freeze base Byrd Station (NBY), beginning in 1957, in the hinterland of Bakutis Coast. Byrd Station was the only major base in the interior of West Antarctica. In 1968, the first ice core to fully penetrate the Antarctic Ice Sheet was drilled here.

The Byrd 1968 core was 2164 m to bedrock and exhibited the post-glacial climatic optimum correlateably well with the Camp Century 1963 core from Greenland.

Dolleman Island

The British Antarctic Survey (BAS) has used Dolleman Island as ice core drilling site in 1976, 1986 and 1993.

Berkner Island

In the 1994/1995 field season the British Antarctic Survey, Alfred Wegener Institute and the Forschungsstelle fur Physikalische Glaziologie of the University of Munster cooperated in a project drilling ice cores on the North and South Domes of Berkner Island.

Cape Roberts Project

Between 1997 and 1999 the international Cape Roberts Project (CRP) has recovered up to 1000 m long drill cores in the Ross Sea, Antarctica to reconstruct the glaciation history of Antarctica.

International Trans-Antarctic Scientific Expedition (ITASE)

The International Trans-Antarctic Scientific Expedition (ITASE) was created in 1990 with the purpose of studying climate change through research conducted in Antarctica. A 1990 meeting held in Grenoble, France, served as a site of discussion regarding efforts to study the surface and subsurface record of Antarctica's ice cores.

Lake Vida

The lake gained widespread recognition in December 2002 when a research team, led by the University of Illinois at Chicago's

Peter Doran, announced the discovery of 2,800 year old halophile microbes (primarily filamentous cyanobacteria) preserved in ice layer core samples drilled in 1996.

Vostok

As of 2003, the longest core drilled was at Vostok station. It reached back 420,000 years and revealed 4 past glacial cycles. Drilling stopped just above Lake Vostok. The Vostok core was not drilled at a summit, hence ice from deeper down has flowed from upslope; this slightly complicates dating and interpretation. Vostok core data are available.

EPICA/Dome C

The ice thickness is 3,309 +/-22 m and the core was drilled to 3,190 m. It is the longest ice core on record, where ice has been sampled to an age of 800 kyr BP (Before Present).

Present-day annual average air temperature is-54.5 °C and snow accumulation 25 mm/y. Information about the core was first published in Nature on June 10, 2004. The core revealed 8 previous glacial cycles.

Although the major events recorded in the Vostok, EPICA, NGRIP, and GRIP during the last glacial period are present in all four cores some variation with depth (both shallower and deeper) occur between the Antarctic and Greenland cores.

Dome F

The first drilling started in August 1995, reached a depth of 2503 m in December 1996 and covers a period back to 320,000 years. The second drilling started in 2003, was carried out during four subsequent austral summers from 2003/2004 until 2006/2007, and by then a depth of 3,035.22 m was reached. This core greatly extends the climatic record of the first core, and, according to a first, preliminary dating, it reaches back until 720,000 years.

WAIS Divide

The West Antarctic Ice Sheet Divide (WAIS Divide) Ice Core Drilling Project began drilling over the 2005 and 2006 seasons, drilling ice cores up to the depth of 300 m for the purposes of gas collection, other chemical applications, and to test the site for use with the Deep Ice Sheet Coring (DISC) Drill. Sampling with the DISC Drill will begin over the 2007 season and researchers and scientists expect that these new ice cores will provide data to establish a greenhouse gas record back over 40,000 years.

Taldice

TAlos Dome Ice CorE Project is a new 1620 m deep ice core drilled at Talos Dome that provides a paleoclimate record covering at least the last 250,000 years. The TALDICE coring site (159°11'E 72°49'S; 2315 m a.s.l.; annual mean temperature-41°C) is located near the dome summit and is characterised by an annual snow accumulation rate of 80 mm water equivalent.

Non-polar Cores

The non-polar ice caps, such as found on mountain tops, were traditionally ignored as serious places to drill ice cores because it was generally believed the ice would not be more than a few thousand years old, however since the 1970s ice has been found that is older, with clear chronological dating and climate signals going as far back as the beginning of the most recent ice age. Although polar cores have the clearest and longest chronological record, four-times or more as long, ice cores from tropical regions offer data and insights not available from polar cores and have been very influential in advancing understanding of the planets climate history and mechanisms.

Mountain ice cores have been retrieved in the Andes in South America, Mount Kilimanjaro in Africa, Tibet, various locations in the Himalayas, Alaska, Russia and elsewhere. Mountain ice cores are logistically very difficult to obtain. The drilling equipment must be carried by hand, organized as a mountaineering expedition with multiple stage camps, to altitudes upwards of 20,000 feet (helicopters are not safe), and the multi-ton ice cores must then be transported back down the mountain, all requiring mountaineering skills and equipment and logistics and working at low oxygen in extreme environments in remote third world countries. Scientists may stay at high altitude on the ice caps for up 20 to 50 days setting altitude endurance records that even professional climbers do not obtain. American scientist Lonnie Thompson has been pioneering this area since the 1970s, developing light-weight drilling equipment that can be carried by porters, solar-powered electricity, and a team of mountaineering-scientists. The ice core drilled in Guliya ice cap in western China in the 1990s reaches back to 760,000 years before the present — farther back than any other core at the time, though the EPICA core in Antarctica equalled that extreme in 2003.

Because glaciers are retreating rapidly worldwide, some important glaciers are now no longer scientifically viable for taking cores, and many more glacier sites will continue to be lost, the "Snows of Mount Kilimanjaro" (Hemingway) for example could be gone by 2015.

Upper Fremont Glacier

Ice core samples were taken from Upper Fremont Glacier in 1990-1991. These ice cores were analysed for climatic changes as well as alterations of atmospheric chemicals. In 1998 an unbroken ice core sample of 164 m was taken from the glacier and subsequent analysis of the ice showed an abrupt change in the oxygen isotope ratio oxygen-18 to oxygen-16 in conjunction with the end of the Little Ice Age, a period of cooler global temperatures between the years 1550 and 1850. A linkage was established with a similar ice core study on the Quelccaya Ice Cap in Peru. This demonstrated the same changes in the oxygen isotope ratio during the same period.

Nevado Sajama

Ice cores from Sajama in Bolivia span ~25 ka and help present a high resolution temporal picture of the Late Glacial Stage and the Holocene climatic optimum.

Huascarán

Ice cores from Huascarán in Peru like those from Sajama span ~25 ka and help present a high resolution temporal picture of the Late Glacial Stage and the Holocene climatic optimum.

Quelccaya Ice Cap

Although the ice cores from Quelccaya ice cap only go back ~2 ka, others may go back ~5.2 ka. The Quelccaya ice cores correlate with those from the Upper Fremont Glacier.

Mount Kilimanjaro ice fields

Evidence for three periods of abrupt climate change in the Holocene climatic optimum have been recovered from six Kilimanjaro ice cores drilled in January and February 2000.

These cores provide a ~11.7 ka record of Holocene climate and environmental variability including three periods of abrupt climate change at ~8.3, ~5.2 and ~4 ka. These three periods correlate with similar events in the Greenland GRIP and GISP2 cores.

East Rongbuk Glacier

A shallow ice core drilled from the East Rongbuk glacier showed a dramatic increasing trend of black carbon concentrations in the ice stratigraphy since the 1990s.

CONTROLLING FACTORS

Short Term (10^4 to 10^6 Years)

Geologically short-term (<120,000 year) temperatures are believed to be driven by orbital factors amplified by changes in greenhouse gases. The arrangements of land masses on the Earth's surface are believed to influence the effectiveness of these orbital forcing effects.

Medium Term (10^6 to 10^8 Years)

Continental drift affects the thermohaline circulation, which transfers heat between the equatorial regions and the poles, as does the extent of polar ice coverage.

The timing of ice ages throughout geologic history is in part controlled by the position of the continental plates on the surface of the Earth. When landmasses are concentrated near the polar regions, there is an increased chance for snow and ice to accumulate.

Small changes in solar energy can tip the balance between summers in which the winter snow mass completely melts and summers in which the winter snow persists until the following winter.

Comparisons of plate tectonic continent reconstructions and paleoclimatic studies show that the Milankovitch cycles have the greatest effect during geologic eras when landmasses have been concentrated in polar regions, as is the case today. Today, Greenland, Antarctica, and the northern portions of Europe, Asia, and North America are situated such that a minor change in solar energy will tip the balance between year-round snow/ice preservation and complete summer melting.

The presence of snow and ice is a well-understood positive feedback mechanism for climate. The Earth today is considered to be prone to ice age glaciations.

Another proposed factor in long term temperature change is the Uplift-Weathering Hypothesis, first put forward by T. C. Chamberlin in 1899 and later independently proposed in 1988 by Maureen Raymo and colleagues, where upthrusting mountain ranges expose minerals to weathering resulting in their chemical conversion to carbonates thereby removing CO_2 from the atmosphere and cooling the earth. Others have proposed similar effects due to changes in average water table levels and consequent changes in sub-surface biological activity and PH levels.

Long Term (10^8 to 10^9 Years)

It has been proposed that long term galactic motions of the sun have a major influence on the Earth's climate. There are two principal motions, the first and most significant is the orbit of the sun around the galactic centre with a period of the order of 240 million years. Since this period is different from the rotation period of the galactic spiral arms, the sun, and the earth with it, will periodically pass through the arms (estimates of the period are uncertain and vary from 143 million years to 176 million years). The second is an oscillatory bobbing motion, similar to a floating buoy, which will periodically take the sun through the galactic disc. The period of this bobbing motion is 67 million years, so a pass through the galactic plane will occur every 33 million years.

The causal link between these galactic motions and climate is unclear but one (controversial) postulate is the effect that entering a denser region of the galaxy will have on increasing the cosmic ray flux (CRF).

This theory has been criticised, both for overstating the correlation with CRF and for failing to propose a believable mechanism that would allow CRF to drive temperature. The claims by Henrik Svensmark that CRF also strongly affects short term climate changes is even more controversial and has been challenged by many. It has also been suggested that there is some correlation between these galactic cycles and geological periods. The reason for this is postulated to be that the earth experiences many more impact events while passing through high density regions of the galaxy. Both the climate changes and sudden impacts may cause, or contribute to, extinction events.

Very Long Term (10^9 Years or More)

Jan Veizer and Nir Shaviv have proposed the interaction of cosmic rays, solar wind and the various magnetic fields to explain the long term evolution of earth's climate. According to Shaviv, the early sun had emitted a stronger solar wind with a protective effect against cosmic rays. In that early age, a moderate greenhouse effect comparable to today's would have been sufficient to explain an ice free earth and the faint young sun paradox. The solar minimum around 2.4 billion years ago is consistent with an established cosmic ray flux modulation by a variable star formation rate in the Milky Way and there is also a hint of an extinction event at this time. Within the last billion years the solar wind has significantly diminished. It is only within this more recent time that passages of the heliosphere through the spiral arms of the galaxy have been able to gain a strong and regularly modulating influence as described above.

Over the very long term the energy output of the sun has gradually increased, on the order of 5% per billion (10^9) years, and will continue to do so until it reaches the end of its current phase of stellar evolution.

PALEOTEMPESTOLOGY

Paleotempestology is the study of past tropical cyclone activity by means of geological proxies as well as historical documentary records. The term was coined by Kerry Emanuel.

Methods

Sedimentary Proxy Records

Examples of proxies include overwash deposits preserved in the sediments of coastal lakes and marshes, microfossils such as foraminifera, pollen, diatoms, dinoflagellates, phytoliths contained in coastal sediments, wave-generated or flood-generated sedimentary structures or deposits (called tempestites) in marine or lagoonal sediments, storm wave deposited coral shingle, shell, sand and shell and pure sand shore parallel ridges.

The method of using overwash deposits preserved in coastal lake and marsh sediments is adopted from earlier studies of paleotsunami deposits. Both storms and tsunamis leave very similar if not identical sedimentary deposits in coastal lakes and marshes and differentiating between the two in a sedimentary record can be difficult. The first studies to examine prehistoric records of tropical cyclones occurred in Australia and the South Pacific during the late 1970s and early 1980s. These studies examined multiple shore parallel ridges of coral shingle or sand and marine shells. As many as 50 ridges can be deposited at a site with each representing a past severe tropical cyclone over the previous 6,000 years. Tsunamis are not known to deposit multiple sedimentary ridges and therefore these features can be more easily attributed to a past storm at any given site. Coastal sedimentary analyses have been done at the U.S. Gulf coast, the Atlantic coast from South Carolina up to New Jersey and New England, and the Caribbean Sea. Also, studies on pre-historic tropical cyclones hitting Australia have been made. A study covering the South China Sea coast has also been published.

Markers in Coral

Rocks contain certain isotopes of elements, known as natural tracers, which describe the conditions under which they formed.

By studying the calcium carbonate in coral rock, past sea surface temperature and hurricane information can be revealed. Lighter oxygen isotopes (O) are left behind in coral during periods of very heavy rainfall. Since hurricanes are the main source of extreme rainfall in the tropical oceans, past hurricane events can be dated to the days of their impact on the coral by looking at the increased O concentration within the coral.

Speleothems and Tree Rings

Isotope studies in speleothems and tree rings (dendrochronology) offers a means by which higher resolution records of long-term tropical cyclone histories can be attained. Unlike the isotope records, the sedimentary records are too coarse in their resolution to register quasi-cyclic activity at decadal to centennial scales. These higher resolution records therefore offer a means for possibly differentiating between the natural variability of tropical cyclone behaviour and the effects of anthropogenically induced global climate change. Recent studies with stalagmites in Belize shows that events can be determined on a week-by-week basis.

Historical Records

Before the invention of the telegraph in the early to mid 19th century, news was as fast as the fastest horse or stagecoach or ship. Normally, there was no advance warning of a tropical cyclone impact. However, the situation changed in the 19th century as sea-faring people and land-based researchers, such as Father Viñes in Cuba, came up with systematic methods of reading the sky's appearance or the sea state, which could foretell a tropical cyclone's approach up to a couple days in advance.

Michael Chenoweth used 18th century journals to reconstruct the climate of Jamaica. Together with Dmitry Divine, he also created a 318-year (1690–2007) record of tropical cyclones in the Lesser Antilles, using newspaper accounts, ships' logbooks, meteorological journals, and other document sources.

In China, the abundance of historical documentary records in the form of *Fang Zhi* (semiofficial local gazettes) offers an extraordinary opportunity for providing a high-resolution historical dataset for the frequency of typhoon strikes.

DENDROCHRONOLOGY

Dendrochronology or tree-ring dating is the scientific method of dating based on the analysis of patterns of tree-rings. Dendrochronology can date the time at which tree rings were formed, in many types of wood, to the exact calendar year. This has three main areas of application: paleoecology, where it is used to determine certain aspects of past ecologies (most prominently climate); archaeology, where it is used to date old buildings, etc.; and radiocarbon dating, where it is used to calibrate radiocarbon ages. In some areas of the world, it is possible to date wood back a few thousand years, or even many thousands. In most areas, however, wood can only be dated back several hundred years, if at all. Currently, the maximum for fully anchored chronologies is a little over 11,000 years from present.

History

Dendrochronology was developed during the first half of the 20th century originally by the astronomer A. E. Douglass, the founder of the Laboratory of Tree-Ring Research at the University of Arizona. Douglass sought to better understand cycles of sunspot activity and reasoned that changes in solar activity would affect climate patterns on earth which would subsequently be recorded by tree-ring growth patterns (*i.e.*, sunspots ’! climate ’! tree rings).

Growth Rings

Growth rings, also referred to as *tree rings* or *annual rings*, can be seen in a horizontal cross section cut through the trunk of a tree. Growth rings are the result of new growth in the vascular cambium, a lateral meristem, and are synonymous with secondary growth. Visible rings result from the change in growth speed through the seasons of the year, thus one ring usually marks the passage of one year in the life of the tree. The rings are more visible in temperate zones, where the seasons differ more markedly. The inner portion of a growth ring is formed early in the growing season, when growth is comparatively rapid (hence the wood is less dense) and is known as "early wood" or "spring wood" or "late-spring wood". The outer portion is the "late wood" (and has sometimes been termed "summer wood", often being produced in the summer, though sometimes in the autumn) and is denser. "Early wood" is used in preference to "spring wood", as the latter term may not correspond to that time of year in climates where early wood is formed in the early summer (e.g. Canada) or in autumn, as in some Mediterranean species.

Many trees in temperate zones make one growth ring each year, with the newest adjacent to the bark. For the entire period of a tree's life, a year-by-year record or ring pattern is formed that reflects the climatic conditions in which the tree grew. Adequate moisture and a long growing season result in a wide ring. A drought year may result in a very narrow one. Alternating poor and favourable conditions, such as mid summer droughts, can result in several rings forming in a given year.

Missing rings are rare in oak and elm trees—the only recorded instance of a missing ring in oak trees occurred in the year 1816, also known as the Year Without a Summer. Trees from the same region will tend to develop the same patterns of ring widths for a given period. These patterns can be compared and matched ring for ring with trees growing in the same geographical zone and under similar climatic conditions. Following these tree-ring patterns from living trees back through time, chronologies can be built up, both for entire regions, and for sub-regions of the world. Thus wood from ancient structures can be matched to known chronologies (a technique called *cross-dating*) and the age of the wood determined precisely. Cross-dating was originally done by visual inspection, until computers were harnessed to do the statistical matching.

To eliminate individual variations in tree ring growth, dendrochronologists take the smoothed average of the tree ring widths of multiple tree samples to build up a ring history. This process is termed replication. A tree ring history whose beginning and end dates are not known is called a *floating chronology*. It can be anchored by cross-matching a section against another chronology (tree ring history) whose dates are known. Fully anchored chronologies which extend back more than 11,000 years exist for river oak trees from South Germany (from the Main and Rhine rivers) and pine from Northern Ireland. Furthermore, the mutual consistency of these two independent dendrochronological sequences has been confirmed by comparing their radiocarbon and dendrochronological ages. Another fully anchored chronology which extends back 8500 years exists for the bristlecone pine in the Southwest US (White Mountains of California). In 2004 a new calibration curve *INTCAL04* was internationally ratified for calibrated dates back to 26,000 Before Present (BP) based on an agreed worldwide data set of trees and marine sediments.

Sampling and Dating

Timber core samples measure the width of annual growth rings. By taking samples from different sites and different strata within a particular region, researchers can build a comprehensive historical sequence that becomes a part of the scientific record; for example, ancient timbers found in buildings can be dated to give an indication of when the source tree was alive and growing, setting an upper limit on the age of the wood. Some genera of trees are more suitable than others for this type of analysis. Likewise, in areas where trees grew in marginal conditions such as aridity or semi-aridity, the techniques of dendrochronology are more consistent than in humid areas. These tools have been important in archaeological dating of timbers of the cliff dwellings of Native Americans in the arid Southwest.

A benefit of dendrochronology is that it makes available specimens of once-living material accurately dated to a specific year to be used as a calibration and check of radiocarbon dating, through the estimation of a date range formed through the interception of radiocarbon (B.P., or 'B'efore 'P'resent, where present equals 1950-01-01) and calendar years. The bristlecone pine, being exceptionally long-lived and slow growing, has been used for this purpose, with still-living and dead specimens providing tree ring patterns going back thousands of years. In some regions dating sequences of more than 10,000 years are available.

The dendrochronologist faces many obstacles, however, including some species of ant which inhabit trees and extend their galleries into the wood, thus destroying ring structure.

Similar seasonal patterns also occur in ice cores and in varves (layers of sediment deposition in a lake, river, or sea bed). The deposition pattern in the core will vary for a frozen-over lake versus an ice-free lake, and with the fineness of the sediment. Some columnar cactus also exhibit similar seasonal patterns in the isotopes of carbon and oxygen in their spines (acanthochronology). These are used for dating in a manner similar to dendrochronology, and such techniques are used in combination with dendrochronology, to plug gaps and to extend the range of the seasonal data available to archaeologists and paleoclimatologists.

While archaeologists can use the technique to date the piece of wood and when it was felled, it may be difficult to definitively determine the age of a building or structure that the wood is in. The wood could have been reused from an older structure, may have been felled and left for many years before use, or could have been used to replace a damaged piece of wood.

Applications

European chronologies derived from wooden structures found it difficult to bridge the gap in the 14th century when there was a building hiatus which coincided with the Black Death. Other plagues which were less well recorded also appear in the record.

In areas where the climate is reasonably predictable, trees develop annual rings of different properties depending on weather, rain, temperature, soil pH, plant nutrition, CO2 concentration, etc. in different years. These variations may be used to infer past climate variations.

Given a sample of wood, the variation of the tree ring growths provides not only a match by year, it can also match location because the climate across a continent is not consistent. This makes it possible to determine the source of ships as well as smaller artifacts made from wood but which were transported long distances, such as panels for paintings.

Dendrochronology has become important to art historians in the dating of panel paintings, and can also provide information as to the source of the panel-many Early Netherlandish paintings have turned out to be painted on panels of "Baltic oak" shipped from the Vistula region via ports of the Hanseatic League. Since panels of seasoned wood were used, an uncertain number of years has to be allowed for seasoning. Panels were trimmed of the outer rings, and often each panel only uses a small part of the radius of the trunk, so dating studies usually result in a "terminus post quem" (earliest possible) date, and a tentative date for the actual arrival of a seasoned raw panel using assumptions as to these factors.

AGROCLIMATOLOGY

Agroclimatology, often also referred to as agricultural climatology, is a field in the interdisciplinary science of agro-meteorology, in which principles of climatology are applied to agricultural systems. Its origins relate to the foremost role that climate plays in plant and animal production. Formal references to the terms "agrometeorology" and "agroclimatology" date to the beginning of the twentieth century, but use of empirical knowledge can be traced back at least 2000 years (Monteith, 2000). Agroclimatology is sometimes used interchangeably with agrometeorology, but the former refers specifically to the interaction between long-term meteorological variables (i.e. climate) and agriculture. As such, they share common fundamental principles, methods and tools, but specific concepts are applied as described here.

Fundamental principles

Understanding the interactions between atmospheric variables and biological systems in agriculture, and applying this knowledge to increase food production and improve food quality, are the main goals of agrometeorology. Biological systems in agriculture are comprised of crops and forests, including the soil in which these grow; animals; and associated weeds, pests and diseases. Atmospheric variables that may affect these systems range from physical variables, such as solar radiation, precipitation, wind speed and direction, temperature, and humidity, to chemical variables, such as trace gas concentrations (e.g. CO_2, O_3). Agrometeorology is concerned with the characterization of these variables not only in the natural environment, but also in modified environments (e.g. irrigated areas, greenhouses, and animal shelters).

The fundamental principles used in the study of interaction between the atmosphere and agricultural systems are: (1) conservation of mass and energy, (2) radiation exchange, and (3) molecular and turbulent diffusion. The response of biological systems to these interactions draws on principles of soil physics, hydrology, plant and animal physiology, plant and animal pathology, entomology and ecology. Topics of research include water and radiation use efficiency by crops, animal comfort levels as affected by the physical environment, air pollution damage to crops, disease and pest development as a function of environmental conditions, and greenhouse gas emission by agricultural activities.

Methods and tools

Spatial scales in agroclimatology cover a wide range, from <0.1 m (e.g. response of fungi to leaf wetness) to regional and global scales (e.g. drought monitoring). Temporal scales may span past, present or future climate.

Choice of instrumentation and measurement methods for weather and biological variables occurs according to the spatial and temporal scales of interest. Most often, agroclimatologists rely on long-term climate data provided by national meteorological services.

In some countries weather stations originally were established in association with agricultural research institutes, attesting to the importance of weather and climate to agricultural production. Expertise in instrumentation, typically sensors for air temperature and humidity, solar radiation, precipitation, and wind measurement, is required from agroclimatologists in some applications, particularly those involving smaller spatial scales than provided by weather stations (i.e. microclimatological scales). In all cases data describing the condition of the biological system are also needed.

These include observations of developmental stages in crops, weeds, or insects; crop, milk or meat yield; grain or forage quality; and other physical and physiological measurements. Increasingly, remote sensing measurement techniques are being used to obtain regional and global scale estimates of variables, such as vegetation indices, or surface temperature.

Data handling, management, and processing procedures, spatial and temporal interpolation, statistical analyses, mathematical models that simulate the response of biological systems, and geographical information systems (GIS), decision support systems (DSS) are examples of tools used by agroclimatologists.

Concepts

A few concepts widely used in agroclimatology are presented here. The reader is referred to Griffith (1994) for additional background and examples of applications.

Degree-days and length of growing season

The concept of thermal time, or degree-days, was one of the earliest concepts developed, with wide application in agroclimatology. Climate records are matched to cardinal temperatures of crops, and daily mean temperatures above a minimum threshold, but below a maximum threshold, are accumulated over phenological phases.

Stages in development for specific species or varieties occur once a certain degree-day value has been reached. Degree-day requirements for crops to reach maturity are then used to select suitable climatic regions for each crop or variety.

Conversely, climatic records are used to calculate degree-days during the growing season. Provided water is available, length of growing season has its starting date defined by last mean spring frost date and ending date defined by first mean fall frost date.

Probability distributions of historical degree-day records allow for selection of adequate crops and cultivars, those that will reach maturity for an accepted risk level.

The same concept can be applied to predict probability of occurrence of damaging development phases of pests in crops during certain times of the year, guiding the application of control measures.

Evapotranspiration and water balance

Water loss from the soil–plant system, through soil evaporation and plant transpiration, is termed evapotranspiration (ET). This is an energy-consuming process fueled mostly by net radiation absorption by the soil–plant surface.

Soil water availability also determines the ET rate, with additional control provided by plant stomata. Under optimum water conditions plants do not need to control water loss, and carbon dioxide uptake also proceeds without restriction.

Hence, crop yields are optimum under non-limiting water conditions. Numerous models have been developed for prediction of ET from crops given optimal and limiting soil water conditions. The Penman–Monteith equation, and the use of crop coefficients are commonly used (Allen et al., 1998) in agroclimatology to predict seasonal crop water requirements, and to determine irrigation needs of crops.

This is accomplished through a soil water budget, where losses due to ET, runoff and deep percolation are compared to inputs such as precipitation and irrigation. Water stored, as determined by soil characteristics and rooting depth, decreases if outputs exceed inputs, and vice-versa.

Limiting growth conditions, and reductions in yields, occur if soil water available to crops falls below a critical level, signaling the need for irrigation. A water budget applied using climatic data determines crop water requirements for optimal yields, and irrigation viability, contributing to long-range planning of land and water resource use.

Agroclimatic zoning

In contrast to climatic classifications, which are usually based on temperature and precipitation, agroclimatic classifications also take soil type, crop potential productivity and moisture deficiency, among other variables, into consideration (Bishnoi, 1989). Agroclimatic zoning allows for assessment of resources for agriculture, crop planning, and improvement in crop productivity.

This concept was expanded to include environmental impact, resulting in the concept of agroecological zoning (AEZ) by the United Nations Food and Agriculture Organization (FAO). Recently, AEZ has been used to establish a global environmental resource database, including climatic, soil, terrain, and land cover, assessing the agricultural potential of 28 crops at three levels of farming technology (Fischer et al., 2001).

Animal comfort

The energy budget of homeothermic animals determines the level of comfort experienced, and if additional energy needs to be expended in keeping the body in a thermally comfortable zone. Excessive energy spent in thermal regulation results in reduced productivity, and may affect survival in extreme cases. Different livestock species and breeds have contrasting climatic requirements, and resistance to diseases typical to each environment. Consequently, climatic factors play a role in determining the level of success of the livestock enterprise. For example, a temperature–humidity index (THI), calculated from climate normals, has been used to indicate when temperate-evolved Holstein cows become less productive in various climate zones (Johnson, 1994).

Importance

Understanding of interactions between weather and agriculture leads to opportunities to use this knowledge for increased agricultural production, and also for minimizing agriculture's risks and impact on the environment. Thus, agrometeorology can play an important role in promoting sustainable development, through protection of the atmosphere and fresh water resources, desertification and drought control, sustainable agriculture, and education and training (Sivakumar et al., 2000). Knowledge obtained through research in agrometeorology is applied to guide: (1) strategic decisions in long-range planning; (2) tactical decisions in short-term planning, and (3) agrometeorological forecasts (WMO, 1981). Agroclimatology addresses needs in long-range planning, with typical examples of strategic decisions being selection of crops, livestock or forest species that match the existing climatic environment (agroclimatic characterization); and scheduling of agricultural operations (planting, pest control, etc.) that take into account long-term records and minimize risks.

Hence, agroclimatology also provides tools needed in assessing the impact of climate change on agriculture, and strategies for farmers to adapt to a changing climate.

2

Earth's Atmosphere

INTRODUCTION

Earth's atmosphere is so much more than the air we breathe. A trip from the surface of Earth to outer space would result in passing through five different layers, each with very different characteristics.

Earth's atmosphere stretches from the surface of the planet up to as far as 10,000 kilometers (6,214 miles) above. After that, the atmosphere blends into space. Not all scientists agree where the actual upper boundary of the atmosphere is, but they can agree that the bulk of the atmosphere is located close to Earth's surface—up to a distance of around eight to 15 kilometers (five to nine miles).

While oxygen is necessary for most life on Earth, the majority of Earth's atmosphere is not oxygen. Earth's atmosphere is composed of about 78 percent nitrogen, 21 percent oxygen, 0.9 percent argon, and 0.1 percent other gases. Trace amounts of carbon dioxide, methane, water vapor, and neon are some of the other gases that make up the remaining 0.1 percent.

The atmosphere is divided into five different layers, based on temperature. The layer closest to Earth's surface is the troposphere, reaching from about seven and 15 kilometers (five to 10 miles) from the surface. The troposphere is thickest at the equator, and much thinner at the North and South Poles. The majority of the mass of the entire atmosphere is contained in the troposphere—

between approximately 75 and 80 percent. Most of the water vapor in the atmosphere, along with dust and ash particles, are found in the troposphere—explaining why most of Earth's clouds are located in this layer. Temperatures in the troposphere decrease with altitude.

The stratosphere is the next layer up from Earth's surface. It reaches from the top of the troposphere, which is called the tropopause, to an altitude of approximately 50 kilometers (30 miles). Temperatures in the stratosphere increase with altitude. A high concentration of ozone, a molecule composed of three atoms of oxygen, makes up the ozone layer of the stratosphere. This ozone absorbs some of the incoming solar radiation, shielding life on Earth from potentially harmful ultraviolet (UV) light, and is responsible for the temperature increase in altitude.

The top of the stratosphere is called the stratopause. Above that is the mesosphere, which reaches as far as about 85 kilometers (53 miles) above Earth's surface. Temperatures decrease in the mesosphere with altitude. In fact, the coldest temperatures in the atmosphere are near the top of the mesosphere—about -90°C (-130°F). The atmosphere is thin here, but still thick enough so that meteors will burn up as they pass through the mesosphere—creating what we see as "shooting stars." The upper boundary of the mesosphere is called the mesopause.

The thermosphere is located above the mesopause and reaches out to around 600 kilometers (372 miles). Not much is known about the thermosphere except that temperatures increase with altitude. Solar radiation makes the upper regions of the thermosphere very hot, reaching temperatures as high as 2,000°C (3,600°F).

The uppermost layer, that blends with what is considered to be outer space, is the exosphere. The pull of Earth's gravity is so small here that molecules of gas escape into outer space.

ATMOSPHERE

An atmosphere is a layer of gases that may surround a material body of sufficient mass,and that is held in place by the gravity of the body. An atmosphere may be retained for a longer duration, if the gravity is high and the atmosphere's temperature is low. Some planets consist mainly of various gases, but only their outer layer is their atmosphere. The term stellar atmosphere describes the outer region of a star, and typically includes the portion starting from the opaque photosphere outwards.

Relatively low-temperature stars may form compound molecules in their outer atmosphere. Earth's atmosphere, which contains oxygen used by most organisms for respiration and carbon dioxide used by plants, algae and cyanobacteria for photosynthesis, also protects living organisms from genetic damage by solar ultraviolet radiation. Its current composition is the product of billions of years of biochemical modification of the paleoatmosphere by living organisms.

Pressure

Atmospheric pressure is the force of per unit area that is applied perpendicularly to a surface by the surrounding gas. It is determined by a planet's gravitational force in combination with the total mass of a column of air above a location. Units of air pressure are based on the internationally-recognized standard atmosphere which is defined as 101,325 Pa. The pressure of an atmospheric gas decreases with altitude due to the diminishing mass of gas above each location. The height at which the pressure from an atmosphere declines by a factor of e is called the scale height and is denoted by H. For an atmosphere with a uniform temperature, the scale height is proportional to the temperature and inversely proportional to the mean molecular mass of dry air times the planet's gravitational acceleration.

For such a model atmosphere, the pressure declines exponentially with increasing altitude. However, atmospheres are not uniform in temperature, so the exact determination of the atmospheric pressure at any particular altitude is more complex.

Escape

Surface gravity, the force that holds down an atmosphere, differs significantly among the planets. For example, the large gravitational force of the giant planet Jupiter is able to retain light gases such as hydrogen and helium that escape from lower gravity objects. Second, the distance from the sun determines the energy available to heat atmospheric gas to the point where its molecules' thermal motion exceed the planet's escape velocity, the speed at which gas molecules overcome a planet's gravitational grasp.

Thus, the distant and cold Titan, Triton, and Pluto are able to retain their atmospheres despite relatively low gravities. Interstellar planets, theoretically, may also retain thick atmospheres. Since a gas at any particular temperature will have molecules moving at a wide range of velocities, there will almost always be some slow leakage of gas into space.

Lighter molecules move faster than heavier ones with the same thermal kinetic energy, and so gases of low molecular weight are lost more rapidly than those of high molecular weight. It is thought that Venus and Mars may have both lost much of their water when, after being photo dissociated into hydrogen and oxygen by solar ultraviolet, the hydrogen escaped.

Earth's magnetic field helps to prevent this, as, normally, the solar wind would greatly enhance the escape of hydrogen. However, over the past 3 billion years the Earth may have lost gases through the magnetic polar regions due to auroral activity, including a net 2% of its atmospheric oxygen. Other mechanisms that can cause atmosphere depletion are solar wind-induced sputtering, impact erosion, weathering, and sequestration—sometimes referred to as "freezing out"—into the regolith and polar caps.

COMPOSITION OF THE ATMOSPHERE

Initial atmospheric makeup is generally related to the chemistry and temperature of the local solar nebula during planetary formation and the subsequent escape of interior gases. These original atmospheres underwent much evolution over time, with the varying properties of each planet resulting in very different outcomes.

The atmospheres of the planets Venus and Mars are primarily composed of carbon dioxide, with small quantities of nitrogen, argon, oxygen and traces of other gases. The atmospheric composition on Earth is largely governed by the by-products of the very life that it sustains. Earth's atmosphere contains roughly 78.09% nitrogen, 20.95% oxygen, a variable amount water vapour, 0.93% argon, 0.038% carbon dioxide, and traces of hydrogen, helium, and other "noble" gases.

The low temperatures and higher gravity of the gas giants—Jupiter, Saturn, Uranus and Neptune—allows them to more readily retain gases with low molecular masses. These planets have hydrogen-helium atmospheres, with trace amounts of more complex compounds. Two satellites of the outer planets possess non-negligible atmospheres: Titan, a moon of Saturn, and Triton, a moon of Neptune, which are mainly nitrogen. Pluto, in the nearer part of its orbit, has an atmosphere of nitrogen and methane similar to Triton's, but these gases are frozen when farther from the Sun.

Other bodies within the Solar System have extremely thin atmospheres not in equilibrium. These include the Moon (sodium gas), Mercury (sodium gas), Europa (oxygen), Io (sulfur), and Enceladus (water vapour). The atmospheric composition of an extra-solar planet was first determined using the Hubble Space Telescope. Planet HD 209458b is a gas giant with a close orbit around a star in the constellation Pegasus. The atmosphere is heated to temperatures over 1,000 K, and is steadily escaping into space. Hydrogen, oxygen, carbon and sulfur have been detected in the planet's inflated atmosphere. and also there are fuels.

STRATOSPHERIC OZONE

Why is Understanding Stratospheric Ozone Important?

Ozone is one of the most important trace species in the atmosphere. Ozone plays two critical roles: it removes most of the biologically harmful ultraviolet light before the light reaches the surface, and it plays an essential role in setting up the temperature structure and therefore the radiative heating/cooling balance in the atmosphere, especially the stratosphere.

Location of the Ozone Layer and Climatology

Ozone is mainly found in two regions of the atmosphere. Most of the ozone can be found in a layer between 10 and 60 km above the Earth's surface. This ozone region located in the stratosphere is known as the "ozone layer." Some ozone can also be found in the lower atmosphere in the region known as the troposphere. Although chemically identical to stratospheric ozone, tropospheric ozone is quite distinct and geophysically different from stratospheric ozone.

Ozone and UV—Biological Threat

Ozone is produced by the photolysis of molecular oxygen, O_2. The oxygen atom, O, produced by this photolysis recombines with O_2 to form ozone, O3. Ozone formation primarily occurs in the tropical upper stratosphere, where it is transported poleward and downward by the largescale Brewer-Dobson circulation. The 13-year average of the total ozone measurements taken by the Nimbus-7 Total Ozone Mapping Spectrometre instrument.

The formation of ozone by the photolysis of molecular oxygen removes most of the incident sunlight with wavelengths shorter than 200 nm. The wavelengths between 200 and 310 nm are removed by the photolysis of ozone itself. This photolysis of ozone in the stratosphere is the process by which most of the biologically damaging ultraviolet sunlight is filtered out. As this filtering process occurs, the stratosphere is heated. This heating is responsible for the temperature structure of the stratosphere, where the temperature increases as the altitude increases. Without this filtering, larger amounts of UV-B would reach the surface. Numerous studies have shown that excessive exposure to UV-B is harmful to plants, animals, and humans.

Ozone and Climate Change

If ozone in the stratosphere were to be removed, the stratosphere would cool. How a cooler stratosphere affects radiative balance in the rest of the atmosphere has been the subject of many detailed studies. These studies have been reanalysed and integrated into the latest Intergovernmental Panel on Climate Change report, "Climate Change 1994: Radiative Forcing of Climate Change".

The result of that report is that stratospheric ozone loss leads to a "small but non-negligible offset to the total greenhouse forcing from CO_2, N_2O, CH4, CFCs...." It is ironic that the size of the negative radiative forcing from ozone loss is nearly equal to the positive radiative forcing from chlorofluorocarbons, the source of the stratospheric ozone loss. The size of the radiative forcing due to stratospheric ozone loss has also been shown to be very sensitive to the profile shape assumed for that loss.

Observed Ozone Changes

While the global amount of ozone is fairly constant, there are significant local, seasonal, and long-term changes. The seasonal ozone changes are basically determined by the winter-summer changes in the stratospheric circulation. Since ozone has a lifetime of weeks to months in the lower stratosphere, the amount of ozone can strongly vary due to transport by stratospheric wind systems.

Since weather conditions in the stratosphere, as in the troposphere, vary from year to year, there is also interannual variability in ozone amounts. Interseasonal changes in ozone are also linked to the 11-year solar cycle in UV output and the amount of volcanic aerosols in the stratosphere. Changes in ozone have also been linked to anthropogenic pollutants, especially the release of manmade chemicals containing chlorine. We describe the more-significant recent global changes in ozone observed by a variety of instruments.

Polar Ozone Changes

The first ozone measurements in the Antarctic were made during the 1950s. A Dobson instrument was installed at Halley Bay in late 1956 in preparation for the International Geophysical Year in 1957.

One of the first discoveries made by this instrument was that the seasonal cycle of ozone in the south polar region is very different from that which had been observed in the north. Dobson which pointed out that its cause was a difference in the circulation patterns of the Antarctic relative to the Arctic.

In the Arctic, the total ozone amount grew rapidly in the late winter and early spring to about 500 Dobson Units. In contrast, the Antarctic early springtime amounts remained near 300 DU. The Dobson instrument at Halley Bay continued to make measurements each year. Farman showed that the springtime ozone amounts over Halley Bay had declined from nearly 300 DU in the early 1960s to about 180 DU in the early-to-mid 1980s.

This result has been confirmed at a number of other stations and shown using satellite data to occur over an area larger than the Antarctic continent. These large ozone changes implied that losses must be taking place in the lower stratosphere where most of the ozone exists. This was shown to be true in a series of ozonesonde measurements. More-recent sonde measurements have shown instances of near-zero concentrations of ozone over a 5-km altitude range.

The 1996-1997 Northern Hemisphere winter experienced a significant ozone depletion over the Arctic and subsequent total ozone values achieved record low values in the spring. Long term records of the Total Ozone Mapping Spectrometre and ground based observations show a downward change over the past several years occurring mostly in February and March and confined to the lower stratosphere similar to the depletion over the Antarctic. Total ozone changes for both polar regions. Chlorine radicals have been conclusively identified as the causes of ozone depletion now in both hemispheres. Measurements of ClO by the UARS MLS instrument observed elevated levels in late February over Northern Hemisphere polar regions. The winter of 1996-1997, showed extremely low temperatures in the Stratosphere.

These cold temperatures led to the formation of polar stratospheric clouds whose particles shift chlorine gas away from its HCl reservoir to active ClO through heterogeneous chemistry. This is the primary mechanism producing the Antarctic ozone hole. Although the buildup of chlorine has occurred approximately uniformly in both hemisphere the unusually low temperatures reached in high northern latitudes mostly likely precipitated the concurrent ozone losses over the Arctic.

Midlatitude Ozone Loss

Midlatitude ozone loss estimates must be extracted from long time series using statistical models. The longest time series of total ozone data is from Arosa in Switzerland. This time series, which dates back to 1926. The Arosa data show a relatively constant amount of ozone for over 4 decades and a decrease in the last decade and a half. Analysis of a more-extensive network of 30+ stations which have been in operation for about 35 years shows negative trends over the last 1.5 decades, especially in the winter and early spring. High-quality global satellite data records began in November 1978 with the launch of the Nimbus-7 Solar Backscatter Ultraviolet radiometre and TOMS instruments.

These data show midlatitude trends in the Northern Hemisphere which are largest in the winter and early spring, peaking at about 6-8% per decade at 40°-50° N in February. These satellite trends are in the process of being updated with a version 7 algorithm for the TOMS and SBUV instruments. Changes in the profile of ozone with altitude can be deduced from sonde data or from the Stratospheric Aerosol and Gas Experiment satellite measurements. Analyses of sonde data show ozone decreases between the tropopause and about 24- km altitude. Analyses of SAGE data show larger decreases than those derived from sondes. SAGE results show negative ozone trends in the lower stratosphere in the tropics.

Column ozone changes deduced from SBUV and TOMS show only small downward trends. Hollandsworth used SBUV profile and total ozone trends to deduce that ozone in the tropics below 32 hPa has increased slightly over the last decade. The resolution of the uncertainty in the magnitude of lower stratospheric and upper tropospheric ozone trends is an important measurement and analysis issue for the coming years.

THE STRATOSPHERIC OZONE DISTRIBUTION

Chemical Processes

Ozone is being continuously created and destroyed by the action of ultraviolet sunlight. The overall amount of ozone in the global stratosphere is determined by the magnitude of the production and loss processes and by the rate at which air is transported from regions of net production to those of net loss. Production of ozone requires the breaking of an O_2 bond, with the extra or "odd" oxygen atom attaching to another O_2 to form O_3 This most frequently occurs via the photo dissociation of O_2 by solar ultraviolet radiation. In the lower stratosphere and troposphere ozone can also be produced by photochemical smog-like reactions. In these reactions H or CH_3 or higher hydrocarbon radicals attach to an O_2 forming HO_2 or CH_3O_2, etc., which then react with NO. This reaction breaks the O_2 bond by forming NO_2. When NO_2 is photolyzed an O atom is formed which reacts with O_2 to form O_3.

Loss of ozone occurs when an O atom reacts with O_3 to re-form the O_2 bond. More importantly, this loss process is catalyzed by the oxides of hydrogen, nitrogen, chlorine, and bromine. These oxides are produced in the stratosphere from longlived, unreactive molecules released at the surface of the Earth. The major source molecules for HOx are methane and water vapour. The main source of NOx is nitrous oxide.

The major sources of chlorine are industrially-produced CFCs and naturallyoccurring methyl chloride. The major sources of bromine are methyl bromide and the halons. These source molecules are transported to the stratosphere where they react or are photodissociated to produce the catalytically-active oxide radicals. The catalytic efficiency of hydrogen, nitrogen, chlorine, and bromine oxides is determined by a set of interlocking reactions which convert the active oxides to catalytically-inactive temporary reservoirs, such as HNO_3, HCl, $ClONO_2$, H_2O, HOCl, HOBr, and $BrONO_2$, and vice versa. In the lower stratosphere, the balance between catalytic oxides and temporary reservoirs is strongly affected by reactions on the surfaces of stratospheric aerosols.

The balance is even more profoundly affected in the polar winter by reactions on the surface of Polar Stratospheric Cloud particles. In the early spring, the chlorine balance is shifted to almost 100% ClOx. This shift in the chemical balance results in a large calculated chemical sensitivity of ozone towards chlorine perturbations and a relatively small calculated sensitivity of ozone towards nitrogen oxide perturbations. Although the basic outline of the chemistry controlling stratospheric ozone is now known, many important aspects of the problem remain to be solved. The primary difference between the Northern and Southern Hemispheric polar ozone loss regions appears to be a result of the "denitrification" that occurs in the Antarctic winter. Denitrification means the removal of nitrogen oxides and HNO3 by large particles which fall into the troposphere. Denitrification takes place when temperatures are cold enough to form large stratospheric ice crystals. When springtime comes there are no nitrogen oxides to convert ClOx to $ClONO_2$ and slow down the rate of ozone depletion. There is some evidence for denitrification when temperatures are not cold enough to form ice crystals. Under those conditions the mechanism for denitrification is not completely understood.

Transport

Much of the currently-observed ozone interannual variability in the stratosphere is controlled by dynamical processes. In particular, this variability is driven by such processes as the quasi-biennial oscillation, El Niño-Southern Oscillation, tropospheric weather systems which extend into the stratosphere, and long-term fluctuations in planetary wave activity.

The annual cycle of total ozone is largely driven by transport effects. Relatively low values of ozone are observed in the tropics and high values are observed in the extratropics. These low tropical ozone values occur in spite of the large ozone production rates in the tropics. If ozone production were precisely balanced by ozone loss everywhere, total ozone would have extremely high values in the tropics.

The observed tropical low values result from vertical advection of low-ozone air from the tropical troposphere into the tropical stratosphere, and the subsequent transport of this air poleward and downward into the extratropics and polar regions. This advective circulation is known as the Brewer-Dobson circulation.

The redistribution of ozone from the production region at low latitudes to extratropical latitudes is modulated by a variety of processes. Foremost among these processes is the annual cycle in the circulation. It is now recognized that the Brewer-Dobson circulation is primarily controlled by large-scale waves in the winter stratosphere.

As these waves propagate through the westerly winds that dominate the winter stratosphere, they exert a westward zonal drag, which through the Coriolis force leads to a poleward and downward transport circulation, which in turn drives the temperatures away from radiative equilibrium. The large-scale waves breaking in the winter upper stratosphere also produce lifting in the tropics.

Since the lifetime of ozone increases with pressure, the poleward downward circulation causes ozone to accumulate in the lower stratosphere over the course of the winter. Since the large-scale waves are not present in the summer, the poleward and downward circulation is significantly weakened, and ozone amounts which have built up during winter begin to decrease due both to transport into the troposphere and to photochemistry. The exchange of mass between the troposphere and the stratosphere is the focus of considerable current research. Stratosphere-troposphere exchange is important for the budget of ozone in the lower stratosphere as well as the ozone budget in the troposphere. Upward transport occurs in the tropics, but the exact mechanism controlling the transport is not clear. Current research is focussing on the role of subvisible cirrus and the radiative impact of infrared heating of subvisible cirrus. Downward transport takes place in midlatitudes through jet stream folds—but the frequency and amount of mass irreversibly moving through these folds is still not understood.

Aerosols and Polar Stratospheric Clouds

It is now known that knowledge of stratospheric aerosols and PSCs is very important to our understanding of stratospheric ozone. The surfaces of aerosols and PSCs are sites for heterogeneous reactions which can convert chlorine from reservoir to radical forms. Likewise, radical nitrogen forms can be sequestered as nitric acid to shift the chemical loss process.

Aerosols

The long-term stratospheric aerosol record reveals at least three components: episodic volcanic enhancements, PSCs and clouds just above the tropical tropopause, and a background aerosol level. At normal stratospheric temperatures, aerosols are most likely super-cooled solution droplets of H_2SO4-H_2O, with an acid weight fraction of 55 to 80%. The primary source of stratospheric aerosols is volcanic eruptions that are strong enough to inject SO_2 buoyantly into the stratosphere.

Aerosol sizes range from hundredths of a micrometre to several micrometres. Although there is some variability, especially just after a volcanic eruption, a log-normal size distribution of spherical particles appears to aptly describe the aerosol. Just after an eruption, the size distribution becomes bimodal, and some particles are nonspherical because of the addition of crustal material. After an eruption, the SO_2 is converted to H_2SO4, which condenses to form stratospheric sulfuric acid aerosols, with a time scale of about 30 days. Subsequently, aerosol loading decreases due to a combination of sedimentation, subsidence, and exchange through tropopause folds. The loading decreases with an e-folding time of 9-to-12 months, although this appears quite variable with altitude and latitude.

The net effect of this post-volcanic dispersion and natural cleansing is a greatly enhanced aerosol concentration in the upper troposphere after a major eruption, especially poleward of about 30° latitude. Except immediately after an eruption, stratospheric aerosol droplets tend to be concentrated into 3 distinct latitudinal bands—one over the equatorial region and the other over each high-latitude region, 50° to 90° N and S. Following a low-latitude eruption, aerosol is dispersed into both hemispheres, whereas following a mid-to-high-latitude eruption, aerosols tend to stay primarily in the hemisphere of the eruption.

Potential sources of a background aerosol component include carbonyl sulfide from the oceans, low-level SO_2 emissions from volcanoes, and various anthropogenic sources, including industrial and aircraft emissions. Also, it is not clear whether there is an upward trend in this background aerosol, as has been hypothesized and linked to increasing aircraft emissions, since any increase may be due to incomplete removal of past volcanic aerosol. Stratospheric aerosol loading in 1979 was approximately 0.5, 10^{12}g, thought to be representative of background aerosol conditions.

The present status of the aerosol is one of enhancement due to the June 1991 eruption of Pinatubo which produced on the order of 30 ´ 10^{12}g of new aerosol in the stratosphere, about 3 times that of the 1982 eruption of El Chichón. This perturbation appears to be the largest of the century, perhaps the largest since the 1883 eruption of Krakatoa. By early 1993, stratospheric loading decreased to approximately 13 Mt, about equal to the peak loading values after El Chichón. Measurements in 1995 showed that the aerosol levels were approaching background levels.

Polar Stratospheric Clouds

The interannual variability in PSC sightings has been addressed by Poole and Pitts, who analysed more than a decade of data from the spaceborne Stratospheric Aerosol Measurement II sensor. They found noticeable variability in PSC sightings in the Antarctic from year to year, even though the southern polar vortex is typically quite stable and long-lived. This variability was found to occur late in the season and can be explained qualitatively by temperature differences. Poole and Pitts found even more year-to-year variability in SAM II Arctic PSC sighting probabilities.

This was expected since the characteristics and longevity of the northern polar vortex vary greatly from one year to the next. The year-to-year variability in Arctic sighting probabilities can also be explained qualitatively by differences in temperature, *e.g.*, zonal mean lower stratospheric temperatures in February 1988 were as much as 20 K colder than those one year earlier.

Solar Ultraviolet and Energetic Particles

Since ozone formation is fundamentally linked to the levels of ultraviolet radiation reaching the Earth, natural variations in that radiation must be understood in order to detect trends. The ultraviolet comprises only one-to-two per cent of the total solar radiation, but it displays considerably more variation than the longer wavelength visible radiation.

For example, from 1986 to 1990 the solar UV increased with onset of the 11-year solar cycle and resulted in an increase of global total ozone of almost 2%. This natural increase in ozone is comparable to the suspected anthropogenic decrease and needs to be understood in order to totally separate the anthropogenic decrease from this natural change. Studies of total ozone trends typically subtract solar cycle and other natural changes from the total ozone record in trend resolution.

Thus, more-quantitative knowledge of this natural solar-cycle-induced total ozone change would be especially valuable. Changes in energetic particle flux from the sun penetrate into the middle atmosphere and may also drive the natural ozone variations. A series of solar flares in 1989 spewed solar particles into the Earth's polar cap regions and led to polar ozone depletion. Further studies related to the very large solar particle events of October 1989 have predicted ozone depletions lasting for several months after the SPEs.

Although SPEs of this magnitude occur infrequently, they need to be understood more completely to be able to separate natural from anthropogenic ozone effects. Relativistic electron precipitations have been predicted to contribute substantially to the odd nitrogen budget of the stratosphere and, therefore, have been predicted to play a large role in controlling ozone in this region.

Another investigation has failed to find any REP-caused ozone depletion. Further work determined that REPs in May 1992, the largest measured relativistic electron flux precipitating in the atmosphere between October 1991 and July 1994, added only about 0.5 to 1% of the global annual source of odd nitrogen to the stratosphere and mesosphere. The actual importance of REPs in regulating ozone is thus not well understood nor characterized, and further work on REPs is required to thoroughly determine their importance regarding modulation of stratospheric ozone.

MODELING THE OZONE DISTRIBUTION, ASSESSMENTS

Models of the stratosphere provide the only means to attempt quantitative prediction of global change, or to evaluate the impact of natural or anthropogenic changes in composition on the stratospheric ozone and climate. In addition, models provide a means to integrate observations and theory, to provide tests of mechanisms for chemical, dynamical, or radiative changes, and to enable interpretation of observations from different platforms.

Two-Dimensional Models

Two-dimensional models are used by several research groups. The models predict the behaviour of ozone and other trace gases in reasonable agreement with measurements. Because of these favourable comparisons to measurements, these models have been utilized recently in many atmospheric studies, for example: the response of the middle atmosphere due to solar variability was studied by Brasseur, Huang and Brasseur, and Fleming; the influence of the Mt. Pinatubo eruption on the stratosphere was studied by Kinnison and Tie; and the effects of proposed stratospheric aircraft on atmospheric constituents were studied by Pitari, Weisenstein, Considine, and Considine.

These 2-D models have also been used to produce multi-year simulations of the response of stratospheric ozone to perturbations of the source gases such as CFCs from which chlorine radicals are produced. An outstanding issue regarding simulations of the stratospheric ozone response to chlorine increases is the lack of ability of 2-D models to accurately predict the ozone trend in the middle and high northern latitudes over the 1980-to-1990 time period. Since the 2-D models predict a smaller trend than observed, it is believed that the models do not adequately model all of the relevant processes and thus require further development.

Three-Dimensional Models

The three-dimensional model with full interaction between chemical, dynamical, and radiative processes remains elusive. The present generation of general circulation models generates unrealistic temperature fields which, in turn, alter the photochemistry. The unrealistic temperatures are related to problems with the model transport circulation. For example, the polar regions are persistently cold in GCMs, which suggests that there is insufficient adiabatic heating in the winter polar region.

Correspondingly, there will be insufficient ascent in the tropics, which weakens the transport from the troposphere into the stratosphere. Subtle changes in the general circulation of the atmosphere in 3-D models can alter and distort the chemical feedbacks. For example, Rasch report on a two-year simulation using version 2 of the National Centre for Atmospheric Research Middle Atmosphere Community Climate Model. A chemical scheme for 24 reactive species, or families, is run as part of this simulation. This model is partially coupled in that the water vapour predicted by MACCM2 is connected to the chemical source of water through oxidation of methane. Prescribed ozone is used in the radiative calculation. In this simulation, the calculated upper stratospheric ozone is substantially lower than is observed; much of the difference is attributed to the lower CH4, compared to observations by the UARS Halogen Occultation Experiment.

This bias leads to excessive ClO and excessive destruction of O3. In effect, the error in this longlived trace gas, which results from the weak transport circulation, leads to noticeable errors in ozone. The difficulties show why most 3-D modeling efforts have focused on "off-line" calculations, *i.e.*, use of chemistry and transport models in which the wind and temperature fields are input from a GCM or from a data assimilation system.

For either approach, there are computational advantages, as the same set of winds and temperatures is used many times. Furthermore, the effects of modifications to the chemical scheme can be isolated, and their effects understood, without the complications caused by feedback processes. A further advantage of the use of assimilated winds and temperatures is that the results of constituent simulations may be compared directly with observations with no temperature biases such as those found in GCMs.

This is particularly important for the study of processes which have a temperature threshold, such as heterogeneous reactions on

PSC surfaces. The most information is gleaned when the model is sampled in a manner consistent with the satellite sampling. The "off-line" approach has been used successfully for many years and is used to test chemical and transport mechanisms, as well as to interpret observations.

These tests include:

- Assessment of the importance of transport of air with high levels of reactive chlorine to middle latitudes;
- Assessment of the rate of ozone loss within the Northern Hemisphere vortex and identification of the variables to which the calculation is sensitive;
- Determination of the importance of upper tropospheric synoptic-scale systems on the vortex temperature, as well as their influence on the transport and mixing of air which has experienced temperatures cold enough for PSC formation; and
- Examination of the impact of ozone transport following the breakup of the Antarctic polar vortex on the global ozone budget.

These 3-D studies provide a picture of the important physical processes which control polar ozone loss. However, because of computer resource restrictions, it is not yet possible to make full 3-D model long-range predictions, including possible influence of the ozone loss on lower stratospheric temperature and climate. For example, future temperature changes may have a significant impact on the Northern Hemisphere vortex. The full 3-D model, with all relevant chemical, dynamical, and radiative processes and feedbacks among them, has yet to be developed.

THE EARTH'S GREENHOUSE EFFECT

The realisation that Earth's climate might be sensitive to the atmospheric concentrations of gases that create a greenhouse effect is more than a century old. Fleming and Weart provided an overview of the emerging science. In terms of the energy balance of the climate system, Edme Mariotte noted in 1681 that although the Sun's light and heat easily pass through glass and other transparent materials, heat from other sources does not. The ability to generate an artificial warming of the Earth's surface was demonstrated in simple greenhouse experiments such as Horace Benedict de Saussure's experiments in the 1760s using a 'heliothermometre' to provide an early analogy to the greenhouse effect.

It was a conceptual leap to recognise that the air itself could also trap thermal radiation. In 1824, Joseph Fourier, citing Saussure, argued 'the temperature [of the Earth] can be augmented by the interposition of the atmosphere, because heat in the state of light finds less resistance in penetrating the air, than in repassing into the air when converted into non-luminous heat'. In 1836, Pouillit followed up on Fourier's ideas and argued 'the atmospheric stratum…exercises a greater absorption upon the terrestrial than on the solar rays'.

There was still no understanding of exactly what substance in the atmosphere was responsible for this absorption. In 1859, John Tyndall identified through laboratory experiments the absorption of thermal radiation by complex molecules.

He noted that changes in the amount of any of the radiatively active constituents of the atmosphere such as water or CO_2 could have produced 'all the mutations of climate which the researches of geologists reveal'. In 1895, Svante Arrhenius followed with a climate prediction based on greenhouse gases, suggesting that a 40% increase or decrease in the atmospheric abundance of the trace gas CO_2 might trigger the glacial advances and retreats. One hundred years later, it would be found that CO_2 did indeed vary by this amount between glacial and interglacial periods. However, it now appears that the initial climatic change preceded the change in CO_2 but was enhanced by it.

G. S. Callendar solved a set of equations linking greenhouse gases and climate change. He found that a doubling of atmospheric CO_2 concentration resulted in an increase in the mean global temperature of 2°C, with considerably more warming at the poles, and linked increasing fossil fuel combustion with a rise in CO_2 and its greenhouse effects: 'As man is now changing the composition of the atmosphere at a rate which must be very exceptional on the geological time scale, it is natural to seek for the probable effects of such a change. From the best laboratory observations it appears that the principal result of increasing atmospheric carbon dioxide… would be a gradual increase in the mean temperature of the colder regions of the Earth.' In 1947, Ahlmann reported a 1.3°C warming in the North Atlantic sector of the Arctic since the 19th century and mistakenly believed this climate variation could be explained entirely by greenhouse gas warming.

Similar model predictions were echoed by Plass in 1956: 'If at the end of this century, measurements show that the carbon dioxide content of the atmosphere has risen appreciably and at the same time the temperature has continued to rise throughout the world, it will be firmly established that carbon dioxide is an important factor in causing climatic change'. In trying to understand the carbon cycle, and specifically how fossil fuel emissions would change atmospheric CO_2, the interdisciplinary field of carbon cycle science began. One of the first problems addressed was the atmosphere-ocean exchange of CO_2. Revelle and Suess explained why part of the emitted CO_2 was observed to accumulate in the atmosphere rather than being completely absorbed by the oceans. While CO_2 can be mixed rapidly into the upper layers of the ocean, the time to mix with the deep ocean is many centuries. By the time of the TAR, the interaction of climate change with the oceanic circulation and biogeochemistry was projected to reduce the fraction of anthropogenic CO_2 emissions taken up by the oceans in the future, leaving a greater fraction in the atmosphere.

In the 1950s, the greenhouse gases of concern remained CO_2 and H_2O, the same two identified by Tyndall a century earlier. It was not until the 1970s that other greenhouse gases–CH_4, N_2O and CFCs–were widely recognised. By the 1970s, the importance of aerosol-cloud effects in reflecting sunlight was known and atmospheric aerosols were being proposed as climate-forcing constituents. Charlson and others built a consensus that sulphate aerosols were, by themselves, cooling the Earth's surface by directly reflecting sunlight. Moreover, the increases in sulphate aerosols were anthropogenic and linked with the main source of CO_2, burning of fossil fuels. Thus, the current picture of the atmospheric constituents driving climate change contains a much more diverse mix of greenhouse agents.

Past Climate Observations, Astronomical Theory and Abrupt Climate Changes

Throughout the 19th and 20th centuries, a wide range of geomorphology and palaeontology studies has provided new insight into the Earth's past climates, covering periods of hundreds of millions of years. The Palaeozoic Era, beginning 600 Ma, displayed evidence of both warmer and colder climatic conditions than the present; the Tertiary Period was generally warmer; and the Quaternary Period showed oscillations between glacial and interglacial conditions. Louis Agassiz developed the hypothesis that Europe had experienced past glacial ages, and there has since been a growing awareness that long-term climate observations can advance the understanding of the physical mechanisms affecting climate change. The scientific study of one such mechanism–modifications in the geographical and temporal patterns of solar energy reaching the Earth's surface due to changes in the Earth's orbital parametres–has a long history. The pioneering contributions of Milankovitch to this astronomical theory of climate change are widely known, and the historical review of Imbrie and Imbrie calls attention to much earlier contributions, such as those of James Croll, originating in 1864.

The pace of palaeoclimatic research has accelerated over recent decades. Quantitative and well-dated records of climate fluctuations over the last 100 kyr have brought a more comprehensive view of how climate changes occur, as well as the means to test elements of the astronomical theory. By the 1950s, studies of deep-sea cores suggested that the ocean temperatures may have been different during glacial times. Ewing and Donn proposed that changes in ocean circulation actually could initiate an ice age. In the 1960s, the works of Emiliani and Shackleton showed the potential of isotopic measurements in deepsea sediments to help explain Quaternary changes. In the 1970s, it became possible to analyse a deep-sea core time series of more than 700 kyr, thereby using the last reversal of the Earth's magnetic field to establish a dated chronology. This deep-sea observational record clearly showed the same periodicities found in the astronomical forcing, immediately providing strong support to Milankovitch's theory.

Ice cores provide key information about past climates, including surface temperatures and atmospheric chemical composition. The bubbles sealed in the ice are the only available samples of these past atmospheres. The first deep ice cores from Vostok in Antarctica provided additional evidence of the role of astronomical forcing. They also revealed a highly correlated evolution of temperature changes and atmospheric composition, which was subsequently confirmed over the past 400 kyr and now extends to almost 1 Myr. This discovery drove research to understand the causal links between greenhouse gases and climate change.

The same data that confirmed the astronomical theory also revealed its limits: a linear response of the climate system to astronomical forcing could not explain entirely the observed fluctuations of rapid ice-age terminations preceded by longer cycles of glaciations. The importance of other sources of climate variability was heightened by the discovery of abrupt climate changes. In this context, 'abrupt' designates regional events of large amplitude, typically a few degrees celsius, which occurred within several decades–much shorter than the thousand-year time scales that characterise changes in astronomical forcing. Abrupt temperature changes were first revealed by the analysis of deep ice cores from Greenland.

Oeschger et al. recognised that the abrupt changes during the termination of the last ice age correlated with cooling in Gerzensee and suggested that regime shifts in the Atlantic Ocean circulation were causing these widespread changes. The synthesis of palaeoclimatic observations by Broecker and Denton invigourated the community over the next decade. By the end of the 1990s, it became clear that the abrupt climate changes during the last ice age, particularly in the North Atlantic regions as found in the

Greenland ice cores, were numerous, indeed abrupt and of large amplitude. They are now referred to as Dansgaard-Oeschger events. A similar variability is seen in the North Atlantic Ocean, with north-south oscillations of the polar front and associated changes in ocean temperature and salinity. With no obvious external forcing, these changes are thought to be manifestations of the internal variability of the climate system.

The importance of internal variability and processes was reinforced in the early 1990s with analysis of records with high temporal resolution. New ice cores new ocean cores from regions with high sedimentation rates, as well as lacustrine sediments and cave stalagmites produced additional evidence for unforced climate changes, and revealed a large number of abrupt changes in many regions throughout the last glacial cycle. Long sediment cores from the deep ocean were used to reconstruct the thermohaline circulation connecting deep and surface waters and to demonstrate the participation of the ocean in these abrupt climate changes during glacial periods.

By the end of the 1990s, palaeoclimate proxies for a range of climate observations had expanded greatly. The analysis of deep corals provided indicators for nutrient content and mass exchange from the surface to deep water, showing abrupt variations characterised by synchronous changes in surface and deep-water properties. Precise measurements of the CH_4 abundances in polar ice cores showed that they changed in concert with the Dansgaard-Oeschger events and thus allowed for synchronisation of the dating across ice cores. The characteristics of the antarctic temperature variations and their relation to the Dansgaard-Oeschger events in Greenland were consistent with the simple concept of a bipolar seesaw caused by changes in the thermohaline circulation of the Atlantic Ocean. This work underlined the role of the ocean in transmitting the signals of abrupt climate change. Abrupt changes are often regional, for example, severe droughts lasting for many years have changed civilizations, and have occurred during the last 10 kyr of stable warm climate. This result has altered the notion of a stable climate during warm epochs, as previously suggested by the polar ice cores. The emerging picture of an unstable oceanatmosphere system has opened the debate of whether human interference through greenhouse gases and aerosols could trigger such events. Palaeoclimate reconstructions cited in the FAR were based on various data, including pollen records, insect and animal remains, oxygen isotopes and other geological data from lake varves, loess, ocean sediments, ice cores and glacier termini.

These records provided estimates of climate variability on time scales up to millions of years. A climate proxy is a local quantitative record that is interpreted as a climate variable using a transfer function that is based on physical principles and recently observed correlations between the two records.

The combination of instrumental and proxy data began in the 1960s with the investigation of the influence of climate on the proxy data, including tree rings, corals and ice cores.

Phenological and historical data are also a valuable source of climatic reconstruction for the period before instrumental records became available. Such documentary data also need calibration against instrumental data to extend and reconstruct the instrumental record. With the development of multi-proxy reconstructions, the climate data were extended not only from local to global, but also from instrumental data to patterns of climate variability. Most of these reconstructions were at single sites and only loose efforts had been made to consolidate records.

Mann et al. made a notable advance in the use of proxy data by ensuring that the dating of different records lined up. Thus, the true spatial patterns of temperature variability and change could be derived, and estimates of NH average surface temperatures were obtained. The Working Group I WGI FAR noted that past climates could provide analogues. Fifteen years of research since that assessment has identified a range of variations and instabilities in the climate system that occurred during the last 2 Myr of glacial-interglacial cycles and in the super-warm period of 50 Ma. These past climates do not appear to be analogues of the immediate future, yet they do reveal a wide range of climate processes that need to be understood when projecting 21stcentury climate change.

Solar Variability and the Total Solar Irradiance

Measurement of the absolute value of total solar irradiance is difficult from the Earth's surface because of the need to correct for the influence of the atmosphere. Langley attempted to minimise the atmospheric effects by taking measurements from high on Mt. Whitney in California, and to estimate the correction for atmospheric effects by taking measurements at several times of day, for example, with the solar radiation having passed through different atmospheric pathlengths. Between 1902 and 1957, Charles Abbot and a number of other scientists around the globe made thousands of measurements of TSI from mountain sites. Values ranged from 1,322 to 1,465 W m-2, which encompasses the current estimate of 1,365 W m-2. Foukal et al. deduced from Abbot's daily observations that higher values of TSI were associated with more solar faculae.

In 1978, the Nimbus-7 satellite was launched with a cavity radiometre and provided evidence of variations in TSI. Additional observations were made with an active cavity radiometre on the Solar Maximum Mission, launched in 1980. Both of these missions showed that the passage of sunspots and faculae across the Sun's disk influenced TSI. At the maximum of the 11-year solar activity cycle, the TSI is larger by about 0.1% than at the minimum. The observation that TSI is highest when sunspots are at their maximum is the opposite of Langley's hypothesis. As early as 1910, Abbot believed that he had detected a downward trend in TSI that coincided with a general cooling of climate. The solar cycle variation in irradiance corresponds to an 11-year cycle in radiative forcing which varies by about 0.2 W m^{-2}.

There is increasingly reliable evidence of its influence on atmospheric temperatures and circulations, particularly in the higher atmosphere. Calculations with three-dimensional models suggest that the changes in solar radiation could cause surface temperature changes of the order of a few tenths of a degree celsius. For the time before satellite measurements became available, the solar radiation variations can be inferred from cosmogenic isotopes and from the sunspot number. Naked-eye observations of sunspots date back to ancient times, but it was only after the invention of the telescope in 1607 that it became possible to routinely monitor the number, size and position of these 'stains' on the surface of the Sun. Throughout the 17th and 18th centuries, numerous observers noted the variable concentrations and ephemeral nature of sunspots, but very few sightings were reported between 1672 and 1699.

This period of low solar activity, now known as the Maunder Minimum, occurred during the climate period now commonly referred to as the Little Ice Age. There is no exact agreement as to which dates mark the beginning and end of the Little Ice Age, but from about 1350 to about 1850 is one reasonable estimate.

During the latter part of the 18th century, Wilhelm Herschel noted the presence not only of sunspots but of bright patches, now referred to as faculae, and of granulations on the solar surface. He believed that when these indicators of activity were more numerous, solar emissions of light and heat were greater and could affect the weather on Earth.

Heinrich Schwabe published his discovery of a '10-year cycle' in sunspot numbers. Samuel Langley compared the brightness of sunspots with that of the surrounding photosphere. He concluded that they would block the emission of radiation and estimated that at sunspot cycle maximum the Sun would be about 0.1% less bright than at the minimum of the cycle, and that the Earth would be 0.1°C to 0.3°C cooler.

These satellite data have been used in combination with the historically recorded sunspot number, records of cosmogenic isotopes, and the characteristics of other Sun-like stars to estimate the solar radiation over the last 1,000 years. These data sets indicated quasi-periodic changes in solar radiation of 0.24 to 0.30% on the centennial time scale. These values have recently been re-assessed. The TAR states that the changes in solar irradiance are not the major cause of the temperature changes in the second half of the 20th century unless those changes can induce unknown large feedbacks in the climate system. The effects of galactic cosmic rays on the atmosphere and those due to shifts in the solar spectrum towards the ultraviolet range, at times of high solar activity, are largely unknown. The latter may produce changes in tropospheric circulation via changes in static stability resulting from the interaction of the increased UV radiation with stratospheric ozone. More research to investigate the effects of solar behaviour on climate is needed before the magnitude of solar effects on climate can be stated with certainty.

EARTH'S ATMOSPHERE: COMPOSITION, CLIMATE & WEATHER

Earth is the only planet in the solar system with an atmosphere that can sustain life. The blanket of gases not only contains the air that we breathe but also protects us from the blasts of heat and radiation emanating from the sun. It warms the planet by day and cools it at night.

Earth's atmosphere is about 300 miles (480 kilometers) thick, but most of it is within 10 miles (16 km) the surface. Air pressure decreases with altitude. At sea level, air pressure is about 14.7 pounds per square inch (1 kilogram per square centimeter). At 10,000 feet (3 km), the air pressure is 10 pounds per square inch (0.7 kg per square cm). There is also less oxygen to breathe.

Composition of Air

According to NASA, the gases in Earth's atmosphere include:

Nitrogen — 78 percent

Oxygen — 21 percent

Argon — 0.93 percent

Carbon dioxide — 0.04 percent

Trace amounts of neon, helium, methane, krypton and hydrogen, as well as water vapor

Atmosphere layers

Earth's atmosphere is divided into five main layers: the exosphere, the thermosphere, the mesosphere, the stratosphere and the troposphere. The atmosphere thins out in each higher layer until the gases dissipate in space. There is no distinct boundary between the atmosphere and space, but an imaginary line about 62 miles (100 kilometers) from the surface, called the Karman line, is usually where scientists say atmosphere meets outer space.

The troposphere is the layer closest to Earth's surface. It is 4 to 12 miles (7 to 20 km) thick and contains half of Earth's atmosphere. Air is warmer near the ground and gets colder higher up. Nearly all of the water vapor and dust in the atmosphere are in this layer and that is why clouds are found here.

The stratosphere is the second layer. It starts above the troposphere and ends about 31 miles (50 km) above ground. Ozone is abundant here and it heats the atmosphere while also absorbing harmful radiation from the sun. The air here is very dry, and it is about a thousand times thinner here than it is at sea level. Because of that, this is where jet aircraft and weather balloons fly.

The mesosphere starts at 31 miles (50 km) and extends to 53 miles (85 km) high. The top of the mesosphere, called the mesopause, is the coldest part of Earth's atmosphere, with temperatures averaging about minus 130 degrees F (minus 90 C). This layer is hard to study. Jets and balloons don't go high enough, and satellites and space shuttles orbit too high. Scientists do know that meteors burn up in this layer.

The thermosphere extends from about 56 miles (90 km) to between 310 and 620 miles (500 and 1,000 km). Temperatures can get up to 2,700 degrees F (1,500 C) at this altitude. The thermosphere is considered part of Earth's atmosphere, but air density is so low that most of this layer is what is normally thought of as outer space. In fact, this is where the space shuttles flew and where the International Space Station orbits Earth. This is also the layer where the auroras occur. Charged particles from space collide with atoms and molecules in the thermosphere, exciting them into higher states of energy. The atoms shed this excess energy by emitting photons of light, which we see as the colorful Aurora Borealis and Aurora Australis.

The exosphere, the highest layer, is extremely thin and is where the atmosphere merges into outer space. It is composed of very widely dispersed particles of hydrogen and helium.

Climate and weather

Earth is able to support a wide variety of living beings because of its diverse regional climates, which range from extreme cold at the poles to tropical heat at the Equator. Regional climate is often described as the average weather in a place over more than 30 years. A region's climate is often described, for example, as sunny, windy, dry, or humid. These can also describe the weather in a certain place, but while the weather can change in just a few hours, climate changes over a longer span of time.

Earth's global climate is an average of regional climates. The global climate has cooled and warmed throughout history. Today, we are seeing unusually rapid warming. The scientific consensus is that greenhouse gases, which are increasing because of human activities, are trapping heat in the atmosphere.

Earth, Venus and Mars

To better understand the formation and composition of Earth, scientists sometimes compare our planet with Venus and Mars. All three of these planets are rocky in nature and are part of the inner solar system, meaning that they are in between the sun and the asteroid belt.

Venus has an almost fully carbon dioxide atmosphere, with traces of nitrogen and sulfuric acid. The planet, however, also has a runaway greenhouse effect on its surface. Spacecraft have to be heavily reinforced to survive the crushing pressure (90 times heavier than Earth), and the oven-like temperatures (872 Fahrenheit or 467 Celsius), found at its surface. The clouds are also so thick that the surface is invisible in visible light. Because not much sun reaches the surface, this means that Venus has no significant seasonal temperature changes.

Mars also has a mostly carbon dioxide atmosphere, with traces of nitrogen, argon, oxygen, carbon monoxide and some other gases. On this planet, the atmosphere is about 100 times thinner than Earth's — a very different situation from the ancient past, when geological evidence shows that water used to flow on the surface more than 4.5 billion years ago. Scientists suggest that the Martian atmosphere may have thinned over time, either because the sun stripped away the lighter molecules in the atmosphere, or because a huge impact by an asteroid or comet catastrophically stripped the atmosphere. Mars undergoes temperature swings influenced by how much sunlight reaches the surface, which also affects its polar ice caps (another great influence on the atmosphere.)

Scientists routinely compare small, rocky exoplanets to Earth, Venus and Mars to get a better sense of their habitability. The routinely accepted definition of "habitability" is that a planet is close enough to the star for liquid water to exist on its surface. Too far, and the water turns icy; too close, and the water evaporates. However, habitability not only depends on the star-planet distance, but also the planet's atmosphere, the star's variability, and other factors.

LAYERS OF EARTH'S ATMOSPHERE

Earth's atmosphere has a series of layers, each with its own specific traits. Moving upward from ground level, these layers are named the troposphere, stratosphere, mesosphere, thermosphere and exosphere. The exosphere gradually fades away into the realm of interplanetary space.

Troposphere

The troposphere is the lowest layer of our atmosphere. Starting at ground level, it extends upward to about 10 km (6.2 miles or about 33,000 feet) above sea level. We humans live in the troposphere, and nearly all weather occurs in this lowest layer. Most clouds appear here, mainly because 99% of the water vapor in the atmosphere is found in the troposphere. Air pressure drops, and temperatures get colder, as you climb higher in the troposphere.

Stratosphere

The next layer up is called the stratosphere. The stratosphere extends from the top of the troposphere to about 50 km (31 miles) above the ground. The infamous ozone layer is found within the stratosphere. Ozone molecules in this layer absorb high-energy ultraviolet (UV) light from the Sun, converting the UV energy into heat. Unlike the troposphere, the stratosphere actually gets warmer the higher you go! That trend of rising temperatures with altitude means that air in the stratosphere lacks the turbulence and updrafts of the troposphere beneath. Commercial passenger jets fly in the lower stratosphere, partly because this less-turbulent layer provides a smoother ride. The jet stream flows near the border between the troposphere and the stratosphere.

Mesosphere

Above the stratosphere is the mesosphere. It extends upward to a height of about 85 km (53 miles) above our planet. Most meteors burn up in the mesosphere. Unlike the stratosphere, temperatures once again grow colder as you rise up through the mesosphere. The coldest temperatures in Earth's atmosphere, about -90° C (-130° F), are found near the top of this layer. The air in the mesosphere is far too thin to breathe; air pressure at the bottom of the layer is well below 1% of the pressure at sea level, and continues dropping as you go higher.

Thermosphere

The layer of very rare air above the mesosphere is called the thermosphere. High-energy X-rays and UV radiation from the Sun are absorbed in the thermosphere, raising its temperature to hundreds or at times thousands of degrees. However, the air in this layer is so thin that it would feel freezing cold to us! In many ways, the thermosphere is more like outer space than a part of the atmosphere. Many satellites actually orbit Earth within the thermosphere! Variations in the amount of energy coming from the Sun exert a powerful influence on both the height of the top of this layer and the temperature within it. Because of this, the top of the thermosphere can be found anywhere between 500 and 1,000 km (311 to 621 miles) above the ground. Temperatures in the upper thermosphere can range from about 500° C (932° F) to 2,000° C (3,632° F) or higher. The aurora, the Northern Lights and Southern Lights, occur in the thermosphere.

Exosphere

Although some experts consider the thermosphere to be the uppermost layer of our atmosphere, others consider the exosphere to be the actual "final frontier" of Earth's gaseous envelope. As you might imagine, the "air" in the exosphere is very, very, very thin, making this layer even more space-like than the thermosphere. In fact, air in the exosphere is constantly - though very gradually - "leaking" out of Earth's atmosphere into outer space. There is no clear-cut upper boundary where the exosphere finally fades away into space. Different definitions place the top of the exosphere somewhere between 100,000 km (62,000 miles) and 190,000 km (120,000 miles) above the surface of Earth. The latter value is about halfway to the Moon!

Ionosphere

The ionosphere is not a distinct layer like the others mentioned above. Instead, the ionosphere is a series of regions in parts of the mesosphere and thermosphere where high-energy radiation from the Sun has knocked electrons loose from their parent atoms and molecules. The electrically charged atoms and molecules that are formed in this way are called ions, giving the ionosphere its name and endowing this region with some special properties.

3

Solar Radiation and Terrestrial Radiation

SOLAR RADIATION

Solar radiation, electromagnetic radiation, including X-rays, ultraviolet and infrared radiation, and radio emissions, as well as visible light, emanating from the Sun. Of the 3.8 × 1033 ergs emitted by the Sun every second, about 1 part in 120 million is received by its attendant planets and their satellites.

The small part of this energy intercepted by Earth (the solar constant, on average 1.4 kilowatts per square metre) is of enormous importance to life and to the maintenance of natural processes on Earth's surface (see also sunlight).

The energy output of the Sun has its peak at a wavelength of 0.47 micrometre (0.000019 inch; a micrometre is 10-6 metre), and the Sun radiates about 8 kilowatts per square cm of its surface.

Solar radiation, often called the solar resource or just sunlight, is a general term for the electromagnetic radiation emitted by the sun.

Solar radiation can be captured and turned into useful forms of energy, such as heat and electricity, using a variety of technologies. However, the technical feasibility and economical operation of these technologies at a specific location depends on the available solar resource.

Basic Principles

Every location on Earth receives sunlight at least part of the year. The amount of solar radiation that reaches any one spot on the Earth's surface varies according to:

- Geographic location
- Time of day
- Season
- Local landscape
- Local weather.

Because the Earth is round, the sun strikes the surface at different angles, ranging from 0° (just above the horizon) to 90° (directly overhead). When the sun's rays are vertical, the Earth's surface gets all the energy possible. The more slanted the sun's rays are, the longer they travel through the atmosphere, becoming more scattered and diffuse. Because the Earth is round, the frigid polar regions never get a high sun, and because of the tilted axis of rotation, these areas receive no sun at all during part of the year.

The Earth revolves around the sun in an elliptical orbit and is closer to the sun during part of the year. When the sun is nearer the Earth, the Earth's surface receives a little more solar energy. The Earth is nearer the sun when it is summer in the southern hemisphere and winter in the northern hemisphere. However, the presence of vast oceans moderates the hotter summers and colder winters one would expect to see in the southern hemisphere as a result of this difference.

The 23.5° tilt in the Earth's axis of rotation is a more significant factor in determining the amount of sunlight striking the Earth at a particular location. Tilting results in longer days in the northern hemisphere from the spring (vernal) equinox to the fall (autumnal) equinox and longer days in the southern hemisphere during the other 6 months. Days and nights are both exactly 12 hours long on the equinoxes, which occur each year on or around March 23 and September 22.

Countries such as the United States, which lie in the middle latitudes, receive more solar energy in the summer not only because days are longer, but also because the sun is nearly overhead. The sun's rays are far more slanted during the shorter days of the winter months. Cities such as Denver, Colorado, (near 40° latitude) receive nearly three times more solar energy in June than they do in December.

The rotation of the Earth is also responsible for hourly variations in sunlight. In the early morning and late afternoon, the sun is low in the sky. Its rays travel further through the atmosphere than at noon, when the sun is at its highest point. On a clear day, the greatest amount of solar energy reaches a solar collector around solar noon.

Diffuse And Direct Solar Radiation

As sunlight passes through the atmosphere, some of it is absorbed, scattered, and reflected by:

- Air molecules
- Water vapor
- Clouds
- Dust
- Pollutants
- Forest fires
- Volcanoes.

This is called diffuse solar radiation. The solar radiation that reaches the Earth's surface without being diffused is called direct beam solar radiation. The sum of the diffuse and direct solar radiation is called global solar radiation. Atmospheric conditions can reduce direct beam radiation by 10% on clear, dry days and by 100% during thick, cloudy days.

Measurement

Scientists measure the amount of sunlight falling on specific locations at different times of the year. They then estimate the amount of sunlight falling on regions at the same latitude with similar climates. Measurements of solar energy are typically expressed as total radiation on a horizontal surface,or as total radiation on a surface tracking the sun.

Radiation data for solar electric (photovoltaic) systems are often represented as kilowatt-hours per square meter (kWh/m2). Direct estimates of solar energy may also be expressed as watts per square meter (W/m2).

Radiation data for solar water heating and space heating systems are usually represented in British thermal units per square foot (Btu/ft2).

Distribution

The solar resource across the United States is ample for photovoltaic (PV) systems because they use both direct and scattered sunlight. Other technologies may be more limited. However, the amount of power generated by any solar technology at a particular site depends on how much of the sun's energy reaches it. Thus, solar technologies function most efficiently in the southwestern United States, which receives the greatest amount of solar energy.

EFFECTS DUE TO SOLAR VARIATION

Interaction of solar particles, the solar magnetic field, and the Earth's magnetic field, cause variations in the particle and electromagnetic fields at the surface of the planet. Extreme solar events can affect electrical devices. Weakening of the Sun's magnetic field is believed to increase the number of interstellar cosmic rays which reach Earth's atmosphere, altering the types of particles reaching the surface. It has been speculated that a change in cosmic rays could cause an increase in certain types of clouds, affecting Earth's albedo.

Geomagnetic Effects

The Earth's polar aurorae are visual displays created by interactions between the solar wind, the solar magnetosphere, the Earth's magnetic field, and the Earth's atmosphere. Variations in any of these affect aurora displays. Sudden changes can cause the intense disturbances in the Earth's magnetic fields which are called geomagnetic storms.

Solar Proton Events

Energetic protons can reach Earth within 30 minutes of a major flare's peak. During such a solar proton event, Earth is showered in energetic solar particles (primarily protons) released from the flare site. Some of these particles spiral down Earth's magnetic field lines, penetrating the upper layers of our atmosphere where they produce additional ionization and may produce a significant increase in the radiation environment.

Galactic Cosmic Rays

An increase in solar activity (more sunspots) is accompanied by an increase in the "solar wind," which is an outflow of ionized particles, mostly protons and electrons, from the sun. The Earth's geomagnetic field, the solar wind, and the solar magnetic field deflect galactic cosmic rays (GCR). A decrease in solar activity increases the GCR penetration of the troposphere and stratosphere. GCR particles are the primary source of ionization in the troposphere above 1 km (below 1 km, radon is a dominant source of ionization in many areas).

Levels of GCRs have been indirectly recorded by their influence on the production of carbon-14 and beryllium-10. The Hallstatt solar cycle length of approximately 2300 years is reflected by climatic Dansgaard-Oeschger events. The 80–90 year solar Gleissberg cycles appear to vary in length depending upon the lengths of the concurrent 11 year solar cycles, and there also appear to be similar climate patterns occurring on this time scale.

Cloud Effects

Changes in ionization affect the abundance of aerosols that serve as the nuclei of condensation for cloud formation. As a result, ionization levels potentially affect levels of condensation, low clouds, relative humidity, and albedo due to clouds. Clouds formed from greater amounts of condensation nuclei are brighter, longer lived, and likely to produce less precipitation. Changes of 3–4% in cloudiness and concurrent changes in cloud top temperatures have been correlated to the 11 and 22 year solar (sunspot) cycles, with increased GCR levels during "antiparallel" cycles. Global average cloud cover change has been found to be 1.5–2%. Several studies of GCR and cloud cover variations have found positive correlation at latitudes greater than 50° and negative correlation at lower latitudes.

However, not all scientists accept this correlation as statistically significant, and some that do attribute it to other solar variability (e.g. UV or total irradiance variations) rather than directly to GCR changes. Difficulties in interpreting such correlations include the fact that many aspects of solar variability change at similar times, and some climate systems have delayed responses.

Solar Variation Theory

There have been proposals that variations in solar output explain past climate change and contribute to global warming. The most accepted influence of solar variation on the climate is through direct radiative forcing, but this is too small to explain significant temperature change. Various hypotheses have been proposed to explain the apparent solar correlation with temperatures that some assert appear to be stronger than can be explained by direct irradiation and the first order positive feedbacks to increases in solar activity. The meteorological community has responded with skepticism, in part because theories of this nature have come and gone over the course of the 20th century.

Sami Solanki, the director of the Max Planck Institute for Solar System Research in Katlenburg-Lindau, Germany said:

The sun has been at its strongest over the past 60 years and may now be affecting global temperatures... the brighter sun and higher levels of so-called "greenhouse gases" both contributed to the change in the Earth's temperature, but it was impossible to say which had the greater impact.

Nevertheless, Solanki agrees with the scientific consensus that the marked upswing in temperatures since about 1980 is attributable to human activity.

"Just how large this role [of solar variation] is, must still be investigated, since, according to our latest knowledge on the variations of the solar magnetic field, the significant increase in the Earth's temperature since 1980 is indeed to be ascribed to the greenhouse effect caused by carbon dioxide."

The theories have usually represented one of three types:

- Solar irradiance changes directly affecting the climate. This is generally considered unlikely, as the amplitudes of the variations in solar irradiance are much too small to have the observed relation absent some amplification process.
- Variations in the ultraviolet component having an effect. The UV component varies by more than the total, so if UV were for some reason having a disproportionate effect, this might explain a larger solar signal in climate.
- Effects mediated by changes in cosmic rays (which are affected by the solar wind, which is affected by the solar output) such as changes in cloud cover.

Although correlations often can be found, the mechanism behind these correlations is a matter of speculation. Many of these speculative accounts have fared badly over time, and in a paper "Solar activity and terrestrial climate: an analysis of some purported correlations." Peter Laut demonstrates problems with some of the most popular, notably those by Svensmark and by Lassen. Damon and Laut report in Eos that *the apparent strong correlations displayed on these graphs have been obtained by incorrect handling of the physical data. The graphs are still widely referred to in the literature, and their misleading character has not yet been generally recognized.*

In 1991, Knud Lassen of the Danish Meteorological Institute in Copenhagen and his colleague Eigil Friis-Christensen found a strong correlation between the length of the solar cycle and temperature changes throughout the northern hemisphere. Initially, they used sunspot and temperature measurements from 1861 to 1989, but later found that climate records dating back four centuries supported their findings. This relationship appeared to account for nearly 80 per cent of the measured temperature changes over this period.

Damon and Laut, however, show that when the graphs are corrected for filtering errors, *the sensational agreement with the recent global warming, which drew worldwide attention, has totally disappeared.*

Nevertheless, the authors and other researchers keep presenting the old misleading graph. Note that the prior link to "graph" is one such example of this.

Sallie Baliunas, an astronomer at the Harvard-Smithsonian Centre for Astrophysics, has been among the supporters of the theory that changes in the sun "can account for major climate changes on Earth for the past 300 years, including part of the recent surge of global warming."

On May 6, 2000, however, *New Scientist* magazine reported that Lassen and astrophysicist Peter Thejll had updated Lassen's 1991 research and found that while the solar cycle still accounts for about half the temperature rise since 1900, it fails to explain a rise of 0.4 °C since 1980. "The curves diverge after 1980," Thejll said, "and it's a startlingly large deviation. Something else is acting on the climate.... It has the fingerprints of the greenhouse effect."

Later that same year, Peter Stott and other researchers at the Hadley Centre in the United Kingdom published a paper in which they reported on the most comprehensive model simulations to date of the climate of the 20th century.

Their study looked at both "natural forcing agents" (solar variations and volcanic emissions) as well as "anthropogenic forcing" (greenhouse gases and sulphate aerosols). They found that "solar effects may have contributed significantly to the warming in the first half of the century although this result is dependent on the reconstruction of total solar irradiance that is used.

In the latter half of the century, we find that anthropogenic increases in greenhouses gases are largely responsible for the observed warming, balanced by some cooling due to anthropogenic sulphate aerosols, with no evidence for significant solar effects." Stott's team found that combining all of these factors enabled them to closely simulate global temperature changes throughout the 20^{th} century. They predicted that continued greenhouse gas emissions would cause additional future temperature increases "at a rate similar to that observed in recent decades". It should be noted that their solar forcing included "spectrally-resolved changes in solar irradiance" and not the indirect effects mediated through cosmic rays for which there is still no accepted mechanism — these ideas are still being fleshed out. In addition, the study notes "uncertainties in historical forcing" — in other words, past natural forcing may still be having a delayed warming effect, most likely due to the oceans.

A graphical representation of the relationship between natural and anthropogenic factors contributing to climate change appears in "Climate Change 2001: The Scientific Basis", a report by the Intergovernmental Panel on Climate Change (IPCC). Stott's 2003 work mentioned in the model section above largely revised his assessment, and found a significant solar contribution to recent warming, although still smaller (between 16 and 36%) than that of the greenhouse gases.

Physicist and historian Spencer R. Weart in *The Discovery of Global Warming* (2003) writes:

The study of [sun spot] cycles was generally popular through the first half of the century. Governments had collected a lot of weather data to play with and inevitably people found correlations between sun spot cycles and select weather patterns. If rainfall in England didn't fit the cycle, maybe storminess in New England would. Respected scientists and enthusiastic amateurs insisted they had found patterns reliable enough to make predictions. Sooner or later though every prediction failed.

An example was a highly credible forecast of a dry spell in Africa during the sunspot minimum of the early 1930s. When the period turned out to be wet, a meteorologist later recalled "the subject of sunspots and weather relationships fell into dispute, especially among British meteorologists who witnessed the discomfiture of some of their most respected superiors." Even in the 1960s he said, "For a young [climate] researcher to entertain any statement of sun-weather relationships was to brand oneself a crank.")

Orbital Variations

In their effect on climate, orbital variations are in some sense an extension of solar variability, because slight variations in the

Earth's orbit lead to changes in the distribution and abundance of sunlight reaching the Earth's surface. These orbital variations, known as Milankovitch cycles, directly affect glacial activity. Eccentricity, axial tilt, and precession comprise the three dominant cycles that make up the variations in Earth's orbit.

The combined effect of the variations in these three cycles creates changes in the seasonal reception of solar radiation on the Earth's surface. As such, Milankovitch Cycles affecting the increase or decrease of received solar radiation directly influence the Earth's climate system, and influence the advance and retreat of Earth's glaciers. Subtler variations are also present, such as the repeated advance and retreat of the Sahara desert in response to orbital precession.

Volcanism

Volcanism is the process of conveying material from the depths of the Earth to the surface, as part of the process by which the planet removes excess heat and pressure from its interior. Volcanic eruptions, geysers and hot springs are all part of the volcanic process and all release varying levels of particulates into the atmosphere.

A single eruption of the kind that occurs several times per century can affect climate, causing cooling for a period of a few years or more. The eruption of Mount Pinatubo in 1991, for example, produced the second largest terrestrial eruption of the 20th century (after the 1912 eruption of Novarupta)and affected the climate substantially, with global temperatures dropping by about 0.5 °C (0.9 °F), and ozone depletion being temporarily substantially increased.

Much larger eruptions, known as large igneous provinces, occur only a few times every hundred million years, but can reshape climate for millions of years and cause mass extinctions. Initially, it was thought that the dust ejected into the atmosphere from large volcanic eruptions was responsible for longer-term cooling by partially blocking the transmission of solar radiation to the Earth's surface. However, measurements indicate that most of the dust hurled into the atmosphere may return to the Earth's surface within as little as six months, given the right conditions.

Volcanoes are also part of the extended carbon cycle. Over very long (geological) time periods, they release carbon dioxide from the Earth's interior, counteracting the uptake by sedimentary rocks and other geological carbon dioxide sinks. According to the US Geological Survey, however, estimates are that human activities generate more than 130 times the amount of carbon dioxide emitted by volcanoes.

Ocean Variability

On a timescale often measured in decades or more, climate changes can also result from the interaction between the atmosphere and the oceans. Many climate fluctuations, including the El Niño Southern oscillation, the Pacific decadal oscillation, the North Atlantic oscillation, and the Arctic oscillation, owe their existence at least in part to the different ways that heat may be stored in the oceans and also to the way it moves between various 'reservoirs'.

On longer time scales (with a complete cycle often taking up to a thousand years to complete), ocean processes such as thermohaline circulation also play a key role in redistributing heat by carrying out a very slow and extremely deep movement of water, and the long-term redistribution of heat in the oceans.

The term thermohaline circulation (THC) refers to the part of the large-scale ocean circulation that is driven by global density gradients created by surface heat and freshwater fluxes. The adjective thermohaline derives from *thermo-* referring to temperature and *-haline* referring to salt content, factors which together determine the density of sea water. Wind-driven surface currents (such as the Gulf Stream) head polewards from the equatorial Atlantic Ocean, cooling all the while and eventually sinking at high latitudes (forming North Atlantic Deep Water). This dense water then flows into the ocean basins.

While the bulk of it upwells in the Southern Ocean, the oldest waters (with a transit time of around 1600 years) upwell in the North Pacific. Extensive mixing therefore takes place between the ocean basins, reducing differences between them and making the Earth's ocean a global system. On their journey, the water masses transport both energy (in the form of heat) and matter (solids, dissolved substances and gases) around the globe. As such, the state of the circulation has a large impact on the climate of the Earth.

The thermohaline circulation is sometimes called the ocean conveyor belt, the great ocean conveyor, or the global conveyor belt. On occasion, it is used to refer to the meridional overturning circulation (often abbreviated as MOC). The term MOC, however, is more accurate and well defined, as it is difficult to separate the part of the circulation which is actually driven by temperature and salinity alone as opposed to other factors such as the wind. Temperature and salinity gradients can also lead to a circulation which does not add to the MOC itself.

The movement of surface currents pushed by the wind is intuitive: we have all seen wind ripples on the surface of a pond. Thus the deep ocean — devoid of wind — was assumed to be perfectly static by early oceanographers. However, modern instrumentation shows that current velocities in deep water masses can be significant (although much less than surface speeds). In the deep ocean, the predominant driving force is differences in density, caused by salinity and temperature (the more saline the denser, and the colder the denser).

There is often confusion over the components of the circulation that are wind and density driven. Note that ocean currents due to tides are also significant in many places; most prominent in relatively shallow coastal areas, tidal currents can also be significant in the deep ocean. The density of ocean water is not globally homogeneous, but varies significantly and discretely. Sharply defined boundaries exist between water masses which form at the surface, and subsequently maintain their own identity within the ocean.

They position themselves one above or below each other according to their density, which depends on both temperature and salinity. Warm seawater expands and is thus less dense than cooler seawater. Saltier water is denser than fresher water because the dissolved salts fill interstices between water molecules, resulting in more mass per unit volume. Lighter water masses float over denser ones (just as a piece of wood or ice will float on water).

This is known as "stable stratification". When dense water masses are first formed, they are not stably stratified. In order to take up their most stable positions, water masses of different densities must flow, providing a driving force for deep currents. The thermohaline circulation is mainly triggered by the formation of deep water masses in the North Atlantic and the Southern Ocean and Haline forcing caused by differences in temperature and salinity of the water.

Formation of Deep Water Masses

The dense water masses that sink into the deep basins are formed in quite specific areas of the North Atlantic and the Southern Ocean. In these Polar Regions, seawater at the surface of the ocean is intensively cooled by the wind. Wind moving over the water also produces a great deal of evaporation, leading to a decrease in temperature, called evaporative cooling. Evaporation removes only molecules of pure water, resulting in an increase in the salinity of the seawater left behind, and thus an increase in the density of the water mass.

In the Norwegian Sea evaporative cooling is predominant, and the sinking water mass, the North Atlantic Deep Water (NADW), fills the basin and spills southwards through crevasses in the submarine sills that connect Greenland, Iceland and Great Britain. It then flows very slowly into the deep abyssal plains of the Atlantic, always in a southerly direction. Flow from the Arctic Ocean Basin into the Pacific, however, is blocked by the narrow shallows of the Bering Strait.

The formation of sea ice also contributes to an increase in seawater salinity; saltier brine is left behind as the sea ice forms around it (pure water preferentially being frozen). Increasing salinity depresses the freezing temperature of seawater, so cold liquid brine is formed in inclusions within a honeycomb of ice.

The brine progressively melts the ice just beneath it, eventually dripping out of the ice matrix and sinking.

This process is known as brine exclusion. By contrast in the Weddell Sea off the coast of Antarctica near the edge of the ice pack, the effect of wind cooling is intensified by brine exclusion. The resulting Antarctic Bottom Water (AABW) sinks and flows north into the Atlantic Basin, but is so dense it actually underflows the NADW. Again, flow into the Pacific is blocked, this time by the Drake Passage between the Antarctic Peninsula and the southernmost tip of South America.

The dense water masses formed by these processes flow downhill at the bottom of the ocean, like a stream within the surrounding less dense fluid, and fill up the basins of the polar seas. Just as river valleys direct streams and rivers on the continents, the bottom topography steers the deep and bottom water masses. Note that, unlike fresh water, saline water does not have a density maximum at 4°C but gets denser as it cools all the way to its freezing point of approximately “1.8°C.

Movement of Thermohaline Circulation

Formation and movement of the deep water masses at the North Atlantic Ocean, creates sinking water masses that fill the basin and flows very slowly into the deep abyssal plains of the Atlantic. This high latitude cooling and the low latitude heating drives the movement of the deep water in a polar southward flow.

The deep water flows through the Antarctic Ocean Basin around South Africa where it is split into two routes: one into the Indian Ocean and one past Australia into the Pacific.

At the Indian Ocean, some of the cold and salty water from Atlantic — drawn by the flow of warmer and fresher upper ocean water from the tropical Pacific — causes a vertical exchange of dense, sinking water with lighter water above. It is known as overturning. In the Pacific Ocean, the rest of the cold and salty water from the Atlantic undergoes Haline forcing and slowly becomes warmer and fresher.

The out-flowing undersea of cold and salty water makes the sea level of the Atlantic slightly lower than the Pacific and salinity or halinity of water at the Atlantic higher than the Pacific. This generates a large but slow flow of warmer and fresher upper ocean water from the tropical Pacific to the Indian Ocean through the Indonesian Archipelago to replace the cold and salty Antarctic Bottom Water. This is also known as Haline forcing (net high latitude freshwater gain and low latitude evaporation).

This warmer, fresher water from the Pacific flows up through the South Atlantic to Greenland, where it cools off and undergoes evaporative cooling and sinks to the ocean floor, providing a continuous thermohaline circulation. Hence, a recent and popular name for the thermohaline circulation, emphasizing the vertical nature and pole-to-pole character of this kind of ocean circulation, is the meridional overturning circulation.

Quantitative Estimation

The deep water masses that participate in the MOC have chemical, temperature and isotopic ratio signatures and can be traced, their flow rate calculated, and their age determined. These include ^{231}Pa / ^{230}Th ratios.

Gulf Stream

The North Atlantic Current, warm ocean current that continues the Gulf Stream northeast, is largely driven by the global thermohaline circulation to further east and north from the North American coast, across the Atlantic and into the Arctic Ocean.

Upwelling

All these dense water masses sinking into the ocean basins displace the water below them, so that elsewhere water must be rising in order to maintain a balance. However, because this thermohaline upwelling is so widespread and diffuse, its speeds are very slow even compared to the movement of the bottom water masses.

It is therefore difficult to measure where upwelling occurs using current speeds, given all the other wind-driven processes going on in the surface ocean.

Deep waters do however have their own chemical signature, formed from the breakdown of particulate matter falling into them over the course of their long journey at depth; and a number of authors have tried to use these tracers to infer where the upwelling occurs. Wallace Broecker, using box models, has asserted that the bulk of deep upwelling occurs in the North Pacific, using as evidence the high values of silicon found in these waters. However, other investigators have not found such clear evidence.

Computer models of ocean circulation increasingly place most of the deep upwelling in the Southern Ocean, associated with the strong winds in the open latitudes between South America and Antarctica. While this picture is consistent with the global observational synthesis of William Schmitz at Woods Hole and with low observed values of diffusion, not all observational syntheses agree. Recent papers by Lynne Talley at the Scripps Institution of Oceanography and Bernadette Sloyan and Stephen Rintoul in Australia suggest that a significant amount of dense deep water must be transformed to light water somewhere north of the Southern Ocean.

Effects on Global Climate

The thermohaline circulation plays an important role in supplying heat to the polar regions, and thus in regulating the amount of sea ice in these regions. Changes in the thermohaline circulation are thought to have significant impacts on the earth's radiation budget. Insofar as the thermohaline circulation governs the rate at which deep waters are exposed to the surface, it may also play an important role in determining the concentration of carbon dioxide in the atmosphere.

While it is often stated that the thermohaline circulation is the primary reason that Western Europe is so temperate, it has been suggested that this is largely incorrect, and that Europe is warm mostly because it lies downwind of an ocean basin, and because of the effect of atmospheric waves bringing warm air north from the subtropics. However, the underlying assumptions of this particular analysis have been challenged.

Large influxes of low density meltwater from Lake Agassiz and deglaciation in North America are thought to have led to a disruption of deep water formation and subsidence in the extreme North Atlantic and caused the climate period in Europe known as the Younger Dryas. For a discussion of the possibilities of changes to the thermohaline circulation under global warming.

Human Influences

Anthropogenic factors are human activities that change the environment. In some cases the chain of causality of human influence on the climate is direct and unambiguous (for example, the effects of irrigation on local humidity), whilst in other instances it is less clear. Various hypotheses for human-induced climate change have been argued for many years though, generally, the scientific debate has moved on from scepticism to a scientific consensus on climate change that human activity is the probable cause for the rapid changes in world climate in the past several decades.

Consequently, the debate has largely shifted onto ways to reduce further human impact and to find ways to adapt to change that has already occurred.

Of most concern in these anthropogenic factors is the increase in CO_2 levels due to emissions from fossil fuel combustion, followed by aerosols (particulate matter in the atmosphere) and cement manufacture. Other factors, including land use, ozone depletion, animal agriculture and deforestation, are also of concern in the roles they play - both separately and in conjunction with other factors - in affecting climate.

DISPOSAL OF RADIATION OF SURFACE OF THE EARTH

There are just three things that can happen to radiation incident onto any extended object. It must be reflected, transmitted or absorbed. When the object is extremely small it more or less scatters incident radiation, and radiation that just grazes the boundary of an object suffers still another effect which we call diffraction. However, neither scattering nor diffraction occurs when the object is large and its edges are not involved. They do not, therefore, occur in the case of radiation incident on the surface of the earth.

Here then, the incident radiation is all used up by two processes, reflection and absorption, since there is no transmission—no passage of radiation through the earth and out at the other side. The portion reflected is about 70% for snow-covered regions, and 7% for the rest of the world. The remainder is absorbed, that is, 30% wherever there is snow, and 93% at all other places, both land and water.

That which is reflected is lost except in so far as it is absorbed by the air above. The absorbed portion goes largely to heating the upper layers of the soil or water, but not all of it, since a considerable part is consumed in maintaining evaporation, and a much smaller part in effecting plant growth and development. Another relatively small part merely melts snow and ice without raising their temperature above the freezing point.

The heated surface in turn heats the soil or rock by conduction, but appreciably to a depth of only a few feet. The heating of water extends to a greater depth owing partly to the penetration of the rays to some distance below the surface, and partly to the mixing of the water by wave action. The heated surface also warms the air above it both by direct contact and by radiation. Furthermore, the heated air through convection shares its warmth with other and colder air above; and the heat consumed in evaporation at one place is liberated, that is made sensible or temperature-producing, some other place, usually in mid-air, where condensation occurs, and far away. Practically every bit of this heating of earth, ocean and air, and supply of energy for evaporation, plant growth, ice melting, and what not else, comes from the sun—all directly except about one part in half a million that reaches us after reflection by the full moon and the planets.

A negligibly small amount comes from the fixed stars, enough to keep the average temperature of the out-doors air about two millionths of a degree Fahrenheit higher than it otherwise would be. Finally, another very small amount comes from the heated interior of the earth.

Quantity and effects of heat from the interior of the earth

If the earth had no atmosphere, and if there were no sun or stars to send us a flood of radiation, the supply of heat from the interior (of which four-fifths, roughly, is from radioactive material) alone would keep up the surface temperatures to about 60° absolute, on the Fahrenheit scale, that is, -400° F., approximately.

Hence the flow of heat from the interior of the earth per square foot of surface is sufficient to raise the temperature of a gallon of water about 1° F. in 16 days, and the total flow through the whole surface out to space sufficient to heat 92,000 tons of water from the freezing to the boiling point every second of time; or enough, starting at room temperatures, to melt 1,000,000 tons of lead per second. These are big figures, and yet all this flow of heat keeps the actual temperature of the surface of the earth only about 1/25 of a degree F. higher than it otherwise would be.

The figures also tell us the surprising story that if 10,000 times as much heat came from the interior of the earth as now actually does come, or, what amounts to the same thing, if everywhere there was a sea of molten cast iron covered over with a layer of rock and dirt only 10 to 12 feet thick, the oceans above could rest thereon serene with no close approach to the boiling point, so excellent an insulator, or poor a conductor, is this material; and that if the dirt and rock crust were 20 feet thick we could go about over it ourselves in perfect comfort!

Outgoing Radiation

On the average, the earth loses to space, or emits to space, by radiation very approximately the same amount of heat each year that it absorbs of incoming radiation during the same time, plus, of course, the supply of heat that reaches the surface from the interior. We know that the loss is substantially equal to the gain because otherwise the surface would be growing warmer from year to year, and we know that this loss is by radiation as there is no other way for the loss to occur—there being no such thing as conduction to empty space.

The amount of this loss, or radiation of the entire earth to space, can be estimated from our knowledge of the incoming radiation and the fraction of it that is ineffective through scattering and reflection, especially by clouds. It can not be measured directly because we have no means of getting out beyond the atmosphere and from that ideal place pointing our heat-gathering apparatus towards the earth.

But, as implied above, we can make a pretty close estimate of the average rate at which radiation is going out from the earth as a whole, and the conclusion is that it is very nearly the same as that from a perfect radiator, or "black body," at the absolute temperature 454° on the Fahrenheit scale, or -6° F. This is sufficient, per square foot of surface, to heat a gallon of water from the freezing to the boiling point in about 20 1/2 hours.

The radiation from the surface of the earth often is very much greater than this value, even twice as great, or more, because the temperature of the surface frequently is far higher than the -6° F., here assumed. On the other hand, at times and places, owing to very low temperatures, it is much less. On the average, however, the radiation from the surface is much in excess of that which finally gets away to space-greater by the amount of return radiation it absorbs (nearly all of it) from clouds and the atmosphere. The surface of the earth radiates at a relatively high temperature, hence in comparative abundance. Some of this radiation goes directly through the atmosphere, but ordinarily most of it is absorbed by the water vapour and clouds in the lower air, and a little by other things, especially ozone (when the sky is clear, for it is above all clouds) and carbon dioxide. That which is absorbed in the lowest layers is, in general, reradiated, but at a lower temperature than that of the surface.

This reradiation is in every direction, half of it downward, some of which is absorbed on the way, and the rest by the surface whose initial temperature and radiation it thus helps to maintain; and half upwards to the next higher layers; and so on up and up from layer to layer, but always with decreasing absorption by the air still above and increasing absorption by that below until the entire atmosphere is left behind.

Clouds Do not Check Radiation

It is a well-known fact that during still clear nights the surface of the earth, and, through it, the adjacent air, cool to a much lower temperature, especially over level land and in valleys and bowl-like depressions, than they do when either the sky is clouded, or the wind is strong.

Furthermore, the lower the clouds, other things being equal, the less the cooling. This interesting and important fact is clearly "explained" in many elementary books and numerous articles on the assumption that clouds and winds check the radiation of the surface of the earth, that is, make it radiate more slowly.

That seems very simple, and would be but for one little fly in the ointment—there isn't a word of truth in it. It might do perhaps as a dose of mental paregoric for a kid with the quizz colic, but it is no good for anything else. Just one thing alone ever reduces radiation, and that is decrease of temperature. No! Clouds do not check in the least radiation from the surface below.

They are, however, themselves good radiators; and as their temperature, when they are low, is nearly that of the earth, they send down to it almost as much radiation as it itself emits, and as practically all this cloud radiation is absorbed by the earth, it follows that the surface temperature remains substantially constant. It isn't that the radiation from the earth is checked in the least, but that it receives from the cloud canopy and absorbs wellnigh as much as it itself gives out. Neither is the approximately constant temperature maintained by an appreciable wind owing to any check whatever exerted by it on surface radiation, but to the fact that the net loss of heat thus sustained, and on clear nights it is considerable, is distributed by turbulence through such a deep layer and great quantity of air that the fall in temperature is small even when the total loss of heat is large.

Temperature of Surface Air

When we talk about the "surface" air it often is advisable to explain just what air we have in mind, for this term is quite flexible. We might mean only that air which is in actual molecular contact with the surface, or that which at most is within a few inches of it, or finally, all below the height of eight or ten feet, the air to which we chiefly are exposed while outdoors.

The temperature of the surface air, in any one of these senses, is determined mainly by that of the surface itself. Whatever the temperature of the surface, that also is the temperature of the contact air, and very nearly the temperature of all that air which by turbulence or otherwise is frequently brought into contact with the surface.

On the stillest of nights this layer at places may be only a few thick. During the daytime, however, especially when there is sunshine to induce thermal convection, and whenever, day or night, there is a measurable movement of the air, it is certain to be at least a good many feet thick. The essential point in this: The temperature of the air near the surface (how near varies with the circumstances) depends more on contacts with that surface than it does on the amount of solar radiation to which it may be exposed.

The surface temperature of course does vary with the intensity and duration of the sunshine, and so therefore does also that of the surface air, but indirectly through contact with the surface and not directly by absorption of solar energy.

Relation between Surface Temperature and Temperature of Surface Air

The temperature of the actual contact air must be the same as that of the surface (of the substance, ground or what not, at its surface) against which it rests. If this surface air remained fixed, as we often are told that it does, then it would seem that the air next in contact with it also should become fixed in position, and so on indefinitely.

But we know that fixity of position of the air molecules does not extend to a measurable distance from any solid, for we can blow smoke past it and see the motion of the air. We therefore are forced to the conclusion that fixity of position does not apply, at least not for any appreciable length of time, even to the contact molecules.

The actual contact molecules of the air are at rest, like a liquid film, but they do not stay at rest. They evaporate, and as they leave the surface others condense thereon—are adsorbed—a continuous process the details of which are not yet all known. In this way the contact molecules, during the extremely brief interval of their contact, are fixed in position, but they are continuously reverting to the gaseous state, and therefore the atmosphere at ordinary temperatures is always fluid however measurably near it may be to the surface in question.

Since the air is directly heated chiefly by contact with the surface of the earth, and indirectly by the sharing of this heat, through convection, with colder air above, it follows that in general wherever the temperature of the lower atmosphere is increasing, that is, over nearly all snow-free land, and particularly during the day time, there the average temperature of the surface is higher than that of the free surface air. This is in accordance with what physicists call the second law of thermodynamics, and what everybody else knows without calling it anything, namely, that the temperature of the heater is higher than the temperature of the thing heated. Similarly, where the lower air commonly is cooled, as it is over snow-covered regions, there the average temperature of the surface is lower than that of the surface air—the cooler is colder than the thing cooled.

Maximum and Minimum Temperatures

Obviously if the heater changes temperature, so also will the heated, and the heater will be the first to change and the first to reach its extreme values—maxima and minima. There is no surprise, therefore, in the fact that the daily maximum temperature of a snow-free land surface occurs earlier, about 1 o'clock P.M., than that of the air above it, which is delayed until around 3 o'clock.

The air and surface minima, occurring near daybreak, are much closer together, owing partly to the slow cooling of the soil through the night. Over the ocean the temperature of the air normally is a little higher, a degree or so, than that of the water, and the time of its maximum value, near 1 o'clock P.M., a little earlier than that of the water. This is due to the fact that here the surface air is humid and also "dusty" with salt particles and therefore absorbs a large amount of radiation, so much indeed that its daily range of temperature is more dependent on this direct absorption than it is on conduction and convection from the surface. The minimum temperatures of air and water occur simultaneously, or nearly so.

Periodic Temperature Changes

Nearly 150 different periods of temperature and other weather changes ranging from 24 hours to 744 years. Nearly all of them, however, have a shorter period than 40 years, and half of them a period of 8 years or less. Of this great number of periods there are only two, the 24-hour or daily period, and the 12-month or annual period, that everybody accepts. There is one other, the so-called 11-year or sunspot period, that is widely, though not universally accepted; and still another the 35-year, or Brückner, period that many believe to be real.

No credence was ever given to any of the others save perhaps by their discoverers, and in most cases even that must have been half-hearted. The daily period is everywhere conspicuous (save for part of the time in polar regions, when the sun is continuously above or continuously below the horizon), and in respect to temperature, gives, on the average, a maximum in the early to mid afternoon and a minimum shortly before sunrise.

Over the oceans this diurnal range is only 1° F. to 3° F. as a rule. It also is small in the humid and cloudy portions of the continental tropics, owing to the large amount of return radiation from the clouds and water vapour. In desert regions, especially at high altitudes, where the sky is clear and the humidity very low the diurnal range of temperature is at its maximum. In extreme cases this range is of the order of 100° F., from distinctly below freezing to decidedly over 100° F.—both in the shade. The annual range also is extremely conspicuous in most parts of the world.

In this case the exception does not occur at and near the poles, but at and for some distance on either side of the equator. At the equator the "year," as it were, counted from the time the sun is overhead at noon until farthest away (23 1/2°), and then back again is 6 months, not 12. Next beyond the equator on either side there obviously are two such "years" but of unequal length. At one distance they are 5 and 7 months, at another 4 and 8, and so on until at the Tropics, Capricorn and Cancer, only one is left, and that one 12 months in duration, the same as from there on to the pole.

The times of occurrence of the annual maxima and minima vary widely from place to place, but always they are after maximum and minimum reception of heat from the sun. The delays are least over inland deserts and greatest over mid to high latitude portions of the oceans. The sunspot period, approximately 11.1 years, is most pronounced at high levels within the tropics. Here the average temperature during the year or two around sunspot minima, or when the spots are fewest and smallest, is about 2° F. higher than the average temperature during the time of spot maxima. The same relation appears to hold, in general, for the middle and higher latitudes but with decidedly less contrast.

The Brückner period is very irregular in length, varying from roughly 20 years to perhaps 50, and the amplitude of its temperature range uncertain but always small. Its irregularity in length deprives it of practically all forecasting value, and indeed makes its very existence as anything other than a fortuitous recurrence highly doubtful. There are two other known and real periods in respect to average temperatures and other climatic elements, but they are far too long to consider in any business affairs.

One concerns the slow change of the season of the year when the earth is nearest the sun, due to the combined effect of the motion of the perihelion and the precession of the equinoxes. Just at present the earth is nearest the sun the first week of January and farthest away the first week of July; and this difference in distance is sufficient, if long continued, to vary the average temperature of the earth by at least 7° or 8° F. That is, at present the winters of the northern hemisphere are shorter and milder, and the summers longer and less hot, than they would be if we were nearest the sun the first week of July and farthest from it the first week of January, as we were about 10,500 years ago, and, in the same length of time, will be again.

This is one reason, and the unequal distribution of land and water another, why the average temperature for the year is about 2° F. higher in the northern hemisphere than in the southern, and why the thermal equator is north of the geographic equator.

Beyond question this particular period is of great climatic importance, but we know all about its course and its cause and the changes it effects come about so slowly that practically they do not concern us at all. The other period referred to is that of the changes in the ellipticity of the earth's orbit or variations in the difference between the annual maximum and minimum distances of the earth from the sun. But the length of this period, roughly 100,000 years, keeps it out of every business equation however prudently constructed.

Just to make the list complete it may be worthwhile to mention a few utterly unimportant but entirely real temperature periods. Those are the periods of the changes in light and heat received from the moon-maximum at full moon, minimum at new moon; changes in the distance of the earth from the sun due to the pull of the moon in its orbit about the earth; and similar but far less changes of and by the planets. The sum total of the effects of all the planets is about equal to that of the moon alone, that is, a change in the average temperature of the earth of about.02° F., due almost wholly to variations in our distance from the sun, or, as we say, to perturbations in the earth's orbit. But, as already stated, this change in temperature is too small to bother about.

Temperature Lag

It was stated that the hottest time of the day is not noon, when the sun is most effective, but two to four hours later; and similarly, that the coldest weather does not come with the shortest days, but generally a month or so later. In proverb form: "As the days grow longer the cold grows stronger." In the early morning of a clear day following a cloudless night, say, the earth and surface air are relatively cool.

Then with sunrise they begin to warm up, but not rapidly, even when there is no wind, because it requires an appreciable amount of heat to warm even a pound of soil 1° F., and several times as much to equally warm a pound of water. But as the sunshine continues, the soil at first gets hotter and hotter, and as its temperature rises the rate at which it loses heat by radiation rapidly increases.

However, since the soil, including of course its covering, warms slowly, owing to its large capacity for heat, its loss by radiation falls more and more behind its gain by absorption as the sun rises higher in the heavens, and therefore catches up with the latter only in the afternoon when the insolation is distinctly less than it was at midday.

Hence the diurnal maximum temperature, whether of the lower air or of the surface of the earth (an earlier phenomenon) necessarily lags behind the maximum intensity of the sunshine. Similarly, the annual maximum temperature occurs several weeks after the days are longest and the heating strongest. Very similarly too, because the earth can give off stored up heat when the supply becomes deficient, the minimum temperature comes several weeks after the shortest days. During this period, as the days grow longer the cold grows stronger.

The diurnal and annual heating and cooling, and lagging of temperature extremes, may be likened to the alternate rise and fall of the water level in a reservoir having a continuously open drain pipe at the bottom and a periodically variable inflow, now greater, now less, than the then rate of outflow, but so regulated that the reservoir may never become empty.

Day Degrees

Not only are we interested in the values and times of occurrence of maximum and minimum temperatures but also, and even more, concerned in the occurrence of certain critical temperatures. For instance, we are very much interested in the temperature at which frost can occur until it does occur, after which, if it has been a "killing" one, we are no longer much concerned as there is nothing left for the next one to injure. Another critical temperature is 42° F. as that closely marks the boundary between growth and dormancy for most vegetation of the temperate zones.

In fact it is customary to call the difference between the average temperature of a given day, if higher than this value, and 42° F., its day degrees. The sum of these daily values over a week, month or season, is the number of day degrees for that period, and is an important index to what might have been the vegetable growth during the time in question. Similarly, engineers and others interested in artificial heating of buildings, count day degrees relative to a temperature of 65° F.

Occasional Extremes

Once in a while an exceptional combination of conditions brings to a given place an abnormally high or low temperature, usually for only an hour or two, or a day at most, but sometimes for several days together, and even a month or longer. It is always easy to know from the current maps of weather distribution exactly what caused the extreme in question, but it never is possible to trace them farther back than two or three steps at most, nor very long to foresee their coming. Some of them one never forgets, and a few continue for a century or more to put disconcerting humps or depressions on our statistical curves.

Wind Direction and Temperature

The effect of wind direction on the temperature of a place depends on its location. Well within the Tropics, and also near the poles, the effect of wind direction obviously is small because the temperature is pretty nearly the same round about in every direction. In middle latitudes, however, the situation is quite different, partly because here the temperatures commonly are not the same in every direction, and partly also, in fact mainly, because here each section of the cyclone and of the anticyclone has its own wind direction, and some of them a wind system entirely distinct from that of the others.

In the forward or eastern portion of the anticyclone the winds are from the region of higher latitudes, and having come a long ways often are distinctly cool to cold for the place and time of year. Similarly, the winds of the western segment, having come from much nearer the equator, usually are relatively warm. In the cyclone, or widespread disturbance, all that segment of 90°, more or less, lying between one line running east, to southeast, from the storm centre and another generally south to southwest (in the northern hemisphere; east to northeast, and north to northwest, in the southern hemisphere) is occupied by a great current of warm air from low latitudes.

The rest of the storm area is covered with cold winds from the east, north, and northwest, in succession as one in the northern hemisphere passes from the front to the rear of the storm centre on the poleward side; from the east, south, and southwest, in the southern hemisphere. In general, all these cold winds in each hemisphere are of polar, that is, high latitude origin.

Clearly then, the temperature of the air in a cyclonic region is likely to change with the direction of the wind. In one portion of this disturbance, namely, along a narrow strip that meteorologists call the cold front, or wind shift line, and which commonly runs west of south (west of north in the southern hemisphere) from the storm centre, the wind direction rapidly changes from southwesterly to northwesterly, with, as a rule, a sharp drop in temperature as the tropical breezes give way to polar blasts. Hence in middle latitudes air temperature is closely dependent upon wind direction, both in cyclones and anticyclones; and that means the greater portion of the time, for usually we are in the midst of one or the other of these disturbances.

OVERFLOW STRATA

Any layer of air that is thoroughly mixed up has a certain rate of decrease of temperature with increase of height, and such that the temperature of an isolated mass of like air rising or falling through it will change at the same rate. In such a layer vertical convection is as easy as horizontal gliding over a smooth surface.

Left to itself, though, and if unclouded, its vertical temperature gradient or lapse rate (lapse, for short) gradually changes, owing largely to gain and loss of heat by radiation, until it becomes decidedly less, and the layer thereby impenetrable to dry or unsaturated air. Saturated air, however, may be buoyed up to considerable heights, as already explained, because the heat of condensation, or heat rendered sensible as a result of condensation, so reduces the lapse rate that the ascending air is warmer and therefore lighter than the air surrounding it. The extent to which this rising air is warmer and lighter than the air through which it is passing depends, of course, on the amount of condensation.

This in turn depends on the amount of water vapour present, and that depends on the temperature. Hence, in general, and starting from the same level, saturated warm air is pushed up to greater heights in the process of convection than is saturated cold air.

In any case, though, the amount of condensation per given increase of height, and therefore the quantity of heat available for further convection, becomes less and less with gain of altitude, and finally, at one level or another, insufficient to induce further ascent. At this level then, whatever it is, the rising air spreads out in a sheet or stratum that differs in humidity, temperature and lapse rate from the atmosphere of every other level, both higher and lower.

In this way, that is, from convections, great and small, including the over- and underrunning associated with general or cyclonic storms, the troposphere is largely built up of overflow strata.

They are not, of course, the same from day to day nor from place to place, but everywhere they are always more or less distinct and numerous. They often mark the levels of cloud layers of the sheet or stratus forms, of alto-cumuli and of the windrow or billow clouds, due to the waves caused by the flow of one stratum over another, much as water waves are induced by wind. Occasionally, two adjacent strata differ from each other so radically that a balloon can float a long distance with the bag in the one and the basket in the other.

Of course the identity of each particular stratum ultimately is lost through mixing with others above and below it, whether caused by the vigorous stirring incident to a general storm, or by virtue of the diffusion and prevalent turbulence over every interface. But so long as it does exist it may be pushed up bodily to greater heights by underrunning air, or depressed to lower levels by an overflow current, with, in either case, a change in the temperature gradient or lapse rate (except in the very unusual case when it initially is that of completely stirred-up air) and a corresponding change in its stability and resistance to penetration by convection in either direction, upward or downward.

If the layer is pushed down, without lateral contraction or expansion, by an overflow of air above it the pressure on it will be increased by the same amount throughout, but it will be compressed most, and thereby heated most, on the upper side where the initial pressure is least, and compressed and heated least on the under side where the initial pressure is greatest. This changes the temperature gradient in the stratum so depressed, except rarely, and in such manner as to render the layer increas ingly difficult of penetration—a firmer floor and a more rigid ceiling. On the other hand, if the stratum is lifted to a higher level it is cooled most on top and least at the bottom and its effectiveness as a barrier to convection corresponding decreased.

It should be noted that in atmospheric convection, and the consequent production of air strata, water vapour plays a most important role. With increase of humidity, under constant temperature and pressure, the density of the air steadily decreases, just as it would with increase of temperature at constant pressure.

This fact probably accounts for many small waterspouts starting at the surface—starting there because the lower air becomes relatively light through high humidity, analogous to the starting of dust whirls over a dry region due to decrease of density incident to increase of temperature. There is, however, a fundamental difference between the surface waterspout and the dust whirl.

The latter consists of dry air made light by increase of temperature, and can ascend (be pushed up) until it has lost a certain amount of its original heat and no further. Not so with saturated or highly humid air. It, too, like the dry air, ascends because it is lighter than the adjacent air around about, but it maintains this relative lightness a much longer time through the latent heat rendered sensible by progressive condensation, and thereby reaches far higher levels.

In respect to convection dry air and humid air are like unto two men in business, one with a working capital but no reserve assets; the other having, in addition to his ordinary current needs a much greater fund that may be drawn upon whenever required. The one, like dry air, may start well, but his power to expand soon is exhausted. The other, by drawing on his reserve, can take advantage of every opportunity and thus rise to a far higher level of success. Just as it takes money to rise high in the business world, so too it requires water vapour to make any considerable ascent in the atmosphere.

The konisphere (dust sphere) and its layers. Not all we breathe is air. With every breath we inhale a million microsticks and -stones and a host of other things that are no part of a pure atmosphere. "Where do they come from?" The heavens above and the earth beneath.

Every wind that sweeps a desert catches up tons, and sometimes millions of tons, of pulverized rock to spread far and wide. Fragments of vegetable fibre litter the soil the world over and are wafted hither and yon as even the gentlest breeze may blow. Pollen of conifers, ragweeds, and a thousand other trees and plants we must take into our lungs from spring to fall every day we breathe the open air. And our bronchial tubes need chimney sweeps (luckily provided by Nature) to get rid of their coatings of soot from kitchens, factories and forest fires. Even the ocean, through its evaporated spray, makes a salt mine of the air that we breathe.

Then, too, lightning sprays nitrogen acids into the atmosphere, while soft coal and volcanic vents similarly add the sulphur acids—but all are too dilute really to bother us. Spores and microbes of many kinds we just have to inhale, for they are everywhere. And as if all this were not enough the earth, every now and then, explodes at some great volcano and hurls tons upon tons of rock powder into the air where it drifts far away for weeks, months or years, according to its degree of fineness and initial height attained. Finally, in addition to all this dust of its own the world stirs up, the atmosphere to its outermost limits is filled with the ashes, so to speak, of daily millions of incinerated meteors, or shooting stars. That is how the earth got its konisphere (dust shell). If it had no atmosphere it would have no konisphere, but having an atmosphere it must also have a coexistent and coextensive konisphere. But this konisphere is not uniform; it has distinct layers that, like other phenomena, show structure in the atmosphere. Few layers are mentioned below.

The Turbulence Layer

The turbulence layer, that is, the layer of air next to the earth that, owing to surface friction, any appreciable wind fills with turbulence. Incident to this churning up of the air there also is a stirring up of the dust. At such times this is thedustiest portion of all the atmosphere, and it carries the largest particles. Its depth is, of course, that of the turbulence, and therefore may be anything from two or three hundred feet up to two or three thousand; while the amount of dust, as determined by the strength of the wind and condition of the surface, can vary from practically nothing at all, as over snow fields, to that of the terror of the desert—the blinding and stifling sand storm. Its upper boundary is rather sharply marked and often distinctly visible from any higher level.

This layer includes also the city pall, that 4 tons a day, per square mile (average for Chicago, and there are worse places) smudge of soot and dirt that shuts out so much of the health-giving radiation of the sun.

The Convection Layer

The convection layer, or stratum of diurnal convection, marked by the dust carried up from near the surface by warm ascending currents. It therefore is deepest and dustiest during summer droughts, and over arid regions. Its upper surface, perhaps two miles high, frequently is seen by the aviator or balloonist almost as distinctly as the surface of an ocean, and even to resemble that surface through the emergence above it here and there of cumulus clouds that look like so many islands. This is the next dustiest of the shells of the konisphere, but even so its burden seldom is heavy enough to bother in any way those who move about in its densest portion—at the surface of the earth.

The Tropic Layer

The tropic layer, or layer coincident with the troposphere, and therefore 6 to 7 miles deep in middle latitudes and two or three miles deeper, on the average, in tropical regions. Its top is the limit of even occasional convection, and the dust of its upper levels relatively both sparse and fine. Its upper surface rarely has been observed, since aeroplanes and balloons seldom pass that level. We know where that surface is, however, because every ascending current of air necessarily carries with it some of the dust of the lower levels, and therefore dust of terrestrial origin must extend to the upper limit of convection, that is, to the top of the troposphere, and no farther. We also have observational evidence of this upper surface through the effect of the dust particles on sunlight.

The Stratic Layer

The stratic layer, or the region of all the atmosphere of appreciable density beyond the troposphere. The dust of this region is of two parts; one roughly constant in amount, the other extremely variable. The first comes from the myriads of meteors that hourly enter the atmosphere.

The second is due to occasional volcanic explosions of great violence which, like those of Asama in 1783, Krakatoa in 1883, Katmai in 1912, and many others, hurl powdered rock far beyond the levels of the highest clouds. The heavier dust particles of the turbulence layer quickly settle of their own weight.

To a less extent that is true also of the dust in the convection layer. In the main, however, the finest particles in the troposphere are carried down by condensation, either of water vapour directly onto them or as a result of being picked up by falling drops or drifting snow flakes. In this way the whole lower atmosphere from the surface of the earth to the tops of the highest clouds—a layer 6 to 10 miles thick—is literally washed, or scrubbed, as such processes are called, by rains and snows. If the air were so dry that there could be no precipitation it quickly would become suffocatingly filled with fine dust. In fact, it is believed by some that our sister planet, Venus, has just such a waterless, dust-filled atmosphere.

The dust of the stratosphere is not so fortunate, if we may put it that way, as that of the lower levels. It must get down through this region by itself, for there is not enough vapour up there to lend it any aid on its earthward journey. Often it is years in getting out of this arid realm, but once it has covered that part of its course the rest of the trip is quick and easy by way of the snowflake and raindrop routes. Even the dust of the earth, therefore, reveals a considerable structure of the atmosphere—at least four distinct layers. And it falls into still other great divisions according to this or that basis of separation. Nearly all the foregoing concerns the troposphere. The little that follows relates to the stratosphere, about which our knowledge still is very slight.

Upper Trades

Since the stratosphere is much warmer, 30° F. to 40° F., in the polar regions, than in the equatorial, it would seem that there must be an upper interzonal circulation of the atmosphere somewhat like a mirrored image of the lower—towards the poles in its under portion and from them in its upper levels. The rotation of the earth obviously would affect this upper interzonal circulation in the same manner that it does that of the troposphere, and therefore lead to east winds over the tropical and adjacent regions, and west winds over the higher latitudes. Furthermore, a little calculation based on the temperature of the atmosphere at various levels in high and low latitudes shows that this upper circulation must begin at the height of 10 to 12 miles. This calculation further shows that at that level the average pressure must be nearly the same everywhere, and therefore the average wind at this level very light. Both these conclusions, namely, that the winds 10 to 12 miles above the surface of the earth must be light, and above that level from the east in tropical and adjacent regions, are supported by all the observational data (a fair amount) we have on the subject.

Twilight Top

We see the course of a small beam of sunshine in a darkened room because there are dust particles in the air that scatter the light. That is why a few whisks over the floor with a dry broom makes the beam brighter. This explains, too, why we can see the shaft from the searchlight, and the focusing "streams" when the sun is "drawing water." In each case the contrast is between the myriads of illuminated motes and the shaded, hence darker, portions of the surrounding air. Not only the dust particles, but also, though to a far less degree, all the gas molecules of the atmosphere, are luminous in sunshine. This air luminosity is the chief factor in the blue of the sky, and an important factor in other sky colours. It accounts also for the twilight arch—the visible boundary between the shadow of the earth and the illuminated atmosphere—that rises above the eastern horizon as the sun sinks beneath the western.

Since the observer is himself within the earth's shadow it is obvious that by noting the exact time this arch is directly over head, say, he may know, from certain astronomical tables, just how many degrees the sun is then below the horizon; and that from this value, in turn, and the radius of the earth he can compute the height of this arch, that is, the greatest height at which the density of the air still is sufficient to scatter a perceptible amount of incident sunshine. Numerous measurements of this kind have been made, and that height thus found to be about 44 miles. In respect, then, to its efficiency as a light-scattering agent also the atmosphere has structure, an inner shell about 44 miles thick in which the scattering is appreciable, and an outer in which it is imperceptible.

Auroral Base

The polar lights, both northern (aurora borealis) and southern (aurora australis) divide the atmosphere into distinct parts, an inner, about 62 miles deep, into which auroras generally do not penetrate, and an outer of unknown thickness, but certainly hundreds of miles, in which they commonly do occur.

Kennelly-Heaviside Layer

When radio-telegraphy over long distances was first attained we were much puzzled to know how it could be, for surely radio waves are just greatly magnified light waves, and light doesn't bend to the curvature of the earth in such manner that an object can be seen a thousand miles away. But if the air were highly transparent, and both the earth and the encircling sky excellent reflectors, then a powerful light at London say, might well be seen from New York, or any other place on the globe.

The light could not get through either reflector and therefore would keep on traveling between them until finally absorbed. The same is true also of electric waves, and for these the earth is a reflector. If therefore, the sky reflected them too we would expect long distance radio communication to be possible, but not otherwise. Hence, when such communication was accomplished, Heaviside and Kennelly told us that the sky must be a reflector of electric waves, that is, an electric conductor, and everybody answered. "Why, of course, it is a conductor." And then came the long search to find how it is made a conductor and at what level its conductivity is adequate to account for the phenomena observed.

We believe now that this conductivity is owing essentially to the presence in the upper atmosphere, 30 to 60 miles or more above the surface, of a million or so free electrons per cubic inch, due to solar radiation in the far ultra violet. The height of the under surface of this reflecting region, or Kennelly-Heaviside layer, as it commonly is called, is not sharply determined since it appears to vary with the wave-length of the incident wireless wave; nor is it constant for any given wave-length, but varies from day to night and from season to season. But despite these variations the atmosphere, in respect to its electric state, and its relation therefore to wireless waves, always consists of two parts: a highly ionized, conducting and wave-reflecting outer shell, the Kennelly-Heaviside side; and a relatively non-conducting, but wave-transmitting inner shell. Electrically, also, the atmosphere has structure.

Ozone Layer

The composition of the atmosphere, there is very little ozone (triatomic oxygen) in the lower air up at least to the level of the highest clouds, but certainly very much more somewhere beyond that height. From spectroscopic observations several persons have computed the height of the centre of gravity of the ozone to be

25 to 30 miles above the surface of the earth. But whatever the correct value of this Iteight, surely in respect to ozone also the atmosphere has its structure—an intermediate shell rich in ozone, and an inner one and an outer that contain practically none at all.

TERRESTRIAL RADIATION

Terrestrial radiation refers to sources of radiation that are in the soil, water, and vegetation. The major isotopes of concern for terrestrial radiation are potassium, uranium and the decay products of uranium, such as thorium, radium, and radon.

Low levels of uranium, thorium, and their decay products are found everywhere. Some of these materials are ingested with food and water, while others, such as radon, are inhaled. The dose from terrestrial sources also varies in different parts of the world. Locations with higher concentrations of uranium and thorium in their soil have higher dose levels. The average dose rate that originates from terrestrial nuclides (except radon exposure) is about 0.057 μGy/hr. The maximum values have been measured on monazite sand in Guarapari, Brazil (up to 50 μGy/hr and in Kerala, India (about 2 μGy/hr), and on rocks with a high radium concentration in Ramsar, Iran (from 1 to 10 μGy/hr).

The major isotopes of concern for terrestrial radiation are uranium and the decay products of uranium, such as thorium, radium, and radon. Radon is usually the largest natural source of radiation contributing to the exposure of members of the public, sometimes accounting for half the total exposure from all sources. It is so important, that is is usually treated separately. The average annual radiation dose to a person from radon and its decay products is about 2 mSv/year and it may vary over many orders of magnitude from place to place.

It is important to note that radon is a noble gas, whereas all its decay products are metals. The main mechanism for the entry of radon into the atmosphere is diffusion through the soil. As a gas, radon diffuses through rocks and the soil. When radon disintegrates, the daughter metallic isotopes are ions that will be attached to other molecules like water and to aerosol particles in the air. Therefore all discussions of radon concentrations in the environment refer to radon-222. While the average rate of production of radon-220 (thoron) is about the same as that of radon-222, the amount of radon-220 in the environment is much less than that of radon-222 because of significantly shorter half-life (it has less time to diffuse) of radon-222 (55 seconds, versus 3.8 days respectively). Simply radon-220 has lower chance to escape from bedrock.

Radon-222 is a gas produced by the decay of radium-226. Both are a part of the natural uranium series. Since uranium is found in soil throughout the world in varying concentrations, also dose from gaseous radon is varying throughout the world. Radon-222 is the most important and most stable isotope of radon. It has a half-life of only 3.8 days, making radon one of the rarest elements since it decays away so quickly. An important source of natural radiation is radon gas, which seeps continuously from bedrock but can, because of its high density, accumulate in poorly ventilated houses. The fact radon is gas plays a crucial role in spreading of all its daughter nuclei. Simply radon is a transport medium from bedrock to atmosphere (or inside buildings) for its short-lived decay products (Pb-210 and Po-210), that posses much more health risks.

Radioactive cascade significantly influences radioactivity (disintegrations per second) of natural samples and natural materials. All the descendants are present, at least transiently, in any natural sample, whether metal, compound, or mineral. For example, pure uranium-238 is weakly radioactive (proportional to its long half-life), but a uranium ore is about 13 times more radioactive than the pure uranium-238 metal because of its daughter isotopes (e.g. radon, radium etc.) it contains. Not only are unstable radium isotopes significant radioactivity emitters, but as the next stage in the decay chain they also generate radon, a heavy, inert, naturally occurring radioactive gas. Radon itself is a radioactive noble gas, but the main issue is that it is a transport medium from bedrock to atmosphere (or inside buildings) for its short-lived decay products (Pb-210 and Po-210), that posses much more health risks.

4

Atmospheric Pressure and Winds

ATMOSPHERIC PRESSURE

Atmospheric pressure, also called barometric pressure, force per unit area exerted by an atmospheric column (that is, the entire body of air above the specified area). Atmospheric pressure can be measured with a mercury barometer (hence the commonly used synonym barometric pressure), which indicates the height of a column of mercury that exactly balances the weight of the column of atmosphere over the barometer. Atmospheric pressure is also measured using an aneroid barometer, in which the sensing element is one or more hollow, partially evacuated, corrugated metal disks supported against collapse by an inside or outside spring; the change in the shape of the disk with changing pressure can be recorded using a pen arm and a clock-driven revolving drum.

Atmospheric pressure is commonly measured with a barometer. In a barometer, a column of mercury in a glass tube rises or falls as the weight of the atmosphere changes. Meteorologists describe the atmospheric pressure by how high the mercury rises.

An atmosphere (atm) is a unit of measurement equal to the average air pressure at sea level at a temperature of 15 degrees Celsius (59 degrees Fahrenheit). One atmosphere is 1,013 millibars, or 760 millimeters (29.92 inches) of mercury.

Atmospheric pressure drops as altitude increases. The atmospheric pressure on Denali, Alaska, is about half that of Honolulu, Hawai'i. Honolulu is a city at sea level. Denali, also known as Mount McKinley, is the highest peak in North America.

As the pressure decreases, the amount of oxygen available to breathe also decreases. At very high altitudes, atmospheric pressure and available oxygen get so low that people can become sick and even die.

Mountain climbers use bottled oxygen when they ascend very high peaks. They also take time to get used to the altitude because quickly moving from higher pressure to lower pressure can cause decompression sickness. Decompression sickness, also called "the bends", is also a problem for scuba divers who come to the surface too quickly. Aircraft create artificial pressure in the cabin so passengers remain comfortable while flying.

Air has weight

If a person were to climb a tall mountain, like Mauna Kea on the Big Island of Hawaii, where the summit reaches to 13,796 feet (4,206 meters), contracting altitude sickness (hypoxia) is a high probability. Before ascending to the summit, visitors must stop at the Information Center, located at an altitude of 9,200 feet (2,804 m) where they are told to acclimatize to the altitude before proceeding further up the mountain. “Well, of course,” you might say, “After all, the amount of available oxygen at such a high altitude is considerably less as compared to what is present at sea level.”

In fact, 21 percent of Earth's atmosphere consists of life-giving oxygen (78 percent is composed of nitrogen and the remaining 1 percent a number of other gases). And the proportion of that 21 percent is virtually the same at sea level as well as at high-mountain altitudes.

That oft-used analogy of comparing air with water (“an ocean of air”) is a good one, for we are all literally swimming through air. Now picture this: A tall plastic bucket is filled to the brim with water. Now, take an ice pick and poke a hole near the top of the bucket. The water will slowly dribble out. Now take the pick and punch another hole down near the bottom of the bucket. What happens? Down there the water will rapidly squirt out in a sharp stream. The reason is the difference in pressure. The pressure that is exerted by the weight of the water down near the bottom of the bucket is greater than up near the top, so the water is "squeezed out" of the hole at the bottom.

Similarly, the pressure of all the air above our heads is the force that pushes air into our lungs and squeezes oxygen out of it and into our bloodstream. As soon as that pressure diminishes (such as when we ascend a high mountain) less air is pushed into the lungs, hence less oxygen reaches our bloodstream and hypoxiation results; again, not due to a lessening of the amount of available oxygen, but to the lessening of atmospheric pressure.

Highs and lows

So how does atmospheric pressure relate to daily weather patterns? No doubt you've seen weather forecasts presented on television; the on-camera weathercaster making reference to high pressure and low pressure systems. What is that all about?

Basically, in a nutshell, every day the heat of the sun varies all over the Earth. Because of unequal solar heating, temperatures vary over the entire globe; the air at the equator is much warmer than at the poles. So the warm, light air rises and spreads toward the poles and the colder, heavier air sinks toward the equator.

But we live on a planet that rotates, so this simple wind pattern is distorted to such a degree that the air is twisted to the right of its direction of motion in the Northern Hemisphere and to the left in the Southern Hemisphere. Today we know this effect as the Coriolis Force and as a direct consequence, great wind spirals are produced which we know as high and low pressure systems.

In the Northern Hemisphere, the air in low pressure areas spirals counterclockwise and inward — hurricanes, for instance, are Coriolis mechanisms, circulating air counterclockwise. In contrast, high pressure systems the air spirals clockwise and outward from the center. In the Southern Hemisphere the direction of the spiraling of the air is reversed. High pressure systems are "domes of density" that press down, while low pressure systems are akin to "atmospheric valleys" where the density of the air is less. Since cool air has less of a capacity to hold water vapor as opposed to warm air, clouds and precipitation are caused by cooling the air. So by increasing the air pressure, the temperature rises; underneath those high pressure domes, the air tends to sink (called "subsidence") into the lower levels of the atmosphere where temperatures are warmer and can hold more water vapor. Any droplets that might lead to the formation of clouds would tend to evaporate. The end result tends to be a clearer and drier environment. Conversely, if we decrease the air pressure, the air tends to rise into the higher levels of atmosphere where temperatures are colder. As the capacity to hold water vapor diminishes, the vapor rapidly condenses and clouds (which are composed of countless billions of tiny water droplets or, at very high altitudes, ice crystals) will develop and ultimately precipitation will fall. Of course, we could not forecast zones of high and low pressure without employing some sort of device to measure atmospheric pressure.

Enter the barometer

Atmospheric pressure is the force per unit area exerted by the weight of the atmosphere. To measure that weight, meteorologists use a barometer. It was Evangelista Torricelli, an Italian physicist and mathematician who proved in 1643 that he could weigh the atmosphere against a column of mercury. He actually measured pressure converting it directly to weight. The instrument Torricelli designed was the very first barometer. The open end of a glass tube is placed in an open dish of mercury. Atmospheric pressure forces the mercury to rise up the tube. At sea level, the column of mercury will rise (on average) to a height of 29.92 inches or 760 millimeters. Why not use water instead of mercury? The reason is that at sea level, the water column would be about 34 feet high! Mercury on the other hand, is 14 times denser than water and is the heaviest substance available that remains a liquid at ordinary temperatures. That permits the instrument to be of a more manageable size.

How NOT to use a barometer

Right now you might have a barometer hanging on the wall of your home or office, but in all likelihood it's not a tube of mercury but rather a dial with an arrow that points to the current barometric pressure reading. Such an instrument is called an aneroid barometer, which consists of a partially evacuated metal cell that expands and contracts with changing pressure and is attached to a coupling mechanism that drives an indicator (the arrow) along a scale graduated in pressure units, either in inches or millibars.

Usually on the indicator dial you will also see words such as "Sunny," "Dry," "Unsettled," and "Stormy." Supposedly, when the arrow points toward these words it is supposed to be an indication of the expected weather ahead. "Sunny," for instance, can usually be found in the range of high barometric pressure — 30.2 or 30.3 inches. "Stormy," on the other hand would be found in the range of low barometric pressure — 29.2 or lower, perhaps even on occasion below 29 inches.

This would all seem logical, except it is all rather simplistic. There can be times, for example, when the arrow will be pointing to "Sunny," and the sky instead is completely overcast. And on other occasions, the arrow will be suggesting "Stormy," and yet what you might see is sunshine mixed with blue sky and fast-moving, puffy clouds.

How to properly use a barometer

That's why along with the black indicator arrow, you should also pay attention to another arrow (usually gold) that can be manually adjusted to any part of the dial. When you check your barometer, first tap the front of the barometer lightly to remove any internal friction and then align the gold arrow with the black one. Then check back some hours later to see how the black arrow has changed relative to the gold one. Is the pressure rising or falling? If it's falling, is it doing so rapidly (perhaps dropping several tenths of an inch)? If so, a storm could be approaching. If a storm has just passed by and the skies have cleared, the barometer might still be indicating "Stormy" weather, but if you had set the gold arrow some hours ago, you almost certainly would see that the pressure is now rapidly rising, suggesting that — in spite of the indication of storminess — that fair weather is on the way.

And your forecast can be improved still further by combining your record of changing barometric pressure with the changing direction of the winds. As we have already learned, air circulates in a clockwise fashion around high pressure systems and counterclockwise around low pressure systems. So if you see a trend toward rising pressure and a northwesterly wind, you might expect generally fair weather to move in, as opposed to a falling barometer and an east or northeast wind which could eventually lead to clouds and precipitation.

IMPORTANCE OF ATMOSPHERE

From the perspective of the planetary geologist, the atmosphere is an evolutionary agent essential to the morphology of a planet. The wind transports dust and other particles which erodes the relief and leaves deposits (eolian processes). Frost and precipitations, which depend on the composition, also influence the relief. Climate changes can influence a planet's geological history. Conversely, studying surface of earth leads to an understanding of the atmosphere and climate of a planet - both its present state and its past. For a meteorologist, the composition of the atmosphere determines the climate and its variations. For a biologist, the composition is closely dependent on the appearance of the life and its evolution.

STEADY STATE

In chemistry, a steady state is a situation in which all state variables are constant in spite of ongoing processes that strive to change them. For an entire system to be at steady state, *i.e.* for all state variables of a system to be constant, there must be a flow through the system (compare mass balance). One of the most simple examples of such a system is the case of a bathtub with the tap open but without the bottom plug: after a certain time the water flows in and out at the same rate, so the water level stabilizes and the system is at steady state.

The steady state concept is different from chemical equilibrium. Although both may create a situation where a concentration does not change, in a system at chemical equilibrium, the net reaction rate is zero while no such limitation exists in the steady state concept. Indeed, there does not have to be a reaction at all for a steady state to develop.] The term steady state is also used to describe a situation where some, but not all, of the state variables of a system are constant. For such a steady state to develop, the system does not have to be a flow system. Therefore such a steady state can develop in a closed system where a series of chemical reactions take place. Literature in chemical kinetics usually refers to this case, calling it steady state approximation.

In simple systems the steady state is approached by state variables gradually decreasing or increasing until they reach their steady state value. In more complex systems state variable might fluctuate around the theoretical steady state either forever or gradually coming closer and closer. It theoretically takes an infinite time to reach steady state, just as it takes an infinite time to reach chemical equilibrium. Both concepts are, however, frequently used approximations because of the substantial mathematical simplifications these concepts offer.

Whether or not these concepts can be used depends on the error the underlying assumptions introduce. So, even though a steady state, from a theoretical point of view, requires constant drivers, the error introduced by assuming steady state for a system with non-constant drivers may be negligible if the steady state is approached fast enough.

Validity

The analytical and approximated solutions should now be compared in order to decide when it is valid to use the steady state approximation. The analytical solution transforms into the approximate one whenbecause then and Therefore it is valid to apply the steady state approximation only if the second reaction is much faster than the first one because that means that the intermediate forms slowly and reacts readily so its concentration stays low.

The equilibrium approximation can be used sometimes in chemical kinetics to yield similar results as the steady state approximation: it consists in assuming that the intermediate is at chemical equilibrium. Normally the requirements for applying the steady state approximation are laxer: the concentration of the intermediate is only needed to be low and more or less constant but it is not needed to be at equilibrium, which is usually difficult to prove and involves heavier assumptions.

PRESSURE

The pressure of the atmosphere, its push per unit area, or weight of a vertical column of it of, say, one square inch cross section where the pressure is measured, extending from that level to its outer limit, is one of the most important of all meteorological measurements, especially as an aid, when known at many places, to weather forecasting. And yet our senses do not make us aware of its magnitude or of its changes.

Mercurial Barometer

The action of an ordinary suction pump, the drinking of cider through straw (they didn't have soda water at the time we are talking about), and a lot of things of like kind, used to b explained by the learned as due to Nature's abhorrence of a vacuum. Galileo is said to have remarked, ironically, it must have been, that in the case of water Nature didn't abhor a vacuum beyond about 30 feet, that being the limit to which water could be pulled up with a suction pump.

Torricelli in experimenting on this problem took a long glass tube, closed at one end and open at the other, filled it with mercury and then stood it upside down with its lower end dipping into a basin of the same substance.

The mercury in the tube dropped until it stood about 30 inches above that in the basin, a height that is the same fraction of that of which water can be sucked as its density is of that of mercury. This showed that Nature abhorred a vacuum only to the extent of the weight of the air then above the place of abhorrence. At sea level this generally is around 14 pounds per horizontal square inch. It decreases with increase of height by the weight of the air left below, a fact that affords a ready means of determining height without directly measuring them.

It also generally varies with the kind of weather and strength of the winds. The simple tube and basin of Torricelli are the essential elements of the mercurial barometer or measurer of the weight, or pressure, of the air by means of a balancing column of mercury. Ordinarily they are provided with a convenient supporting frame, scales, thermometers and adjustment screws; but, simple or complex, the purpose is to mea- sure the pressure of the air at the level of the basin surface in terms of the height of a balancing column of mercury.

Aneroid Barometer

The pressure element of the aneroid barometer is a flexible metal shell, fully exhausted and sealed, with top and bottom held apart by means of a suitable spring. With change of atmospheric pressure the shell correspondingly contracts or expands, as the case may be, which movement commonly is translated into the travel of a pointer over a dial marked in terms of the corresponding readings of a mercurial barometer.

Altimeters

An altimeter is only an aneroid barometer with a scale in terms of heights instead of pressures—decrease of pressure indicating an increase of height.

Barograph

The most common barograph, is only an aneroid barometer so constructed that the traversing arm writes an continuous record on a uniformly moving and properly graduated sheet of paper. The mercurial barometer also can be made to keep a continuous record, graphic or photographic, of its own height.

WIND

Wind Velocity

The velocity of the wind can be measured more or less accurately by many different means. However, the device commonly used for this purpose is the Robinson cup anemometer, the best form of which consists of three or four equally spaced hemispherical cups that open horizontally, face in the same direction around their common path and are attached through short arms to a vertical shaft which, through proper gearing, momentarily closes an electric circuit at the end of every so many revolutions.

At each closing of the circuit a record is made, in the office or elsewhere, on a uniformly moving sheet of paper. The instrument may be so geared that a contact occurs for each mile, approximately, travel of the wind, or other distance as desired. This value, together with the known rate of movement of the record paper, gives the average speed of the wind over whatever appreciable interval, and at whatever time, one may wish to know it.

Quick Acting Anemometers

As usually constructed the Robinson cup anemometer just described is too sluggish, has too much inertia, to respond promptly to changes of wind velocity and therefore is not adapted to the measurement of wind gusts and eddies. However, there are various more or less elaborate ways by which such measurements can be made.

A tube, for instance, can be kept facing the wind and the pressure inside registered through the resulting height of a balancing liquid column; or the cooling of an electrically heated wire, exposed to the wind, may be measured by its resistance and, by proper calibration, the wind velocity and its changes thus determined quite accurately. Various other methods of making a quick acting anemometer readily suggest themselves, but, although greatly needed, none has yet come into use that at once is simple, sturdy, accurate, continuously recording and inexpensive.

Wind Direction

The direction of the wind is best and most easily obtained by the use of a wind vane, a horizontal arrow with a vertically broad tail so mounted on an upright rod as to turn easily with the wind. The supporting rod may turn with the vane, and may carry a pointer at its lower end moving just beneath a dial on the ceiling of a room or portico beneath, from which one may see the pointing of the vane and thus know the direction of the wind without going out of doors.

This ingenious device was one of the luxuries of Roman villas 2000 years ago, and also one of Jefferson's many conveniences at Monticello. However, as commonly used at meteorological stations, the turning rod carries certain electrical contacts by which registration to eight directions is secured on a moving sheet of paper at any desired location. It also is practicable to obtain a continuous record of wind direction by means of a vertically moving pen over a sheet of paper on a cylindrical drum carried by the vane rod.

Recently the vane has been rendered much more sensitive by making the tail blade relatively short and of stream-line shape — blunt and rounded in front and tapering to a vertical edge in the rear—instead of long and flaring, with the edge in front.

HUMIDITY

There are many ideas about humidity, mostly vague and erroneous. The correct ideas are but few. The humidity, or water vapour content, of the air is expressed in various ways according to the purpose in view.

Absolute Humidity

This expression means the actual mass of water vapour present per unit volume; and it makes no difference whether the other gases of the air are present or not. They have nothing to do with the humidity. We often say, too, that the absolute humidity is the actual pressure per unit exerted by the water vapour present. This is not the weight of the water vapour in a vertical column of unit cross section, but just that fraction of the total atmospheric pressure, at the place in question, that the number of water molecules per unit volume at that place is of the total number of molecules of all kinds in that unit volume. At any given temperature this vapour pressure per unit area is directly proportional to the mass of water vapour per unit volume. Hence, if we know what we are doing, we may express humidity in terms of either mass or pressure. Absolute humidity seldom is measured directly in meteorological observations.

Relative Humidity

Relative humidity is that fraction, or percentage, which the actual water vapour present per unit volume is of the total amount that unit volume could contain at the same temperature, and in the presence of a flat surface of pure water. Commonly one says, "per unit volume of air," "total amount the air could contain," and the like, but that is all wrong.

The water vapour is not contained by the air, it is a part of the air; neither do the other gases of the atmosphere soak up water vapour, as a sponge takes up liquid water. In fact, they do not appreciably affect the amount of water vapour per given volume necessary to produce saturation, or any particular fraction of saturation. That is determined by temperature alone. In meteorological work relative humidity is determined indirectly and by reference to tables or graphs previously established by elaborate experiments under known conditions.

Psychrometer

The psychrometer or instrument generally used for determining the humidity, consists of a pair of thermometers, one dry, the other wet (having a closely fiting jacket of wet muslin) at the time of observation and provide with some means of obtaining ample ventilation.

The dry thermometer gives the current temperature of the air, while the wet one indicates the cooling due to evaporation, a quantity that depends on both the actual temperature of the air and its relative humidity, or, if we prefer, humidity deficit (difference between saturation and the current humidity), or dryness. These numerical relations have been determined empirically by numerous careful experiments and recorded in convenient tables, so that all we now have to do is to read both the wet bulb and the dry bulb thermometers and then look up in printed tables the relative humidity, absolute humidity and dew point, or temperature at which the vapour actually present would produce saturation.

Hair Hygrograph

It has been found experimentally that at all ordinary temperatures of the atmosphere the human hair, when clean and oilless, responds but little to temperature changes but appreciably lengthens with any considerable increase of the relative humidity. Hence, although the rate of this gain in length to a given increase of the relative humidity decreases as saturation is approached, it is possible, with a small strand of hairs, acting through a suitable cam, to make a tracing pen so traverse a moving graduated sheet of paper as to produce a reliable record of the varying humidity.

CLOUDS

Amount

The amount of cloudiness, or portion of the sky covered with clouds, is a very important weather element, but ordinarily it is only crudely estimated in terms of tenths of the sky that appears to be covered. Usually, too, no distinction is made between very thin and very thick clouds, although they differ greatly in respect to the amount of sunlight that gets through them.

Height

The height of a cloud canopy, now an exceedingly important matter in connection with aviation, can be determined fairly closely by any one of several methods. Evidently all aviator can fly up to the cloud and note the reading of the altimeter when he gets there. Or, without going up oneself, one can set free a pilot balloon (a small rubber balloon), so inflated as to have a known rate of ascent, and note the interval from the time of launching to disappearance in the cloud. This interval multiplied by the rate of ascent gives the approximate height of the cloud above the observer. Greater accuracy is obtained by following the balloon with two theodolites at known positions some distance apart. From the pointing of the two instruments at the instant the balloon disappeared the exact height of the cloud is readily computed. Evidently, too, the height of a cloud could be gotten fairly well in many cases, with a good range finder.

A simple way of getting this height at night is to throw a parallel beam of light, search light or "ceiling light," onto the cloud at a known angle of elevation and then sight the illuminated spot, noting also this angle of elevation, from a point at a known distance from the light, at its level and in the same vertical plane as the light and spot.

The height of the cloud may then be read on the pointer, if the other values have been properly chosen, or in any case, easily computed. If the beam is turned up 45° then clearly the height of the cloud is equal to the distance from the light to that point on the ground which is directly beneath the illuminated spot.

Velocity

There are also several methods of measuring the velocity of a cloud. We can follow some recognizable feature with one or two theodolites and compute the velocity by triangulation. If it is an isolated cloud casting a shadow we may be able to get the velocity of the shadow, which, evidently, is the same as the velocity of the cloud casting it.

Or we may watch through a fixed peep-hole the travel of a cloud image over a horizontal mirror—a nephoscope, as it is called —and compute the value desired from the obvious fact that the velocity of the cloud is to the speed of its image across the mirror as its height above the observer is to the height of the peep-hole above the mirror. This assumes that the height of the cloud is known. The nephoscope also, and always, gives the direction of travel of the cloud.

Kind of Cloud

The kind of cloud, as well as its amount, velocity and direction of travel, is noted in meteorological observations because, in connection with other things, it is significant of the coming weather, and because it is a climatic element, though one of quite secondary importance. And the kind can be recorded either by its recognized name, or by letter or number corresponding to a particular cloud picture.

The names that refer to the primary or fundamental forms are: cirrus, or curl cloud; stratus, or layer cloud; cumulus, or pile cloud; and nimbus, or rain cloud. But these few names are not enough to go around, so we have also combination names; cirrostratus, cirro-cumulus, strato-cumulus and cumulo-nimbus; names with "alto" as prefix, signifying high, that is, high for that particular kind of cloud; alto-stratus, and alto-cumulus; and names beginning with " fracto" to indicate a ragged state of the cloud; fracto-stratus, fracto-cumulus, and fracto-nimbus; and, finally, names descriptive, usually, of appearance of a number of clouds that differ in some important respect from any of the above forms, such as the mammato-cumulus, a sort of a festooned cloud; the funnel cloud, characteristic of the tornado; the banner cloud, a cloud floating like a banner from a mountain peak; crest cloud, the cloud that envelops and stretches along a mountain ridge; and an indefinite number of still other names, generally self-explanatory.

PRECIPITATION

The amount of snowfall and rainfall, and the time, rapidity and frequency of its occurrence all are important matters. Systematic observations of the amount of rain caught in properly exposed open-mouthed vessels have been made, at least sporadically and in various countries, for many centuries. In Palestine, for instance, quantitative measurements were made in the first century A.D. Today they are made in great numbers in every progressive country, and the method of making them is everywhere, and always has been, essentially the same.

Rain Gage

Every rain gage, however much it may differ from others (and there are many kinds), is an open-mouthed vessel set upright where the amount of precipitation is not appreciably affected by the surroundings. To be at all accurate the mouth of the vessel must be horizontal and sharp lipped so that all the precipitation that hits the mouth, and none other, will be caught. To decrease the errors of measurement the catch may be funneled to a cylindrical vessel whose cross section is a known fraction, a tenth, say, of that of the exposed mouth. The rainfall then, in terms of the horizontal layer of water it would have produced if none had run off, or otherwise disappeared, is one tenth, or whatever the fraction may be, of the depth of the water in the smaller vessel.

A continuous record of the rainfall may be obtained by allowing the water, as it is caught, to run through a small pipe into a little rocking bucket, divided into two equal parts by a partition, and so adjusted that as soon as water representing one one-hundredth of an inch of rainfall, or any other predetermined amount, has been caught in one compartment it tips over, thus emptying that half and catching in the other, and so on as long as the rain lasts.

The bucket can be so connected in an electric circuit that each tip it makes, hence every hundredth, say, of an inch of rainfall, is automatically recorded on a moving sheet of paper. In this way the quantity, intensity and time of occurrence of each rain is permanently recorded.

If desired the catch can be piped into a suitable vessel supported by a weighing device adapted to either occasional reading or continuous recording. Evidently, too, a record may be obtained on a moving sheet of paper by a pen attached to a float in the rain catch vessel. But whatever the kind of device used for measuring rainfall and however various the details, the fundamentals remain the same.

Snow Gage

When the precipitation is snow it is more difficult to obtain a true catch, owing to the tendency of the wind disturbances, induced by the vessel itself, to keep the snow from falling into it, or even to whisk from the gage a portion of that already captured. Also drifting snow may be caught and thus precipitations recorded when actually there were none at all—only shiftings of snow that previously had fallen.

These troubles are reduced by placing a downward deflecting shunt around the gage at the level of the catchment mouth. The snow thus caught may be melted and the resulting water measured, or it may be weighed and the equivalent rainfall computed.

Snow Tube

Often it is desirable to know the mass or water equivalent of the snow at a particular time over a considerable mountain area, especially for computing the probable run-off available later on for power and irrigation. This information can be obtained by pushing a thin-walled, open-minded, metal tube vertically through the snow and thus picking up a column of the snow of known cross section.

The lower end of the tube should be rather sharp lipped and slightly smaller in diameter than the main body to prevent packing. The difference in weight between tube plus snow core, and tube alone, gives the weight of a column of snow of the known cross section. This is readily reduced to equivalent water depth, or, for convenience, the scales may be graduated in terms of water depth and set to read zero when the tube is empty. By taking many such samples properly distributed the available run-off from a given reservoir of snow can be fairly closely determined.

Hail Gage

If the catch of a rain gage is immediately passed over a sloping section of wire gauze in such manner as to shunt the hail stones, when there are any, into a special compartment the volume, or weight, of the water they produce on melting is, of course, a measure of the hail fall. To insure a catch fairly representative of the fall, especially if it happens to be sparse, the mouth of the gage must be quite large. If the water that drains through the gauze is also caught, separately, the combination might be called a rain-hail gage.

Drosometer

This instrument that weighs the amount of dew that collects on a given area of any chosen material is very little used since the quantitative values thus obtained have practically no current use. However, the frequency of the occurrence of dew, month by month, or week by week, and its relative amounts—light, moderate, heavy—is an important climatic factor in respect to certain crops and their diseases. Grapes, for instance, may succeed admirably at one place, where the dews are few and light, and fail at another, less than a mile away, owing to fungus diseases that are serious where dews are frequent and heavy.

EVAPORATION

Evaporation occurs from growing leaves, damp soil and free water surfaces. However, measurements of evaporation generally are concerned with the latter, or free water surface, only. And even in this case it is not the evaporation that is measured but the net evaporation, that is, the difference between the total amount of water that left the surface under consideration as vapour, and the amount that in the same time, condensed onto it as water from the vapour in the adjacent space.

This difference indeed may be either positive, evaporation greater than condensation; zero, evaporation and condensation equal; or negative, condensation greater than evaporation, according as the temperature of the water is greater than, equal to, or less than the dew point of the nearby air. The common method of measuring net evaporation is to note the rate of fall, or rise, of the surface of water in a suitably exposed vessel, usually a cylinder 2 to 4 feet in diameter and a foot deep. Innumerable measurements of this kind have been made under various conditions, from which we know that evaporation varies with the temperature and salinity of the water, fresh water evaporating about 5% faster than sea water; condition of the surface (clean or foul); temperature and humidity of the air; area of the surface;

and wind velocity. Nevertheless, no exact, useful and all-inclusive equation has been found for evaporation. Indeed, it is practically certain that no such equation can rigidly apply to natural bodies of water, such as lakes and ponds.

Of course the amount of evaporation by a given body of water is proportional to the energy it absorbs from all sources during the time in question minus the energy it loses in every way except by evaporation. But this does not necessarily help much since each of the quantities of energy involved may be more difficult to determine than the evaporation directly.

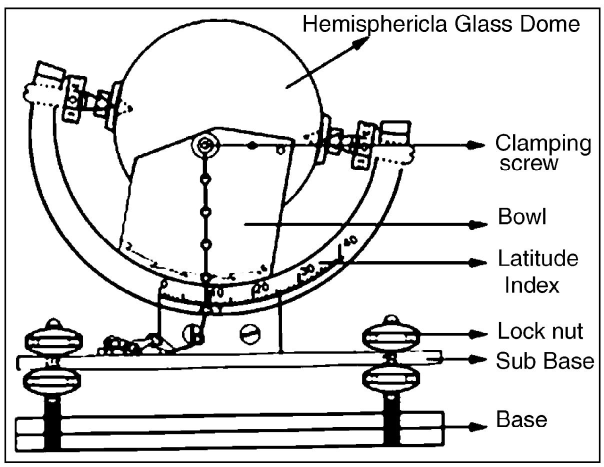

Fig. ***Sunshine Recorder.***

Insolation

Fundamentally, sunshine is the most important of all meteorological and climatological elements, and it is measured in respect to different factors.

Duration

One measure of sunshine is just its duration, either in hours and minutes or as a percentage of the total possible, and the times of its occurrence. This can be done by using a glass sphere as a burning lens to sear a line, during sunshine, on a curved and properly ruled sheet of paper. The same measurements are made also with a vacuum-jacketed differential air thermometer, one bulb blackened, the other bright with a plug of mercury in the connecting stem, and so adjusted that while the sun is shining, and only then, the position of the mercury is such as to be in touch with a couple of sealed-in wires.

This affords the means of obtaining, with an electric current, an automatic record of the duration, but not the intensity or quality, of the sunshine. Various other devices could be used to accomplish the same ends. Those outlined are only the two in widest use.

Intensity

The intensity of the sunshine is measured in terms of the amount of energy it delivers per unit of time, over a unit area that either directly faces the sun or is horizontal. The radiation is caught by some form of a black body, or perfect absorber, that is protected from disturbances from other sources than the sun, and its intensity indicated by the rate of heating of the receiver by a thermometer, mercurial or electrical; by change of voltage of a thermo-couple, or otherwise. Also every pyrheliometer, as this instrument is called, is provided with a number of necessary and convenient accessories.

Quality

The quality of the radiation is determined both as to kind, or wave length, and relative intensity, by traversing its spectrum across a thin, narrow strip of blackened platinum which continuously records, automatically, its changes in electrical resistance, incident to variations of temperature; and, finally, correcting these values for the inequality of the dispersion or spreading out of the spectrum (the width of the strip being constant), and for the previous loss of energy by reflection and absorption within the instrument.

Sky Light

The light of the sky is measured in terms of its brightness over various portions, colour, and polarization—a state caused by scattering and reflection, but not ordinarily perceptible to the eye.

Earth Radiation

Evidently the earth loses heat by radiation at rates that vary with the temperature and the nature of the radiating surfaces. Many efforts have been made to measure this radiation, but with indifferent success. In fact, the measurements actually made usually are of the difference between the energy lost in a given time, per unit horizontal area of a black body, to the sky, and the energy simultaneously absorbed from the sky by that same area.

If the temperature of the black body is known, as it may be, its total radiation is easily computed; and if the black body is so protected as to lose and gain heat by radiation alone, this total computed loss, minus the measured difference between loss and gain, gives the energy received per unit horizontal area from the sky. The radiation from the surface of the earth probably can be computed more reliably from its temperature than actually measured, since it loses heat nearly as does a black body.

ATMOSPHERIC DUST

The dust in the atmosphere may be classified in many different ways. One kind is the relatively gross particles. These are most abundant in the lower air, often 10,000 to 50,000 per cubic inch, especially in cities and over desert regions. These affect our health and our industries and are largely responsible for the haze that limits the vision of the aviator.

It therefore is important that the quantity of this dust be measured, and the things it consists of determined. Both can be done by first humidifying the air, then drawing it rapidly into an expansion chamber through a small opening—a narrow slit is especially convenient—in such manner that it will impinge squarely against a flat object, such as a microscope cover glass, and finally examining with a suitable microscope the material thus caught sticking to the baffle surface owing to the water condensed on the particles incident to the cooling of the expanding humid air.

Another method of catching the dust is to cause the air, in its current condition, to impinge gently against a surface made sticky with a very thin film of vaseline, or other suitable substance. The material of this dust includes many kinds of rocks and minerals, soot, sea salt, pollen (in season), spores and every sort of vegetable fibre. Indeed one feels like wearing a dust mask, when he recalls the fact that at a single breath he may inhale a million sticks and stones of a hundred different kinds!

But this is not the whole story. There is another, though overlapping, kind of dust that consists of particles often far more numerous than those just discussed; most of them smaller and many of them liquid or even gaseous. This "dust" consists of all those particles, solid, liquid or gaseous, upon which water vapour readily condenses as the temperature falls below the dew point—the condensation nuclei of fogs and clouds. It will not pass through a plug of raw cotton or similar material.

The number of these particles, ranging from a few hundred to more than a million per cubic inch, may easily be estimated by saturating the air to be examined in a closed vessel, then giving it two or three sharp, or fog-producing, expansions and counting the number of droplets caught per unit area on the lower of two horizontal surfaces a known short distance apart.

For convenience the surface on which the droplets fall may be of polished silver lightly ruled in squares of known area. There generally are other conveniences also, microscope, lamp, pump, filters, et cetera, but the above are the essentials of the Aitken dust counter, named after John Aitken, the first to give much attention to this subject.

This sort of "dust," essential to the formation of every fog and cloud, has been under investigation now for more than 50 years. We know that nonhygroscopic particles of whatever origin are no part of it. We know, too, that particles of sea salt, gotten into the air by evaporating spray, are a large portion of it; and that many other substances, even certain gases, act in this capacity. Nevertheless, the problem of the condensation nuclei still is but imperfectly solved.

Vlsibility

The degree of transparency of the air is very important to the aviator. It means, as the case may be, good seeing and knowing all the time where he is, or bad seeing and getting lost. Poor seeing generally is due chiefly to glare, caused by the reflection and scattering of light by particles in the line of sight. This glare commonly is reduced by a passing cloud. Also it can be reduced by looking through yellow to reddish glasses since light of this colour is not nearly so much scattered by dust as is the blue, for instance.

If the poor seeing is owing to actual fog it is but little improved by screens of kind since fog scatters all colours about equally. The usual method of measuring visibility is the crude one of noting the maximum distance at which houses, trees, et cetera, can be recognized—seen well enough to know what they are.

Electrical State

The electrical of the atmosphere may be measured, but seldom is at meteorological stations, since it has but little weather significance. The factors most frequently measured are the difference in voltage between two horizontal levels, a known distance apart; the number of ions or electrified particles, both positive and negative, per unit volume; and the electrical conductivity.

Free Air And Upper Air Measurements

All the foregoing concerns measurements made near the surface of the earth. If similar measurements are to be made at considerable heights then evidently we must take the necessary instruments there ourselves in aeroplanes or balloons, or send up self-registering, or automatic wireless reporting instruments with free balloons of adequate size. All these means are in abundant use, and the basic instruments employed are fundamentally the same as those already described. They embody many ingenious details, to be sure, but like most other details, they may not be tomorrow what they are today, and therefore will not be further considered.

URBAN ATMOSPHERES

Urban Atmospheres captures a unique, synergistic moment-expanding urban populations, rapid adoption of Bluetooth mobile devices, tiny ad hoc sensor networks, and the widespread influence of wireless technologies across our growing urban landscapes. The United Nations recently reported that 48 per cent of the world's population current live in urban areas and that this number is expected to exceed the 50 per cent mark world wide by 2007.

In developed nations the number of urban dwellers is even more dramatic-expected to exceed 75%. Current studies project Bluetooth-enabled devices to reach 5.4 billion units by 2005 -five times the number of mobile phones or Internet connections. Mobile phone penetration already exceeds 80% of the population in places like the European Union and parts of Asia. WiFi hardware is being deployed at the astonishing rate of one every 4 seconds globally.

We argue that now is the time to initiate inspirational research into the very essence of these newly emerging technological urban spaces. We desire to move towards an improved understanding of the emotional experience of urban life. We are currently conducting a number of Urban Probes - a lightweight, provocative, intervention methodology designed to rapidly deconstruct urban situations, reveal new opportunities for technology in urban spaces, and guide future long term research in urban computing. In fact the very essence of person, place, and community are being redefined by personal wireless digital tools that transcend traditional physical constrains of time and space.

New metaphors for visualizing, interacting, and interpreting the real-time ebb and flow of urban places will emerge. Urban Atmospheres is focused on exposing, deconstructing, and understanding the challenges of this newly emerging moment in urban history and its dramatic influence on technology usage and adoption.

THE PLANETARY PATTERN OF WINDS

Just as there are on this great globe fairly definite zones of *climate*, so there are even more definite belts of *prevailing wind*. Some of these round-the-world prevailing winds are quite steady and uniform, such as the trades and anti-trades. Others, such as the prevailing westerlies of temperate latitudes, are extremely variable from day to day and week to week. But the *average* wind directions observed in temperate latitudes prove that here, too, there is a great round-the-world wind system from west to east—winds that blow generally across continents and oceans, across desert sands, sawtooth mountains, fields of rippling grain, waving trees in virgin forests, and wide expanses of heaving, restless, white-capped blue water. The winds of the world are in essence a heat-pressure mechan- ism similar to that which causes a sea breeze, though enormously larger and more complex. The driving force behind all earthly winds, and all other earthly motions, is the enormous outpouring of radiant energy from the sun, our celestial ancestor and gigantic power-house ninetythree million miles out in space.

This radiant energy falls nearvertically on equatorial regions, and very obliquely on polar regions. Hence the equatorial regions are warmer, and are receiving more heat than they lose by earth-radiation. The poles, on the other hand, are radiating away more heat than they receive by sun-radiation. Hence, unless the tropics are to become unbearably hot and the arctics are to chill down to absolute zero (about -460°F. or -273°C), there must be some atmospheric circulation between the two.

A heated region usually develops low pressure at the surface and high pressure aloft, that a cooled region develops high pressure at the surface and low pressure aloft. The result, where only short distances are involved, is that lower winds blow generally from cold to heat, and upper winds blow generally from heat to cold. Hence we might expect, as part of the planetary wind picture, some tendency towards an equatorward drift at the surface together with a poleward drift aloft.

However, that any largescale wind is gradually turned aside, to the right in the northern hemisphere and to the left in the southern, by the earth's slow rotation under the wind in question. In other words this 'earth's deflective force,' so called, tends to turn any equatorward-drifting air westward, and to turn any poleward-drifting air eastward.

In the upper troposphere, say at altitudes above two and a half miles, the *average* world wind pattern is a grandly simple thing that stems directly and understandably out of these elementary heat-pressure-wind principles. Upper pressure is relatively high over the equator, and relatively low over the poles, with a fairly uniform average horizontal pressure gradient between the two.

Either of these average drifts can be temporarily upset by regional vagaries, such as a southward surge of cold upper air in the north hemisphere. But as a norm or average condition, the two great upper drifts perennially move from the equator spirally eastward and poleward in each hemisphere.

The planetary pattern of lower winds (below the two-and-ahalf-mile level) is more complex, even in idealized form. For the lower winds are influenced by surface friction, and the earth's deflective force acts on them as a more drastic limitation on poleward or equatorward progress. Thus the planetary lowerwind pattern includes, not one wind belt in each hemisphere, but three. Moreover, these wind belts, and the pressure belts which separate them and regulate them, are in fact much modified and distorted (at least in the northern hemisphere) by the alternating surfaces of great continents and great oceans.

Perhaps the easiest and best introduction to them would be a voyage through ideally typical weather at just 5° of latitude (three hundred sea miles) each day down the whole 'length' of the world, along the ocean-parting meridian of long. 30°W from arctic to antarctic. Any such imaginary meridianal voyage would cut across the following idealized or average wind-andweather belts: a zone of arctic easterlies above lat. 60°N a stormy 'polar frontal' zone (actually variable in both width and location) around lat. 60°N or south of this; prevailing southwesterlies-to-westerlies (the familiar prevailing winds of temperate latitudes) between lat. 60°N and lat. 30°N a dry, clear calm belt called the 'horse latitudes' (probably because horses often died here on becalmed sailing ships) around lat. 30°N northeast 'trades' (so called by reason of their steady, perennial character and consequent importance to commercial sailing ships) from south of lat. 30°N down towards the equator: a humid, oppressive, and showery calm belt called the 'doldrums' (because of their dispiriting or dulling or 'dolting' effect on human sailors unwillingly becalmed in them) around the equator; southeast trades from near the equator down to somewhere near lat. 30°S the southern horse latitudes around lat. 30°S; the southern prevailing westerlies between lat. 30°S and lat. 60°S; the southern polar-frontal zone around lat. 60°S or north of here; and the antarctic easterlies south of lat. 60°S.

The steady poleward air transfer of the prevailing westerlies and southwesterlies in our latitudes is compensated by occasional large and violent southward outbreaks of polar air (cold waves), especially in winter, when the polar anti-cyclone extends down over Alaska and northern Canada. Southward outbreaks sometimes also occur far aloft, with 100-m.p.h. north winds in the stratosphere. Similar equatorward (northward) outbreaks occur in the southern hemisphere.

Considering first the northern hemisphere: There is an average high-pressure area, or anti-cyclone, around the 'cold pole' (despite the fact that many arctic observers have noted low pressures). This polar high-pressure area drives the cold, arctic lower air spirally outward (or southward and westward) as the arctic easterlies (mostly northeasterlies) down towards lat. 60°N. Along lat. 30°N (the horse latitudes) there is a subtropical high-pressure belt, the northern slope of which drives the prevailing westerlies (mostly southwesterlies) up towards lat. 60°N.

Around lat. 60°N or south of there, these two great and conflicting wind systems meet along the low-pressure belt of the polar-frontal zone—a stormy region which may actually swing far north or south, and may extend as separate and vari-formed fronts, between various unlike air masses, through a wide sweep of latitude. The subtropical high-pressure belt in lat. 30°N (or its southern slope) also drives the northeasterly trades, as a general southwestward-to-westward drift around the earth, down into the equatorial low-pressure belt of the doldrums.

In the southern hemisphere, the subtropical high-pressure belt around lat. 30°S drives the southeast trades up into the doldrums, and also drives the southern prevailing westerlies (mostly northwesterlies) down into the low-pressure polar-frontal zone around lat. 60°S or north of there, where they meet and battle with the antarctic easterlies (mostly southeasterlies) from the south-polar high-pressure area.

In temperate latitudes, both north and south, lower winds have the same *general* or *average* direction as upper winds, and the prevailing westerlies form a deep *average* pattern (seldom fully duplicated in actual existing winds) that extends with increasing force up through the troposphere, moderating again in the lower stratosphere, and then mostly reversing, in light to moderate intensity, above twelve miles or so. Within the tropics, on the 'winter side' of the equator, the easterly trades give way, at about two and a half miles altitude, to the planetary upper westerlies.

Called the counter-trades or anti-trades, these tropical upper winds blow mostly from southwesterly in the northern hemisphere and northwesterly in the southern. The same planetary upper westerlies are present, to some extent, in arctic regions above the arctic easterlies.

5

Atmospheric Moisture

INTRODUCTION

Everyday, invisible plumes of water vapor circulate through the atmosphere, and when conditions are right, they form clouds and precipitation. To see these plumes, scientists rely on satellite sensors with spectral bands capable of detecting this airborne moisture. This map plots the average total precipitable water in the atmosphere -- a measurement of how much moisture could theoretically precipitate given the right conditions. Notice the bands of moisture along the equatorial regions, which is where most moisture evaporates from the ocean into the air. Also visible are the so-called “atmospheric rivers” that transport moisture from the equatorial regions into the upper latitudes.

ATMOSPHERIC HUMIDITY AND PRECIPITATION

Atmospheric humidity, which is the amount of water vapour or moisture in the air, is another leading climatic element, as is precipitation. All forms of precipitation, including drizzle, rain, snow, ice crystals, and hail, are produced as a result of the condensation of atmospheric moisture that forms clouds in which some of the particles, by growth and aggregation, attain sufficient size to fall from the clouds and reach the ground.

Atmospheric humidity

At 30 °C (86 °F), 4 percent of the volume of the air may be occupied by water molecules, but, where the air is colder than -

40 °C (-40 °F), less than one-fifth of 1 percent of the air molecules can be water. Although the water vapour content may vary from one air parcel to another, these limits can be set because vapour capacity is determined by temperature. Temperature has profound effects upon some of the indexes of humidity, regardless of the presence or absence of vapour.

The connection between an effect of humidity and an index of humidity requires simultaneous introduction of effects and indexes. Vapour in the air is a determinant of weather, because it first absorbs the thermal radiation that leaves and cools Earth's surface and then emits thermal radiation that warms the planet. Calculation of absorption and emission requires an index of the mass of water in a volume of air. Vapour also affects the weather because, as indicated above, it condenses into clouds and falls as rain or other forms of precipitation. Tracing the moisture-bearing air masses requires a humidity index that changes only when water is removed or added.

Absolute humidity is the vapour concentration or density in the air. If mv is the mass of vapour in a volume of air, then absolute humidity dv is simply dv = mv/ V, in which V is the volume and dv is expressed in grams per cubic metre. This index indicates how much vapour a beam of radiation must pass through. The ultimate standard in humidity measurement is made by weighing the amount of water gained by an absorber when a known volume of air passes through it; this measures absolute humidity, which may vary from 0 gram per cubic metre in dry air to 30 grams per cubic metre (0.03 ounce per cubic foot) when the vapour is saturated at 30 °C. The dv of a parcel of air changes, however, with temperature or pressure even though no water is added or removed, because, as the gas equation states, the volume V increases with the absolute, or Kelvin, temperature and decreases with the pressure.

Specific humidity

The meteorologist requires an index of humidity that does not change with pressure or temperature. A property of this sort will identify an air mass when it is cooled or when it rises to lower pressures aloft without losing or gaining water vapour. Because all the gases will expand equally, the ratios of the weight of water to the weight of dry air, or the dry air plus vapour, will be conserved during such changes and will continue identifying the air mass.

The mixing ratio r is the dimensionless ratio r = mv/ ma, where ma is the mass of dry air, and the specific humidity q is another dimensionless ratio q = mv/ (ma + mv). Because mv is less than 3 percent of ma at normal pressure and temperatures cooler than 30 °C, r and q are practically equal. These indexes are usually expressed in grams per kilogram because they are so small; the values range from 0 grams per kilogram in dry air to 28 grams per kilogram in saturated air at 30 °C. Absolute and specific humidity indexes have specialized uses, so they are not familiar to most people.

Relative humidity

Relative humidity (U) is so commonly used that a statement of humidity, without a qualifying adjective, can be assumed to be relative humidity. U can be defined, then, in terms of the mixing ratio r that was introduced above. U = 100r/ rw, which is a dimensionless percentage. The divisor rw is the saturation mixing ratio, or the vapour capacity. Relative humidity is therefore the water vapour content of the air relative to its content at saturation. Because the saturation mixing ratio is a function of pressure, and especially of temperature, the relative humidity is a combined index of the environment that reflects more than water content. In many climates the relative humidity rises to about 100 percent at dawn and falls to 50 percent by noon. A relative humidity of 50 percent may reflect many different quantities of vapour per volume of air or gram of air, and it will not likely be proportional to evaporation.

An understanding of relative humidity thus requires a knowledge of saturated vapour, which will be discussed later in the section on the relation between temperature and humidity. At this point, however, the relation between U and the absorption and retention of water from the air must be considered. Small pores retain water more strongly than large pores; thus, when a porous material is set out in the air, all pores larger than a certain size (which can be calculated from the relative humidity of the air) are dried out.

The water content of a porous material at air temperature is fairly well indicated by the relative humidity. The complexity of actual pore sizes and the viscosity of the water passing through them makes the relation between U and moisture in the porous material imperfect and slowly achieved. The great suction also strains the walls of the capillaries, and the consequent shrinkage is used to measure relative humidity.

The absorption of water by salt solutions is also related to relative humidity without much effect of temperature. The air above water saturated with sodium chloride is maintained at 75 to 76 percent relative humidity at a temperature between 0 and 40 °C (32 and 104 °F).

In effect, relative humidity is a widely used environmental indicator, but U does respond drastically to changes in temperatures as well as moisture, a response caused by the effect of temperature upon the divisor rW in U.

Humidity and climate

The small amount of water in atmospheric vapour, relative to water on Earth, belies its importance. Compared with one unit of water in the air, the seas contain at least 100,000 units, the great glaciers 1,500, the porous earth nearly 200, and the rivers and lakes 4 or 5. The effectiveness of the vapour in the air is magnified, however, by its role in transferring water from sea to land by the media of clouds and precipitation and that in absorbing radiation.

The vapour in the air is the invisible conductor that carries water from sea to land, making terrestrial life possible. Fresh water is distilled from the salt seas and carried over land by the wind. Water evaporates from vegetation, and rain falls on the sea too, but the sea is the bigger source, and rain that falls on land is most important to humans. The invisible vapour becomes visible near the surface as fog when the air cools to the dew point. The usual nocturnal cooling will produce fog patches in cool valleys.

Or the vapour may move as a tropical air mass over cold land or sea, causing widespread and persistent fog, such as occurs over the Grand Banks off Newfoundland. The delivery of water by means of fog or dew is slight, however.

When air is lifted, it is carried to a region of lower pressure, where it will expand and cool as described by the gas equation. It may rise up a mountain slope or over the front of a cooler, denser air mass. If condensation nuclei are absent, the dew point may be exceeded by the cooling air, and the water vapour becomes supersaturated. If nuclei are present or if the temperature is very low, however, cloud droplets or ice crystals form, and the vapour is no longer in the invisible guise of atmospheric humidity.

The invisible vapour has another climatic role—namely, absorbing and emitting radiation. The temperature of Earth and its daily variation are determined by the balance between incoming and outgoing radiation. The wavelength of the incoming radiation from the Sun is mostly shorter than 3 μm (0.0001 inch). It is scarcely absorbed by water vapour, and its receipt depends largely upon cloud cover. The radiation exchanged between the atmosphere and Earth's surface and the eventual loss to space is in the form of long waves. These long waves are strongly absorbed in the 3- to 8.5-μm band and in the greater than 11-μm range, where vapour is either partly or wholly opaque. As noted above, much of the radiation that is absorbed in the atmosphere is emitted back to Earth, and the surface receipt of long waves, primarily from water vapour and carbon dioxide in the atmosphere, is slightly more than twice the direct receipt of solar radiation at the surface. Thus, the invisible vapour in the atmosphere combines with clouds and the advection (horizontal movement) of air from different regions to control the surface temperature.

The world distribution of humidity can be portrayed for different uses by different indexes. To appraise the quantity of water carried by the entire atmosphere, the moisture in an air column above a given point on Earth is expressed as a depth of liquid water. It varies from 0.5 mm (0.02 inch) over the Himalayas and 2 mm (0.08 inch) over the poles in winter to 8 mm (0.3 inch) over the Sahara, 54 mm (2 inches) in the Amazon region, and 64 mm (2.5 inches) over India during the wet season. During summer the air over the United States transports 16 mm (0.6 inch) of water vapour over the Great Basin and 45 mm (1.8 inches) over Florida.

The humidity of the surface air may be mapped as vapour pressure, but a map of this variable looks much like that of temperature. Warm places are moist, and cool ones are dry; even in deserts the vapour pressure is normally 13 mb (0.01 standard atmosphere), whereas over the northern seas it is only about 4 mb (0.004 standard atmosphere). Certainly the moisture in materials in two such areas will be just the opposite, so relative humidity is a more widely useful index.

Average relative humidity

The average relative humidity for July reveals the humidity provinces of the Northern Hemisphere when aridity is at a maximum. At other times the relative humidity generally will be higher. The humidities over the Southern Hemisphere in July indicate the humidities that comparable regions in the Northern Hemisphere will attain in January, just as July in the Northern Hemisphere suggests the humidities in the Southern Hemisphere during January. A contrast is provided by comparing a humid cool coast to a desert. The midday humidity on the Oregon coast, for example, falls only to 80 percent, whereas in the Nevada desert it falls to 20 percent. At night the contrast is less, with averages being over 90 and about 50 percent, respectively.

Although the dramatic regular decrease of relative humidity from dawn to midday has been attributed largely to warming rather than declining vapour content, the content does vary regularly. In humid environments, daytime evaporation increases the water vapour content of the air, and the mixing ratio, which may be about 12 grams per kilogram, rises by 1 or 2 grams per kilogram in temperate places and may attain 16 grams per kilogram in a tropical rainforest. In arid environments, however, little evaporation moistens the air, and daytime turbulence tends to bring down dry air; this decreases the mixing ratio by as much as 2 grams per kilogram.

Humidity also varies regularly with altitude. On the average, fully half the water in the atmosphere lies below 0.25 km (about 0.2 mile), and satellite observations over the United States in April revealed 1 mm (0.04 inch) or less of water in all the air above 6 km (4 miles). A cross section of the atmosphere along 75° W longitude shows a decrease in humidity with height and toward the poles. The mixing ratio is 16 grams per kilogram just north of the Equator, but it decreases to 1 gram per kilogram at 50° N latitude or 8 km (5 miles) above the Equator. The transparent air surrounding mountains in fair weather is very dry indeed.

Closer to the ground, the water vapour content also changes with height in a regular pattern. When water vapour is condensing on Earth's surface at night, the content is greater aloft than at the ground; during the day the content is, in most cases, less aloft than at the ground because of evaporation.

ATMOSPHERIC MODEL

An atmospheric model is a mathematical model constructed around the full set of primitive dynamical equations which govern atmospheric motions. It can supplement these equations with parameterizations for turbulent diffusion, radiation, moist processes (clouds and precipitation), heat exchange, soil, vegetation, surface water, the kinematic effects of terrain, and convection. Most atmospheric models are numerical, i.e. they discretize equations of motion.

They can predict microscale phenomena such as tornadoes and boundary layer eddies, sub-microscale turbulent flow over buildings, as well as synoptic and global flows. The horizontal domain of a model is either *global*, covering the entire Earth, or *regional* (*limited-area*), covering only part of the Earth. The different types of models run are thermotropic, barotropic, hydrostatic, and nonhydrostatic. Some of the model types make assumptions about the atmosphere which lengthens the time steps used and increases computational speed.

Forecasts are computed using mathematical equations for the physics and dynamics of the atmosphere. These equations are nonlinear and are impossible to solve exactly. Therefore, numerical methods obtain approximate solutions. Different models use different solution methods. Global models often use spectral methods for the horizontal dimensions and finite-difference methods for the vertical dimension, while regional models usually use finite-difference methods in all three dimensions. For specific locations, model output statistics use climate information, output from numerical weather prediction, and current surface weather observations to develop statistical relationships which account for model bias and resolution issues.

Types

The main assumption made by the thermotropic model is that while the magnitude of the thermal wind may change, its direction does not change with respect to height, and thus the baroclinicity in the atmosphere can be simulated using the 500 mb (15 inHg) and 1,000 mb (30 inHg) geopotential height surfaces and the average thermal wind between them.

Barotropic models assume the atmosphere is nearly barotropic, which means that the direction and speed of the geostrophic wind are independent of height. In other words, no vertical wind shear of the geostrophic wind. It also implies that thickness contours (a proxy for temperature) are parallel to upper level height contours. In this type of atmosphere, high and low pressure areas are centres of warm and cold temperature anomalies. Warm-core highs (such as the subtropical ridge and Bermuda-Azores high) and cold-core lows have strengthening winds with height, with the reverse true for cold-core highs (shallow arctic highs) and warm-core lows (such as tropical cyclones). A barotropic model tries to solve a simplified form of atmospheric dynamics based on the assumption that the atmosphere is in geostrophic balance; that is, that the Rossby number of the air in the atmosphere is small. If the assumption is made that the atmosphere is divergence-free, the curl of the Euler equations reduces into the barotropic vorticity equation. This latter equation can be solved over a single layer of the atmosphere. Since the atmosphere at a height of approximately 5.5 kilometres (3.4 mi) is mostly divergence-free, the barotropic model best approximates the state of the atmosphere at a geopotential height corresponding to that altitude, which corresponds to the atmosphere's 500 mb (15 inHg) pressure surface.

Hydrostatic models filter out vertically moving acoustic waves from the vertical momentum equation, which significantly increases the time step used within the model's run. This is known as the hydrostatic approximation. Hydrostatic models use either pressure or sigma-pressure vertical coordinates. Pressure coordinates intersect topography while sigma coordinates follow the contour of the land. Its hydrostatic assumption is reasonable as long as horizontal grid resolution is not small, which is a scale where the hydrostatic assumption fails. Models which use the entire vertical momentum equation are known as nonhydrostatic. A nonhydrostatic model can be solved anelastically, meaning it solves the complete continuity equation for air, or elastically, meaning it solves the complete continuity equation for air and is fully compressible. Nonhydrostatic models use altitude or sigma altitude for their vertical coordinates. Altitude coordinates can intersect land while sigma-altitude coordinates follow the contours of the land.

History

The history of numerical weather prediction began in the 1920s through the efforts of Lewis Fry Richardson who utilized procedures developed by Vilhelm Bjerknes. It was not until the advent of the computer and computer simulation that computation time was reduced to less than the forecast period itself. ENIAC created the first computer forecasts in 1950, and more powerful computers later increased the size of initial datasets and included more complicated versions of the equations of motion. In 1966, West Germany and the United States began producing operational forecasts based on primitive-equation models, followed by the United Kingdom in 1972 and Australia in 1977. The development of global forecasting models led to the first climate models. The development of limited area (regional) models facilitated advances in forecasting the tracks of tropical cyclone as well as air quality in the 1970s and 1980s.

Because the output of forecast models based on atmospheric dynamics requires corrections near ground level, model output statistics (MOS) were developed in the 1970s and 1980s for individual *forecast points* (locations). Even with the increasing power of supercomputers, the forecast skill of numerical weather models only extends to about two weeks into the future, since the density and quality of observations—together with the chaotic nature of the partial differential equations used to calculate the forecast—introduce errors which double every five days. The use of model ensemble forecasts since the 1990s helps to define the forecast uncertainty and extend weather forecasting farther into the future than otherwise possible.

Initialization

The atmosphere is a fluid. As such, the idea of numerical weather prediction is to sample the state of the fluid at a given time and use the equations of fluid dynamics and thermodynamics to estimate the state of the fluid at some time in the future. On land, terrain maps, available at resolutions down to 1 kilometre (0.62 mi) globally, are used to help model atmospheric circulations within regions of rugged topography, in order to better depict features such as downslope winds, mountain waves, and related cloudiness which affects incoming solar radiation. The main inputs from country-based weather services are surface observations from automated weather stations at ground level over land and from weather buoys at sea. The World Meteorological Organization acts to standardize the instrumentation, observing practices and timing of these observations worldwide. Stations either report hourly in METAR reports, or every six hours in SYNOP reports. Models are *initialized* using this observed data. The irregularly spaced observations are processed by data assimilation and objective analysis methods, which perform quality control and obtain values at locations usable by the model's mathematical algorithms. The grid used for global models is geodesic or icosahedral, spaced by latitude, longitude, and elevation. The data are then used in the model as the starting point for a forecast.

A variety of methods are used to gather observational data for use in numerical models. Sites launch radiosondes, which rise through the troposphere and well into the stratosphere. Information from weather satellites is used where traditional data sources are not available. Commerce provides pilot reports along aircraft routes and ship reports along shipping routes. Research projects use reconnaissance aircraft to fly in and around weather systems of interest, such as tropical cyclones. Reconnaissance aircraft are also flown over the open oceans during the cold season into systems which cause significant uncertainty in forecast guidance, or are expected to be of high impact 3–7 days into the future over the downstream continent. Sea ice began to be initialized in forecast models in 1971. Efforts to involve sea surface temperature in model initialization began in 1972 due to its role in modulating weather in higher latitudes of the Pacific.

Computation

A model is a computer program that produces meteorological information for future times at given locations and altitudes. Within any model is a set of equations, known as the primitive equations, used to predict the future state of the atmosphere. These equations are initialized from the analysis data and rates of change are determined. These rates of change predict the state of the atmosphere a short time into the future, with each time increment known as a time step. The equations are then applied to this new atmospheric state to find new rates of change, and these new rates of change predict the atmosphere at a yet further time into the future. *Time stepping* is repeated until the solution reaches the desired forecast time. The length of the time step chosen within the model is related to the distance between the points on the computational grid, and is chosen to maintain numerical stability. Time steps for global models are on the order of tens of minutes, while time steps for regional models are between one and four minutes. The global models are run at varying times into the future. The UKMET Unified model is run six days into the future, the European Centre for Medium-Range Weather Forecasts model is run out to 10 days into the future, while the Global Forecast System model run by the Environmental Modeling Centre is run 16 days into the future.

The equations used are nonlinear partial differential equations which are impossible to solve exactly through analytical methods, with the exception of a few idealized cases. Therefore, numerical methods obtain approximate solutions. Different models use different solution methods: some global models use spectral methods for the horizontal dimensions and finite difference methods for the vertical dimension, while regional models and other global models usually use finite-difference methods in all three dimensions. The visual output produced by a model solution is known as a prognostic chart, or *prog*.

Parameterization

Weather and climate model gridboxes have sides of between 5 kilometres (3.1 mi) and 300 kilometres (190 mi). A typical cumulus cloud has a scale of less than 1 kilometre (0.62 mi), and would require a grid even finer than this to be represented physically by the equations of fluid motion. Therefore the processes that such clouds represent are *parameterized*, by processes of various sophistication. In the earliest models, if a column of air in a model gridbox was unstable (i.e., the bottom warmer than the top) then it would be overturned, and the air in that vertical column mixed. More sophisticated schemes add enhancements, recognizing that only some portions of the box might convect and that entrainment and other processes occur. Weather models that have gridboxes with sides between 5 kilometres (3.1 mi) and 25 kilometres (16 mi) can explicitly represent convective clouds, although they still need to parameterize cloud microphysics. The formation of large-scale (stratus-type) clouds is more physically based, they form when the relative humidity reaches some prescribed value. Still, sub grid scale processes need to be taken into account. Rather than assuming that clouds form at 100% relative humidity, the cloud fraction can be related to a critical relative humidity of 70% for stratus-type clouds, and at or above 80% for cumuliform clouds, reflecting the sub grid scale variation that would occur in the real world.

The amount of solar radiation reaching ground level in rugged terrain, or due to variable cloudiness, is parameterized as this process occurs on the molecular scale. Also, the grid size of the models is large when compared to the actual size and roughness of clouds and topography. Sun angle as well as the impact of multiple cloud layers is taken into account. Soil type, vegetation type, and soil moisture all determine how much radiation goes into warming and how much moisture is drawn up into the adjacent atmosphere. Thus, they are important to parameterize.

Domains

The horizontal domain of a model is either *global*, covering the entire Earth, or *regional*, covering only part of the Earth. Regional models also are known as *limited-area* models, or LAMs. Regional models use finer grid spacing to resolve explicitly smaller-scale meteorological phenomena, since their smaller domain decreases computational demands. Regional models use a compatible global model for initial conditions of the edge of their domain. Uncertainty and errors within LAMs are introduced by the global model used for the boundary conditions of the edge of the regional model, as well as within the creation of the boundary conditions for the LAMs itself.

The vertical coordinate is handled in various ways. Some models, such as Richardson's 1922 model, use geometric height (z) as the vertical coordinate. Later models substituted the geometric z coordinate with a pressure coordinate system, in which the geopotential heights of constant-pressure surfaces become dependent variables, greatly simplifying the primitive equations. This follows since pressure decreases with height through the Earth's atmosphere. The first model used for operational forecasts, the single-layer barotropic model, used a single pressure coordinate at the 500-millibar (15 inHg) level, and thus was essentially two-dimensional. High-resolution models—also called *mesoscale models*—such as the Weather Research and Forecasting model tend to use normalized pressure coordinates referred to as *sigma coordinates*. This coordinate system receives that name since the independent variable σ is used to represent a pressure level (p) scaled with the surface pressure (p_0) and in some cases the pressure at the top of the domain (p_T).

HISTORY OF THE ATMOSPHERE

Earliest Atmosphere

The outgassings of the Earth was stripped away by solar winds early in the history of the planet until a steady state was established, the first atmosphere.

Based on today's volcanic evidence, this atmosphere would have contained 60% hydrogen, 20% oxygen (mostly in the form of water vapour), 10% carbon dioxide, 5 to 7% hydrogen sulfide, and smaller amounts of nitrogen, carbon monoxide, free hydrogen, methane and inert gases. A major rainfall led to the buildup of a vast ocean, enriching the other agents, first carbon dioxide and later nitrogen and inert gases.

A major part of carbon dioxide exhalations were soon dissolved in water and built up carbonate sediments.

Second Atmosphere

As early as 3.8 billion years ago, water related sediments have been found. About 3.4 billion years ago, nitrogen was the major part of the then stable second atmosphere. An influence of life has to be taken into account rather soon since hints on early life forms are to be found as early as 3.5 billion years ago. The fact that this is not in line with the — compared to today 30% lower — solar radiance of the early sun has been described as the faint young Sun paradox.

The geological record, however, shows a continually relatively warm surface during the complete early temperature record of the earth with the exception of one cold glacial phase about 2.4 billion years ago.

In the late Archean Era an oxygen containing atmosphere began to develop from photosynthesizing algae. The early basic carbon isotopy is very much in line with what is found today As Jan Veizer assumed that not only did we have life as far back as we had rocks, but there was as much life then as today and the fundamental features of the carbon cycle were established as early as 4 billion years ago.

Third Atmosphere

The accretion of continents about 3.5 billion years ago added plate tectonics, constantly rearranging the continents and also shaping long-term climate evolution by allowing the transfer of carbon dioxide to large land-based carbonate storages. Free oxygen did not exist until about 1.7 billion years ago and this can be seen with the development of the red beds and the end of the banded iron formations.

This signifies a shift from a reducing atmosphere to an oxidising atmosphere. O_2 showed major ups and downs until reaching a steady state of more than 15%. The following time span was the Phanerozoic, during which oxygen-breathing metazoan life forms began to appear.

CLIMATE DURING GEOLOGICAL AGES

Precambrian Climate

In the first three quarters of the Earth's history, only one major glaciation is to be found in the geological record. Since about 950 million years ago, the Earth's climate has varied regularly between large-scale or just polar cap wide glaciation and extensively tropical climates. The time scale for this variation is roughly 140 million years and may be related to Earth's motion into and out of galactic spiral arms and compared to the previous time, significantly reduced solar wind. The climate of the late Precambrian showed some major glaciation events spreading over much of the earth. At this time the continents were bunched up in the Rodinia supercontinent. Massive deposits of tillites are found and anomalous isotopic signatures are found, which gave rise to the Snowball Earth hypothesis. As the Proterozoic Eon drew to a close, the Earth started to warm up. By the dawn of the Cambrian and the Phanerozoic, life forms were abundant in the Cambrian explosion with average global temperatures of about 22 °C.

Phanerozoic Climate

Major drivers for the preindustrial ages have been variations of the sun, volcanic ashes and exhalations, relative movements of the earth towards the sun and tectonically induced effects as for major sea currents, watersheds and ocean oscillations. In the early Phanerozoic, increased atmospheric carbon dioxide concentrations have been linked to driving or amplifying increased global temperatures. Royer et al. 2004 found a climate sensitivity for the rest of the Phanerozoic which was calculated to be similar to today's modern range of values.

The difference in global mean temperatures between a fully glacial Earth and an ice free Earth is estimated at approximately 10 °C, though far larger changes would be observed at high latitudes and smaller ones at low latitudes.

One requirement for the development of large scale ice sheets seems to be the arrangement of continental land masses at or near the poles.

The constant rearrangement of continents by plate tectonics can also shape long-term climate evolution. However, the presence or absence of land masses at the poles is not sufficient to guarantee glaciations or exclude polar ice caps.

Evidence exists of past warm periods in Earth's climate when polar land masses similar to Antarctica were home to deciduous forests rather than ice sheets.

The relatively warm local minimum between Jurassic and Cretaceous goes along with widespread tectonic activity, e.g. the breakup of supercontinents. Superimposed on the long-term evolution between hot and cold climates have been many short-term fluctuations in climate similar to, and sometimes more severe than, the varying glacial and interglacial states of the present ice age. Some of the most severe fluctuations, such as the Paleocene-Eocene Thermal Maximum, may be related to rapid climate changes due to sudden collapses of natural methane clathrate reservoirs in the oceans.

A similar, single event of induced severe climate change after a meteorite impact has been proposed as reason for the Cretaceous-Tertiary extinction event. Other major thresholds are the Permian-Triassic, and Ordovician-Silurian extinction events with various reasons suggested.

Quaternary Sub-era

The Quaternary sub-era includes the current climate. There has been a cycle of ice ages for the past 2.2–2.1 million years (starting before the Quaternary in the late Neogene Period).

Note in the graphic on the right the strong 120,000-year periodicity of the cycles, and the striking asymmetry of the curves. This asymmetry is believed to result from complex interactions of feedback mechanisms. It has been observed that ice ages deepen by progressive steps, but the recovery to interglacial conditions occurs in one big step.

6

Stable and Unstable Atmosphere

STABILITY OF AIR

Adiabatic temperature change is an important factor in determining the stability of the air. We can think of air stability as the tendency for air to rise or fall through the atmosphere under its own "power". Stable air has a tendency to resist movement. On the other hand, unstable air will easily rise. What gives air "power" to rise? The tendency for air to rise or fall depends on the adiabatic and environmental lapse rates.

Rising air cools, sinking air warms- no exceptions!

The actual stability of an air parcel is determined by the orientation of the environmental lapse rate in comparison with either the dry or moist adiabatic lapse rates. The environmental lapse rate is simply what it says- the rate of change of the temperature of the environment (atmosphere) with changing altitude.

It is important to relaize that because the atmosphere (environment), on average, is not rising or sinking, the environmental lapse rate can look much different than the dry or moist adiabatic lapse rates. In fact, it is those differences that allow us to determine whether a particular part of the atmosphere is stable or unstable.

Atmospheric Stability & Lapse Rates

Adiabatic Processes

When discussing stability in atmospheric sciences, we typically think about air parcels, or imaginary blobs of air that can expand and contract freely, but do not mix with the air around them or break apart. The key piece of information is that movement of air parcels in the atmosphere can be estimated as an adiabatic process. Adiabatic processes do not exchange heat and they are reversible.

Imagine you have a parcel of air at the Earth's surface. The air parcel has the same temperature and pressure as the surrounding air, which we will call the environment. If you were to lift the air parcel, it would find itself in a place where the surrounding environmental air pressure is lower, because we know that pressure decreases with height. Because the environmental air pressure outside the parcel is lower than the pressure inside the parcel, the air molecules inside the parcel will effectively push outward on the walls of the parcel and expand adiabatically. The air molecules inside the parcel must use some of their own energy in order to expand the air parcel's walls, so the temperature inside the parcel decreases as the internal energy decreases. To summarize, rising air parcels expand and cool adiabatically without exchanging heat with the environment.

Now imagine that you move the same air parcel back to Earth's surface. The air parcel is moving into an environment with higher air pressure. The higher environmental pressure will push inward on the parcel walls, causing them to compress, and raise the inside temperature.

The process is adiabatic, so again, no heat is exchanged with the environment. However, temperature changes in the air parcel can still occur, but it is not due to mixing, it is due to changes in the internal energy of the air parcel.

DETERMINING STABILITY

How do you know if an air parcel will be stable after some initial displacement? Stability is determined by comparing the temperature of a rising or sinking air parcel to the environmental air temperature. Imagine the following: at some initial time, an air parcel has the same temperature and pressure as its environment. If you lift the air parcel some distance, its temperature drops by 9.8 K·km-1, which is the dry adiabatic lapse rate. If the air parcel is colder than the environment in its new position, it will have higher density and tend to sink back to its original position. In this case, the air is stable because vertical motion is resisted. If the rising air is warmer and less dense than the surrounding air, it will continue to rise until it reaches some new equilibrium where its temperature matches the environmental temperature. In this case, because an initial change is amplified, the air parcel is unstable. In order to figure out if the air parcel is unstable or not we must know the temperature of both the rising air and the environment at different altitudes.

One way this is done in practice is with a weather balloon. We can get a vertical profile of the environmental lapse rate by releasing a radiosonde attached to a weather balloon. A radiosonde sends back data on temperature, humidity, wind, and position, which are plotted on a thermodynamic diagram. This vertical plot of temperature and other variables is known as a sounding.

Dry Stability

If an air parcel is dry, meaning unsaturated, stability is relatively straightforward. An atmosphere where the environmental lapse rate is the same as the dry adiabatic lapse rate, meaning that the temperature in the environment also drops by 9.8 K·km-1, will be considered neutrally stable. After some initial vertical displacement, the temperature of the air parcel will always be the same as the environment so no further change in position is expected.

If the environmental lapse rate is less than the dry adiabatic lapse rate, some initial vertical displacement of the air parcel will result in the air parcel either being colder than the environment (if lifted), or warmer than the environment (if pushed downward). This is because if lifted, the temperature of the air parcel would drop more than the temperature of the environment. This is a stable situation for a dry air parcel and a typical scenario in the atmosphere. The global average tropospheric lapse rate is 6.5 K·km-1, which is stable for dry lifting.

STABILITY & CLOUD DEVELOPMENT

Cloud development is linked closely with the concept of stability, i.e., the tendency of air to rise. Although several factors determine whether or not clouds will form, the stability of the atmosphere is far and away the single greatest indicator of cloud formation.

Air tends to cool and condense as it rises, and to become warm and dry as it sinks. A Parcel of Air is an imaginary mass of air that doesn't exchange properties with surrounding air masses. In reality air masses do exchange properties, but this often occurs very slowly, especially if the air masses are large. An adiabatic process is one where no heat is exchanged between an air parcel and the surrounding air. When we talk about an adiabatic process in the current context we are talking about a rising (or sinking) parcel of air that is not exchanging any heat with its surroundings. When air rises it cools at a relatively constant rate. If the air is unsaturated, this rate, called the dry adiabatic rate, is 10°C per 1000m (5.5°F per 1000ft), i.e., a parcel of unsaturated air cools by 100C every 1000 meters if it doesn't exchange heat with its surroundings.

As air rises and cools its relative humidity increases. At some point the dew point equals the air temperature and the air becomes saturated. Further lifting results in condensation and cloud formation with an accompanying release of latent heat into the rising parcel of air (remember that condensation is a warming process). Because the heat liberated by condensation partially offsets the cooling due to expansion, the parcel now cools at a lesser rate as it rises. This rate is known as the moist adiabatic rate. The moist adiabatic rate applies to saturated air.

On average, the moist adiabatic rate is less than the dry adiabatic rate. The moist adiabatic rate is not constant but varies with temperature and moisture content. For cool air the moist adiabatic rate ~ dry adiabatic rate. For warmer air the moist adiabatic rate is less than the dry adiabatic rate. An average value of 6°C per 1000m (3.3°F per 1000ft) is commonly used.

To determine the stability of an air parcel, one compares its temperature to the temperature of the surrounding air mass. If the air parcel's temperature is less than the temperature of the surrounding air mass, it is denser than the surrounding air and therefore has a tendency to sink. Air that has a tendency to sink is known as a stable air. If the air parcel's temperature is greater than the temperature of the surrounding air mass, the air parcel is less dense and tends to rise. Rising air, as we have already learned, is known as unstable.

For stable air, the environmental lapse rate is 4°C per 1000m (2°F per 1000ft). When the environmental lapse rate is less than the moist adiabatic rate an air parcel cools more quickly than the surrounding air mass. This is known as absolute stability. ln this case the air parcel strongly resists lifting. If the parcel is forced to lift by mechanical means (such as orographic uplift or uplift along a frontal boundary), it will spread out horizontally. Any clouds that form as a result will be thin and horizontal such as cirrostratus, altostratus, nimbostratus, and stratus clouds. All of these cloud types are associated with stable air.

Since the moist adiabatic rate must be less than the environmental lapse rate for stable conditions to exist, a moderate to small environmental lapse rate enhances stability in the atmosphere. Warm air aloft (caused by warm advection) and cool air at the surface (caused by nighttime radiational cooling, cold advection, or a cold surface) result in a moderate to small environmental lapse rate.

Fog and haze form in stable atmospheric conditions because of the large scale sinking of air. This can form an inversion condition, known as a subsidence inversion.Subsidence inversions are often associated with large high-pressure systems. Inversions are absolutely stable because the air beneath the inversion is physically impeded from moving upward. This traps large numbers of particulates close to the ground that serve as fog-forming condensation nuclei.

Neutral Stability is an atmospheric condition that occurs when the environmental lapse rate is equal to the dry adiabatic rate. Absolute instability occurs when dry adiabatic rate is less than the environmental lapse rate. In this situation, an air parcel will be warmer and less dense than the surrounding air and will rise due to buoyant forces. Clouds with extensive vertical development are indicative of absolute instability.

Conditional instability is a state of instability that depends upon whether or not the rising air is saturated. Conditional stability occurs when the environmental lapse rate is between the moist and dry adiabatic rates. The atmosphere is normally in a conditionally unstable state.

Many factors lead to instability. One is a steep environmental lapse rate resulting from cool air aloft (brought on by cold advection, the environmental lapse rate or both) coupled with warm air at the surface (caused by daytime solar heating, warm advection, or a warm surface).

Mixing is another factor that affects instability. Mixing increases warming below and cooling higher up in the atmosphere. Another factor that enhances instability is lifting. When a layer of air is forced to rise it tends to become more unstable because the top layer cools more rapidly than the bottom. This steepens the environmental lapse rate. This effect is enhanced even more when the lower layer of the lifted parcel is moist and the upper layer is dry. In this case, less lifting is require to steepen the environmental lapse rate. This is referred to as convective instability and is associated with severe storms.

UNSTABLE AIR

Air is unstable when the environmental lapse rate is greater than the dry adiabatic rate. Under these conditions, a rising parcel of air is warmer and less dense than the air surrounding it at any given elevation.

THE VERTICAL STRUCTURE OF THE ATMOSPHERE

Most of the gaseous constituents are well mixed throughout the atmosphere. However, the atmosphere itself is not physically uniform but has significant variations in temperature and pressure with altitude. shows the structure of the atmosphere, in which a series of layers is defined by reversals of temperature. The lowest layer, often referred to as the lower atmosphere, is called the troposphere. It ranges in thickness from 8km at the poles to 16km over the equator, mainly as the result of the different energy budgets at these locations. Although variations do occur, the average decline in temperature with altitude (known as the lapse rate) is approximately 6.5°C per kilometre. The troposphere contains up to 75% of the gaseous mass of the atmosphere, as well as nearly all of the water vapour and aerosols, whilst 99% of the mass of the atmosphere lies within the lowest 30km. Owing to the temperature structure of the troposphere, it is in this region of the atmosphere where most of the world's weather systems develop. These are partly driven by convective processes that are established as warm surface air (heated by the Earth's surface) expands and rises before it is cooled at higher levels in the troposphere.

The tropopause marks the upper limit of the troposphere, above which temperatures remain constant before starting to rise again above about 20km. This temperature inversion prevents further convection of air, thus confining most of the world's weather to the troposphere. The layer above the tropopause in which temperatures start to rise is known as the stratosphere. Throughout this layer, temperatures continue to rise to about an altitude of 50km, where the rarefied air may attain temperatures close to 0°C. This rise in temperature is caused by the absorption of solar ultraviolet radiation by the ozone layer. Such a temperature profile creates very stable conditions, and the stratosphere lacks the turbulence that is so prevalent in the troposphere.

The stratosphere is capped by the stratopause, another temperature inversion occurring at about 50km. Above this lies the mesosphere up to about 80km through which temperatures fall again to almost -100°C. Above 80km temperatures rise continually (the thermosphere) to well beyond 1000°C, although owing to the highly rarefied nature of the atmosphere at these heights, such values are not comparable to those of the troposphere or stratosphere.

Radiation Laws

The Earth's atmosphere has an important influence on the energy budget of the global climate system. This is determined by the thermodynamic processes involved in solar and terrestrial energy transfers. The Earth's principal source of energy is the Sun, which produces electromagnetic radiation from nuclear fusion reactions involving hydrogen in its core. Radiation emitted from its surface has a temperature of approximately 5800 Kelvin (K). The radiation is emitted over a spectrum of wavelengths, with a specific quantity of energy for each wavelength, calculated using Planck's Law:

$$E_{\lambda} = a/ [\lambda^{5}\{e^{(b/\lambda T)} - 1\}]$$

Where E_{λ} is the amount of energy ($Wm^{-2}\ \mu m^{-1}$) emitted at wavelength λ (μm) by a body at temperature T (K), with a and b as constants (Henderson-Sellers and Robinson, 1986). This assumes that the Sun is a perfectly radiating (black) body.

By differentiating Equation (Planck's Law), it is possible to determine the wavelength of maximum radiation emission from the Sun:

$$\lambda = 2897/ T$$

This is Wien's Law and for T = 5800K (the solar surface temperature) the wavelength of maximum energy is approximately 0.5μm. This represents radiation in the visible part of the spectrum.

By integrating Equation 1, one can determine the total energy emitted by the Sun, given by the Stefan-Boltzmann Law:

$$E_{Total} = \sigma T^{4}$$

where σ is the Stefan-Boltzmann constant. Solving Equation 3 for the solar temperature of 5800K reveals a total energy output of about 64 million Wm^{-2}. The solar radiation disperses uniformly in all directions. After travelling 93 million miles only a tiny fraction of the energy emitted by the Sun is intercepted by the Earth †. Therefore, the energy flux arriving at the top of the Earth's atmosphere is many orders of magnitude smaller than that leaving the Sun. The latest satellite measurements indicate a value of $1368Wm^{-2}$ for the energy received at the top of the atmosphere on a surface perpendicular to the solar beam. This is known as the solar constant.

COMPONENTS OF ATMOSPHERE

Troposphere and stratosphere

At about the close of the last century, soundings of the atmosphere were begun with the aid of small balloons carrying light devices that automatically registered the temperature and pressure throughout both the ascent and the descent—a kind of exploration that soon was taken up at various places. From these records, in turn, the heights corresponding to given points on the traces were readily computed.

In this way we gradually have come to know the average temperature and pressure of the air at every height from the surface of the earth up to at least 12 to 15 miles, under different weather conditions, for all the seasons, and in many parts of the world. Pretty soon a means of registering the humidity was further added to the apparatus carried, so that we now have a fair knowledge also of the average vertical distribution of water vapour that corresponds to each particular type of weather. Obviously, too, these sounding balloons, as they are called, afford some knowledge of the direction and velocity of the wind at various levels for the particular time and place at which a flight is observed.

From the data thus obtained several interesting generalizations soon became evident. The most conspicuous of all, and for a long while the most doubted, because it was not understood, was the fact that at 6 or 7 miles above sea level, in middle latitudes, and generally 8 to 10 in tropical regions, the temperature no longer decreases with increase of height, but remains substantially constant.

How far this equal temperature, or isothermal condition, extends is, of course, unknown, but it does go at least to the greatest altitudes yet attained, that is, 15 to 20 miles. It may extend to the limit of the atmosphere of appreciable density, or it may not. We have no direct and positive evidence of either alternative. This much we know. From the surface up to a considerable distance above sea level, generally 6 to 10 miles, depending mainly on the latitude, the temperature decreases at the average rate of about 1° Fahrenheit in 300 feet; then almost abruptly, as a rule, practically ceases to change with further ascent.

Clearly, therefore, this unsuspected and striking phenomenon divides the atmosphere into two great parts; a lower portion in which temperature rapidly decreases with increase of height, and an upper in which the temperature is nearly independent of height. In the lower, convection, or ascent and descent of the air, can occur under certain conditions, because although ascending air cools by expansion (the pressure on it being decreased by the weight of the air left below) its ascent brings it into air that also is colder. Sometimes the ascending air, especially when saturated with water vapour, cools less rapidly with increase of height than does the air through which it is passing, in which case convection is certain and often vigorous.

Similarly, a local mass of air, particularly when cloud-laden, may warm less rapidly on sinking, as such masses do just after sun down, than the air it is falling through, whereupon convection again is inevitable. In short, this lower, cooling-with-altitude portion of the atmosphere is a region of convections and overturnings—not at all times, but under humidity conditions that will be explained later. It therefore has been called the troposphere, the sphere, or spherical shell, in which turning (going up and then going down), or convection, occurs.

In the region above the troposphere, where the temperature is constant with height, marked or vigorous convection does not and can not occur. It can not occur here because rising air rapidly cools with ascent, owing to expansion incident to decrease of pressure on it by the weight of the air passed through. Such cooling would keep the rising air all the time colder and therefore heavier, volume for volume, than the surrounding air and thereby quickly reduce its motion first to zero and then reverse it.

Formerly this was called the isothermal region, and to some extent it is still so called, owing to the fact that vertically, so far as explored, the temperature is substantially constant. Since convection here is impossible masses of air forced into this region would spread out in horizontal strata, and indeed there commonly are evidences of the existence of just such strata, though we seldom if ever are sure of their origin. This filler structure suggested the other, and now the all but exclusive, name, "stratosphere," of this region.

Thus the troposphere and stratosphere are quite distinct from each other. Clouds and every sort of precipitation, rain, snow, graupel, sleet and hail, and every kind of storm, are forever agitating the former, but never for a moment disturb the serenity or overcome the stability of the latter. They are sharply separated, the one from the other, along an approximately horizontal but invisible surface called the tropopause, and so called because that is where convection ceases, as above explained.

Perhaps it will be interesting to recall here that sometimes we use the expression, "the sphere," to mean the earth as a whole; that in our first approach to particulars we divide the sphere into the lithosphere, or solid portion of the earth, the hydrosphere, or water portion, and the atmosphere; and that the atmosphere in turn consists of two great concentric shells, the troposphere and the stratosphere. The lithosphere and the hydrosphere also have interesting structures about which many a fascinating tale has been told, but this is not the place to repeat them; ours is the story of the atmosphere and of it alone.

Trades (trade winds). When we examine the troposphere closely we find that it too has structure. Some of these parts are fleeting, but others are at least semi-permanent. One of the more nearly constant is the trade wind, or, better, the trade winds, as there are several winds properly so designated. Advantage has been taken of them, of course, in shipping, especially by sailing vessels, but this is not the source of their name. It does not derive from any idea of commerce, but from the fact that their characteristic is persistent blowing along a particular trade, that is, tread, track or way.

The trades are east winds (winds from easterly points) over the tropical oceans, or, more exactly, between the latitudes 30° N.

and 30° S. There really are five such winds that are well defined, one over the north Atlantic Ocean between latitude 30° N., roughly, and the Atlantic doldrums, or region of calms near the equator, and another over the south Atlantic between latitude 30° S., also roughly, and the same doldrums. Two other trade winds are similarly located over the Pacific Ocean, and there is one over the Indian Ocean, south of the equator, between Australia and Madagascar.

These wind currents, the trades, are very shallow near their poleward boundaries or edges, but grow gradually deeper and deeper to a maximum of at least 4 or 5 miles as their equatorial borders along the doldrums are approached. They are well defined and important structures of the troposphere.

Antitrades

Immediately over the trade winds are the antitrades, or winds that blow away from the equator and at the same time turn more and more nearly eastward and come down lower and lower with increase of latitude until they merge with, and become a part of, the prevailing westerlies-the winds beyond latitude 30° N. or 30° S., as the case may be, that from the surface up to at least well into the stratosphere usually are from westerly points.

The antitrade (there is one over each trade wind) represents a part of the return to higher latitudes of its accompanying trade, which through most of its route approaches closer and closer to the equator.

The rest of the trade, joined by a greater or less amount of tropical air, turns around the western end of a high pressure ridge, or ocean area of prevailing light and variable winds and clear skies. The frequent and deep southerly winds onto the United States from the Gulf of Mexico and the Atlantic Ocean are of one of those great joint trade and tropical currents.

Here then are two major features or elements in the structure of the troposphere, the trade and the antitrade, the east wind and the west wind, that all who pilot the argosies of the skies must know, and know how to use to proper advantage.

And the troposphere has many other parts. Two of the strangest and most important of these, strange because though apparently impossible companions they yet are always together, are the cold wind and the warm wind that peacefully flow beside each other—that, in technical terms, are each in dynamical equilibrium with the other. We know that a column of warm air will not stand up in equilibrium with an adjacent column of cold air, and we know that we know it by the way cold out-doors air pushes up and away the warm air in a heated chimney or smoke stack.

Hence we are unwilling at first to believe that a cold wind can blow alongside of a warm wind and not drive it away—that the twain can meet on equal terms and each hold its own. But they can and do, because an object, such as a mass of air, whenever moving over the earth tends always, because of the earth's rotation, to turn to one side or the other (the right, going with the wind, in the northern hemisphere, the left in the southern) of the course it is on. And the magnitude of this tendency (the force the moving object would exert on a frictionless vertical surface that would hold it to a fixed geographic direction) is proportional both to its mass and to its velocity. In the case, therefore, of adjacent stationary columns of warm and cold air only gravity is operative and the heavier (because colder) column underruns and buoys up the lighter.

This is why air rushes up heated chimneys and smoke stacks—why they draw, and the higher the better. In the case of the winds, however, the situation is quite different, for in proportion to their mass and their velocity they now exert (or would, against a suitable restraining wall) a horizontal or deflective force that also must be considered.

If, then, the winds were in the right direction, and had the proper positions with reference to each other, the colder from the east, say, and the warmer just south thereof and from the west, or, more generally, if they were passing each other counter clockwise (in the northern hemisphere, clockwise in the southern) they might have such velocities that the resulting deflective force would just balance the difference in pressure due to difference in density.

Ordinarily this condition of equilibrium implies a very gently sloping interface between the warm current and the cold, with the former overrunning the latter.

Owing, however, to surface friction and various obstacles, such as islands, and mountains, to the free flow of the winds, and to a change, for one reason or another, in the relative amounts of air on the two sides, perfect equilibrium never is long maintained. It does not break down abruptly and completely, but gives way quite slowly, and in so doing often leads to the development of the general cyclonic storm of the middle and higher latitudes.

Cold Front

When the breakdown between the warm and the cold currents is well developed, with the mass of cold and, of course, polar air pushing its way equatorwards while the warmer air flows beside it poleward, we speak of the interface between them, or, rather, of the intersection of that interface with the surface of the earth, as a cold front. We also call this locus a windshift line because here the direction of the wind changes, as one system, the warm winds, departs and another, the cold, comes on.

It likewise is called a squall line because all along it the winds are turbulent and squally and, in summer, often accompanied by thunder storms. This cold front, then, is the boundary, along the surface, between distinct wind systems—between a system of relatively cold and dry winds under skies and from higher latitudes, and a system of comparatively warm and humid winds under clouded skies and from lower latitudes—two more of the great parts in the structure of the troposphere.

Warm Front

The warmer of the two passing winds just mentioned soon reaches, in the course of its poleward travel, the colder air of higher latitudes. But here, as elsewhere, two masses of air, having different temperatures can not be in equilibrium with each other at rest and standing side by side. They must be, and are, in motion, and in such manner as to retain for a time, and to a greater or less extent, their individuality. Here again the colder air is wedged under the warm which flows not along this wedge, but up its slope—slantingly as a rule, but up, nevertheless. The line along which this part of the interface between the cold and the warm winds cuts the surface is called the warm front.

Thus the cold and the warm sectors of the traveling cyclone maintain their independence as polar winds and tropical airs, respectively, until gradually separated from their source, after which they soon are brought into like conditions each with the air of its new location and merged with it. In this way polar air becomes tropical air, and tropical winds polar winds, back and forth ceaselessly and indefinitely; but always as entities in their travels—as elements in the structure of the stratosphere—just as rivers and lakes are entities in the ceaseless round of water from oceans to continents and continents to oceans.

Inversions

Normally the temperature of the lower atmosphere, or troposphere, decreases at every level with increase of height, but there are exceptions to this rule; and, besides, the rate of increase always is more or less irregular. A very common exception occurs next to the surface. Indeed, during still clear nights this exception is itself the rule, and practically without exception. At such times, owing to the rapid net loss of heat by the surface through radiation, the adjacent air, by contact with the cold surface, becomes cooled to a distinctly lower temperature than the air at a slightly higher level.

The height to which this cooling extends depends upon the amount of air movement and consequent turbulence or mixing. If this is slight the cooling is restricted to near the surface, but is all the more pronounced. If the movement of the air is appreciable the cooling is less pronounced but extends higher.

Finally if there is considerable wind the mixing so distributes the loss of heat as to be small at any level, even at the surface. When, however, the cooling is marked and restricted to a shallow layer the temperature increases through this layer with increase of height instead of decreasing, as is the rule, from the ground up.

This is an inversion of the usual temperature gradient, or a temperature inversion, or, for short, just an "inversion."

For a while the air next above a surface inversion layer can slowly blow along over it without rapidly mixing with the colder air or carrying it away thus providing another case of structure in the troposphere. Appreciable winds, however, do wear away such a layer rather rapidly. They also commonly tend to produce, and frequently do produce, a temperature inversion some distance, perhaps a thousand feet or more, above the ground. The surface inversion occurs, as explained, on still clear nights and is due to loss of heat by radiation, and therefore might be called a radiation inversion. The inversion now under consideration may occcur any time the wind blows, because it is due to the turbulence or vigorous mixing of the lower air incident to surface friction and the interference to free flow by trees, house and other irregularities.

Turbulence produces a temperature inversion —the turbulence inversion—in this way: Before the wind sets in, the temperature of the lower thousand feet or more of the air may and often does, decrease slowly with increase of height, perhaps only one degree Fahrenheit in 400 to 500 feet. It even may increase with height, especially of early mornings, through the lowest levels, as just explained. In either case a complete mixing of the air from the surface up to the level of 1000 feet, says, brings all the air, not to a common temperature, as the stirring of water smooths out any thermal inequalities it may have had, but to such temperature that wherever, within that layer, a portion of its air may be taken, up, down or sidewise, it will, on arrival, have precisely the same temperature as the then surrounding air at the same level. That is, it will come to such temperature that nowhere will a moving portion of its either give heat to, or take heat from, the air with which it at any time is in contact. Obviously, because if the agitated air is not yet in a state of temperature equilibrium with a moving portion of itself, all that need be done to make it so is to stir it up further and mix it more thoroughly.

Now an isolated mass of air that neither gains heat from nor loses heat to the atmosphere through which it passes, obviously must cool with ascent, owing to its loss of heat incident to its expansion against the decreasing pressure. Where the air is unsaturated, so that no fog or cloud is formed, and thoroughly stirred up, the rate of this decrease, and therefore the rate of temperature decrease in and of the layer itself is, approximately, 1° F. per 190 feet, allowing for the average amount of humidity.

This, as explained, is a much faster rate of decrease of temperature than usually exists in the lower air. Hence the mixing of the lower air by turbulence so redistributes its heat as to make the lower portion warmer than it was, or otherwise would be, and the upper portion colder. Immediately above the topmost portion of the turbulent layer the air is undisturbed by vertical convection and therefore distinctly warmer than the upper portion of the agitated stratum. The turbulence inversion (inversion caused by turbulence) is a sharp partition between two portions of the troposphere, one the lower, full of irregularities, the other smooth-flowing.

Owing to its comparatively low temperature the top of the turbulence layer often is covered with a broad but shallow cloud of the stratus type which, because located at an inversion level, is warmest over its upper surface. But whether clouded or clear this inversion level is difficult of passage by air from either side. If air should pass this level going up it would at once be surrounded by other air much lighter, because warmer, than itself and therefore it would drop back.

Similarly, air passing it from above would be promptly pushed up again by the denser, because colder, atmosphere it was replacing. In short, this inversion stratum, though very thin and commonly invisible, is an impassable ceiling to rising air from below and an impenetrable floor to falling air from above. Gradually, however, through heat conduction, thermal convection, and in other ways, the inversion is smoothed away, and interchange across this level thus made possible wherever the air has within and of itself an adequate supply of water or, really, steam power—wherever its water vapour is sufficient to give abundant condensation, as in a cumulus cloud, and thereby a quantity of heat sufficient to keep it all the way to great altitudes continuously warmer and lighter than the surrounding medium.

It is interesting that as the temperature changes almost abruptly, perhaps several degrees, with change of height at the level of this turbulence inversion, so also must the density of the air change abruptly. And, furthermore, if the pressure gradient or push that causes the winds, is practically the same, as it seems to be, on either side of, and close to, this level, then there also must be here a nearly sudden jump in the wind velocity, from slower in the under, denser air to faster in the lighter air immediately above. However, this change of velocity, seldom more than one per cent of the whole, always is too small to be of any practical importance.

DISTRIBUTION OF TEMPERATURE

The pull, as we call it, of gravity makes water run down hill. It also makes a heavy liquid underrun a lighter one in the same level; both are drawn in the direction of the bottom, but the pull on the heavier, or denser, is greater than on the lighter, and the stronger pull prevails. Gravity also makes an isolated mass of liquid or gas in a heavier one go up, not down; it is pushed or buoyed up by a force equal to the difference between the weight of the lighter and that of an equal volume of the heavier.

Clearly, then, whenever two masses of air of unequal density come into free contact with each other the lighter is pushed up and away, except in the case of properly adjusted winds. Now air rapidly increases in volume, and correspondingly decreases in density, with increase of temperature—roughly 1% per 5° F. at ordinary temperatures.

Hence the hot air in a chimney is lighter than an equal volume of the cold air on the outside, and therefore is pushed up by the latter which, in turn, is heated and itself pushed up, and so on as long as there is a fire in the grate to supply the heat. To be sure, the combustion alters the composition of the air (makes it richer in carbon dioxide if coal is used, and in both carbon dioxide and water vapour if wood or gas is the fuel, and poorer in oxygen) in such manner as to render that in the chimney heavier, at the same temperature, than that outside, but this increase in density through change in composition is small in comparison to its decrease in density by heating.

At most it could balance or offset a temperature increase of only about 40° F. over coal, or 10° F. over wood, while ordinarily the effect if much less, since commonly only part of the oxygen is consumed; hence the heating, being several times this maximum value, has, in any case, the best of the argument, as it were, and the chimneys keep on drawing. Similarly, air in the open is underrun and pushed up by even slightly cooler adjacent air of the same composition, unless, as already explained, the two masses happen to be flowing past each other in the right positions and directions and with the proper velocities.

Actually, the heated air expands as its temperature rises, and overflows above wherever its pressure is thus made greater than that of the adjacent atmosphere. This overflow, or outflow, decreases the pressure at the bottom, and in the lower portions, of the heated air, and at the same time increases the pressure round about under the places of overflow—mass, hence weight, is removed from one place and added to others.

This disturbs the balance. Gravity tends to restore it and thereby induces winds in the direction, initially at least, of higher to lower pressure. If the heated region is very small, equilibrium is quickly established, unless the heating is maintained. But where the higher temperature covers a large the winds no longer flow directly towards the centre of lowest pressure but more or less round about it, owing to the rotation of the earth, in a manner seemingly most contrarious.

This heating in innumerable cases is very local and of only a few hours' duration; in many others it is quite extensive and lasts days Weeks, and even all season long; while its greatest manifestation is the year after year and age after age continuously higher temperature in tropical realms and lower in the frigid zones.

This perpetual heating of the atmosphere over one great region, and its ceaseless cooling over another, or rather, two others, keeps it continuously out of balance and makes the winds, especially the trades and the westerlies, forever to blow—to blow dizzily over a rotating earth, and time and again violently and confusedly incident to the rapid, the all but explosive, delivery, by condensation, to a limited region of vast quantities of heat that had been slowly accumulated by evaporation from others afar off.

The whole of the atmosphere to the tops of the highest clouds, that is, the whole of the troposphere, is a huge convection system, greatly complicated by the rotation of the earth and all but hopelessly confused by evaporation and condensation. The stratosphere, too, has its circulation, but as yet not much is known about it.

Of course it is difference in pressure *at the same level* that pushes the air about, or makes the winds to blow, but, as explained, this difference in pressure depends, in turn, mainly (water vapour has a little to do with it) on the distribution of temperature. That is one reason, but not the only one, why this distribution is so important.

Perhaps some good physicist will insist that it really isn't difference of pressure at the same height above sea level that makes the winds blow, but difference of pressure over an "isentropic surface," or surface of "equal entropy."

Well, he would be right in respect to appreciable heights above the surface, because for the free air the isentropic surface is the "level" surface. But nothing short of a surgical operation can get the idea of entropy into the other fellow's head, and there is no rivet, weld, or hermetic seal that will keep it there. Besides, commonly (not always), there isn't much difference between the two after all—"same level" and "isentropic level"—and so we will stick to the one everybody knows and no one forgets, that is, "same level."

Source of Heat

When we think of the source of heat, especially in the wintertime, we are likely to have in mind some sort of combustion, for that is the cause of the tropical climate we have indoors at that season. But indoors is a mighty small place in comparison with all outdoors; and outdoors is heated, too, often very hot in summer, and always far above the 460° below zero Fahrenheit that would be its temperature if there were no heating at all. Almost every bit of this enormous amount of heating comes from just one source, the sun.

Incoming Radiation

The radiation from the sun is so great that if it all got through the atmosphere enough would fall on each square foot directly facing it to heat a gallon of water from the freezing point to the boiling point in just three and a half hours. But it does not all get through, and what does get through always comes in slopingly except wherever the sun happens for the moment to be directly overhead.

In fact, owing to the reflecting power of clouds, especially, and the surface of the earth, and to the scattering (not reflection) of light by the molecules of the air and by the myriads of dust motes, one-third, roughly, of the incoming radiation is thrown off to space without producing any effect whatever on the temperature of the atmosphere or of the earth beneath. Another one-third, again roughly, of the incoming solar radiation is absorbed by the atmosphere, and the remaining portion by the earth.

These statements apply to the earth as a whole. The ratios between loss by reflection and scattering, air absorption, and earth absorption, vary widely from place to place and season to season, owing mainly to differences in humidity, cloudiness, and elevation of the sun above the horizon, and differences in the character of the surface of the earth—whether land, water, snow or ice, bare soil or vegetation.

Clear Sky Radiations

It is interesting to note that the amount of radiation reaching the earth from a clear sky is equal to a considerable fraction of that which reaches it from the sun directly. At sea level the amount of sky radiation onto a horizontal surface of any particular size, a square foot, say, is equal to about 7.8% of the amount of unaffected, or direct, solar radiation onto an equal area squarely facing the sun at the same time and place. When the sun is directly overhead its supply of heat to a horizontal surface at sea level is nearly 13 times as great as that from the sky. When it is one-third of the way down from the zenith to the horizon its contribution of heat to the earth is only a little more than 6 times that from the sky, and each is then decidedly less than it is when the sun is in the zenith. Finally, the two sources are equal, though both are still further enfeebled, when the sun is above the horizon about one-twelfth the distance to the zenith.

The brightness of the clear sky is greatest near the sun, as even casual observations readily show, and decreases gradually with increase of distance therefrom over a large part of the whole area. Hence the total of sky radiation received per minute on a horizontal surface is greatest at noon, as is also the direct solar radiation. The intensity of sky radiation decreases, in general, with increase of height above sea level, while that of the direct solar radiation increases.

RADIATION FROM AN OVERCAST SKY

When the sky is overcast, neither its brightness nor the total amount of radiation received on a given horizontal area is at all constant, even for the same height of the nun, because the clouds in question may be of any kind from the thinnest cirrus that just dims the sun, to the darkest nimbus that reduces even noonday brilliance to twilight.

If, however, the sky is completely overcast by an approximately uniform cloud layer dense enough, but not greatly more than enough, to prevent the position of the sun from showing, then the total radiation from this cloud layer onto a horizontal surface is, on the average, slightly greater than that from a clear sky.

Evidently, too, the amount received of this cloud-transmitted radiation generally must increase with increase of height above sea level of the place of reception.

The brightness of the cloud layer is surprisingly close to uniform. It is greatest nearly overhead (just a little way off in the direction of the sun), but still nine-tenths as bright half way to the horizon, and half as bright almost at the horizon.

LOCATION AND TEMPERATURE

Every one knows that the average temperature of the tropical regions is higher than that of the polar areas, but it is not a familiar fact that, nevertheless, int he course of a year the temperature reaches 90° F. or over (in the shade) on more days in central Alaska than at Panama.

It also is a surprising fact to most of us that the average temperature through January at St. Louis is the same as that in southern Iceland; and that the average temperature for the entire winter, December, January and February, at Sitka, Alaska, is about the same as that at Washington, D. C. And there are lots of other similar surprises, as, for instance, the fact that semi-tropical vegetation that would be killed by frost is now growing wild on the Scilly Islands in the latitude of northern Newfoundland. Clearly, then, difference in longitude may be accompanied by nearly as great a variation in temperature as is difference in latitude. During winter especially the lines of equal temperature run far poleward over the oceans, and equatorward over the continents.

Another matter of great influence on the local temperature is the nature of the surrounding area, both near and distant. A fair inland point, for instance, becomes much hotter in summer and greatly colder in winter than does a mid-sea island at the same latitude. Also a coast where the prevailing winds are on-shore has a more nearly even temperature than one of the same latitude with off-shore winds. The first is bathed in ocean breezes of relatively equable temperature; the second in winds that have traveled far over land and that therefore are characterized by its temperature irregularities and extremes.

It is interesting in this connection also to note that the temperature is a little higher, and in some cases quite noticeably higher, in a city than in the adjacent country; and further that in the country the forest is cooler in summer, and slightly warmer in winter, than the open fields. In some cases the foot of a high mountain occasionally is very peculiar in respect to temperature.

It may happen that the air is quite cold when all of a sudden there comes across and down the mountain a roaring hot wind that within a few hours clears away every trace of even a deep snow. This wind went up the other side of the mountain saturated and rainy, hence it cooled relatively little with ascent owing to the latent heat of vaporization there rendered sensible by condensation. As it came down, however, it was dry and therefore heated rapidly with descent and consequent increase of pressure. Such are the famous foehn winds on the northern side of the Alps, and the chinook winds of the Rocky Mountains.

South Pole Colder than the North

We often are asked which is the colder, the north pole or the south. One answer might be that, as we have no records longer than a few hours at either, we don't know.

That is true enough so far as bare statistics are concerned, but in this case we can reason the matter out from sure and simple premises:

- Normally, the greater the height the lower the temperature, other things being equal;
- The south pole is 10,000 feet, roughly, above, and the north pole at, sea level. Hence we should expect the south pole to be the colder of the two.
- The greater the height the less, in general, the amount of cloud, water vapour and other gas to radiate back to the earth and thereby help to keep up its temperature;
- The south pole is much higher than the north. The south pole, therefore, because it gets less return radiation than does the north, should have the lower temperature.
- The faster heat is supplied by the surface to the lower air the higher the temperature of that air, other things being equal;
- By actual measurement the ice over the Arctic Ocean gives off enough heat in 24 winter hours, coming from the relatively warm water below, to increase the temperature of a layer of air 450 feet thick by 20° F., while that given off at the south pole is only a small fraction of this amount, owing to the much poorer conductivity of the snow and the far greater depth to a temperature equal to that of the

arctic water. Then for this reason also the south pole must be colder than the north. It seems therefore that even without the actual observations we may feel reasonably certain that the temperatures of the two poles are not the same, especially their night temperatures since these occur when they can be but little affected by the variations in our distance from the sun, and that the south pole is distinctly the colder of the two.

Relation of Surface Temperature to Height

Those who live in mountains regions know by personal experience that, in general, the higher up the mountain the lower the temperature. Instrumental records show that this relation is true not only for mountains but also for hills and even plateaus, and that the approximate numerical values are 1° F. decrease per 330 feet ascent on a mountain, 365 among hills and 455 on plains.

The reason for this difference in favour of the plateau is the fact that the air is heated mainly by the surface of the earth which, in turn, is heated by the sunshine. That is, in the case of the plateau the surface which is the heater of the air is at the height in question all around as far as the level area extends, while the air on the mountain is affected in part by the temperature of the free air, especially when there is an appreciable wind, and this free-air temperature is lower, as we shall see presently, except on still clear nights, than surface air at the same altitude.

Relation of Temperature to Height in the Free Air

According to observations the temperature of the air normally decreases with increase of latitude from the surface of the earth up to the height of several mile, roughly 6 to 7 in middle latitudes. Beyond this level the temperature remains substantially constant up to the greatest heights yet attained, probably around 18 miles.

The average rate of decrease of temperature of the free air with increase of height is about 1° F. per 300 feet, from the surface up to the level at which appreciable decrease ceases—up to the "tropopause," or limiting reach of convection. In general this temperature decrease is most rapid in the upper half to two-thirds of the depth under consideration and least rapid in the lower one-third.

Immediately above a land surface the change of temperature with height is widely variable, from a *decrease*, perhaps ten fold the above average value, to an *increase* ten fold that rate, at least through the first 20 feet or so, above suitable regions, such as plains, valley bottoms and bowl-shaped depressions, and, of course, during still clear nights. Few phenomena of the atmosphere are as often "explained" as is the fact that, in general, temperature decreases with increase of height, and hardly any other as inadequately, not to say erroneously, explained.

The facts are:

- The ascending (pushed-up) warm air comes under less and less pressure with increase of height by the weight of the air left below. It therefore continuously expands while rising, and all the time against pressure—the weight of whatever air is still above it. But this expansion against pressure is work, and work at the expense of the only supply of energy the rising air has—its heat. Hence as it rises it must and does become cooler.
- This cooling of the air by convection does not go on to absolute zero, nor, as explained, does the surface air keep on getting hotter and hotter. And these limitations are owing to the fact that the atmosphere loses heat by radiation. The free air therefore is all the time gaining heat by absorption of radiation and losing heat by emission of radiation, while the surface air is gaining heat also by contact with the warmed earth.
- But loss of heat by radiation decreases very rapidly with fall of temperature, while the power to absorb radiation does not change. Hence as the air ascends higher and higher, and thereby gets colder and colder, it presently comes to a temperature at which its loss by radiation is equal to its gain by absorption. Beyond this level it can not rise, because if it did so it instantly would become colder and denser than its environment and fall back again.

From this level on up the temperature of the air must remain roughly constant except as modified, perhaps, by change of composition. This is the isothermal region, or stratosphere, which every planet must have whatever the extent and composition of its atmosphere.

- Half, roughly, of the sunshine that is not lost by reflection or by scattering, that is, half of it that does any heating at all, gets entirely through the air and is absorbed by the surface of the earth, which, on being thus heated, heats in turn the adjacent or lowest air.
- The other half of the effective radiation from the sun (portion used and not immediately lost) is absorbed mainly by the water vapour, and as the density of this vapour generally decreases very rapidly with increase of height it follows that by the direct absorption of sunshine the heating of the air likewise is greatest in its lowest levels.
- The surface of the earth loses heat (it doesn't keep on getting hotter and hotter indefinitely) not only by conduction to the adjacent air, but also by radiation, a kind of radiation greatly absorbed by water vapour. Therefore in this third way, too, as in each of the others, the atmosphere is more and more strongly heated with decrease of height.
- Since the cold air of the stratosphere is losing heat by radiation at the same rate that it is gaining it by absorption it follows that the warmer air of lower levels is losing by radiation faster than it is gaining by absorption, the net difference at each *level* (not moment to moment for the moving air) being made up by heat brought there by convection from the surface, either directly as such, or indirectly through evaporation and subsequent condensation.

Thus the whole of the troposphere, or, in other words, the whole of that portion of the atmosphere in which clouds can and do occur, is continuously being heated below and cooled above. In this way, that is, by heating below and cooling above, convection and, in general, a decrease of temperature from bottom to top of the convective layer, is maintained without the air as a whole getting either warmer or colder.

- Although the surface air is most heated it is not equally heated everywhere. Hence the warmest and lightest portions are pushed up—forced to rise—by the cooler and denser air round about as a cork is bobbed up when let go of under water.

Temperature Changes in the Stratosphere

Although the stratosphere, that portion of the atmosphere beyond the highest clouds, has no immediate contact with the warming and cooling surface, nevertheless its temperature at any given locality often varies 10° F. to 20° F., and even more, not vertically, as a rule, but horizontally, or from day to day.

Thus the temperature of the stratosphere commonly is distinctly higher over the forward and central portions of a cyclonic area than it is over the corresponding portions of an anticyclone. The cause of this change is not definitely known.

Temperature Inversions

The structure of the atmosphere, during still clear nights the surface air becomes so cold over level land areas, in valleys and in basins, that often there is a rapid rise of temperature with increase of height through the first 10 to 100 feet, or more. This is the surface inversion, so favorable to production of frost, and without which orchard heating commonly would be unnecessary and, moreover, in general impracticable. Every wind of appreciable strength so thoroughly mixes up the lower air by turbulence that the temperature of the top portion of the agitated stratum is decidedly lower than that of the undisturbed air immediately above it. This is the turbulence inversion, which, because of its low temperature often is accompanied by a stratiform cloud.

It might seem that a similar inversion should occur at every interface between over- and under-running air currents, but such inversions, so far as they exist at all, are too slight to be of any particular importance. This is owing to the smallness of the friction between free air currents and the consequent all but complete absence of turbulence.

The highest temperature inversion in the atmosphere, of which we have any actual and direct measurement, is that at the base of the stratosphere during the passage of an anticyclone. As already stated the cause of this particular, and often very pronounced, inversion is not yet definitely known.

Thermal belt, or green belt

There are various ways of protecting fruit from frost, but the best of them all is the proper selection of the orchard site. In a hilly or mountainous region that best location is neither on the floor of the valley nor, generally, on the top of a ridge, but, as a rule, some distance, not too far, up one or the other of the slopes. A strip along a hillside or mountainside at this level is known as the thermal belt, green belt, verdant belt, frostless belt, etc., because on still, cloudless nights this level is warmer than any other above or below it, hence least likely to have frost, and most likely to show green and uninjured vegetation.

After sundown, when the sky is clear and there is no wind, the surface of the earth everywhere cools much more rapidly than the free atmosphere, and in turn correspondingly chills the nearby air either directly through actual contact, or indi- rectly by turbulence mixing with that which had been so chilled. On the side of a hill, then, this air, because it is denser (being colder) than that at the same level over the adjacent valley, flows down slope much as would a sheet of water. However, as this drainage air reaches lower levels it evidently is subjected to increase of pressure to the extent of the weight of the air passed below.

It therefore is compressed—work is done upon it—and its temperature made higher than it otherwise would have been. In the early evening, and well up on the hill, this heating of the down flowing air causes it to become warmer and warmer with descent, almost to the valley bottom. Here the gain of heat through increase of pressure not only makes up for all that was lost by contact with the cold surface, but actually warms the air to a higher temperature than it had before it was first chilled. But this heating is pronounced only where the descent is rapid. Air already at the bottom of the hill is not thus heated. It does not come under any greater pressure for there is no lower place to which it can rapidly drain. It therefore just gets colder and colder as the surface temperature continues to fall incident to the net loss of heat by radiation.

It also gets colder, but not so rapidly, near the bottom where the slope is gentle and the current sluggish; and less and less rapidly with increase of height and gain in the speed of flow. There must therefore be some level along either valley wall at which, for the time being, the heating of the descending air by compression is just equal to its cooling by contact with the chilled surface.

Above this level the air evidently gets colder with ascent and below it colder with descent. During most of the night the flood of cold air grows gradually deeper, carrying the level of maximum temperature—the level of its crest-higher and higher up the valley sides. Near morning, however, it becomes practically stationary, and where it then is frost obviously is least likely to occur. This is the place to plant your orchard—here along the thermal belt where the temperature is highest and the chance of frost the very least; where vegetation may pass through the night unharmed, while all above is frozen stiff and all below white with frost.

7

Air Masses & Fronts

AIR MASS

An air mass is a large volume of air in the atmosphere that is mostly uniform in temperature and moisture. Air masses can extend thousands of kilometers in any direction, and can reach from ground level to the stratosphere—16 kilometers (10 miles) into the atmosphere.

An air mass is a large volume of air in the atmosphere that is mostly uniform in temperature and moisture. Air masses can extend thousands of kilometers across the surface of the Earth, and can reach from ground level to the stratosphere—16 kilometers (10 miles) into the atmosphere.

Air masses form over large surfaces with uniform temperatures and humidity, called source regions. Low wind speeds let air remain stationary long enough to take on the features of the source region, such as heat or cold. When winds move air masses, they carry their weather conditions (heat or cold, dry or moist) from the source region to a new region. When the air mass reaches a new region, it might clash with another air mass that has a different temperature and humidity. This can create a severe storm.

Meteorologists identify air masses according to where they form over the Earth. There are four categories for air masses: arctic, tropical, polar and equatorial. Arctic air masses form in the Arctic region and are very cold. Tropical air masses form in low-latitude areas and are moderately warm. Polar air masses take shape in high-latitude regions and are cold. Equatorial air masses develop near the Equator, and are warm.

Air masses are also identified based on whether they form over land or over water. Maritime air masses form over water and are humid. Continental air masses form over land and are dry.

Therefore, an air mass that develops over northern Canada is called a continental polar air mass and is cold and dry. One that forms over the Indian Ocean is called a maritime tropical air mass and is warm and humid.

AIR MASS FORMATION

Source regions of common air masses. The arctic front is along the northern border of Alaska and Canada. The polar front is along the northern border of the United States.Where an air mass receives it's characteristics of temperature and humidity is called the source region. Air masses are slowly pushed along by high-level winds, when an air mass moves over a new region, it shares its temperature and humidity with that region. So the temperature and humidity of a particular location depends partly on the characteristics of the air mass that sits over it. Storms arise if the air mass and the region it moves over have different characteristics. For example, when a colder air mass moves over warmer ground, the bottom layer of air is heated. That air rises, forming clouds, rain, and sometimes thunderstorms. How would a moving air mass form an inversion? When a warmer air mass travels over colder ground, the bottom layer of air cools and, because of its high density, is trapped near the ground.

In general, cold air masses tend to flow toward the equator and warm air masses tend to flow toward the poles. This brings heat to cold areas and cools down areas that are warm. It is one of the many processes that act towards balancing out the planet's temperatures. Air masses are slowly pushed along by high-level winds. When an air mass moves over a new region, it shares its temperature and humidity with that region. So the temperature and humidity of a particular location depends partly on the characteristics of the air mass that sits over it. Air masses are classified based on their temperature and humidity characteristics. Below are examples of how air masses are classified over North America.

Maritime tropical (mT): moist, warm air mass

Continental tropical (cT): dry, warm air mass

Maritime polar (mP): moist, cold air mass

Continental polar (cP): dry, cold air mass

Storms arise if the air mass and the region it moves over have different characteristics. For example, when a colder air mass moves over warmer ground, the bottom layer of air is heated. That air rises, forming clouds, rain, and sometimes thunderstorms. How would a moving air mass form an inversion? When a warmer air mass travels over colder ground, the bottom layer of air cools and, because of its high density, is trapped near the ground.

In general, cold air masses tend to flow toward the equator and warm air masses tend to flow toward the poles. This brings heat to cold areas and cools down areas that are warm. It is one of the many processes that act towards balancing out the planet's temperatures.

AIR POLLUTION AND HEALTH

An important reason for controlling air pollutants such as particulate matter or sulfur dioxide is the damaging effects they have on human health. These effects include premature death, as well as increases in the incidence of chronic heart and lung disease.

Estimates of the health damages associated with air pollution are important because they can provide both an impetus for environmental controls and a means of evaluating the benefits of specific pollution control policies. To estimate the health damages associated with air pollution in developing countries, policy makers are often forced to extrapolate results from studies conducted in industrialized countries. These extrapolations, however, may be inappropriate for two reasons.

First, it is not clear that the relationships found between pollution and health at the relatively low levels of pollution experienced in industrialized countries hold for the extremely high pollution levels witnessed in developing countries.

Levels of particulate matter, for instance, are often three to four times higher in developing countries than in industrialized ones. Second, in developing countries, people die at younger ages and from different causes than in industrialized countries, implying that extrapolations of the impacts of air pollution on mortality may be especially misleading. This document reports the results of a study relating levels of particulate matter to daily deaths in Delhi, India between 1991 and 1994. We focus on Delhi, the capital of India, because it is one of the world's most polluted cities. During the study period, the average total suspended particulate level in Delhi was 378 micrograms per cubic metre—approximately five times the World Health Organiza-tion's annual average standard.

Furthermore, TSP levels in Delhi during this time period exceeded the WHO 24-hour standard on 97 per cent of all days on which readings were taken. In addition, the distributions of deaths by age and by cause in Delhi are very different from those in the U.S. In the U.S., over 70 per cent of all deaths occur after the age of 65. In Delhi, over 70 per cent of all deaths occur before the age of 65, with over 20 per cent occurring before the age of five.

Furthermore, 46 per cent of all non-trauma deaths in the U.S. are attributable to cardiovascular disease compared to only 23 per cent in Delhi.' Because the main effects of acute exposure to air pollution on daily deaths occur through impacts on cardiovascular and respiratory disease, for which age is a known risk factor, we expect these differences to affect the relationship between pollution and mortality.

Our estimates of health damages have policy implications for pollution control in Delhi, and permit us to compare extrapolations from U.S. studies with actual pollution impacts. We find that a given reduction in TSP reduces non-trauma deaths in Delhi by a smaller percentage than predicted by U.S.. studies. Indeed, the percentage decrease in deaths corresponding to a 100 microgram reduction in TSP is 2.3 per cent—about one-third of the effect found in the U.S.

On the other hand, because the age distribution of impacts vary from Delhi to the U.S., so do the number of life-years saved. The largest impact of particulates on daily deaths in the U.S. occurs among persons 65 and older. In Delhi, the largest impact occurs in the 15 to 44 age group, implying that for each death associated with air pollution, more life-years will be saved in Delhi than in the U.S. on average.

Estimating Concentration-response Relationships For Delhi

The relationship between air pollution and premature mortality is most often studied using time-series analysis of daily observations of the number of deaths and pollution levels. These studies capture the effects of short-term exposure to pollution on the probability of dying. The underlying assumption is that there is a distribution of susceptibility to the effects of air pollution in any population.

People who are in a weakened physical state or who have a history of chronic obstructive pulmonary disease or cardio-pulmonary problems are thought to be the most vulnerable. In the case of a sharp rise in pollution, the most vulnerable people are more likely to die. Clearly, this type of analysis does not capture all of the effects of pollution exposure. Long-term exposure to pollution can also reduce life expectancy by altering lung function and making people more susceptible to COPD. However, measuring the effects of chronic exposure requires a long-term prospective study in which a sample population is followed long enough for the chronic effects to manifest themselves.

Two such studies have been completed to date: the Harvard Six Cities study which followed participants over the course of 15 years and the American Cancer Society study which lasted approximately eight years. Due to cost considerations and time constraints, our work focusses on the acute effects of air pollution.

Because time-series studies focus on a given geographic location over a number of years, factors that are often thought to influence the health of the population, such as percentage of smokers, income level, occupational exposure to pollutants, access to medical care and age distribution, do not need to be incorporated into the analysis as they are considered to remain relatively constant within the study area over time.

Typically, the only other factors aside from pollution included in these models are weather variables and seasonal controls. In measuring the effects of air pollution, most attention has been focussed on particulates, especially those particles measuring less than 10 microns in diametre which penetrate the lungs more readily. Even though particulate matter tends to be the pollutant most strongly associated with premature mortality, the presence of other pollutants may be important as well.

Data

Mortality data for years 1991 through 1994 were obtained from the New Delhi Municipal Committee one of the three distinct regions which comprise the National Capital Territory. Because the NDMC houses a large concentration of Delhi's hospitals, approximately one-fourth of the 60,000 deaths in Delhi each year occur in the NDMC, in spite of the fact that only 3.6 per cent of the population resides there.

Although the NDMC data represent only 25 per cent of all deaths occurring in Delhi, the geographic distribution of the Delhi residents who died due to non-traumatic causes in the NDMC mirrors the geographic distribution of the population. We compare the distributions of deaths by cause and by age for the NDMC and the National Capital Territory of Delhi.

The distributions of deaths by cause are similar, with between 20 and 25 per cent of all medically certified deaths attributable to causes associated with air pollution. Roughly the same percentage of deaths are attributable to infectious diseases and to perinatal causes. The distribution of deaths by age group differs somewhat between the NDMC and the National Capital Territory.

Forty-three per cent of deaths in the NDMC occur before the age of five, compared to 33 per cent of deaths in Delhi. For this reason, we estimate impacts by age group, as well as for specific causes of death. We also estimate the impact of air pollution on all non-trauma deaths. Because of problems associated with pinpointing the precise cause of an individual's death, counts of total non-trauma deaths have often been used as the basis of time-series analysis of this type, even though the links are strongest between air pollution and both cardiac and respiratory disorders. The data display a marked seasonal pattern, with the highest number of deaths occurring during the rainy monsoon seasons. The Central Pollution Control Board provided daily data on air pollution levels collected at the nine monitoring stations located throughout the city. Six of these monitors have been in operation since 1987 and are operated directly by the CPCB.

The other three monitors, added to the monitoring network in 1990, are operated on behalf of the CPCB by the National Environmental Engineering Research Institute. Readings of total suspended particulates, sulfur dioxide and nitrogen oxides are taken at each station on a rotational basis approximately every three days. The data include average, maximum and minimum daily observations of these pollutants at each monitor.

The monitors were not in operation on weekends or holidays during the study period. Daily means of TSP, SO, and NOR, were calculated using all available readings on a given day. Total Suspended Particulate levels for Delhi over the four years of the study. WHO established a guideline range several years ago of 150 to 230 micrograms per cubic metre per 24-hour period. TSP levels routinely fall well above this 24-hour guideline and, at times, reach as high as six times the guideline levels.

While two coal-fired power plants, a large motor vehicle fleet and chemical and cement industries located within the city all contribute to the high particulate levels in Delhi, we suspect that much of the TSP is currently attributable to resuspended dust and natural sources, primarily because sulfur dioxide and nitrogen oxide levels remain well below the WHO guidelines.

Even so, available data on the size composition of particles suggest that the ratio of particles less than or equal to 10 microns in diametre to TSP is about the same as in the United States. Daily meteorological data covering all four years of the study period were obtained from the National Climatic Data Centre in Asheville, NC. These data, collected on a daily basis at Delhi's International Airport located in the southwest of city, include average daily temperature, maximum and minimum temperature, mean daily dew point temperature, rainfall and visual range. Temperatures never fall below freezing and typically range between 60 and 100 degrees Fahrenheit. During the study period, fewer than five per cent of the days fell below 55 degrees and fewer than one per cent exceeded 100 degrees.

The Econometric Model

Daily mortality figures are considered counts of rare events and are therefore often modeled using Poisson regression analysis. We fit Poisson regressions to the Delhi mortality data using the method of maximum likelihood.

Formally, the log likelihood function is:

where y_t, is the count of deaths occurring on day t, $\lambda_t = \exp(X_t\beta)$is both the mean and the variance of deaths, X_t, is a matrix of covariates on day t, and β is a vector of regression coefficients. The predicted count of deaths on day t is therefore , implying that, if pollution is entered in a linear fashion in the right-hand side of the model, the percentage change in mortality following a given change in pollution levels is—holding all other factors unchanged—$\Delta C.\beta c$, where C is the pollution concentration variable and βc is the coefficient associated with this variable. The maximum likelihood estimate of β is obtained by applying the method of iteratively re-weighted least squares to minimize the quadratic form:

$$[Y-\lambda]-\Omega^{-1}[Y-\lambda]$$

where Y is the T x 1 vector of daily deaths, $\lambda = \exp(X\beta)$, X is the Tx1 matrix of covariates, and $\Omega = \mathrm{diag}\{\exp(X_t\ \beta)\}$. We exploit this equivalence to modify expression to correct for serial correlation. Specifically, we assume that daily deaths follow a serially correlated process of the first order with autocorrelation paramneter p, and implement the generalized estimating equation approach devised by Liang and Zeger, replacing the matrix Ω in equation with the matrix $\Lambda=A^{1/2}PA^{1/2}$ where $A=\text{diag}\{\exp(X_t\beta)\}$, $P=\{r_{ts}\}$, and $r_{ts}=\rho^{|t-s|}$, for t, s = 1, 2..., T. Coefficient estimates are obtained from an IRLS step nested within a grid search over values of p ranging between -1 and 1. As is often the case with Poisson models, we are also concerned about the possible presence of over dispersion, which may signal omitted variables and tends to result, when ignored, in unrealistically low standard errors.

To avoid such a problem, the Poisson standard errors can be multiplied by a correction factor equal to

where x^2 is the value of the Pearson chi square goodness of fit statistic and dof is the number of degrees of freedom, *i.e.*, the number of observations minus the number of independent variables.

Model Specification

Development of the Core Model

We estimated the autoregressive Poisson model for total non-trauma deaths, as well as for deaths by selected cause and age group. Our modeling strategy was carried out in four steps. We began by accounting for the variability in the number of total non-trauma deaths using seasonal/cyclical terms, a daily time trend and year dummy variables. Then, we added weather variables to control for the impact of temperature and humidity.

Next, we added pollution variables to see if they had any additional explanatory power. Because of the complexity of the autoregressive Poisson model, we used a log-linear specification and ordinary least squares to build the models, using the Akaike Information Criterion at each step to select additional regressors. Finally, we corrected the model for serial correlation using IRLS.

Following Schwartz, we have controlled for the seasonality in the data by including trigonometric terms for cycles ranging from one year to 2.4 months in length. A daily time trend and dummy variables for the year of the study account for population increases and other unobserved factors thought to influence the number of deaths. We considered using indicator variables for both the month of the study and the season to control for the seasonality in the data but found that the trigonometric terms provided the best fit. We accounted for the effects of weather on daily mortality by adding temperature and dew point temperature variables to the model.

We considered both maximum and mean temperatures with various lag structures up to seven days in length as well as indicator variables for the hottest and most humid one, five and ten per cent of the days during the study period. In the end, we found that contemporaneous mean weather variables and dummy variables for the most extreme 10 per cent of the days were the best predictors.

With season and weather accounted for, we incorporated particulate pollution into the model. Again, we considered daily maximum, minimum and mean values of TSP as well as values lagged up to three days. While we did not find significant effects of contemporaneous levels of TSP on mortality, we did find that mean TSP lagged two days had a significant effect.

Sensitivity Analyses

To examine the robustness of the core model, we conducted several sensitivity tests. First, we added dummy variables to the model to indicate which monitoring stations were included in the calculation of the TSP variable for a given day. The addition of these variables did not affect the significance of the pollution variable nor did it affect the magnitudes of the coefficients dramatically.

An F-test of the joint significance of these variables, however, did not allow us to reject the null hypothesis that all the coefficients were equal to zero. Since other studies have shown statistically significant impacts of SO_2 on daily mortality, we examined the impacts of SO_2 on daily deaths in Delhi. First, we considered the impact of SO_2 on mortality alone and then included it in the model with TSP. When considered alone, SO_2 was found to have a negative coefficient for every mortality endpoint except cardiovascular disease and was significant for all endpoints except respiratory deaths, cardiovascular deaths and deaths in the 0 to 4 age group. When considered jointly with SO_2, the effect of TSP on daily mortality retained its significance in all cases and, in general, did not change in magnitude. While the negative coefficient on SO_2 is somewhat puzzling, we do not give much weight to these results. SO_2 levels during the study period were very low and the correlation coefficient between SO_2 and TSP, while positive, is not large. In studies where SO_2 was found to have a significant and positive effect on daily mortality, SO_2 and TSP were much more highly correlated and SO_2 levels were in general much higher.

We do not report the effects of NO_x on daily deaths since levels during the study period were similarly low and the link with daily mortality is not well established in the medical literature. We tested the model's sensitivity to outliers by removing the highest five per cent of the TSP values and repeating the regressions. While admittedly the magnitudes of the coefficients were affected in some specifications of the model by the removal of these extremev alues, in general, the significance of the TSP variablew as not.

In some instances, such as cardiovascular deaths, deaths among children aged 5 to 14, and among the elderly, the impact actually increased once the extreme values were removed. Finally, to test for the effects of over dispersion, we estimated models for each endpoint using a serially independent Poisson model and then applied the correction used by McCullagh and Nelder and Agresti. Correcting for over dispersion in this manner had only slight effects on the magnitudes of the coefficients on TSP and did not affect their significance.

AIR POLLUTION EFFECTS

We can distinguish between short-term acute effects and long-term chronic effects of air pollution. Acute effects may be brought on, for instance, by pollution episodes, while chronic effects may generally affect a greater proportion of the population. It is important to remember that air pollutants are transboundary, *i.e.* they know no borders and travel easily from their sources towards other locations spreading pollution throughout the world. The respiratory system is by far the most important route for entry of air pollutants into the body and thus becomes their primary target. Air pollution affects virtually everyone but is considered to affect poorer people more than wealthier ones.

Air Pollution Effects on Humans

Air Pollution Effects: Sulfur Dioxide

Sulfur dioxide is a colorless gas with a pungent, suffocating odour. SO_2 is corrosive to organic materials and it irritates the eyes, nose and lungs; therefore it is quite a dangerous air pollutant. Sulfur is contained within all fossil fuels, and is released in the form of sulfur dioxide during fossil fuel combustion.

Because of the widespread use of fossil fuels, sulfur dioxide is among the most common air pollutants produced in every part of the planet. Sulfur dioxide may often act in synergy with other pollutants to produce certain air pollution effects. Let's summarize some sulfur dioxide pollution effects on human health:

- Irritation of eyes, nose, throat; damage to lungs when inhaled
- Acute and chronic asthma
- Bronchitis and emphysema
- Lung cancer

Air Pollution Effects: Nitrogen Dioxide

Nitrogen dioxide is a gas of reddish-brown color with a distinct sharp, biting odour. It is often analysed in conjunction with another nitrogen gas–nitric oxide. Together these two gases are referred to as NO_x. Combustion of fossil fuels always produces both NO_2 and NO. But almost 90% of the NO_x combustion product is in the form of NO which is then oxidized to NO_2 in the air. NO_x can destroy organic matter, ex. human tissue. Exposure to high concentrations of NO_x can make living organisms more susceptible to bacterial infections and lung cancer. Just like other pollutants, nitrogen dioxide affects people with existing medical conditions more severely than healthy people. For example, asthma sufferers may experience enhanced sensitivity after short-term NO_2 exposure as compared to those without any asthmatic problems. Another group at higher risk is children. For instance, children aged 12 and younger who are exposed to NOx suffer more respiratory illness than the ones who are not exposed.

Depending on different NO_2 concentrations in the air, nitrogen dioxide pollution effects may be as follows:

- Increased incidence of respiratory illness
- Increased airway resistance
- Damage to lung tissue
- Chronic obstructive pulmonary disease, or COPD
- Emphysema
- Pulmonary edema
- Infant and cardiovascular death

It is important to note that nitrogen dioxide is a major component of the photochemical smog and thus a contributor to the formation of ozone which is another serious air pollutant

Air Pollution Effects: Carbon Monoxide

Carbon monoxide is an extremely toxic gas which affects the ability of the body to receive oxygen. Hemoglobin which transports oxygen in the blood is bound by carbon monoxide, which leads to the shortage of oxygen in the body. Carbon monoxide is the most common type of fatal poisoning in many countries around the world. Unlike many other air pollutants, carbon monoxide does not directly affect eyes, nasal passages or lungs.

Exposure to carbon monoxide may lead to:

- Toxicity of the central nervous system and heart
- Headaches, dizziness, nausea and unconsciousness
- Loss of vision
- Decreased muscular coordination
- Abdominal pain
- Severe effects on the baby of a pregnant woman

- Impaired performance on simple psychological tests and arithmetic; loss of judgment of time
- In cases of prolonged exposure to high CO concentrations, unconsciousness, convulsions and death would occur

Carbon monoxide pollution effects are exacerbated in weaker people, specifically in those with heart and lung diseases. What makes this gas even more dangerous is the fact that it is invisible and odourless, and thus not immediately detectable as a hazard.

Air Pollution Effects: Ozone Effects

Ozone is a poisonous gas with a sharp and cold odour. It can be found in the stratosphere where it occurs naturally and plays a beneficial role by protecting the Earth from ultraviolet sunlight; and in the troposphere where it occurs naturally and also forms part of the human-caused photochemical smog. It is of course the tropospheric ozone that we are interested in as an important air pollutant. The chemical reaction between nitrogen oxides and volatile organic compounds in the presence of sunlight results in the photochemical smog; the tropospheric ozone is an end product of this reaction and a component of the smog itself. Because the photochemical smog requires a lot of sunshine to form, it occurs mostly in sunny and heavily polluted places, such as Los Angeles, Mexico City and even Athens. Ozone's main "victim" within the human body is its respiratory system. Once in the lungs, ozone burns through cell walls. The immune system fails to protect the lungs because ozone pushes the defensive cells back. As cellular fluid starts seeping into the lungs, breathing becomes rapid, shallow and painful. Exposure to ozone over long periods of time leads to a stiffening of the lungs and a reduced ability to breathe. As an example, a study conducted in California in the 1980s shows that children living in ozone-polluted areas have smaller than normal lungs and adults lose up to 75% of their lung capacity.

So, exposure to the tropospheric ozone may cause:

- Burning nose and watering eyes
- Tightening of the chest
- Coughing, wheezing and throat irritation

- Rapid, shallow, painful breathing
- Susceptibility to respiratory infections
- Inflammation and damage to the lining of the lungs
- Aggravation of asthma
- Fatigue
- Cancer

Air Pollution Effects: Ammonia Effects

Ammonia is a colorless, pungent, hazardous caustic gas composed of nitrogen and hydrogen. Though ammonia is used for different applications in many sectors, agriculture is its largest consumer and producer. Livestock farming, animal waste and fertilizer application are the biggest sources of atmospheric ammonia emissions within the agricultural sector. Gaseous ammonia is a dangerous air pollutant. Breathing in large amounts can cause death.

Ammonia Pollution Effects

On the respiratory system:

- Nose and throat irritation and burns
- Swelling of the throat and airways; airways destruction
- Pulmonary edema
- Chronic lung disease
- Cough
- Asthma
- Lung fibrosis
- Inhaling large amounts of ammonia can be fatal

On the skin and eyes:

- Skin burns
- Skin conditions, ex. dermatitis
- Burning sensation in the eyes
- Ulceration and perforation of the cornea; blindness
- Cataracts and glaucoma

Air Pollution Effects: Volatile Organic Compounds

Volatile organic compounds are defined as organic compounds which easily evapourate and enter the atmosphere. VOCs are important pollutants for two reasons. First, they are precursors to the formation of ozone; second, they include compounds which are carcinogenic and mutagenic in their own right. Typical VOCs include propane, benzene, ethanol, methanol, ether, carbon tetrachloride and vinyl chloride; substances such as petrol and resins contain many individual VOCs, and many others are produced during combustion processes.

Toxicity of some VOCs and ensuing health effects are no doubt issues of serious concern. For example, exposure to benzene and 1,3-butadiene is a suspected cause of around 10% of leukemia incidence in the UK. Exposure to toluene – another dangerous VOC – may lead to the dysfunction of the central nervous system resulting in behavioural problems, memory loss and disturbance of the circadian rhythm; toluene is also suspected to cause cancer.

Some other VOCs, ex. carbon tetrachloride and PCBs, are believed to produce abnormal changes in fetus development and consequently lead to birth defects. Carbon tetrachloride also leads to liver damage. Vinyl chloride causes Raynaud's phenomenon, necrosis of the small bones of the hand, liver damage, and a rare, highly malignant tumor of the liver.

Benzene pollution effects on human health:

- Short-term breathing of low levels of benzene can cause drowsiness, dizziness, rapid heart rate, headaches, tremors, confusion and unconsciousness; exposure to high levels of benzene may result in death
- *Chronic effects*:
  - Damage to the bone marrow
  - Decrease in red blood cells
  - Excessive bleeding and depression of the immune system increasing the risk of infection
  - Leukemia

Butadiene pollution effects on human health:

- Short-term exposure to high levels of 1,3-butadiene may lead to distorted blurred vision, vertigo, general tiredness, decreased blood pressure, headache, nausea, decreased pulse rate, and fainting
- Long-term exposure to high levels can increase the risk of cardiovascular diseases and cancer

Let us summarize some effects of VOCs on human health:

- Tiredness, vertigo, drowsiness, dizziness, nausea, confusion, unconsciousness
- Anemia
- Bone marrow damage
- Liver damage
- *Dysfunction of the central nervous system*:
  - Behavioural problems
  - Memory loss
  - Disturbance of the circadian rhythm
- Cardiovascular diseases
- Cancer; specifically leukemia
- Abnormal changes in fetus development, birth defects
- VOCs also contribute to sick building syndrome indoors
- As facilitators in ozone formation, VOCs may indirectly contribute to respiratory problems

Air Pollution Effects: Airborne Particles

Airborne particles are tiny fragments of solid or liquid nature suspended in the air. Particles may be primary – when emitted into the atmosphere directly by sources, or secondary – when formed in the atmosphere through the interaction of primary pollutants. For example, sulfur oxides and nitrogen oxides are primary pollutants which may transform in the air into sulfates and nitrates.

Dust particles are solid particles between 1 and 100 ìm in diametre; fumes, or smoke, are solid particles less than 1 ìm in diametre. Experts believe that dust is the most damaging among all widely measured air pollutants. Smaller dust particles are more dangerous than larger ones because they can penetrate deep into the lungs being deposited on areas where the body's natural clearance mechanisms such as coughing cannot remove them. Combustion of fossil fuels by road transport, power plants etc. is a major source of particulate air pollution. Particles may come in a whole variety of chemical compositions including heavy metals such as cadmium, mercury and lead. They often act in synergy with other air pollutants, ex. sulfur dioxide, to produce negative health effects in humans.

As particulate pollution levels rise, it is common to see an increase in hospital admissions for:

- Stuffy noses, sinusitis
- Sore throats
- Wet cough, dry cough, phlegm
- Head colds
- Burning eyes
- Wheezing; shortness of breath
- Lung disease
- Chest discomfort or pain

As with other pollutants, children are of course more susceptible to particulate pollution.

Specific children's disorders caused by airborne particles may include:

- Infant death
- Low birth weight
- Reduced lung function

Here is a summary of the particulate pollution effects provided by the US Environmental Protection Agency:

- Increased respiratory symptoms
- Decreased lung function
- Aggravated asthma
- Chronic bronchitis

- Irregular heartbeat
- Nonfatal heart attacks
- Premature death in people with heart or lung disease

Indeed, particulate pollution affects both developed and developing countries:

— For example, a major study conducted in the US between 1982 and 1989 found that people living in American cities with the highest levels of particulate air pollution were 15% to 17% more likely to die one to three years early

— In another example, particulate air pollution is linked to nearly 350,000 premature deaths in China and India every year

Air Pollution Effects on Animals

We don't know a lot about air pollution effects on animals.

- Probably one of the best examples here is that of acid rain and how it affects freshwater animal life.

Sulfur dioxide and nitrogen dioxide are transformed in the atmosphere to produce acid compounds–sulfuric and nitric acids. These compounds then fall back on to the ground as particulates or raindrops—in other words, acid rain. So acid rain also falls on streams and lakes, acidifies them and destroys fish life in these freshwater ecosystems. For example, in Sweden acid rain made over 18,000 lakes so acidic that all the fish died out. Salmon species appear to be particularly sensitive to acidity. Some other populations of animals in Europe and North America that have also been declining due to acid rain are brown trout, mayfly larvae, beetle larvae, mollusks, and aquatic bird species.

- Pollution may also affect animals through plants on which they feed.

For example, pea aphids feed on pea plants exposed to sulfur dioxide in the air. High exposure to sulfur dioxide negatively affects the health of the pea plants, and therefore, the health of the aphids as well.

Some other examples of air pollution effects on animals:

- Excessive ultraviolet radiation coming from the sun through the ozone layer in the upper atmosphere which is eroded by some air pollutants, may cause skin cancer in wildlife
- Tropospheric ozone may damage lung tissues of animals

It is also probably logical to assume that many higher order animals experience air pollution effects similar to those experienced by humans.

Air Pollution Effects on Forests, Trees and Plants

Air pollution can have both long-term and short-term effects on plants.

- Physical injury to leaves is the immediate effect of air pollution on plants. Here is how leaves are affected by different air pollutants:
  - Ozone produces a speckle of brown spots, which appear on the flat areas of leaf between the veins
  - *Sulfur dioxide*: Larger bleached-looking areas
  - *Nitrogen dioxide*: Irregular brown or white collapsed lesions on intercostal tissue and near the leaf edge
  - *Ammonia*: Unnatural green appearance with tissue drying out
- Of all main air pollutants, sulfur dioxide often comes up as the one that most negatively affects plants and trees.

Here is a very illustrative example of how destructive sulfur dioxide can be to vegetation. The development of ore deposits in Canada in the middle of the 20 th century was a major source of sulfur dioxide emissions in high concentrations. Sulfur emissions of one specific iron smelting plant in Wawa, Ontario, caused the destruction of large tracts of native boreal forest in a belt stretching some 30 km from the smelter. Another good example is that of the Black Triangle – the area where the borders of Germany, Poland and the Czech Republic meet, with significant brown coal deposits. Industrial activity in this area led to high sulfur emissions which resulted in devastation of many square kilometres of forest, particularly above an altitude of 750 m. However, it has been noted that sulfur dioxide concentrations in the atmosphere have recently been significantly reduced almost throughout the world and do not carry such dramatic effects as they used to during the 20th century. Lichens are considered to be most sensitive to sulfur dioxide. During the period of high levels of sulfur pollution, large parts of Europe lost many species of lichen and became known as "lichen deserts". But as sulfur pollution levels dropped, many lichen species re-appeared. Sulfur dioxide may also affect higher plants, including wild species, crops and trees.

These effects may be:

- Cell metabolism disruption
- Leaf injury and loss
- Reduced growth and reproduction
- Increase in susceptibility of plants to attacks by insect herbivores
- Nitrogen dioxide, another air pollutant, may act in synergy with sulfur dioxide to produce a negative effect on plants' photosynthesis.
- Tropospheric ozone can prevent plant respiration by blocking stomata and negatively affecting plants' photosynthesis rates which will stunt plant growth; ozone can also decay plant cells directly by entering stomata.
- Particles, just like ozone, often affect plants and trees via blocking of leaf stomata through which plants undertake the gas exchange necessary for photosynthesis and respiration.

Andrew Farmer notes that:

- In sufficient quantities, dusts may form a smothering layer on leaves, reducing light and hence lowering photosynthetic rates. Many dusts are inert and so only act by shading.

However, some dusts are also chemically active. Thus cement dust will also dissolve leaf tissue, resulting in additional injury. Coal dust may also contain toxic compounds. Dusts may also affect ecosystems through their action on soil. Thus the alkaline chemistry of limestone dusts can raise the soil pH of acid and neutral habitats, resulting in the loss of plant and animal species."

So, particulate air pollution effects on plants and trees may be as follows:

- Blocked stomata
- Increased leaf temperature
- Reduced photosynthesis
- Reduced fruit set, leaf growth, pollen growth
- Reduced tree growth
- Leaf necrosis and chlorosis, bark peeling
- Acid rain severely affects trees and plants as well.

It can kill trees, destroy the leaves of plants, can infiltrate soil by making it unsuitable for purposes of nutrition and habitation. It is also associated with the reduction in forest and agricultural yields.

Air Pollution Effects on Wider Environment

The effects of acid rain have been known for a long time. Though experts admit that the problem of acid rain is generally under control, it's still worth reiterating the effects brought on by acid rain on the wider environment.

Acid rain has adverse effects on:

- Forests and other vegetation
- Freshwater lakes and streams destroying aquatic life
- Soil
- Buildings and materials

Air pollution also has a negative impact on visibility.

Economic Losses as Air Pollution Effects

Apart from direct health-related and other environmental issues, air pollution brings with it economic losses as well.

Some of the economic losses caused by air pollution are as follows:

- Direct medical losses

- Lost income from being absent from work
- Decreased productivity
- Travel time losses due to reduced visibility
- Losses from repair of damage to buildings
- Increased costs of cleaning
- Losses due to damage to crops and plants

For example, experts calculated that California is losing $28 billion every year in health care costs, school absences, missed work and lost income potential from premature deaths, because of air pollution in Southern California and in the San Joaquin Valley.

FRONTS

Front is the transition zone between air masses with distinctly different properties.

The differences in density are most often caused by temperature differences. Separate air masses with different humidities as well. We identify fronts by the movement of this transition zone and the properties that move over a geographical location. What weather changes do you expect when TV weather person says a cold front is moving through the area?

How do you identify a front on a surface weather map or by your own weather observations? Look for:

- Sharp temperature changes over a relatively short distance
- Change in moisture content
- Rapid shifts in wind direction
- Pressure changes
- Clouds and precipitation patterns

A weather front is a transition zone between two different air masses at the Earth's surface. Each air mass has unique temperature and humidity characteristics. Often there is turbulence at a front, which is the borderline where two different air masses come together. The turbulence can cause clouds and storms.

Instead of causing clouds and storms, some fronts just cause a change in temperature. However, some storm fronts start Earth's largest storms. Tropical waves are fronts that develop in the tropical Atlantic Ocean off the coast of Africa. These fronts can develop into tropical storms or hurricanes if conditions allow.

Fronts move across the Earth's surface over multiple days. The direction of movement is often guided by high winds, such as Jet Streams. Landforms like mountains can also change the path of a front.

There are four different types of weather fronts: cold fronts, warm fronts, stationary fronts, and occluded fronts.

The term "front" was suggested by the Bjerkneses because the collisions of two air masses reminded them of a battlefront during a military operation. Fronts develop when two air masses with different temperatures and, in most cases, different moisture contents come into contact with each other. The result depends on the relative temperature and moisture content of the two air masses and the relative movement of the two masses.

When a mass of cold air moving across Earth's surface comes into contact with a warm air mass, the denser cold air mass may force its way under the lighter warm air mass. The boundary formed between these two air masses is a cold front. Cold fronts are generally accompanied by a decrease in atmospheric pressure and the development of large cumulonimbus clouds that bring rain and thunderstorms. Cold fronts are represented on weather maps as solid lines with solid triangles at regular distances along them. The direction in which the triangles point shows the direction in which the cold front is moving.

COLD FRONT

A cold front forms when a cold air mass pushes into a warmer air mass. Cold fronts can produce dramatic changes in the weather. They move fast, up to twice as fast as a warm front.

As the cold front passes, winds become gusty. There is a sudden drop in temperature, and also heavy rain, sometimes with hail, thunder, and lightning. Atmospheric pressure changes from falling to rising at the front. After a cold front moves through your area, you may notice that the temperature is cooler, the rain has stopped, and the cumulus clouds are replaced by stratus and stratocumulus clouds or clear skies.

On weather maps, a cold front is represented by a solid blue line with filled-in triangles along it, like in the map on the left. The triangles are like arrowheads pointing in the direction that the front is moving. Notice on the map that temperatures at the ground level change from warm to cold as you cross the front line.

8

Classification of Climates

CLIMATE CLASSIFICATION

The Köppen Climate Classification System is the most widely used system for classifying the world's climates. Its categories are based on the annual and monthly averages of temperature and precipitation. The Köppen system recognizes five major climatic types; each type is designated by a capital letter.

Climate classification, the formalization of systems that recognize, clarify, and simplify climatic similarities and differences between geographic areas in order to enhance the scientific understanding of climates. Such classification schemes rely on efforts that sort and group vast amounts of environmental data to uncover patterns between interacting climatic processes. All such classifications are limited since no two areas are subject to the same physical or biological forces in exactly the same way. The creation of an individual climate scheme follows either a genetic or an empirical approach.

Tropical Moist Climates

Tropical moist climates extend northward and southward from the equator to about 15 to 25° of latitude. In these climates all months have average temperatures greater than 18° Celsius. Annual precipitation is greater than 1500 mm. Three minor Köppen climate types exist in the A group, and their designation is based on seasonal distribution of rainfall. Af or tropical wet is a tropical climate where precipitation occurs all year long. Monthly temperature variations in this climate are less than 3° Celsius. Because of intense surface heating and high humidity, cumulus and cumulonimbus clouds form early in the afternoons almost every day. Daily highs are about 32° Celsius, while night time temperatures average 22° Celsius. Am is a tropical monsoon climate. Annual rainfall is equal to or greater than Af, but most of the precipitation falls in the 7 to 9 hottest months. During the dry season very little rainfall occurs. The tropical wet and dry or savanna (Aw) has an extended dry season during winter. Precipitation during the wet season is usually less than 1000 millimeters, and only during the summer season.

Dry Climates

The most obvious climatic feature of this climate is that potential evaporation and transpiration exceed precipitation. These climates extend from 20 - 35° North and South of the equator and in large continental regions of the mid-latitudes often surrounded by mountains.

Moist Subtropical Mid-Latitude Climates

This climate generally has warm and humid summers with mild winters. Its extent is from 30 to 50° of latitude mainly on the eastern and western borders of most continents. During the winter, the main weather feature is the mid-latitude cyclone. Convective thunderstorms dominate summer months. Three minor types exist: Cfa - humid subtropical; Cs - Mediterranean; and Cfb - marine. The humid subtropical climate (Cfa) has hot muggy summers and frequent thunderstorms. Winters are mild and precipitation during this season comes from mid-latitude cyclones. A good example of a Cfa climate is the southeastern USA. Cfb marine climates are found on the western coasts of continents. They have a humid climate with short dry summer. Heavy precipitation occurs during the mild winters because of the continuous presence of mid-latitude cyclones. Mediterranean climates (Cs) receive rain primarily during winter season from the mid-latitude cyclone. Extreme summer aridity is caused by the sinking air of the subtropical highs and may exist for up to 5 months. Locations in North America are from Portland, Oregon to all of California.

Moist Continental Mid-latitude Climates

Moist continental mid-latitude climates have warm to cool summers and cold winters. The location of these climates is pole ward of the C climates. The average temperature of the warmest month is greater than 10° Celsius, while the coldest month is less than -3° Celsius. Winters are severe with snowstorms, strong winds, and bitter cold from Continental Polar or Arctic air masses. Like the C climates there are three minor types: Dw - dry winters; Ds - dry summers; and Df - wet all seasons.

Polar Climates

Polar climates have year-round cold temperatures with the warmest month less than 10° Celsius. Polar climates are found on the northern coastal areas of North America, Europe, Asia, and on the landmasses of Greenland and Antarctica. Two minor climate types exist. ET or polar tundra is a climate where the soil is permanently frozen to depths of hundreds of meters, a condition known as permafrost. Vegetation is dominated by mosses, lichens, dwarf trees and scattered woody shrubs. EF or polar ice caps has a surface that is permanently covered with snow and ice.

General Considerations

The climate of an area is the synthesis of the environmental conditions (soils, vegetation, weather, etc.) that have prevailed there over a long period of time. This synthesis involves both averages of the climatic elements and measurements of variability (such as extreme values and probabilities). Climate is a complex, abstract concept involving data on all aspects of Earth's environment. As such, no two localities on Earth may be said to have exactly the same climate.

Nevertheless, it is readily apparent that, over restricted areas of the planet, climates vary within a limited range and that climatic regions are discernible within which some uniformity is apparent in the patterns of climatic elements. Moreover, widely separated areas of the world possess similar climates when the set of geographic relationships occurring in one area parallels that of another. This symmetry and organization of the climatic environment suggests an underlying worldwide regularity and order in the phenomena causing climate (such as patterns of incoming solar radiation, vegetation, soils, winds, temperature, and air masses). Despite the existence of such underlying patterns, the creation of an accurate and useful climate scheme is a daunting task.

APPROACHES TO CLIMATIC CLASSIFICATION

The earliest known climatic classifications were those of Classical Greek times. Such schemes generally divided Earth into latitudinal zones based on the significant parallels of 0°, 23.5°, and 66.5° of latitude (that is, the Equator, the Tropics of Cancer and Capricorn, and the Arctic and Antarctic circles, respectively) and on the length of day. Modern climate classification has its origins in the mid-19th century, with the first published maps of temperature and precipitation over Earth's surface, which permitted the development of methods of climate grouping that used both variables simultaneously.

Many different schemes of classifying climate have been devised (more than 100), but all of them may be broadly differentiated as either empiric or genetic methods. This distinction is based on the nature of the data used for classification. Empirical methods make use of observed environmental data, such as temperature, humidity, and precipitation, or simple quantities derived from them (such as evaporation). In contrast, a genetic method classifies climate on the basis of its causal elements, the activity and characteristics of all factors (air masses, circulation systems, fronts, jet streams, solar radiation, topographic effects, and so forth) that give rise to the spatial and temporal patterns of climatic data. Hence, while empirical classifications are largely descriptive of climate, genetic methods are (or should be) explanatory. Unfortunately, genetic schemes, while scientifically more desirable, are inherently more difficult to implement because they do not use simple observations. As a result, such schemes are both less common and less successful overall. Moreover, the regions defined by the two types of classification schemes do not necessarily correspond; in particular, it is not uncommon for similar climatic forms resulting from different climatic processes to be grouped together by many common empirical schemes.

Genetic classifications

Genetic classifications group climates by their causes. Among such methods, three types may be distinguished: (1) those based on the geographic determinants of climate, (2) those based on the surface energy budget, and (3) those derived from air mass analysis.

In the first class are a number of schemes (largely the work of German climatologists) that categorize climates according to such factors as latitudinal control of temperature, continentality versus ocean-influenced factors, location with respect to pressure and wind belts, and effects of mountains. These classifications all share a common shortcoming: they are qualitative, so that climatic regions are designated in a subjective manner rather than as a result of the application of some rigorous differentiating formula.

An interesting example of a method based on the energy balance of Earth's surface is the 1970 classification of Werner H. Terjung, an American geographer. His method utilizes data for more than 1,000 locations worldwide on the net solar radiation received at the surface, the available energy for evaporating water, and the available energy for heating the air and subsurface. The annual patterns are classified according to the maximum energy input, the annual range in input, the shape of the annual curve, and the number of months with negative magnitudes (energy deficits). The combination of characteristics for a location is represented by a label consisting of several letters with defined meanings, and regions having similar net radiation climates are mapped.

Probably the most extensively used genetic systems, however, are those that employ air mass concepts. Air masses are large bodies of air that, in principle, possess relatively homogeneous properties of temperature, humidity, etc., in the horizontal. Weather on individual days may be interpreted in terms of these features and their contrasts at fronts.

Two American geographer-climatologists have been most influential in classifications based on air mass. In 1951 Arthur N. Strahler described a qualitative classification based on the combination of air masses present at a given location throughout the year. Some years later (1968 and 1970) John E. Oliver placed this type of classification on a firmer footing by providing a quantitative framework that designated particular air masses and air mass combinations as "dominant," "subdominant," or "seasonal" at particular locations. He also provided a means of identifying air masses from diagrams of mean monthly temperature and precipitation plotted on a "thermohyet diagram," a procedure that obviates the need for less common upper-air data to make the classification.

CLIMATES, TROPIC AND ARCTIC

Variability — that is the outstanding characteristic of temperate-zone weather. Even the brief sea log just quoted shows how weather in the tropics is much the same from place to place and day to day; temperate-zone weather, conversely, is diverse and changeable. But had our voyage extended far enough northward, it would have reached another region of comparative weather sameness — cold sameness instead of warm sameness -the arctic zone.

The temperate zone can have no stable and typical weather of its own. To southward (in the northern hemisphere) is the steady, warm, moist climate of the tropics. To northward is the fairly steady, mostly cold, fairly dry climate of the arctic. (This picture is of course repeated, with reverse directions, in the southern hemisphere.) The temperate zone is a buffer region between tropic and arctic climates, a battleground on which the gigantic atmospheric forces of heat and moisture advance and retreat, an unstable compromise between irreconcilable extremes, swayed this way and that in its evanescent weather as tropic forces (warm air masses) or arctic forces (cold air masses) momentarily gain ascendancy.

Tropic Climate

Tropic wind-and-climate belts migrate somewhat, north and south, with the apparent north-south seasonal swing of the sun. To get within-the-tropics weather in its most typical form we must go well south of the Tropic of Cancer, nearly down to the equator — say to Panama in about lat. 8° N. In these latitudes there are only two main seasons — a winter dry season under the influence of the steady northeast trades, and a summer wet season brooded over by the fitful calms of the doldrums.

The outstanding features of tropical-marine climate are that the average temperature is practically constant from month to month (nor does it ever vary, by more than a few degrees, from day to day); and that the cloudiness and rainfall, always appreciable, reach high values during the wet-season doldrums invasion of summer and early fall.

Upper winds over Panama, at half-mile altitude, average northeast about 8 m.p.h. in the rainy season and north-northeast about 20 m.p.h. in the dry. At two miles altitude there is less seasonal change — the average wind only varies from east 10 m.p.h. (rainy) to cast-northeast 12 m.p.h. (dry). Upper-air characteristics over the Caribbean Sea, quite typical of tropical-marine climate in general, were studied during the 1939 hurricane season by a United States Weather Bureau expedition to Swan Island.

This four-mile-long islet, named after a nineteenth-century pirate, lies almost exactly in the middle of the Sea at about lat. 16° N., between Jamaica and British Honduras. Within four months, observers Rhamlow and Leech released some two hundred and fifty pilot balloons and one hundred and twenty radio-sounding balloons. They found that the trade wind extended well aloft, though with diminishing force, here as in Panama — prevailing easterlies all the way up to about two and a half miles altitude. But above this level, the upper winds are mostly out of the west and increase with altitude.

Surface temperatures at Swan Island, and all over the Caribbean area, average about +75° to 80° F. the year round. But at the base of the stratosphere, ten miles up in these latitudes, it is cold indeed — about — 105° to — 110° F. Thus in the thick tropical troposphere there is an average lapse rate of something like — 3½° F/1000 ft (about — 6° C/Km).

Together with plenty of warmth and moisture in the lower levels, this lapse rate means that conditional instability must extend through considerable air layers, particularly in the vertical-sun-warmed rainy season of summer and fall — and it is this conditional instability, together with inexhaustible supplies of moisture from the warm ocean, that produces the daily torrential showers and thundershowers, the regular afternoon downpours and the moist, misty mornings which can make equatorial regions such a continual sweat-bath for ill-adapted white men.

On or close to the ocean this deep-tropical weather is bearable, or perhaps even comfortable. But the steaming jungles inland, where temperatures can rise unchecked and where all breeze is choked off by the tangled uprearing of dense vegetation, are sometimes aptly described as 'green hells.'

As Julian Duguid wrote '... a rich eternal garment of green, dappled with golden sunspots... a dense, fever-stricken thicket shimmering in the heat with a perpetual glassy haze....'

However, by no means all the land areas within the tropics are covered with jungle — some of them are sandy and arid in the extreme, and doubly hot though not so muggy. (The word 'jungle' itself, originally, meant only 'waste land' in India.) And even in the fecund jungle areas of the tropics, such as Panama and northern South America, vegetation waxes lush and green with the rainy season, wanes sparse and tawny with the dry season.

There are also irregular year-to-year variations in moisture and prevailing winds that change the face of tropical nature. I got off a ship on the Colombian coast in August, 1939, and walked inland for a while, expecting to wade through wetseason green tangles of vines and trees peopled with tropic creatures — but the soil proved to be dry and sandy, the jungle a dried-up shrinkage of scrub, the creatures hidden away from the glaring vertical sun.

Arctic Climate

Before Admiral Peary of the United States Navy marched five hundred icy miles on snowshoes from Ellesmere Land to the North Pole in the early spring of 1909, and since his time, many expeditions have entered the northern arctic regions. No latter-day explorer, incidentally, has ever surpassed Peary's record of intelligent planning and courageous, competent performance. But the airplane has rung startling changes in arctic travel.

In May, 1937, four large air transports set down four Russian scientists and a dog, together with a little portable cabin and food and fuel for a year or more, on one of the uncertain drifting ice floes that cover the Arctic Ocean around the pole. Leader Papanin, astronomer-aerologist Federov, hydrologist Shirsov, and radioman Krenkel had hoped to remain at or near the earth's axis for a whole year of observations, but their ten-foot-thick ice floe soon began to drift (as Peary or Fridtjof Nansen could have foretold), at first very slowly, across the two-mile-deep polar basin towards Greenland. Then it moved, less gradually, southward along the east coast of that high continent, and finally sailed at twenty-five miles a day on southward. The floe-campers were rescued by icebreakers and airplanes twelve hundred miles south of the pole in February, 1938.

Papanin's expedition found the summer temperatures of the polar basin hovering mostly around + 30° to 46° F.; clear spells of a day or two alternated with dull, overcast, rainy-to-wetsnowy weather. Winds varied from calm to light or moderate, mostly out of the north or northeast. Much of the ice was melting in the warmer spells, adding to the troubles and perils of the ice-campers. There was always the danger that their floe would disintegrate and dump them unceremoniously into the Arctic Ocean.

When winter came they had drifted several hundred miles south of the pole itself, but winter climate around the polar basin is probably pretty much the same everywhere. Temperatures ranged all the way from 50° F. below zero to 30° or more above, in accordance with the sweep of unlike air masses and cyclones, which occasionally extends well into the Arctic.

For several months in midwinter the polar region is, of course, in continual darkness, with the sun visible at first in the spring only as a pale horizon glow around noon. On the thirty-first of December, strangely enough, the Russian wind-powered radio reported a temperature of + 43° F. over the arctic ice! But within a few hours, the mercury had tobogganed to a more usual value of -31° F.

The general north-polar region is, of course, about midway on the most direct (great-circle) air routes between Europe and Asia, and between eastern United States and Asia. Some years hence, perhaps, great 500-mile-an-hour strato-transports will be winging daily across these icy-watery wastes. Compared with other climatic regions, the polar area offers good flying conditions — during at least two or three days out of a week, that is, and particularly at higher levels. Spring, when the polar high-pressure area lies symmetrically over the geographic pole, and not over the continental 'cold poles' in Greenland and northern Siberia, is the best flying season. A sealed-cabin, supercharged stratoliner, in any case, would be well up in the low arctic stratosphere (above thirty thousand feet) most of the time. Aside from Ellesmere Land at the north tip of America, one of the best fixed bases for observing the climate of the north-polar basin is the small island of Franz Joseph Land in lat. 81° N. long. 60° E., about six hundred land miles from the pole. In 1932-33 the Russian scientist Moltchanov (incidentally one of the pioneers in the development of radio-sounding balloons) made both surface and upper-air observations on Franz Joseph Land for a year. Like the floe-campers, he found average summer surface temperatures around +30° to 40° F. Winter temperatures averaged around -10 F.

Above the ground, for the first mile, there was usually a temperature inversion amounting to 5° F. or so; above that height the lapse rate or vertical temperature gradient was always surprisingly uniform at -3° or -4° F. per thousand feet up to the low polar 'tropopause,' marking the beginning of the stratosphere at about five miles altitude in winter, six miles in summer. Temperatures at the stratosphere base averaged about -75° F. in winter, about -40° F. in summer. July and August were the warmest months, as usual in the northern hemisphere; December was the coldest month at the surface, February coldest aloft.

The southern hemisphere in general, being mostly water, shows more typical wind-and-climate belts than the northern hemisphere, which is all cluttered up with land masses that get too warm in summer and too cold in winter. By the same token, or a reverse one, the antarctic climate, founded upon a high, icecovered continent symmetric around the geographic pole, is more typically 'polar' than the arctic climate.

During the 1933-35 Byrd occupancy of 'Little America' in about lat. 79° S., long. 164° W., aerologists Haines and Grimminger recorded a great deal of antarctic weather. Mostly the average temperature was around +20° to 25° F. in summer (January), and around -35° to -40° in winter (July) — considerably colder, that is, than the north-polar climate. The coldest summer temperature was +12° on January 1; the warmest, +37° on December 22. In winter the mildest reading was +23° on July 27; the coldest was -71° on July 21. During a roaring blizzard in July the temperature fell to -65° with the wind sixty miles an hour out of the southeast — and wind, rather than temperature, is what really hurts in cold climates. These southeasterly blizzards in Antarctica correspond to the northeasterly gales around the north-polar basin.

High winds in Antarctica sometimes come from the north, just as north-polar gales can arise out of the south, but the colder blizzards from somewhere around the South Pole are more usual. Laurence M. Gould, Byrd's geologist and deputy in 1928-30, writes of one terrific southerly gale in the Rockefeller Mountains that first blew away the heavily packed snow wall in front of his firmly anchored airplane, and then finally blew the plane itself away and wrecked it. Some time before the maximum wind came, the airplane's air-speed metre indicated a steady velocity of ninety miles per hour, with gusts well over a hundred·

Conquest of Climate

The decade from 1900 to 1910 signalized an. amazing conquest of the tropics as a livable environment for civilized man. The years after 1940, likewise, may see a parallel conquest of the arctics.

Within the tropics, except in certain areas plagued by extreme and wearisome heat, general living conditions were easy enough — too easy for man's own good, in fact; and aside from air transport, radio communication, mechanical refrigeration, and other new conveniences, the conquest was nearly all medical. Such dread warm-country diseases as typhoid, cholera, and yellow fever were finally vanquished — at least in the case of selected, well policed areas.

The work of Walter Reed, his assistant surgeons, and the heroic American soldiers who voluntarily entered the valley of 'yellow jack's' macabre shadow -all this really conquered the tropics and equatorial regions for white men, and contributed more than engineering and steam shovels to the triumphant completion, in 1913, of the world's greatest inter-ocean canal. Within the arctics and around the poles, on the contrary, disease is unknown unless brought to the scene by human carriers, but living conditions (without many bothersome precautions) are unduly hard and exacting. Aside from the extremes of dark and cold and gale and blinding snow, the chief hazards are: difficulty in finding your way around (particularly in the long, cold, dark arctic night); difficulty in surviving ordinary mischances such as falling into water, badly tearing clothes, and so on, particularly if alone; difficulty in getting adequately balanced diet and sufficient ultra-violet radiation.

Yet even now, with our immense progress in transportation and communication, some of these difficulties can be largely overcome at a properly organized arctic station. Clothing must, of course, be of arctic type — layers of wool or fur inside for warmth, layers of canvas or leather outside for wind-breaking. Adequate shelter, plenty of fuel, and fresh food can now be transported easily by airplane to almost any arctic site. On location, howling arctic gales will provide plenty of electric power for lighting and radio communication, and perhaps for heating as well.

Many of the dark terrors of the outer arctic night, which have heretofore held men prisoned for months in their base camps (except in the full of the moon) can perhaps be conquered by such devices as more powerful and efficient portable searchlights (using super-hot filaments, say, or possibly fluorescent light-sources); portable radio-telephone sets for constant two-way communication with and direction-finding on one or more base stations; more compact and accurate instruments, and simpler methods, for celestial navigation.

Against the white, too-uniform snow-glare of the polar day, improved goggles (perhaps of polarizing type) may be employed. Lighter, stronger and warmer tents and clothing can perhaps be devised. Much of the heavy local leg-work of arctic exploration can now be done with snow-mobiles, either tread-driven or propeller-driven. By means of vitamin diets, ultra-violet sunlamps and radio-diathermy, human beings near the poles will perhaps be enabled to maintain a high state of health during the long arctic night, and to overcome quickly any respiratory diseases that do develop.

Eventually we may even see arctic health resorts, rest camps and play camps. Ellsmere Land in northern Canada (Peary's jumping-off place), less than three hours' flying time from the North Pole, is only twelve hours by present slow air liner from New York City. 'Polar Dude Camp on Ellsmere Land — spend next weekend here in the dark restfulness of the polar night under the flickering curtain of the glorious aurora borealis; pack ice exploring parties for the more active' — and so on. Such an advertisement may not startle our children twenty years hence.

TEMPERATE CLIMATE

We have said that the temperate zone is, in a climatic sense, nothing better than a battleground between tropic and arctic weather forces — between warm and cold air masses. Nevertheless the temperate zone (and in particular, the north-temperate zone) is that part of the earth in which civilized man evolved, and is probably the region most favorable to his continued progress and development.

On the *average*, of course, temperate temperatures are pleasantest and best for all human activities — neither too hot nor too cold — though this average may rarely be matched by the vagaries of daily weather. And even the extreme moodiness and changeability of temperate weather has its advantages. Variety in weather, at least as to the four- or five-day spells and sequences so common in middle latitudes, perhaps satisfies some deep, innate rhythmic urges in man which cannot be allowed to atrophy without intellectual and cultural and physical loss.

However such obscure causes may or may not operate, it is certain that most human progress up from the dumb-ape level — most human art and ingenuity and imagination, most human science and engineering and social organization — has been best nurtured so far, and probably will be best nurtured in the future, in temperate latitudes.

Among all the various meteorological factors that go to make up a climate, temperature is the one that most directly influences human health, happiness, and efficiency. How do various temperatures really feel to average human beings? Different degrees of heat of course feel quite different to different people, or to the same people under different conditions — in particular, under varying conditions of exertion. But it is possible to indicate average human reactions to temperature, and they are somewhat as follows:

High humidity makes warm temperatures feel warmer, by retarding evaporation from the body. High humidity also makes cold temperatures feel colder, by making the clothing conduct heat better away from the body.

Wind makes cold temperatures feel colder, by removing the cushion of warm air that tends to form around the body in the absence of wind. This wind-cooling effect is particularly powerful at below-zero temperatures. On the other hand, wind makes warm temperatures up to 100° F. feel less warm, by increasing evaporation from the body. But if the temperature is over 105°, wind increases the discomfort.

The optimum temperature range, 68°-70° F., feels more comfortable if the air is normally moist. For health and comfort, indoor relative humidity should be around sixty per cent. But in the average home or office, in winter, it is seldom more than twenty per cent. Evaporating pans on radiators help some, but not very much. Constant-duty electric evaporators, preferably combined with electric lights, might well be used in temperate-zone houses during the winter.

Today's temperature anywhere is indicated by the thermometer; tomorrow's (which in temperate or arctic latitudes may be shockingly different) is foretold by a good forecast. Beyond these limits, our best present guide to future temperatures is the 'normal.' To get the normal temperature for any particular day of the year at any particular place, the average temperatures for that day, observed over as long a span of years as possible, are all averaged together. The normal temperature is, of course, merely a fictitious quantity seldom or never duplicated by any actual daily average; for climatically speaking, we dwellers in the temperate zone are always traveling now south, now north with the ebb and flow of warm and cold air masses.

Is our North American temperate-zone climate changing? This question keeps recurring in weather talk, and most of the oft-quoted answers foist off something about 'colder winters in Grandfather's day'— a notion perhaps promoted by oldsters in support of their claims to general superiority. The rising generation, contrariwise, have been inclined to charge the whole idea off to dotage and general boastfulness.

A definitive answer has been given by J. B. Kincer of the United States Weather Bureau: For a good many years up through 1936 there has been a decided tendency to warmer and drier weather, the trends being especially marked during the past quarter of a century, notwithstanding an occasional bad flood or a severely cold winter. Take the winter season, for example: With the exception of those for 1917-18 and 1935-36, the winters for the past twenty-five years or more, considering the country as a whole, have been rather uniformly warmer than normal, and, on an annual basis, every year since 1929 has had above normal temperature.

Also, in the matter of rainfall, there has been an equally marked tendency to droughts in recent years. However, an examination of the longer weather records of the country, going back one hundred years or more, indicates that this does not represent a permanent change of climate, but rather a warm, dry phase of our normal climate, to be followed, doubtless, by a cooler, wetter phase, when there will be more rain in summer and lower temperatures in winter....

Since the turn of the century, Mr. Kincer goes on to say, there has been a distinct yearly trend towards warmer winters and drier summers all over the world — and this in spite of a few exceptionally cold winters here and there. In other words, there are long (say fifty-year) cycles in climate, just as there are short (say five-day) cycles in weather. If all these cycles were only considerate enough to maintain constant length, amplitude, and form, both weather forecasting and climate forecasting would be child's play. But the form, intensity, and length of weather and climate changes are all complexly variable. In any case the semipermanent (say 500-year) picture of our American climate shows, apparently, no appreciable change. But the geologic-time picture, measured in thousands or millions of years, would again show marked climatic cycles — the advance and retreat of the various ice ages.

MIDDLE LATITUDE CLIMATES

The middle latitudes are regions of great atmospheric restlessness and variability, dominated at the surface and in the upper atmosphere by westerly winds (Hare, 1960). The climate is controlled at the surface by a succession of cyclones and anticyclones, normally moving from west to east, that are steered by the upper flow. The fall of temperature toward the poles occurs not in a uniform manner, but with the strong thermal gradients concentrated in one or more narrow latitudinal bands, or fronts. Baroclinic conditions lead to the development of jet streams just below the tropopause (Barry and Carleton, 2001). The equatorward limit of middle latitude climates is often taken as the surface subtropical high-pressure belt.

The poleward limit is more diffuse and variable, although it is often marked by the subpolar lows. The latitudinal extent of the climatic zone will vary from month to month and year to year, depending on changes in the position of the bordering centers of action, but most frequently it occupies the zone between 35 and 56 degrees N and S. As the overall equator-to-pole temperature contrast decreases from winter to summer, so does the strength of the westerlies. Day-to-day weather at a particular location is much affected by the latitude and strength of the westerly current. The upper westerlies vary between two extreme states. At times the upper flow is little deformed by waves, and the flow is parallel to the lines of latitude. At such times of high zonal index the westerlies blow strongly over large longitudinal zones and sweep a succession of depressions eastward at high speeds with frequent rain and gales in middle latitudes. At other times the Long or Rossby waves in the upper westerlies become greatly amplified, resulting in a meridional flow, a condition described as low zonal index. If the tips of the waves become cut off, leaving cold upper pools with associated low pressure at the surface in low latitudes and warm upper highs and associated surface anticyclones at high latitudes, blocking develops.

Blocks have preferred seasonal and geographical incidences, being most common in spring and in the eastern Atlantic and over eastern Asia. Having once formed, blocks frequently are very persistent and may dominate the circulation for several weeks and on occasion for a whole season, introducing large temperature and precipitation anomalies into middle latitudes. Evidence exists of annual variations in blocking frequency and persistent anomalous climatic conditions, e.g. the mid-1970s European drought could have been caused by persistent or repetitive blocking episodes.

The strength of the circulation in any season is often expressed in relation to oscillations in the strength and position of the main centers of action, e.g. the Icelandic Low, and the Azores High. Such pressure oscillations include the North Atlantic Oscillation (NAO) and North Pacific Oscillation (NPO). In recent years the NAO, in particular, has often been strongly positive, implying strong westerlies in the Atlantic-European sector and a succession of mild winters in much of Europe (Perry, 2000).

Great differences occur in the geographical extent of land and sea areas in the two hemispheres. In the northern hemisphere midlatitudes contain the large landmasses of North America and Eurasia that lead to prominent contrasts in regional climate with important longitudinal variations. West coast locations are exposed to the predominant westerly winds and have equable temperature regimes, often with abundant precipitation throughout the year. The interiors of the continents have a more continental climate with greater annual temperature ranges and lower average precipitation totals, most of which falls in the summer season as a result of convective processes. In the southern hemisphere the much greater extent of sea area ensures that the westerlies blow more consistently and strongly (being known by a variety of names, including the Roaring Forties), and the circulation is less disturbed by periods of blocking than in the northern hemisphere. The kinetic energy of the southern westerlies is about 60% larger than the northern hemisphere westerlies. Smith (1967) has suggested that continentality and oceanicity are the principal criteria in defining subdivisions of climate in middle latitudes with general temperature levels forming a secondary subdivision. The adjective "temperate" is often applied to midlatitudes or used as a proxy term (e.g. Trewartha, 1968) but is certainly not suited to all climates in these latitudes as Bailey (1964) has shown.

Marine midlatitude climates

Along the western margins of the continents in the middle latitudes of both hemispheres, the climate is under a considerable maritime influence. In Western Europe the topography allows penetration of maritime airmasses deep into the continent in contrast to the situation in both North America and South America. A distinction is normally made between warm temperate or Mediterranean and cool temperate climates.

Mediterranean type

The approximate latitudinal extent of this type is on the order of 5–10 degrees between latitudes 30/35–40. A large proportion of the total precipitation falls in the winter when day-to-day weather is controlled by the behavior of the westerlies (Perry, 1981). During the hot dry summers the subtropical high-pressure area and its attendant ridges take control. In the type area (the Mediterranean Basin), substantial winter precipitation totals occur during low zonal index phases, when a meandering jet stream with a major trough exists over the Mediterranean, favoring cyclogenesis. Lee depressions form in the Gulf of Genoa and south of the Atlas Mountains and move in a generally eastward direction through the basin, often becoming reinvigo-rated in the east near Cyprus. Whereas the sea acts as a heat sink in summer, in winter it represents a heat source and cold air-masses entering the basin quickly become unstable after passage over the relatively warm water. The high intensity of rainfall is reflected in the small number of rain-days, even in the wetter areas. Regional winds, e.g. the cold Mistral and Bora, are related to meteorological and topographic factors, whereas the persistent northerly winds of summer in the eastern Mediterranean known as the etesians give this area a distinct climatic subtype. High annual sunshine totals are a feature of Mediterranean climates, with totals exceeding 3000 hours being quite common. In the northern hemisphere Southern California has this type of climate, as does the southern hemisphere coast of Chile around Santiago, the West Australian coast around Perth, and the South African coast around Cape Town.

Cool temperate

Poleward of the Mediterranean climate, on the west side of the continents, the climate is changeable throughout the year with well-distributed precipitation, brought by a series of depressions and their associated fronts and a predominance of maritime air masses. In northwestern Europe and on the Oregon and Washington coasts of North America, large positive temperature anomalies occur in winter compared with the average for the latitude, as maritime airmasses cross the warm-water currents in the North Atlantic and North Pacific. Since summers are relatively cool for the latitude, annual temperature ranges of less than 20°C are common. Because depressions are deeper and more vigorous in winter near the coast, there is normally a fall or winter precipitation maximum, whereas spring and early summer have a minimum of precipitation, reflecting both the increased frequency of blocking anticyclones in these seasons and the lower moisture content of maritime airmasses at this period due to the lower sea temperatures. High precipitation totals are a feature of all mountain areas and large lapse rates give such areas a short growing season and a cloudy, damp and often raw climate. A high degree of changeability from day to day is a characteristic of these climates as rapid alternations of airmasses occur, although occasionally persistence of a particular synoptic situation leads to more settled conditions. The coldest winter weather and the warmest summer spells develop when continental airmasses replace, for a time, the more usual airflow from oceanic sources. In the southern hemisphere sizeable belts of this climatic type occur in Chile, Tasmania, and the South Island of New Zealand.

Continental midlatitude climates

In both North America and Scandinavia the transition from maritime to continental climates is rapid due to the barrier effect of the mountains imposed on the invasion of surface maritime airmasses, but across the European Plain the transition is much more gradual. Meridional airmass movement is a particular characteristic of the North American climate and rapid changes of temperature level can occur, especially in winter, as frontal depressions cross the continent. Very severe winters occur at times with disrupting snowstorms, and are the result of strong amplification of the long waves with ridging over western North America and a deep trough in the east. This allows deployment of Arctic airmasses into the eastern states, while often in the west weather is mild and dry. Blizzard conditions develop on the cold polar side of traveling lows. In summer severe heatwaves can develop when warm air from the Gulf of Mexico is advected northward into the central and eastern states.

Over the USSR the winter circulation is dominated by the intense Siberian anticyclone, although the position and the intensity of this cold anticyclone can cause considerable departures of temperature values from the normal in individual years. Mean

January temperatures below −40°C occur in parts of Siberia, accompanied by dry sunny weather. During the short summers, temperatures can rise to 35°C on occasion, even as far north as the Arctic Circle. In these continental climates the transition seasons of spring and fall are very short. Although the seasonal variation of precipitation in the interior and eastern parts of both North America and Asia can be complex, there is normally a summer maximum brought on by instability showers and thunderstorms. Local variations in precipitation totals reflect such factors as the presence of large lakes, e.g. the Great Lakes of North America enhance snowfall totals on their eastern shores. In the United States the 51 cm annual isohyet follows approximately the 100° West Meridian, and in the dry area between the Rocky Mountains and this longitude occasional drought years occur, such as those resulting in the Dust Bowl conditions of the 1930s.

HYDROLOGIC IMPLICATIONS OF CLIMATE

Climate change is likely to alter the hydrologic cycle in ways that may cause substantial impacts on water resource availability and changes in water quality. For example, the amount, intensity and temporal distribution of precipitation are likely to change. Less dramatic but equally important changes in run-off could arise from the fact that the amount of water evaporated from the landscape and transpired by plants will change with changes in soil moisture availability and plant responses to elevated CO_2 concentrations. This will affect stream flows and groundwater elevations. This overview briefly summarizes potential impacts on the most important water resource elements.

Precipitation changes

Along with the projected future warming there will be changes in atmospheric and oceanic circulation, and in the hydrologic cycle, leading to altered patterns of precipitation and run-off. The most likely will be an increase in global average precipitation and evaporation as a direct consequence of warmer temperatures. Evaporation will increase with warming because a warmer atmosphere can hold more moisture and higher temperatures increase the evaporation rate. On average, current climate models suggest an increase of about 1%–2% per degree Celsius from warming forced by CO_2. An increase in global average precipitation does not mean that it will get wetter everywhere and in all seasons. In fact, all climate model simulations show complex patterns of precipitation change, with some regions receiving less and others receiving more precipitation than they do now; changes in circulation patterns will be critically important in determining changes in local and regional precipitation patterns.

Changes in precipitation frequency and intensity

Many have argued that, in addition to changes in global average precipitation, there could be more pronounced changes in the characteristics of regional and local precipitation due to global warming. For example, Trenberth *et al.* hypothesized that, on average, precipitation will tend to be less frequent, but more intense when it does occur, implying greater incidence of extreme floods and droughts, with resulting consequences for water storage. Thus, the prospect may be for fewer but more intense rainfall–or snowfall–events.

Changes in average annual run-off

Run-off changes will depend on changes in temperature and precipitation, among other variables. Arnell used several climate models to simulate future climate under differing emissions scenarios. The study linked these climate simulations to a large-scale hydrological model to examine changes in annual average surface run-off by 2050.

They found that all simulations yield a global average increase in precipitation, but likewise exhibit substantial areas where there are large decreases in run-off. Thus, the global message of increased precipitation clearly does not readily translate into regional increases in surface and groundwater availability.

Hydrological impacts on coastal zones

The IPCC Working Group II TAR identifies several key impacts of sea level rise on water providers located in coastal areas, including:

- Lowland inundation and wetland displacement,
- Altered tidal range in rivers and bays,
- Changes in sedimentation patterns,
- More severe storm surge flooding,
- Increased saltwater intrusion into estuaries and freshwater aquifers, and
- Increased wind and rainfall damage in regions prone to tropical cyclones.

Water quality changes

Where stream flows and lake levels decline, water quality deterioration is likely as nutrients and contaminants become more concentrated in reduced volumes. Warmer water temperatures may have further direct impacts on water quality, such as reducing dissolved oxygen concentrations. Prolonged droughts also tend to allow accumulation of contaminants on land surfaces, which then pose greater risks when precipitation returns.

At the other extreme, heavy precipitation events may result in increased leaching and sediment transport, causing greater sediment and non-point source pollutant loadings to watercourses. Floods, in particular, increase the risk of water source contamination from sewage overflows, agricultural land, and urban run-off.

Water storage and management

An intensified hydrological cycle could make reservoir management more challenging, because there is often a trade-off between storing water for dry period use and evacuating reservoirs before the onset of the flood season to protect downstream communities. Reservoirs have been usually sized to handle a certain amount of stream flow variability, determined from a relatively short historical record.

If the variability increases, reservoirs may be undersized to meet demands or adequately serve as flood protection. Thus, it may become more difficult to meet delivery requirements during prolonged periods between reservoir refilling without also increasing the risk of flooding. Earlier spring run-off from snowmelt is a likely manifestation of global warming. To the extent that adequate reservoir space is available, changing the operation procedures of reservoirs could mitigate some of these effects.

Groundwater changes

In many communities, groundwater is the main source of water for both irrigation and municipal and industrial demands. Generally there are two types of groundwater resources–renewable and non-renewable. Renewable groundwater is directly tied to near-surface hydrologic processes; it is thus intricately tied to the overall hydrologic cycle and could be directly affected by climatic change.

In many places, the overdraft of renewable groundwater aquifers occurs because the rate of withdrawal exceeds the rate of recharge. In fact, renewable groundwater supplies are often thought of being the same resource as surface water because they are so intertwined.

Thus, climate changes could directly affect these recharge rates and the sustainability of renewable groundwater. Non-renewable groundwater supplies are usually derived from deep earth sediments deposited long ago and so have little climatic linkage.

Water demand changes

Future climate change could affect municipal and industrial water demand. Municipal water demand–especially for garden, lawn, and recreational field watering–is affected by climate to a certain extent, but rates of water use are highly dependent on water resource regulations and local user education. Industrial use for processing purposes is relatively insensitive to climate change; it is conditioned by technologies and modes of use. Demand for cooling water would be affected by a warmer climate because increased water temperatures will reduce the efficiency of cooling, perhaps necessitating increased source water abstraction to meet cooling requirements (or, alternatively, changes in cooling technologies to make them more efficient).

Regional changes

Although the preceding sections postulate some expected hydrologic changes from global warming, these generalizations will not be applicable in all places and at all times. Watson *et al.* examined the regional impacts of climate change, with a particular focus on assessing vulnerability.

The report noted that more than 1 billion people do not have access to adequate water supplies, and that some 19 countries, primarily in the Middle East and Northern and Southern Africa, face severe water shortfalls. This number could double by 2025, in large part because of the increased demand caused by economic and population growth. Climate change could exacerbate the situation.

Watson *et al.* noted that many developing countries are particularly vulnerable to climate change because they already experience water shortfalls, being in arid and semi-arid regions. Many people derive their water from single-point systems, such as boreholes or isolated reservoirs.

This lack of "water diversification" increases people's vulnerability to water shortage. These systems do not have the redundancy necessary to minimize the risks during times of shortage.

Also, given the often limited technical, financial and management resources available to developing countries, adjusting to shortages or implementing adaptation measures can impose a burden on national economies. These small water supply systems are found in many parts of the world.

Persistent drops in water level in these systems could adversely affect the quality of water by increasing the concentrations of sewage waste and industrial effluents, thereby increasing the potential for outbreaks of disease and reducing the quantity of potable fresh water available for domestic use. There is evidence that flooding is likely to become a larger problem in many temperate and humid regions, requiring adaptation not only to droughts and chronic water shortages but also to floods and associated damage, raising concerns about dam safety.

Trenberth *et al.* hypothesized that global warming due to enhanced GHGs could increase the intensity of precipitation and reduce its frequency, which would be particularly problematic in regions with rapid changes in land use and land cover, because this would mean changes in surface run-off and groundwater recharge characteristics.

Flooding could be worse, accompanied by rapid drying and less overall water resource availability. The effects on water resources could be sufficient to lead to conflicts among users, regions and countries.

9

Earth's Climate System

Taking Earth's temperature was an early priority. Earth's climate system adjusts to maintain a balance between solar energy that reaches the planetary surface and that which is reflected back to space: a concept known to science as the "radiation budget." Clouds, dust, volcanic ash and airborne particulates also play a major role.

Climate System: "The climate system is the highly complex system consisting of five major components: the atmosphere, the hydrosphere, the cryosphere, the lithosphere and the biosphere, and the interactions between them. The climate system evolves in time under the influence of its own internal dynamics and because of external forcings such as volcanic eruptions, solar variations and anthropogenic forcings such as the changing composition of the atmosphere and land use change."

CLIMATE SYSTEM

The key to understanding global climate change is to first understand what global climate is, and how it operates. At the planetary scale, the global climate is regulated by how much energy the Earth receives from the Sun. However, the global climate is also affected by other flows of energy which take place within the climate system itself. This global climate system is made up of the atmosphere, the oceans, the ice sheets (cryosphere), living organisms (biosphere) and the soils, sediments and rocks (geosphere), which all affect, to a greater or less extent, the movement of heat around the Earth's surface.

The atmosphere plays a crucial role in the regulation of Earth's climate. The atmosphere is a mixture of different gases and aerosols (suspended liquid and solid particles) collectively known as air. Air consists mostly of nitrogen (78%) and oxygen (21%). However, despite their relative scarcity, the so-called greenhouse gases, including carbon dioxide and methane, have a dramatic effect on the amount of energy that is stored within the atmosphere, and consequently the Earth's climate.

These greenhouse gases trap heat within the lower atmosphere that is trying to escape to space, and in doing so, make the surface of the Earth hotter. This heat trapping is called the natural greenhouse effect, and keeps the Earth 33°C warmer than it would otherwise be. In the last 200 years, man-made emissions of greenhouse gases have enhanced the natural greenhouse effect, which may be causing global warming.

The atmosphere however, does not operate as an isolated system. Flows of energy take place between the atmosphere and the other parts of the climate system, most significantly the world's oceans.

For example, ocean currents move heat from warm equatorial latitudes to colder polar latitudes. Heat is also transferred via moisture. Water evaporating from the surface of the oceans stores heat which is subsequently released when the vapour condenses to form clouds and rain. The significance of the oceans is that they store a much greater quantity of heat than the atmosphere. The top 200 metres of the world's oceans store 30 times as much heat as the atmosphere. Therefore, flows of energy between the oceans and the atmosphere can have dramatic effects on the global climate.

The world's ice sheets, glaciers and sea ice, collectively known as the cryosphere, have a significant impact on the Earth's climate. The cryosphere includes Antarctica, the Arctic Ocean, Greenland, Northern Canada, Northern Siberia and most of the high mountain ranges throughout the world, where sub-zero temperatures persist throughout the year. Snow and ice, being white, reflect a lot of sunlight, instead of absorbing it. Without the cryosphere, more energy would be absorbed at the Earth's surface rather than reflected, and consequently the temperature of the atmosphere would be much higher.

All land plants make food from the photosynthesis of carbon dioxide and water in the presence of sunlight. Through this utilisation of carbon dioxide in the atmosphere, plants have the ability to regulate the global climate. In the oceans, microscopic plankton utilise carbon dioxide dissolved in seawater for photosynthesis and the manufacture of their tiny carbonate shells. The oceans replace the utilised carbon dioxide by "sucking" down the gas from the atmosphere. When the plankton die, their carbonate shells sink to the seafloor, effectively locking away the carbon dioxide from the atmosphere. Such a "biological pump" reduces by at least four-fold the atmospheric concentration of carbon dioxide, significantly weakening the Earth's natural greenhouse effect, and reducing the Earth's surface temperature.

COMPONENTS OF THE CLIMATE SYSTEM

The Biosphere

Life may be found in almost any environment existing on Earth. Nevertheless, in a discussion on the climate system, it is convenient to regard the biosphere as a discrete component, like the atmosphere, oceans and cryosphere. The biosphere, both on land and in the oceans, affects the albedo of the Earth's surface. Large areas of continental forest have relatively low albedos compared to barren regions such as deserts. The albedo of deciduous forests is about 0.15 to 0.18 whilst that of coniferous forests is 0.09 to 0.15. Tropical rainforest reflects even less energy, approximately 7 to 15% of that which it receives. In comparison, the albedo of a sandy desert is about 0.3. Clearly, the presence of the continental forests affect the energy budget of the climate system.

The biosphere also influences the fluxes of certain greenhouse gases such as carbon dioxide and methane. Plankton in the surface oceans utilise the dissolved carbon dioxide for photosynthesis. This establishes a flux of carbon dioxide, with the oceans effectively "sucking" down the gas from the atmosphere. On death, the plankton sink, transporting the carbon dioxide to the deep ocean. Such primary productivity reduces by at least four-fold the atmospheric concentration of carbon dioxide, weakening significantly the Earth's natural greenhouse effect.

The biosphere also influences the amount of aerosols in the atmosphere. Millions of spores, viruses, bacteria, pollen and other minute organic species are transported into the atmosphere by winds, where they can scatter incoming solar radiation, and so influence the global energy budget. Primary productivity in the oceans results in the emission of compounds known as dimethyl sulphides (DMSs). In the atmosphere these compounds oxidise to form sulphate aerosols called marine non-sea-salt (nss) sulphate. These nss sulphates act as condensation nuclei for water vapour in the atmosphere, thus allowing the formation of clouds. Clouds have a highly complex effect on the energy budget of the climate system. Thus changes in primary productivity in the oceans can affect, indirectly, the global climate system. There are, of course, many other mechanisms and processes which couple the biosphere with the rest of the climate system, but the discussion has illustrated the major influences of the biosphere upon the global climate system.

The Geosphere

The fifth and final component of the global climate system is the geosphere, consisting of the soils, the sediments and rocks of the Earth's land masses, the continental and oceanic crust and ultimately the interior of the Earth itself. These parts of the geosphere each play a role in the regulation and variation of global climate, to a greater or lesser extent, over varying time scales.

Variations in global climate over tens of millions or even hundreds of millions of years are due to modulations within the interior of the Earth. Changes in the shape of ocean basins and the size of continental mountain chains (driven by plate tectonic processes) may influence the energy transfers within and between the coupled components of the climate system.

On much shorter time scales physical and chemical processes affect certain characteristics of the soil, such as moisture availability and water run-off, and the fluxes of greenhouse gases and aerosols into the atmosphere and oceans. Volcanism, although driven by the slow movement of the tectonic plates, occurs regularly on much shorter timescales. Volcanic eruptions replenish the carbon dioxide in the atmosphere, removed by the biosphere, and emit considerable quantities of dust and aerosols. Volcanic activity can therefore affect the energy budget and regulation of the global climate system.

Oceans

The atmosphere does not respond as an isolated system. Like the atmosphere, the thermodynamic state of the oceans is determined by the transfer of heat, momentum and moisture to and from the atmosphere. Ignoring for the moment the other components of the climate system, these fluxes within this coupled ocean-atmosphere system exist in equilibrium.

Momentum is transferred to the oceans by surface winds, mobilising the global surface ocean currents. Surface ocean currents assist in the latitudinal transfer of sensible heat in a similar fashion to the process occurring in the atmosphere. Warm water moves poleward whilst cold water returns towards the equator. Energy is also transferred via moisture. Water evaporating from the surface of the oceans stores latent heat which is subsequently released when the vapour condenses to form clouds and precipitation.

The significance of the ocean is that it stores a much greater quantity of energy than the atmosphere. This is on account of both its larger heat capacity (4.2 times that of the atmosphere) and its much greater density (1000 times that of air). The vertical structure of the ocean Figure can be divided into two layers which differ in the scale of their interaction with the overlying atmosphere. The lower layer comprises the cold deep water sphere, making up 80% of the oceans' volume.

The upper layer, which has closest contact with the atmosphere, is the seasonal boundary layer, a mixed water sphere extending down only 100m in the tropics but several kilometres in polar regions. The seasonal boundary layer alone stores approximately 30 times as much heat as the atmosphere. Thus for a given change in heat content of the ocean-atmosphere system, the temperature change in the atmosphere will be around 30 times greater than that in the ocean. Clearly then, small changes to the energy content of the oceans could have considerable effects on global climate.

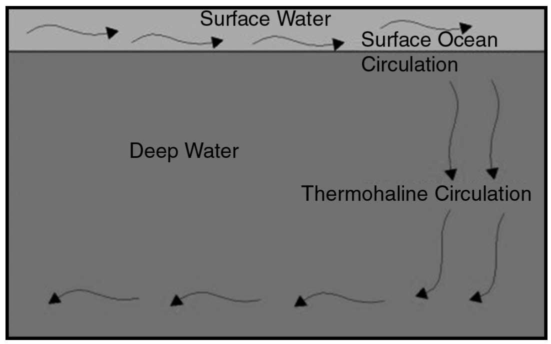

Fig. ***Vertical Structure and Circulation of the Oceans***

Energy exchanges also occur vertically within the oceans, between the mixed boundary layer and the deep water sphere. Sea salt remains in the water during the formation of sea ice in the polar regions, with the effect of increased salinity of the ocean. This cold, saline water is particularly dense and sinks, transporting with it a considerable quantity of energy. To maintain the equilibrium of water (mass) fluxes, a global thermohaline † circulation exists, which plays an important role in the regulation of the global climate.

The Cryosphere

The cryosphere consists of those regions of the globe, both land and sea, covered by snow and ice. These include Antarctica, the Arctic Ocean, Greenland, Northern Canada, Northern Siberia and most of the high mountain ranges throughout the world, where sub-zero temperatures persist throughout the year. The cryosphere plays another important role in the regulation of the global climate system.

Snow and ice have a high albedo † (reflectivity), that is they reflect much of the solar radiation they receive. Some parts of the Antarctic reflect as much as 90% of the incoming solar radiation, compared to a global average of 31%. Without the cryosphere, the global albedo would be considerably lower. More energy would be absorbed at the Earth's surface rather than reflected, and consequently the temperature of the atmosphere would be higher. Indeed, during the Cretaceous Period (120 to 65 million years ago) evidence suggests there was little or no snow and ice cover, even at the poles, and global temperatures were at least 8 to 10°C warmer than today. The cryosphere also acts to decouple the atmosphere and oceans, reducing the transfer of moisture and momentum, so stabilising the energy transfers within the atmosphere. The formation of sea ice in polar regions can initiate global thermohaline circulation patterns in the oceans, which greatly influence the global climate system. Finally, the presence of the cryosphere itself markedly affects the volume of the oceans and global sea levels, changes to which can affect the energy budget of the climate system.

ROLE OF CARBON DIOXIDE

We have already seen that carbon dioxide plays an important role in producing the greenhouse effect, which makes Earth a comfortable place to live, and that it does so by interfering with the return of energy to space from the Earth's surface by infrared radiation. As with most things in life that are necessary, useful and pleasant, we can have too much of a good thing. Carbon dioxide is now 30 per cent higher than it was 150 years ago. Carbon dioxide has never been as high as it is now for the last 400,000 years, as far as we know from the results of drilling polar ice in Antarctica. Also, the climate has never been as unusually warm as it has been in the last 20 years and, as far as we know, for even the last several thousand years. Only extreme skeptics, or the ignorant, would seriously question a connection between these two facts.

CO_2 in the Oceans

What exactly happens to the carbon dioxide introduced by burning wood, coal and hydrocarbons? This question is easy to ask but actually difficult to answer. At some point, some fifty years ago, it was thought that the buildup of carbon dioxide in the atmosphere would be rather slow because there is 50 times more carbon dioxide in the ocean than in the air. Scientists expected that the ocean and the atmosphere would share the newly introduced carbon dioxide, and that by far the greater proportion would be accommodated by the ocean, with its much larger capacity. Although this is not an unreasonable expectation, the fact to keep in mind is that carbon dioxide is first introduced into the atmosphere and from there it has to be passed along to the ocean. Carbon dioxide can only enter the ocean at its surface since the deep ocean is not immediately accessible. To fully reach the deep ocean takes a molecule of carbon dioxide about a thousand years. On the other hand, the uppermost ocean layer, called the "mixed layer," can be accessed on a 10-year time-scale.

These facts were first established by measuring the radiocarbon content of the ocean, from surface waters to great depths, by Hans Suess at UCSD. Radiocarbon (^{14}C) is made in the atmosphere, from bombardment of nitrogen atoms with cosmic rays. This radiocarbon reacts to make up a certain percentage of the carbon dioxide in our atmosphere. Half of a given number of radiocarbon atoms decay back to nitrogen within 6000 years. Radiocarbon in the ocean can only come from the atmosphere. Since it constantly disappears by decay, the amount of radiocarbon within the ocean depends on how fast the carbon dioxide from the atmosphere penetrates into the ocean and replenishes the store of radiocarbon there.

The fact that much of the ocean is not readily available to take up the additional carbon dioxide introduced through human activities raised a warning flag in the scientific community. However, another factor of equal importance emerged when considering how carbon dioxide enters the ocean. Seawater is somewhat reluctant to take on more carbon dioxide than it already has. The reason is that carbon dioxide reacts with water to make carbonic acid. In turn, the presence of this acid tends to discourage acceptance of additional carbon dioxide. This back-pressure effect has the consequence that the ability of the water to hold carbon dioxide increases by only about 1% for each 10% increase of the gas in the atmosphere. This ratio is widely known as the "Revelle buffer factor," after Dr. Roger Revelle who drew attention to this complication in the mid-1950.

There is one way to decrease this carbon dioxide back-pressure and that is to change the ocean's chemistry by dissolving calcium carbonate. Carbonate is plentiful in the shells of marine organisms, including coral reef materials. This has led a number of scientists to raise the yet unanswered question of whether the continued addition of carbon dioxide to the atmosphere (which adds acid to the upper ocean) will eventually result in damage to coral reefs.

Besides radiocarbon (^{14}C), the distribution of the stable isotopes of carbon atoms (^{13}C and ^{12}C) has been very useful for investigating the rate at which carbon dioxide invades the ocean. The materials burned to produce energy (such wood, coal, and hydrocarbons) have more of the carbon-12 isotope than the CO_2 normally dissolved in the ocean. Thus, a change in the ratios of carbon these two stable isotopes (that is, $^{13}C/^{12}C$) in seawater contains clues to the progressive invasion of the industrial carbon dioxide from the atmosphere into the upper ocean. The isotopic change of seawater is recorded in the isotopic composition of the skeletons of marine organisms, seashells, corals and certain sponges. We will learn more about the way carbon isotopes have helped unravel climate change in later lessons.

Human Emissions of Carbon Dioxide

The rise of carbon dioxide in the atmosphere is readily documented by direct measurements (i.e. the "Keeling" curve) and by the results of ice core studies (from the polar regions). However, the precise contribution to this rise from human-caused emissions of carbon dioxide is not so readily determined. Estimation of future carbon dioxide emissions from fossil fuelscan be accomplished by analyzing the production and use of coal, petroleum and hydrocarbon gas as reported by multinational corporations and by various nations and then applying the appropriate statistics. Other smaller contributions to emissions like cement production by heating of carbonate, which also releases carbon dioxide, can also be estimated with some confidence.

Big uncertainties arise from estimating the contributions from burning of wood and dung as fuel and from destruction of grasslands and forests and deforestation in general. The reason is that most of such burning goes on at a small scale in a large number of poorly developed countries and is not reported with a high degree of accuracy. Deforestation, which is proceeding quite rapidly in many countries, can be monitored best by using spacecraft. However, the conversion of images from space of burning forests or changes in forest cover into units of "tons of carbon released" is not straightforward.

Another even more important asymmetry is the use of energy. The industrialised countries use considerably more energy per capita than the less developed countries and have correspondingly higher emissions. Energy use goes parallel with standard of living. Some of the more advanced countries (e.g. Sweden) achieve the same standard of living with less energy use, largely through higher efficiency of use. Not knowing the total emissions very well, and knowing the rates of change even more poorly, makes it difficult to extrapolate into the future.

Again, these uncertainties mainly result from the lack of knowledge about burning of materials other than fossil fuel and about the destruction of forests — the use of coal and hydrocarbons is reasonably well known. Thus, the economic behaviour and emissions of industrial countries can be quite well monitored, while this is not true for many of the less developed countries. This asymmetry in knowledge (and therefore potential control) complicates the process of reaching political agreements on emission controls.

Development of a "New" Scientific Field

Life on our planet is made up of an incredible variety of carbon molecules, and, in essence, life processes are carbon chemistry. Conversely, the carbon cycle, which can be viewed as the movement of carbon atoms through various places of storage on Earth, called "reservoirs," is intimately tied to life processes. In the study of the carbon cycle, biology and geochemistry merged to form a new scientific discipline called "biogeochemistry."

Biogeochemists study the carbon cycle and its interconnections with the cycles of other elements involved in life processes, mainly nitrogen, oxygen and phosphorus, but also sulfur and iron and certain trace metals. In addition, the water cycle helps drive the carbon cycle, and this is where climate and the carbon cycle are most intimately connected.

Biogeochemistry includes the history of the great carbon reservoirs in the crust of the Earth, like limestone rocks and the coal deposits, as well as the distribution of nitrate and phosphate in the ocean. It seeks to explain the composition of the atmosphere (consisting mainly of nitrogen and oxygen, as well as other trace gases) as a result of bacterial action and photosynthesis.

And biogeochemistry records the exchange of matter at the interfaces: the decay of organic matter in soils and resulting gases released into the air; the uptake of oxygen by the ocean and its utilization at depth; and the leaching of nutrients from the soil and their transport into the sea.

The all-important role of life processes in maintaining Earth's environments was stressed early in the 20th century by the Russian mineralogist, Vladimir Vernadsky, who may be considered the father of biogeochemistry. The American limnologist and geochemist G. Evelyn Hutchinson) led the way and first outlined the broad scope and principles of this new field. More recently, the basic elements of the discipline of biogeochemistry have been restated and popularized by the British engineer and science writer,James Lovelock, under the label of the "Gaia Hypothesis." Lovelock emphasizes a concept that life processes regulate the radiation balance of Earth to keep it habitable.

OCEAN CIRCULATION AND CLIMATE

When talking about the greenhouse effect, we have considered the fact that water on Earth occurs in all its three phases — gaseous, fluid and solid. As vapour in the air, water provides the strongest warming effect of all the greenhouse gases. As a fluid, it covers some seventy per cent of the Earth's surface, and if spread evenly over the planet, it would make a layer two miles thick. As a solid, it whitens the polar regions and the high latitudes in winter, and greatly increases the reflectivity of Earth's surface wherever it covers the ground. Ice also forms a protective layer on water wherever it covers the sea surface.

As we learned in the last lesson, the physical properties of water make it an ideal means for conditioning the climate on Earth. It stores heat better than most other substances – that's why tea takes some time to make, but then stays hot for quite a while. It takes up heat when moving from solid to fluid or from fluid to gas, and releases that heat again when moving through the phases in the opposite direction.

The "latent" heat thus stored in water vapour is freed up with precipitation (and can then drive storms). This transfer of energy by latent heat is a hundred times more efficient than transfer by sensible heat, gram for gram.

Quite generally, the ocean surface is a source for heat in the atmosphere. The sea surface tends to stay at the same temperature for a long time, because of the high specific heat off water and the thickness of the mixed layer. Thus, the ocean surface stabilises the atmospheric flow patterns, which depend on heat supply. Without this stabilizing influence of the ocean, the weather would be much more fickle.

The general circulation of the atmosphere discussed esrlier is closely related to the general circulation of the ocean. Winds blowing over the sea surface produce ocean currents. Winds also evaporate water, which precipitates elsewhere as rain. A large amount of heat is transferred to the atmosphere in this process. Evaporation and precipitation also affect the patterns of water density, which are inseparably tied to ocean circulation.

But how will the ocean's circulation respond to climate change? Although this sounds like a simple question, answering it is difficult. The ocean responds largely to winds, and wind fields respond to the distribution of heat, which depends on the ocean circulation. Within a logic loop like this, where the final answer depends on the initial guess, any small errors can be readily amplified into large errors in the prediction game.

THE NATURE OF EARTH SCIENCE

Science may be stimulated by argument and debate, but it generally advances through formulating hypotheses clearly and testing them objectively. This testing is the key to science. In fact, one philosopher of science insisted that to be genuinely scientific, a statement must be susceptible to testing that could potentially show it to be false. In practice, contemporary scientists usually submit their research findings to the scrutiny of their peers, which includes disclosing the methods that they use, so their results can be checked through replication by other scientists. The insights and research results of individual scientists, even scientists of unquestioned genius, are thus confirmed or rejected in the peer-reviewed literature by the combined efforts of many other scientists. It is not the belief or opinion of the scientists that is important, but rather the results of this testing. However, that one opposing scientist would have needed proof in the form of testable results.

Thus science is inherently self-correcting; incorrect or incomplete scientific concepts ultimately do not survive repeated testing against observations of nature. Scientific theories are ways of explaining phenomena and providing insights that can be evaluated by comparison with physical reality.

Each successful prediction adds to the weight of evidence supporting the theory, and any unsuccessful prediction demonstrates that the underlying theory is imperfect and requires improvement or abandonment. Sometimes, only certain kinds of questions tend to be asked about a scientific phenomenon until contradictions build to a point where a sudden change of paradigm takes place. At that point, an entire field can be rapidly reconstructed under the new paradigm.

Despite occasional major paradigm shifts, the majority of scientific insights, even unexpected insights, tend to emerge incrementally as a result of repeated attempts to test hypotheses as thoroughly as possible. Therefore, because almost every new advance is based on the research and understanding that has gone before, science is cumulative, with useful features retained and non-useful features abandoned. Active research scientists, throughout their careers, typically spend large fractions of their working time studying in depth what other scientists have done. Superficial or amateurish acquaintance with the current state of a scientific research topic is an obstacle to a scientist's progress. Working scientists know that a day in the library can save a year in the laboratory. Even Sir Isaac Newton wrote that if he had 'seen further it is by standing on the shoulders of giants'. Intellectual honesty and professional ethics call for scientists to acknowledge the work of predecessors and colleagues. The attributes of science briefly described here can be used in assessing competing assertions about climate change. Can the statement under consideration, in principle, be proven false? Has it been rigorously tested? Did it appear in the peer-reviewed literature? Did it build on the existing research record where appropriate? If the answer to any of these questions is no, then less credence should be given to the assertion until it is tested and independently verified.

The IPCC assesses the scientific literature to create a report based on the best available science. It must be acknowledged, however, that the IPCC also contributes to science by identifying the key uncertainties and by stimulating and coordinating targeted research to answer important climate change questions.

A characteristic of Earth sciences is that Earth scientists are unable to perform controlled experiments on the planet as a whole and then observe the results. In this sense, Earth science is similar to the disciplines of astronomy and cosmology that cannot conduct experiments on galaxies or the cosmos. This is an important consideration, because it is precisely such whole-Earth, system-scale experiments, incorporating the full complexity of interacting processes and feedbacks, that might ideally be required to fully verify or falsify climate change hypotheses.

Nevertheless, countless empirical tests of numerous different hypotheses have built up a massive body of Earth science knowledge. This repeated testing has refined the understanding of numerous aspects of the climate system, from deep oceanic circulation to stratospheric chemistry. Sometimes a combination of observations and models can be used to test planetary-scale hypotheses. For example, the global cooling and drying of the atmosphere observed after the eruption of Mt. Pinatubo provided key tests of particular aspects of global climate models.

Another example is provided by past IPCC projections of future climate change compared to current observations. Figure reveals that the model projections of global average temperature from the First Assessment Report were higher than those from the Second Assessment Report.

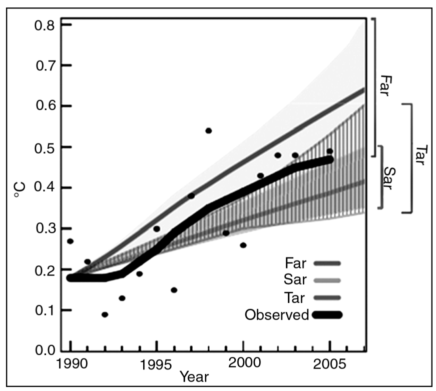

Fig. ***Yearly Global Average Surface Temperature***

Subsequent observations showed that the evolution of the actual climate system fell midway between the FAR and the SAR 'best estimate' projections and were within or near the upper range of projections from the TAR.

Not all theories or early results are verified by later analysis. In the mid-1970s, several substances about possible global cooling appeared in the popular press, primarily motivated by analyses indicating that Northern Hemisphere temperatures had decreased during the previous three decades. In the peer-reviewed literature by Bryson and Dittberner reported that increases in carbon dioxide should be associated with a decrease in global temperatures. When challenged by Woronko, Bryson and Dittberner explained that the cooling projected by their model was due to aerosols produced by the same combustion that caused the increase in CO_2.

However, because aerosols remain in the atmosphere only a short time compared to CO_2, the results were not applicable for long-term climate change projections. This example of a prediction of global cooling is a classic illustration of the selfcorrecting nature of Earth science. The scientists involved were reputable researchers who followed the accepted paradigm of publishing in scientific journals, submitting their methods and results to the scrutiny of their peers and responding to legitimate criticism.

A recurring theme throughout this stage is that climate science in recent decades has been characterised by the increasing rate of advancement of research in the field and by the notable evolution of scientific methodology and tools, including the models and observations that support and enable the research.

During the last four decades, the rate at which scientists have added to the body of knowledge of atmospheric and oceanic processes has accelerated dramatically. As scientists incrementally increase the totality of knowledge, they publish their results in peer-reviewed journals. Between 1965 and 1995, the number of substances published per year in atmospheric science journals tripled.

10

Moisture in Environment

INTRODUCTION

Drought as an environmental factor in crop production is not taken in its -catastrophic connotation. Catastrophic droughts as a problem in agriculture are a dealt with at the political, engineering or the economical levels. For the plant breeding program drought is defined as insufficient moisture supply which causes a reduction in plant production. It is the gap between crop demand for water and the supply of water. The agricultural drought on the larger scale of the region is approached at the meteorological level where the Palmer Drought Severity Index is the classical estimate. It is based on precipitations and temperature and is useful only on a longer time span such as months or weeks.

Fog is a collection of water droplets or ice crystals suspended in the air at or near the Earth's surface. While fog is a type of a cloud, the term "fog" is typically distinguished from the more generic term "cloud" in that fog is low-lying, and the moisture in the fog is often generated locally (such as from a nearby body of water, like a lake or the ocean, or from nearby moist ground or marshes).

Fog is distinguished from mist only by its density, as expressed in the resulting decrease in visibility: Fog reduces visibility to less than 1 km (5/8 statute mile), whereas mist reduces visibility to no less than 1 km (5/8 statute mile). For aviation purposes in the UK, a visibility of less than 2 km but greater than 999 m is considered to be mist if the relative humidity is 95% or greater - below 95% haze is reported.

The foggiest place in the world is the Grand Banks off the island of Newfoundland, the meeting place of the cold Labrador Current from the north and the much warmer Gulf Stream from the south. Some of the foggiest land areas in the world include Argentia, Newfoundland and Labrador and Point Reyes, California, each with over 200 foggy days per year. Even in generally warmer southern Europe, thick fog and localized fog is often found in lowlands and valleys, such as the lower part of the Po Valley and the Arno and Tiber valleys, as well as on the Swiss plateau, especially in the Seeland area, in late autumn and winter.

Other notably foggy areas include coastal Chile (in the south), coastal Namibia, and the Severnaya Zemlya islands. Seattle, Washington, USA, has many foggy days per year. Such ordinary aerological processes as the formation of rain are now understood almost as well as any laboratory experiment. First causes in the sun or elsewhere may still be obscure, but the general earthly mechanism of fogs and clouds and showers is reasonably clear.

DEFINITION OF MOISTURE

Moisture is presence of water in the air. Water is a combination of two molecules of hydrogen and one molecule of oxygen. This bond is so strong that even after billions of years, it will remain intact.

Facts and Importance of Moisture

70% of the Earth is covered with water. Moisture or water in the form of gas is a crucial and necessary element in our Earth's atmosphere.

Moisture is everywhere. It causes humidity, making us feel sticky and hot. It creates clouds due to which rainfalls occur and also hail storms and sleets which make driving impossible. The presence of excessive moisture can result in severe damage by triggering strong storms like Hurricane Katrina; which uprooted houses and had cars flying.

In deserts, plants wither up due to minimal presence of moisture in the air. In rain forests, moisture content is high; leading to intensive growth of vegetation.

Moisture is a form of water vapor. Water through the process of evaporation gets absorbed as moisture in the air. Imagine feeling hot and sweating; your body emits water molecules which are absorbed in the air as moisture and during that process a small amount of heat is gathered from your body making you feel cool. But there is a limit to how much moisture the atmospheric air can hold. After a while, the air gets crammed with moisture resulting in the air being saturated. When it is windy, the air expands and can take in more moisture. Thus we always love windy days, as the moisture makes us feel cool.

Humidity is defined as the amount of moisture in the air. When we say a city like Kolkata has 80% humidity, we mean that the air contains 80% moisture.

CUMULUS

Cumulus clouds are thick, dense, detached 'woolpacks' or 'puffballs,' usually as high as they are wide. Their bases, as we have seen, are usually flat and uniformly fixed at the condensation level, but their upper parts are dome-shaped and variously bulbed and rounded, like cauliflowers. Their colour ranges through every nuance of sunlight and shadow from dazzling white above to dark gray below.

It will bear repeating that a detached cumulus cloud always marks the top of a rising air current, which perhaps originated in the first place from an overheated portion of the earth's surface, or from some updraft in the general air flow.

In any growing cumulus cloud conditional instability is being realised and condensation of moisture is contributing its latent heat to make the updrafts more violent. If moisture is plentiful and the instability extends upward to great heights, the cumulus grows into a cumulo-nimbus, or thunderstorm. But if an ordinary growing cumulus bumps into a very stable layer such as a deep temperature inversion, its top flattens and its growth stops.

Most ordinary fair-weather cumulus clouds, formed under clear skies at levels between 2000 and 8000 feet (600 and 2500 meters), wax and wane daily with the sun (or with surface temperature). Foul-weather cumulus clouds, on the other hand, occur at any time of day along cold fronts and within moist, unstable air masses. A torn, immature, or partly evaporated cumulus is sometimes called fracto-cumulus. Sometimes cumulus clouds on the lee side of a mountain range evaporate in the dynamically heated downdraft so rapidly that they seem to disappear as if by magic.

Strato-cumulus

Strato-cumulus usually occurs in the form of a wide and fairly level layer marked by thick rolls and dark, rounded masses of cloud. It somewhat resembles alto-cumulus, but its cloud masses are usually larger, thicker, and darker. Sometimes there are small or large open breaks between the rolls and masses; sometimes the layer is continuous but steeply undulating, especially along its upper surface as seen from an airplane. Strato-cumulus clouds usually occur in the same levels as fair-weather cumulus — say 3000 to 8000 feet (1000-2500 meters) — and may be formed by the flattening of cumulus clouds, instability in a stratus layer, or wave motion between two differently flowing layers of air. Strato-cumulus can change into, or develop from, stratus.

Nimbo-stratus. This cloud type, formerly called plain 'nimbus,' is in general any low, thick, formless, dark-gray cloud layer from which rain, snow or other precipitation is actually falling. Even if the rain momentarily stops, this cloud has a ragged 'rainy' look, and is still called 'nimbo-stratus.' Below the main cloud deck there may be various ragged 'scud' clouds (fracto-stratus or fracto-cumulus) that occasionally merge into nearly solid lower layers or into the main cloud deck above. Nimbo-stratus transmits a feeble illumination that seems to come from within the cloud itself.

Usually nimbo-stratus develops out of alto-stratus and rainformed lower fracto-stratus or 'scud,' marking the near approach of a warm front in the central region of a cyclone. Less frequently, nimbo-stratus may develop out of heavy and threatening strato-cumulus. In either case the nimbo-stratus cloud, at heights of perhaps 1000 to 5000 feet (300 to 1500 meters), results from widespread condensation in air forced to rise by reason of its movement around or towards the centre of a cyclone. Scattered areas of nimbo-stratus may develop from cumulonimbus, as in the rear parts of a thunderstorm.

Cumulo-Nimbus

This is a great cumulus cloud mass in which condensation of moisture, release of latent heat, and increasing updrafts (fed by a deep layer of moist and unstable air) have grown to gigantic proportions. Its colour varies all the way from alabaster whiteness on the upper turrets to dark gray, black, or dark greenish below.

The whole cloud mountain may tower from some ordinary cumulus level, such as half a mile, up to a height of five, six or even seven miles. In a broad sense any large cumulus mass producing a shower can be called cumulonimbus, and this type is the common shower cloud, with or without thunder, of the tropics. But in temperate latitudes, cumulo-nimbus connotes more specifically a large, towering thunderstorm cloud whose top, blown out into anvil shape and cirrus-like texture by substratospheric winds, consists mostly of ice crystals. Hoods of alto-stratus may mark temperature inversions that the growing cloud mountain breaks through.

Special Cloud Forms

One special cloud form worth mentioning is mammato-cumulus or 'rain balls,' whose cumulus heads bulge downward instead of upward from a high cloud deck. It usually appears as a multitude of small, dark downward bulges in a turbulent layer of alto-stratus, alto-cumulus, or stratocumulus. The pendulous cloud pocks are evidently caused by local cold downdrafts perhaps started by detached showers of falling hail or snow. Mammato-cumulus always indicates very disturbed and unstable conditions in the upper air, seldom occurs far from a thunderstorm, and sometimes precedes tornadoes.

Another interesting cloud form is the detached, lens-shaped ovals that are sometimes formed at considerable heights out of quiescent or slowly dissolving cloud masses (alto-stratus 'lenticalus,' and so on). Sometimes the crest cloud on a mountain summit, caused by the steady flow of air up over the peak, takes this lenticular shape. The lens shape is produced, in any case, by symmetrical air flow and selective evaporation.

When high winds are blowing in mountain country, any tall detached peak is likely to fly a 'banner cloud' that streams out, pennant-like, far to leeward. Sometimes the great plume is formed of blowing snow from the peak itself. But if the humidity is just right, a true banner cloud can form on a perfectly bare peak such as Yosemite's Half Dome in summer, caused by the dynamic cooling effect of lowered pressure in the lee of the peak.

RAIN AND SNOW AND ICE

At ordinary temperatures all lower clouds are composed of minute water droplets perhaps 1/2500 of an inch across. Left to themselves in still air, these cloud droplets would sink slowly, at a speed less than one mile per hour. Yet as the droplets jostled together, somewhat larger droplets might be formed here and there. From a cloud mass (low stratus or high fog) near the ground in very moist air, these larger droplets, perhaps ten times as large as the original cloud droplets, might sprinkle downward in the form of a gentle drizzle.

But such a drizzle does not even approach the wetting power of real rain. Ordinary raindrops are ten times as big again as the drizzle droplets — say 1/25 of an inch across. And rain clouds, considerably less common than ordinary clouds, are distinguished from them by two characteristics: fairly strong rising currents, and considerable depth of cloud. Many rain clouds indeed probably extend, summer or winter, well above the freezing level, and ice particles probably play an important part in their mechanism. In any case the cloud droplets of any active rain cloud are being borne steadily upward, though not quite as fast as the air surrounding them. In the upper parts of the cloud the droplets are probably farther apart, and somewhat larger by reason of longer and freer condensation.

Here and there they are combining into still larger drizzle droplets. Well up in the clouds the rising currents, not very uniform anyway, must somewhere begin to diminish. As a result the larger droplets begin to fall downward, or at least rise slower than most of the cloud droplets, and in either case begin to sweep through the smaller droplets and the moist air surrounding them, gathering unto themselves more and more moisture as they fall. Finally these growing drops perhaps become large enough to continue earthward, even through moderate rising currents, in the form of rain. Much tropical rain and some temperate-zone rain probably originates in this simple way.

But in temperate and arctic latitudes probably most rain, and perhaps all heavy rain and certainly all snow, evolve originally out of ice particles. These ice particles can sublime directly out of water vapour onto sublimation nuclei (dust particles and the like, always present in the atmosphere) very easily at low temperatures around 5° F. (-15° C.) or colder; or, perhaps less frequently, they can be frozen out of ascending water (cloud) droplets that are likewise cooled well below freezing and jostled sufficiently by turbulence. The level at which the ice particles are first formed varies with latitude and the season and the general weather situation — from as low as a mile or less in winter to as high as four miles or more in summer. The upper parts of a typical heavy rainstorm, therefore, are mostly iceparticle clouds — cirro-stratus, alto-stratus, and the like -rather thin-textured clouds with indistinct edges. The lower parts of the same typical rainstorm, at least during its earlier phases, are mostly water-droplet clouds — cumulus, stratus, and the like — opaque-looking and well defined.

As the ice particles from the upper clouds fall slowly (at perhaps two or three miles an hour) through the lower (waterdroplet) clouds, a curious transfer of moisture occurs. By reason of physical laws that need not be elaborated here, at certain temperatures and humidities water vapour is evaporated from liquid droplets, yet sublimed on solid particles. Thus the falling ice particles constantly gain in bulk at the expense of the water droplets around them, and with increasing size fall gradually faster. And in addition to their sublimation-growth, they freeze to themselves any subcooled water droplets that they happen to hit. The ultimate form of the ice crystals depends on the humidity, and varies from simple ice needles to skeleton stars (snow) to snow balls (graupel). As these ice crystals fall, the ice-particle cloud itself thickens and lowers, perhaps encroaching on the water-droplet clouds below — or possibly engulfing them entirely, as in cold-weather nimbo-stratus. But after falling below the freezing level (provided it is well up in the air), the ice crystals soon melt into raindrops.

This is the idealized picture. But in a violent rainstorm -and most particularly, in a turbulent thunderstorm — cloud droplets, ice particles, raindrops, and snowflakes may all be milling around in some parts of the cloud.

Rain

In 'light' rain the drops are very small, averaging perhaps 1/50 of an inch in diameter. Ordinary or 'moderate' raindrops are about twice as big (1/25 inch). Drops of 'heavy' rain are larger still, perhaps 1/16 inch wide. The very largest raindrops in a cloudburst may 1/8 inch or more across. Raindrops splashing on the ground of course look much bigger than the effective spherical sizes just given, and even while falling through the air they are somewhat flattened out.

The largest possible raindrop size is about\! inch, which falls through still air at about twenty miles per hour; any larger drop, falling faster, would be broken into smaller drops by the wind of its fall. Though all rain results from rising moist air, the gentleness and uniformity or violence and variability of the formative updrafts go far to determine what form the precipitation will take. Rain out of stable (stratus) cloud layers, gently rising, say, along a warm front, is light, steady, and continuous. But rain out of unstable cumulus along a cold front (where updrafts are patchy but powerful) is likely to be heavy, variable, and showery.

In the heaviest cloudbursts rain falls at the rate of three or four inches of solid water per hour. (This excessive rainfall rate cannot continue for long, but even an hour of it may cause floods.) A fall of nine or ten inches in a few hours, by no means rare, brings creeks and rivers raging out of their banks and inundates the countryside. The heaviest rainfall so far ever recorded was at Baguio, Philippine Islands, where forty-six inches — nearly four feet of solid water on a level surface — fell in twenty-four hours! Rain is most beautiful when it falls through clear lower air in the near-level rays of a low sun, glistening and sparkling in all directions, and perhaps painting the spectral arch of the rainbow. The primary bow, and perhaps a secondary are, are always centered about the observer's shadow. Hence they can never form more than half a complete circle for anyone on level ground or at sea.

The cause of the rainbow is selective reflection and refraction of the various colours or wavelengths composing white light (at certain critical angles that vary slightly with the wavelength) by the small crystal-clear spheres of the falling raindrops. The primary bow, ranging from a sharp outer edge of red to a diffused band of inner blue or violet, has a radius of about forty degrees from the observer's shadow. The rarer secondary bow, if visible at all, begins rather sharply with red about nine degrees outside the primary and shades away outward into a nebulous greenish or bluish glow. The larger the raindrops and the clearer the lower air, the more striking and beautiful are the rainbow colours. Ground observers, of course, see the largest bows when the sun is low. But from a high mountain-peak, or better still from an airplane flying between or below showery clouds, it is sometimes possible under a high sun to see a strangely unfamiliar but doubly miraculous and brilliant rainbow rounding into more than half a circle and seemingly laid on its side.

Less spectacular and less rare are the faintly colored aureoles, called 'glories' or 'mountain-bows,' that anyone on a sharp peak or in an airplane can see around his shadow on the white surface of a sunlit cloudbank.

If the cloud droplets are near at hand, these glories may resemble faint but orthodox rainbows. Smaller aureoles that appear on more distant cloud-floors — faint colour-rings only two degrees or so in diameter — are probably due to selective diffraction of light rays around the sharp opaque edge of the peak or airplane.

Snow

Few places on the earth are not visited by rain, occasional or frequent, light or heavy, steady or showery. Over the whole earth, indeed, it has been estimated that rain is always falling at the stupendous rate of sixteen million tons per second! But in northern latitudes, through a large part of the year, this never-ending deluge takes the form of feathery-white showers of snow — snow that, unlike rain, accumulates everywhere, in the valleys and on the hills, to store moisture against a future melting. This snow blanket brings to sight general whiteness, and to sound general silence.

Snow only four inches deep absorbs something like nine tenths of ordinary sound energy, being more efficient as a noise muffler than heavy draperies of velvet. A feather-like snowflake floating slowly downward seems to have little in common with a piece of ice. Nevertheless, that is what the snowflake really is, a piece of ice usually crystallized into a very intricate and beautiful pattern. Snowflake patterns are so elaborate, consisting mainly of outgrowths at sixty-degree angles, that each flake presents a large air-catching surface compared with its weight, and falls slowly like a parachute. Snowflakes can be caught on a dark cloth and examined under a magnifying glass, and they have often been photographed with microscopic cameras.

Snow, as already indicated, is probably formed by the same general causative factors as rain, operating however at temperatures entirely below freezing. As the nuclear ice crystals fall, they freeze to themselves as crystalline extensions any subcooled minute water droplets in their path, and grow also by direct sublimation out of invisible water vapour.

If snowflakes fall into temperatures above freezing on their downward path, however, each elaborate crystal melts into a drop of rain. In a climbing airplane it is not uncommon to sound through a region of rain up into a zone of snow (though the exploration is not particularly to be recommended, owing to icing hazards), and much winter rain undoubtedly begins as snow in the higher levels.

Although snow can melt into rain, rain cannot freeze back into snow. If rain from an upper, warmer air layer passes through a colder-than-freezing layer close to the ground, the result is frozen raindrops, or round pellets of clear ice, properly called *sleet*. Sometimes, however, snow falls in the form of round, compacted pellets which resemble sleet, but are more correctly called *graupel*; and occasionally these pellets come through to the ground. The remaining form of aerial ice is *hail* -pellets formed by repeated up-and-down rides in the towering thunderstorm clouds.

If rain falls in liquid drops on ground, houses, trees, or wires that have been previously cooled below freezing, it forms a layer of clear ice called *glaze*. 'Icing' of a somewhat similar sort can occur (to the embarrassment of pilots) on the wings or other parts of an airplane in flight through cold clouds or rain. Another type of ground icing (also noticed on airplanes) is *rime*, where subcooled fog or cloud droplets freeze on exposed projections, gradually building out into the wind. Widespread and continued glaze, which may bring down trees limbs, telephone wires, and power lines by the hundreds, is called an 'ice storm.'

The formation of clouds out of thin air, the birth of precipitation out of clouds — both these processes are physical experiments conducted by Nature in her open-air laboratory on a gigantic scale. Even a local summer shower involves millions of tons of air, hundreds of thousands of tons of water, and energy equivalent to the burning of thousands of tons of coal. A widespread storm, such as occurs along the path of a cyclone, may involve materials and forces many times greater.

VISIBLE AND INVISIBLE MOISTURE

Invisible water vapour is always present in the air, varying in amount from a mere trace up to about four per cent by volume or two and one half per cent by weight (the water vapour being somewhat lighter than dry air). 'Specific humidity' — the actual mass of water vapour present in a given mass of air, measured usually in grams of water vapour per kilogram of air, and varying from less than 1 g/kg in very dry air up to perhaps 20 g/kg in very moist air — is the most conservative and generally useful measure of atmospheric moistures. For some purposes, though, dryness or moistness of air is more conveniently expressed as 'relative humidity' — the ratio of the actual amount of water vapour present (the existing specific humidity) to the greatest possible amount of water vapour that could exist at the existing temperature (the saturation specific humidity).

And for many practical weather purposes, the best measure of atmospheric moisture is the 'dewpoint'-that temperature which has a saturation specific humidity equal to the present, existing specific humidity.

Finally, we saw how any one of three conditions indicates that condensation of water vapour into liquid form is imminent:

- The nearness of the existing specific humidity to the saturation specific humidity;
- The nearness of the dewpoint to the temperature; and
- The nearness of the relative humidity to one hundred per cent.

When water vapour in the air comes out of its invisible form into the visible form of liquid water, however finely divided, it must have some solid nucleus on which to form. Now, on a summer night, say, if the lower air is quite still as the dark earth radiates away its heat, the cooling ground cools only that layer of air immediately next to it. When the temperature of this still, surface air reaches its dewpoint, invisible moisture condenses into liquid dew on sticks and stones, blades of grass, or any other cool and convenient objects — nuclei of condensation, we might call them, albeit large ones. Thus the fairly common idea that 'dew falls' is wrong. If the temperature happens to be below freezing (less than +32° F. or 0° C.), the invisible water vapour sublimes directly into feathery ice crystals, forming hoar-frost.

Invisible water vapour that condenses into minute droplets in the free air, forming fog or clouds, also needs solid nuclei on which to form. But whereas the nuclei for dew and frost are large and evident, nuclei for fog and cloud droplets are of infinitesimal size. Nevertheless these nuclei — minute particles of sea salt, smoke, hygroscopic dust, and the like — are always present in the troposphere.

Given these nuclei, rapid condensation of invisible moisture into minute droplets of liquid water (visible in the aggregate as fog or cloud) begins when, by reason of evaporation, dewpoint increases to existing temperature; or when (as is far more usual), by reason of atmospheric cooling, temperature reaches existing dewpoint. In any case, condensation begins when and where dewpoint and temperature come together, for any reason, at any level in the air ocean.

FOG

Fog can be at best a creation of elfin beauty, is on the average an inconvenience scarcely ever welcomed by honest people, and becomes at worst, in transportation, a great hazard. It may stall land vehicles for hours, force ships along the coast to anchor for days, and even more sorely bedevil airplanes that must keep moving at two or three miles per minute if they are to keep the air at all. In temperate latitudes the gray, creeping fog-mass is usually formed entirely of myriads of minute water droplets, each something like one thousandth of an inch in diameter. Damp haze, which in its denser forms may approach the murkiness of light fog, can (unlike fog) begin forming when temperature and dewpoint are far apart. This denser form of haze sometimes extends for several thousand feet above the earth's surface, and the well-defined top of the haze layer usually marks the base of a temperature inversion. To an aviator above the haze it appears almost as a level, pearly-white sea brilliantly reflecting the sunlight; only as he descends into the haze do ground patterns begin to loom murkily into view. But at night, lights shine upward clearly through this haze, whereas true fog of any considerable thickness blots them entirely from view.

The reason for this is that fog droplets are much larger than damp-haze droplets (though still incredibly small) — large enough, in relation to the wavelength of visible light, to block off almost completely any direct rays. Yet the transition from dense haze to light fog may be gradual, with gradually increasing relative humidity.

As dense fog requires both plenty of vaporous moisture to start with, and then considerable cooling, for its formation, it is found mostly in temperate latitudes. In arctic regions, or at high altitudes, fog (or cloud) may be composed of myriads of minute ice-needles — 'frost smoke' that glitters and sparkles in the sun. In the tropics fog occurs but rarely, because the temperature contrasts that favour it are usually lacking.

Fog is classified, according to the mode of its formation, into three main types:

1. Radiation fog, also called 'land fog' or 'summer fog,' in which air already stagnant on the scene cools until its dewpoint is reached;
2. Advection fog or transportation fog, also called 'winter fog' or 'sea fog,' in which air is brought in bodily from outside regions to the scene of its fogginess; and
3. Frontal fogs, usually temporary and transient, that occur with the passage of fronts between unlike air masses.

Radiation Fog

Radiation fog is most frequent during the long, clear nights of autumn, though it may occur at any season, particularly in rather stagnant air conditions near the centre of a high-pressure area. The fog sometimes appears to rise from the ground; but actually, the only thing that rises is a zone of temperatures lower than the dewpoint of the air in question. As the ground and the moist air immediately above the ground slowly radiate their warmth away into space, the temperature of this lower air falls, approaching its dewpoint. When the two coincide, fog appears in visible form. As the top of the fog layer itself radiates away heat much faster than dry air, the fog usually continues to deepen until sunrise.

On a flat plain, if the wind is very light (say only one or two miles per hour), the fog layer may be very shallow. This 'ground fog,' perhaps only a few feet thick, results from very gentle mixing in the air immediately above the ground; there is not enough eddying and turbulence to raise the surface coldness higher.

But in rolling or mountainous country, even on a calm night, cooled surface air drains into the hollows (as anyone can notice by driving over darkened hill and dale in an automobile, or even walking across broken terrain, with a thermometer). As a result, large pools of valley air may be cooled sufficiently to produce fog 'lakes' many fathoms deep.

If the breeze is somewhat stronger — say two to six miles per hour — the surface coldness is mixed by eddying and turbulence through a considerable layer of the lower air, causing true fog that blots out hillocks and hollows alike. This sort of fog usually begins at or near the surface, working gradually upward, though to an observer on the ground it seems to materialize rather suddenly out of nothing more tangible than a few stray wisps of vapour.

The maximum height of the fog depends on many aerological factors, but it is seldom more than a few hundred feet. Temperature in the fog layer usually decreases with altitude at something like the moist adiabatic rate (-2½°/1000 ft); and just above the top of the fog there is usually, if not always, a temperature inversion.

If the breeze is much stronger than six miles an hour or so, eddying and turbulence diffuse the surface coldness through such an extensive layer of overlying air, and bring down so much potential warmth from aloft, that the air temperature never reaches its dewpoint and fog is entirely prevented.

The conditions favorable to deep and bothersome radiation fog are, therefore: to begin with a fairly high dewpoint for the season (plenty of actual water present in vaporous form, that is), and not much difference (say less than 15° F. at sunset) between temperature and dewpoint (or a corresponding small difference, say less than 7° F., between wet- and dry-bulb temperatures); second, a clear or nearly clear sky, for maximum radiational cooling; and third, through at least part of the night, light breezes registering two to six miles per hour at the surface. Once begun, radiation fog continues to deepen through the hours of darkness and, curiously enough, may increase for a brief time after sunrise. This last effect may be due to either of two factors: the power of sunlight to make floating particles attract water; or the increase of lower-level turbulence and mixing caused by solar heating at the surface. It is a rather frightening experience to take off from a clear airport at dawn, and see it fog in under you a few minutes later.

Within a few hours after sunrise, however, an ordinary radiation fog begins to 'burn off.' This burning off always proceeds from the ground upward as the solar radiation, penetrating through the fog layer, warms first the ground itself and then, progressively, the air above it.

The lower fog does not actually 'lift' in this case any more than it 'rises' out of the ground at night; what actually lifts at first under the morning sun is a zone of temperatures higher than the dewpoint. But before long rising currents from the warmed ground break through the fog layer here and there as the solar radiation continues. And the sun, shining through these breaks with redoubled energy, soon produces enough warm updrafts to raise the whole fog layer slowly upward and sweep it into scattered drifts of cloud.

The worst radiation fogs are found around great cities, where innumerable smoke, oil, and soot particles impair visibility on their own account, serve as ready nuclei for condensation, and hinder the evaporation of fog once formed. Thus the infamous London fogs, which greatly increased during the late nineties with the growth of the great city, may hang on in yellowish 'pea-soup' form, perhaps for days, until the advent of a new air mass and freshening winds.

Though less tenacious, radiation fogs are nearly as common in many regions of the open country. They can also occur over limited areas of calm ocean, but the usual sea fog arises out of different causes. An extreme example of radiation fog occurs in the Great Valley of California, where murky-gray obscurity may hang on for days beneath the 'lid' of a temperature inversion.

A particular type of radiation fog, called 'high fog' or 'stratus fog,' starts in mid-air and builds downward. This high fog, also, is particularly prevalent in the walled-in valleys of California, but I have noticed it occasionally over the eastern seaboard, and it probably occurs widely, though rather rarely, in other regions.

Once we had a formation of army airplanes operating all night from an air base in the District of Columbia. This high fog formed suddenly, around midnight, at a height of about fifteen hundred feet, trapping one of the planes above it. The predisposing aerological condition here is a zone of decreasing temperature and high humidity from the ground upward for the first thousand feet or so, with a pronounced and lid-like temperature inversion above. The air below the inversion is thoroughly mixed, equally moist, and coldest at the base of the inversion in mid-air.

In other words, specific humidity is fairly constant (dewpoint decreasing about 1° F/1000 ft) throughout the moist lower layer, but temperature is lowest at the top. As the whole moist layer cools by radiation at night, the top temperature reaches its dewpoint first, and the fog builds downward from the base of the inversion. Sometimes you can see it beginning in wisps along the side of a steep mountain.

Its downward advance is usually rather sudden at first — perhaps five hundred feet or so in an hour; then it lowers more slowly. But within a few hours, it may reach down to the ground. Some typical temperatures, in a California fog of this sort, are as follows: 600 feet (surface), 57° F.; 900 feet (top of fog), 54°; 1200 feet, 64°; 1500 feet, 66°.

Advection Fog

Whereas radiation fog is due to simple cooling of the air in place, advection fog may be due to either of two causes. One is the horizontal advance of warm, moist air over a cold land or water surface. The other is the advance of cold air over a moist, warm surface. Of these two, the first cause is by long odds the most common.

On land, a warm-air advection fog often results when gentle, moist southerly winds blow in from the ocean, particularly if the land surface is covered with snow. The warm, moist air is chilled by the cold ground below it until its falling tempera- ture, uniformly distributed through several hundred feet of depth by gentle turbulence, reaches its dewpoint. In the United States any large northward outbreak of warm, moist (tropical maritime) air in wintertime may produce widespread advection fog.

The warm sector of a marked temperate cyclone may be completely blanketed by such a fog, and if the pressure gradients are steep, this may be one of the rather exceptional cases where fog accompanies a high wind, usually from some southerly quarter. These warm-air advection fogs are far less common on land than radiation fogs; but the advection fogs are likely to be more widespread and prolonged, forming or advancing at any hour of the day or night.

On the ocean, warm-air advection fogs are by far the most common type, and more frequent in summer than in winter. They are likely to form wherever a cold ocean current or area is surrounded by considerably warmer water — such as off Newfoundland, where the Labrador current projects southward; off Nantucket, where the colder water from the northern New England coast meets the warmer waters from south of Cape Cod; or along the whole line of the Pacific coast, where water upwelling inshore from chill depths below is much colder than the Pacific waters offshore.

The cold-water and fog area may be small and distinct — such as the five-mile-wide cold current in the middle of Santa Barbara Channel. The whole fog area moves with the prevailing winds, which are usually light -say fifteen miles per hour or less on the open water — though in a great temperate cyclone they may be much higher. I remember one case of moderately thick fog on a steamship off the Atlantic coast, with the wind blowing a whole gale out of the southeast and a mountainous sea running. If the wind is much above twenty miles per hour, however, the minute fog droplets tend to coalesce into the larger droplets of mist or drizzle, with somewhat improved visibility.

These sea fogs vary widely in height — sometimes they are so shallow that the mastheads of ships project into clear air above. At other times they are more than a thousand feet deep. Their average depth is perhaps around five hundred feet. In any actual case, of course, this height can be determined by a kite or captive-balloon sounding from a ship in the fog — or from shore elevations along a mountainous coast like California. Sometimes the sea murk may take the form of 'high fog' or low stratus, with good visibility along the water surface.

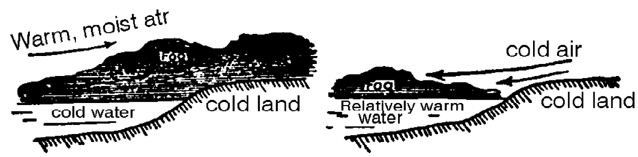

Fig. ***Advection Fogs***

The second type of advection fog occurs when cold air flows over a moist, relatively warm surface such as ocean or lake water, or possibly land wet by recent rain or melting snow. Here the invisible water vapour, evaporating from the water surface, is chilled below its dewpoint by the cold air above so that it visibly 'steams.' On land, the steaming of lakes on autumn mornings, as radiation-cooled land air sweeps across water areas less prodigal with their summertime heat, shows the mechanism of cold-air advection fogs in miniature form. Similar steaming fogs occur in arctic regions when cold, below zero polar air sweeps across leads of open water in the pack ice; though here, of course, the fog is formed of ice crystals. Cold-air advection fog on the ocean is quite rare, but notably hazardous when it does occur. Near the United States it is confined mostly to the cast coast, and arises from the rapid southward and eastward outbreak of a large, cold (polarcontinental) air mass.

When the fresh polar air, at near-zero temperature, rushes southward or southeastward behind a marked cold front over water at temperatures around 50° F. (or even 70° in the Gulf Stream), steaming fog arises on a widespread scale and perhaps masthead high. What with the viewless murk, the piercing cold, and a howling gale out of the north or northwest, even a steel steamship may do well to anchor inshore, despite every modern aid to navigation.

Frontal Fogs

Somewhat different in mechanism from radiation fogs and advection fogs are the fogs that occur near the marked 'fronts' between unlike air masses. These frontal and pre-frontal fogs, however, are neither very common nor very persistent, and only two varieties of them are worth mentioning here.

First there is a type of pre-frontal fog that occurs in advance of a warm front, in a zone perhaps fifty or a hundred miles wide where heavy rain from the warm-front sloping cloud-surface above has saturated the cooler air of the underlying cold wedge, and falling pressure ahead of the front causes further expansional cooling of the cold, moist air. (This pre-frontal fog is distinct, of course, from the warm air mass [advection-type] fog sometimes present in the warm sector behind the warm front.)

The other type of pre-frontal fog may occur a few miles ahead of an advancing cold front, where a rapid fall in pressure encourages the fog-forming proclivities of the warm air mass.

Any fog, however formed, may drift bodily elsewhere and engulf regions well removed from its origin, provided aerological conditions do not favour its rapid evaporation. The front of the fog, particularly at sea, may take the well-defined form of a steeply sloping 'bank,' with a distinct and clear-cut edge. When a sea fog comes ashore, it remains on the ground if the land is colder than the water. If the land is relatively warmer, however, the lower surface of the fog lifts somewhat above the ground.

FORMATION OF CLOUDS

Fogs and clouds are of the same texture, and their different appearances are mainly a matter of viewpoint. High cloudturrets may gleam as white, as apparently solid, as alabaster in the bright upper sunlight; but explore the insides of those seeming-solid vapour-peaks in an airplane, and they become just as ragged, gray-murky, cold, and damp as any fog. Having seen how cooling of humid air near the ground produces fog, we may well guess that cooling of some sort in the upper air produces cloud. How is this large-scale free-air cooling brought about? Not by conduction (for air is a very poor conductor of heat), not usually by radiation (for dry air neither absorbs nor radiates heat readily), but mainly by the expansion cooling of air forced in some way to rise.

In the preceding we examined the underlying mechanism of this rising-refrigeration process — we followed the salient facts of dry-adiabatic cooling, of moist-adiabatic cooling, of dry-air and moist-air stability and instability. It remains now to see how these various aerological conditions play their part in forming clouds out of thin air.

Forced Cloud Formation

If an entire layer of moist air is driven by prevailing winds up the slope of a mountain range, or drifts up the inclined plane of an underlying wedge of colder air, it cools throughout its whole extent at something like the dry adiabatic rate (about -5° F/1000 ft — or 5.5°, to be exact) so long as moisture remains entirely and invisibly vaporous. But at some moderate altitude, if its humidity is high, the rising air layer reaches its dewpoint.

Then moisture begins to condense on myriads of infinitesimal nuclei that are usually present in air — salt particles, haze particles, and so on — into the form of innumerable small water droplets visible in the aggregate as cloud. The individual droplets forming the cloud are about the same size as fog droplets or somewhat smaller — say 1/2500 of an inch in diameter. After this cloud condensation begins, of course, released latent heat retards the further cooling of the rising air to the moist adiabatic rate (say -3° F/1000 ft at moderate altitudes).

If the moist, rising air layer is ill-disposed towards updrafts and downdrafts within itself — if, in a word, it is stable — it forms merely one great, shapeless horizontal mass of flat, quiescent, layer-like cloud in the same way that a groundcooled air layer forms fog.

A large proportion of all clouds are formed, as simply as fog, within such moist but stable air layers lifted in some way by external causes. If all clouds were so formed, the science of aerology would be much simplified. But more complicated mechanisms of cloud formation produce some of the most splendid clouds and spectacular storms.

Free Cloud Formation

It is worth recalling from the stability in the preceding that, in 'absolutely unstable' air (lapse rate steeper than about -5° F/1000 ft), vertical currents tend to continue and to grow, once they are triggered off by commonplace causes such as surface heating, mountain updrafts, and so on. The same may be true of 'conditionally unstable' air (lapse rate between -5° and -2 1/2° F/ 1000 ft), which behaves like absolutely unstable air once condensation has begun.

By the 'beginning of condensation,' of course, we mean the formation of the minute water droplets that constitute visible cloud. The ultimate result of instability in any air layer, dry or moist, is updrafts (together with compensating downdrafts elsewhere). And the result of updrafts, in any moist air, is bound to be clouds.

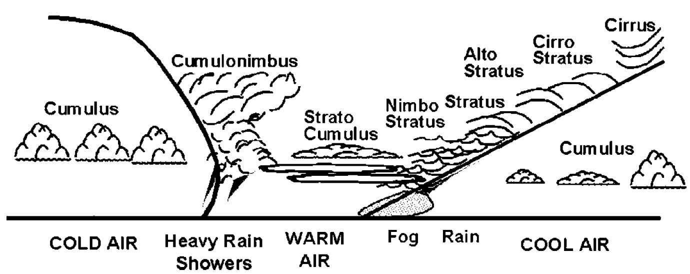

Fig. ***Cloud Formation***

Unlike the flat, quiescent layer-like clouds formed by general air lifting under external forces, these cloud-children of instability are usually lumpy, heaped-up, puffy, and cauliflower-like in appearance. The instability clouds themselves may be very large, occasionally towering higher than the highest mountains, yet they tend towards localized lumps, turrets, and masses, called 'cumulus,' rather than widespread and shapeless layers, called 'stratus.'

In freely rising air, instability cloud-birth always begins at a certain definite level — the *condensation level* where falling temperature meets existing dewpoint. Under some conditions this condensation level is very definite, and readily apparent as a flat cloud base or 'ceiling.' The typical woolly, cream-puff clouds of fair-weather summer skies, for example, all have flat bases (and not the rounded bases that some artists like to paint into them); and these bases are actually all at the same elevation, though more distant clouds may appear lower to ground observ- ers by reason of perspective.

Each detached, puffy cloud in this idealized summer picture represents, of course, the visible crown of an updraft (invisible below the condensation level) that is sustaining the cloud. In the clear spaces between these detached woolpacks there are usually downdrafts more gentle and more extensive than the updrafts that they balance.

There is a rough but useful relationship between 'ceiling,' or height of cloud base, and surface temperature and dewpoint. At moderate temperatures, each degree of temperature-dewpoint difference means about two hundred and fifty feet of ceiling. Thus if the surface temperature is 70° F. and the surface dewpoint 66° F., the ceiling should be around one thousand feet. Occasionally the mechanism of updraft-builds-cloud is revealed in detached and clearly apparent form by a large fire.

The smoke from the fire rises, of course, along with a great column of heated air. Just as the smoke begins to thin out, perhaps, at some moderate elevation, a new and thicker turruted murkiness appears and builds upward.

This upper murkiness is not smoke, but cloud formed by the updraft, and the elevation where it appears is the condensation level.

If a large and deep air layer is sufficiently unstable and sufficiently moist — conditions that exist in tropical or summertime thunderstorm weather — both the updraft and the consequent instability cloud may grow out of all proportion until the final result is vertical currents ballooning upward at possibly one or two hundred miles an hour, and a towering cloud mountain as high and wide as Everest.

For the wholesale condensation of water vapour into cloud droplets releases wholesale latent heat, which intensifies the updraft, which condenses still more moisture and releases still more latent heat, and so on. You can sometimes watch these thunderheads visibly growing against the blue sky — the faster they grow, the stronger the updraft and the more unstable the air to begin with.

The upward momentum of rising air also helps in this cloud-building, which is ultimately checked, in extreme cases, by the great upper temperature inversion that is the stratosphere. Instability clouds are not limited to hot weather. It is all a matter of relative warmth and coldness alow and aloft — of lapsed rate, provided enough moisture for cloud-building is available.

A strong wintertime outbreak of frigid polar air over a relatively warm water surface (say along the Atlantic coast of the United States, or across the Great Lakes) can produce as puffy and squally clouds as anyone could wish. And in winter, with the sun low in the sky, these instability clouds and snow squalls look blacker and more forbidding than most summer storms.

Combined Cloud Formation

Not all clouds are distinctly of layer-like, quiescent stratus type or indisputably of puffy, instability cumulus type. It is possible to have both types present in the same section of sky, perhaps at different levels. Hybrid or combination types, such as strato-cumulus, are also possible.

The kind and prodigality of cloud-building depends on the relative strength and interplay of the three great cloud-building forces — moisture, instability, and external lifting. Any moist air layer bordering on instability tends to become more unstable, the more it is lifted by external forces. Such a layer, externally lifted, may at first form quiescent stratus clouds which, as the lifting goes on, may bulge upward here and there into growing instability cumulus turrets. Nevertheless, every cloud that we see in the sky must result from some chance combination of the three cloud-building forces.

The more unstable the air is, the more moist it is, and the more violently it is lifted by external forces, the greater will be the bad-weather effects in clouds, rain, snow, hail, and the like. We have said earlier that radiation is usually an unimportant factor in cloud formation.

But moist air absorbs and radiates heat better than dry air, and it is possible for a high, moist layer (having a relative humidity close to one hundred per cent) to radiate away enough heat at night so that it forms a layer of high stratus cloud. Possible — but not common. This ever-changing interplay of cloud-building forces in Nature's air ocean produces what, at first sight, seems to be an endless and bewildering variety of cloud forms. But these cloud forms (or the more common ones, at any rate) can be grouped into a few types and classes that are quite easy to remember and identify. The most basic cloud distinction of all, between layer-like stratus and heaped-up cumulus, has indeed already been made.

Parade of the Clouds

The first attempt at naming clouds systematically and completely was made by Luke Howard, an Englishman, in 1803. He classified all clouds by shape or appearance into three general types — cirrus or 'curl' clouds, cumulus or 'heap' clouds, and stratus or 'layer' clouds. Other types of clouds, such as nimbus or rain clouds, Howard associated in some way with his three basic types.

During the nineteenth century other cloud classifications were proposed, but Howard's scheme, from time to time elaborated and improved, gradually grew into a standard system increasingly accepted throughout the world. We are concerned only with the commoner clouds, such as any amateur can learn to name and recognize. And most amateur difficulties in the past have come, I believe, from too-exhaustive or illogical systems of classification. So we shall keep our cloud charts very simple, and as logical as possible. The most fundamental cloud differentiation, already hinted at in the preceding section, is based, like Howard's, on shape or form:

CLOUD CLASSIFICATION BY FORM OR SHAPES

Two Fundamental Shop, Classes

Layer class (from stable air)	Stratus (= 'a spreading out')	A quiescent cloud layer, rather flat and formless, extending horizontally for the most part.
Puffy class (from unstable air)	Cumulus (='a heap')	A piled-up heap of cloud, usually having definite three-dimensional shape, often growing vertically.

Two Special Shape Types

Wispy type	Cirrus (= 'curl')	White, fibrous mares' tails or feathery curls at very high altitude.
Broken type	Such as Fracto-stratus * Fracto-cumulus	Heat-evaporated or wind-torn into separate smaller cloud masses with ragged, wispy edges.

Mixed and Storm Typer

Combination type	Strato-cumulus	Combining some charicteris-tics of two different classes.
Precipitation types	Nimbus (= 'rainstorm') Nimbo-stratus Cumulo-nimbus	Clouds of stratiform or cumu-loform shape, respectively, from which rain, snow, or ice is actually failing.

Hyphens in cloud names have been dropped by professional serologists, but they are retained here for the sake of simplicity and clearness. Another way of classifying clouds, and perhaps a better one for understanding how they fit into the pageant of tropospheric weather, depends on their height above the earth's surface -or, in a more general sense, their altitude above sea level. Beginning with the highest and thinnest clouds, and ending with the lowest and thickest, the more common forms are:

Table. Cloud Classification By Altitude

Upper or citrus-like types	Cirrus Cirro-stratus Cirro-cumulus	Usually occur in the upper troposphere, often not far below the stratosphere, at heights of about 4 to 7 miles (6-10 km.).
Middle or high ('alto') types	Alto-stratus Alto-cumulus	Usually occur in the middle troposphere at heights of 2 to 4 miles (3-6 km.).
Lower types	Stratus Cumulus Strato-cumulus Nimbo-stratus	Usually occur in the lower troposphere, from the surface up to a height of nearly 2 miles (3 km.).
Vertical growth type	Cumulo-nimbus (thunderstorm)	Single cloud mass may extend from near the surface up to heights of 4 to 7 miles.

This table shows at a glance how various types of clouds fit into the general picture of the troposphere. From it one can appreciate why towering cumulo-nimbus clouds can be seen hundreds of miles away in clear air; one can sense why a spreading high deck of cirro-cumulus stands out in the red light of a clear sunset for perhaps three hundred miles to westward, where the sun-rays, receding at ten miles per minute, still light the cloud deck's edges half an hour after they have left the ceiling directly overhead.

Cirrus

True cirrus clouds are fibrous and delicate in appearance, showing as white wisps against the upper blue. Sometimes they arc combed out into mares' tails; sometimes they are formed more thickly into curls on feathery plumes or parallel 'plowed-field' bands.

But always they are the most tenuous clouds of the sky, scarcely dimming the sun's disk, and the highest, floating at five to seven miles above the earth (8 to 11 kilometers). As the temperatures at these levels are always far below zero, cirrus clouds are formed from minute, free-floating crystals of ice (simple hexagonal prisms). In fair weather they may result from general lifting of the upper troposphere above the active, 'cold-dome' portion of a high-pressure area.

As foul-weather harbingers they are 'locks of the approaching storm,' blown out by substratospheric gales far in advance of the general cloud area of a cyclonic warm front.

Sometimes broken or wispy clouds fracto-stratus or (fractocumulus) in the lower levels have a superficial resemblance to cirrus clouds, and are mistaken for them by laymen. But the true cirrus clouds, carefully watched, are very evidently much higher than these fracto-forms.

Cirrus clouds are always interesting to watch because they show the motion of air near the top of the troposphere. Sometimes the long, fibrous tails seem to radiate from a cyclone centre below the horizon, and actually are combed out along the direction of air motion by the upper winds.

At other times cirrus plumes or rays can be seen to move, not along their length but across it, like successive ranks of marching soldiers. Here the cirrus plumes make visible a colossal wave motion in the substratosphere — white crests of upsurging condensation and blue troughs of downdropping evaporation — great, silent aerial surges that may measure ten or twenty miles from crest to crest. Some cirrus clouds, again, are probably nothing more than evaporating wisps of what was originally cirro-stratus -a sort of 'fracto-cirro-stratus,' one might say. After sunset or before sunrise, cirrus clouds often glow with beautiful tints of flame-yellow and rose-pink.

Cirro-stratus

Cirrus-like cloud smoothed into one continuous, thin layer, resembling a whitish veil, is called ciffo-stratus. It may be smooth and milky, or again uneven and somewhat fibrous, forming out of thickening cirrus. Cirro-stratus never casts shadows, nor does it greatly dim either the sun or the moon, though in thickening and lowering form (approaching alto-stratus) it begins to blur their disks somewhat.

Cirro-stratus is almost as high as cirrus (3½ to 7 miles, or 5½ to 11 kilometers), and is likewise formed of minute hexagonal ice crystals. In all citrus-like cloud-forms, but most noticeably in cirro-stratus, these ice crystals may refract and reflect light in such a way as to form halos or luminous rings around the sun or the moon.

The most common halo is a single ring of about twenty-two degrees radius, colored faintly if at all; but one or even two additional outer rings may be seen occasionally, together with rarer 'mock suns,' 'sun pillars,' tangent arcs, and the like. ('Coronas,' or rare, colored sun-andmoon rings that may superficially resemble halos, are usually caused by water-droplet clouds at much lower altitudes.) Cirro-stratus cloud is most frequently formed in the upper and outer warm-front portions of a cyclonic cloud area, thickening to alto-stratus towards the storm centre, thinning out to true cirrus wisps in advance. Hence while quiescent detached cirrus is a frequent fair-weather cloud, cirrus thickening into a cirro-stratus veil is a more definite portent of possible cyclonic approach. The portent is still more definite if cirro-stratus thickens to alto-stratus — but by that time, the rain is not far behind.

Cirro-cumulus

Cirro-cumulus clouds are small, fleecy, white flakes or very small cloud balls — a 'mackerel sky' — perhaps merging into cirrus or cirro-stratus, and composed of supercooled droplets or ice crystals floating at great heights (3½ to 5½ miles or 5½ to 9 kilometers). True cirro-cumulus clouds are not very common, and are easily confused with thinning forms of alto-cumulus — which latter are always well below the true citrus-like cloud levels. The distinguishing feature is, however, that in cirro-cumulus the individual flakes usually appear to be smaller by reason of their great height, and they are never thick enough to cast shadows or show dark undersides. Cirro-cumulus clouds may sometimes be formed by small up-and-down currents along a wide but shallow unstable layer in the upper air. But more often they are arranged, like some types of cirrus, in parallel rows or lines, showing wave-motion between overrunning and underrunning air currents somewhat similar to ocean waves along the sea surface where water meets differently moving air. All these high, cirriform clouds can best be seen, incidentally, through dark (preferably purplish) glasses; and they can best be photographed, like most other clouds, through red filters.

Alto-stratus

This cloud usually takes the form of a thick grayish or bluish veil, always somewhat fibrous in texture. It is thicker and less translucent than cirro-stratus, which often merges gradually into it with the approach of a cyclonic storm. Through typical alto-stratus the sun and the moon gleam only weakly and vaguely, as through ground glass; they have a watery and diffused look, or may be almost wholly obscured, usually casting no shadows. Unlike cirriform clouds, altostratus is composed mostly of minute water droplets plus ice crystals scattered here and there.

The usual height of alto-stratus is 2 to 3½ miles (3 to 6 kilometers). It occurs most frequently along the slightly sloping surface of a warm front, where the overrunning warmer, moister air, ascending the inclined plane of the cold-air wedge, has lost by condensation a good deal of its original warmth and moisture. This frontal alto-stratus often marks a zone where light rain or snow is just beginning to fall from the high clouds, though this outermost precipitation (except cold-weather snow) may evaporate again before reaching the ground. Aside from the typical warm-front cloud-sheet, local alto-stratus of a more level and quiescent type can occasionally result from the radiational cooling of a high and very humid air layer.

Alto-cumulus

Alto-cumulus is quite similar in general appearance to cirro-cumulus, but the individual flakes or cloud-balls of alto-cumulus are always larger and grayer. Usually they cast shadows and show dark undersides. The lower the alto-cumulus is, the larger are its opaque lumps. Certain forms of altocumulus ('castellatus') display little cloud-turrets growing upward from the main moisture-supplying layer.

In common with alto-stratus and most of the lower clouds, alto-cumulus is formed of minute water droplets, which may, however, be sub-cooled well below freezing in the liquid state.

The usual height of alto-cumulus is 2 to 3 miles (3 to 5 kilometers). Like cirro-cumulus, alto-cumulus is often arranged in parallel rows along a wide and level deck, resulting from wave motion between differently moving layers of air. Alto-cumulus usually results from upper-air wave motion of some sort, or instability; it may grow into alto-stratus or nimbo-stratus.

Stratus

Any flat, shapeless, more or less uniform layer of cloud in the lower levels, usually dull gray in colour when viewed from below and whitish when viewed from above, is called stratus. Stratus resting on the ground is simply fog, formed and dissipated in the same way as fog; and high fog is simply low stratus. More usually, however, stratus occurs at heights of 1500 to 3000 feet (500 to 1000 meters). Low stratus is often a local cloud, gradually breaking away into shreds and smaller masses called fracto-stratus or 'scud.' Stratus can produce a fine drizzle, but not appreciable rain or snow.

11

Wind, Climate and Weather Measurements

WIND AND CLIMATE BELTS

This great, rotating ball that is Earth harbors within its atmosphere a thousand varieties of wind and weather over the diverse oceans and continents of its expansive surface, even at any one instant of time; and all these weather vagaries are constantly changing, quickly or slowly, from day to day, week to week, month to month. Nevertheless there is a grand and logical scheme underlying general world-wide weather—a pattern that unmistakably appears in the average weather, *or climate*, experienced at representative places spaced up and down the meridians of latitude; and in the *prevailing winds* that arise out of these belt-like, world-encircling regions of climatic unlikeness.

This *average* planetary wind-and-climate picture is seldom, if ever, duplicated by actual world-wide weather on any single day—particularly in north-temperate regions, where actual weather variability seems to discredit the existence of any such regular scheme. Yet over the oceans—where the sea surface is relatively flat and relatively frictionless to wandering breezes, where sameness of temperature discourages climatic oddities and local weather freaks, where the climate belts and prevailing winds are most evident, typical and regular—it is possible to glimpse in actual weather some prominent features of the grand planetary pattern. You can best sense certain of its gradual but definite and continuing changes, from latitude to latitude, by cruising north or south in a seagoing ship. Here, for example, is the actual record of a March cruise from Panama to Philadelphia in the old but able American freighter *Timber Rush*—a slow two-thousandmile cruise from torrid to temperate latitudes at the comfortable ocean-going dog-trot of ten knots:

March 11th: latitude 10°N, off Panama coast, temperature 79°F, broken clouds, wind ENE moderate (trade wind), caught barracuda on trolling line.

12th: Lat 12°, in Caribbean, temp 79°, scattered to broken clouds, wind NE X N fresh to strong (norther), sea rough, water temperature 78°.

13th: Lat 14½°, in Caribbean, temp 77°, scattered clouds, wind ENE fresh (intensified trade), water very rich blue.

14th: Lat 18°, just southwest of Haiti, temp 78°, scattered, clouds to clear, wind E'ly light (in lee of Hispaniola).

15th: Lat 22°, near Bahamas and Tropic of Cancer, temp 77°, scattered clouds, wind E'ly to variable light (horse latitudes) scattered clouds to clear.

16th: Lat 26½°, in Atlantic, temp 73°, showers and thundershowers along a cold front, wind SW'ly moderate shifting to NW fresh with frontal passage.

17th: Lat 30½°, in Atlantic, temp 67°, scattered to broken clouds, wind NW moderate (prevailing westerlies), water dark steely blue, water temperature 74°.

18th: Lat 35°, off Cape Hatteras in Gulf Stream, temp 60°, broken clouds, wind N'ly fresh and increasing, water temperature 72°, passed school of whales.

19th: Lat 38°, entering Delaware Bay, temp 35°, clear, calm to S'ly light, water temperature 50° after leaving Gulf Stream, 42° near coast, water dirty green.

Thus we passed, in the last four days of slow steaming at two hundred and forty easy sea miles a day, from a tropical climate to a temperate climate, from smiling summer skies and tufted cloudlets to cold, murky storms and driving showers, from moist, warm, enervating sameness to irritating and unexpected variability.

Centers of Action

The whole idealized wind-belt picture described above is based, it will bear repeating, on *average* conditions. Daily departures from this average, in any one region, may be very wide. It is also based on the conception of a *uniform earth*—either all water, or all uniform land. As a matter of actual fact the southern hemisphere is so nearly all water that this idealized pattern actually holds there much of the time; and the pattern may also hold, though less consistently, well out in the Atlantic and Pacific Oceans of the northern hemisphere.

But in general the northern hemisphere is something like equally divided into great land surfaces and great water surfaces, which vary greatly in the frictional drag they oppose to surface winds.

Also, while oceans tend to remain equable in temperature at all hours and all seasons, continents tend to get too warm in summer and too cold in winter. As cold areas tend to develop high surface pressure, and warm areas low surface pressure, the idealized latitudinal pressure belts of the northern hemisphere are in fact somewhat broken up into continental and oceanic 'highs' and 'lows' which change considerably with the seasons.

These semipermanent, or seasonal, lows and highs are called *centers of action*, because their counter-clockwise or clockwise circulations control in some degree the general movements and conflicts of air masses in their neighborhoods, and influence greatly both the position and movement of various segments of the polar front and the development and motion of the traveling cyclones that are largely responsible for most temperatezone bad weather.The centers of action in and around North America—those that have most bearing on our weather in the United States -are as follows:

Winter	*Summer*
Canadian High	Hawaiian High
Aleutian low	Bermuda-Azores high
Icelandic low	(Rokies low - weak)
	(Aleutian and Icelandic lows - weak)

It will be well to recall these centers of action, and to remember the cardinal fact that (in the northern hemisphere) winds blow clockwise about a high-pressure area and counter-clockwise around a low-pressure area—particularly when we come to consider the traveling 'highs' that are usually associated with vagrant air masses, and the traveling 'lows' or cyclones out of which grow warm and cold fronts.

Table. Typical Climate of The Deep Tropics

Month	*Mean Tempera-ture*	*Mean Cloudiness (% of sky)*	*Monthly Rainfall (inches)*	*Average Wind direction and force*
Jan.	80	50	3·5	North, * moderate
Feb.	80	52	1·5	North, moderate
Mar.	81	50	1·5	North, moderate
Apr.	81	61	4·2	North, light
May	81	75	12.6	Variable, light
June	81	81	13·7	Variable, light
July	81	81	15.6	Variable, light
Aug.	80	80	15.2	Variable, light
Sept.	81	77	12.5	Variable, light
Oct.	80	77	15·5	Variable, light
Nov.	79	79	22.2	Variable, light.
Dec.	80	61	11.1	North, light

What are the outdoor conditions of life and thought and feeling near the poles and within the arctics? As a brief and partial answer to this always interesting question, Frederick Crockett, who was also on the 1928-30 Byrd expedition, gave me the following word-picture:

Usually during the months-long arctic 'day' there are low-hanging clouds, but below them, along the ground, visibility may be very good. The view in all directions is extraordinarily uniform—no shadows, no dark objects, no horizon, all gray above blending with chalky white below, 'like walking in a bowl of milk.' On this white, featureless surface it is impossible to see hummocks before stumbling over them, and walking is consequently very tiresome without skis, which more or less 'feel their way' along.

Distances in this milky light are very deceiving. A piece of dog dung a hundred yards away may look like a far-distant house. Because of the uniform white glare from the surface below, from which men's eyes are not naturally protected, sun blindness is likely unless protective goggles are worn. A very painful experience once it is incurred, something like burning hot sand in your eyes.

We had no trouble sleeping safely out on the trail at temperatures around 50°F below zero. Our sleeping bags were caribou hide with the hair turned inward, with thin woollen blankets inside. When crawling out in the morning one had to exhaust the bag of its warm air, lest the vaporous moisture condense and freeze, making a very uncomfortable sleeping place for the next night.

For clothing, we departed somewhat from previous arctic experience and wore separate thin layers of wool (successive sweaters). These sweaters we put on or took off, according to temperature and the degree of exertion needed at the time. On the trail we carried an outer natural-fur parka, but did not usually wear it (even at fifty-below temperatures) until we made a long halt, or when outside in the evening, and so on.

In general, wind was more important than temperature as a factor in comfort. On one winter day in July, 1929, two other men and I started out on skis for a short local exploration. It was 9 A.M., and still completely dark, when we left camp. The temperature was colder than 70 below -actually -72°F, as I remember it. Fortunately, however, there was no wind. It was completely calm;

the sky was clear, with the intensely bright southern-hemisphere stars (Canopus, the Southern Cross, and so on) swinging slowly around over our heads as the morning wore on, interspersed with the pale-green, flickering curtains of the aurora australis.

Starting off cool in thick clothing, we got well warmed up in a half-hour of skiing across the sand-like antarctic snow, which does not melt slightly under your skis like snow in warmer climates, providing a slippery surface. Hence skiing calls for heavy exertion in order to make any progress. Strange as it may seem, though at 9·30 A.M. the temperature was still 70 below and it was still dark, we were going forward barehanded and almost stripped to the waist. The only time we really felt the cold was on slight

downslopes where we could slide along at six or seven miles an hour. Here the lack of heavy exercise, combined with the slight wind of our forward speed, made us immediately feel the bitterness of the Antarctic cold. We continued thus in the darkness until around noon, when the sun appeared for an hour or so as a faint glow on the northern horizon.

To find the day-to-day march of normal temperature in your own particular American region, you must consult a Weather Bureau publication entitled 'Normals of Daily Temperature of the United States.' But by averaging the normals for Washington City with those of certain inland cities in the same latitude, we can obtain something fairly representative of a large (central and eastern) portion of the country. Atlantic seaboard temperatures average two or three degrees Fahrenheit warmer in winter than inland temperatures for the same latitude; in summer the seacoast may average a degree or two cooler. Also, yearly maximums and minimums on the seacoast come several days later than corresponding maximums and minimums inland.

PATTERN OF WINDS

Just as there are on this great globe fairly definite zones of *climate*, so there are even more definite belts of *prevailing wind*. Some of these round-the-world prevailing winds are quite steady and uniform, such as the trades and anti-trades. Others, such as the prevailing westerlies of temperate latitudes, are extremely variable from day to day and week to week.

But the *average* wind directions observed in temperate latitudes prove that here, too, there is a great round-the-world wind system from west to east — winds that blow generally across continents and oceans, across desert sands, sawtooth mountains, fields of rippling grain, waving trees in virgin forests, and wide expanses of heaving, restless, white-capped blue water.

The winds of the world are in essence a heat-pressure mechanism similar to that which causes a sea breeze, though enormously larger and more complex. The driving force behind all earthly winds, and all other earthly motions, is the enormous outpouring of radiant energy from the sun, our celestial ancestor and gigantic power-house ninetythree million miles out in space.

This radiant energy falls nearvertically on equatorial regions, and very obliquely on polar regions. Hence the equatorial regions are warmer, and are receiving more heat than they lose by earth-radiation. The poles, on the other hand, are radiating away more heat than they receive by sun-radiation. Hence, unless the tropics are to become unbearably hot and the arctics are to chill down to absolute zero (about -460° F. or -273° C.), there must be some atmospheric circulation between the two. A heated region usually develops low pressure at the surface and high pressure aloft, that a cooled region develops high pressure at the surface and low pressure aloft. The result, where only short distances are involved, is that lower winds blow generally from cold to heat, and upper winds blow generally from heat to cold. Hence we might expect, as part of the planetary wind picture, some tendency towards an equatorward drift at the surface together with a poleward drift aloft.

However, that any largescale wind is gradually turned aside, to the right in the northern hemisphere and to the left in the southern, by the earth's slow rotation under the wind in question. In other words this 'earth's deflective force,' so called, tends to turn any equatorward-drifting air westward, and to turn any poleward-drifting air eastward. In the upper troposphere, say at altitudes above two and a half miles, the *average* world wind pattern is a grandly simple thing that stems directly and understandably out of these elementary heat-pressure-wind principles. Upper pressure is relatively high over the equator, and relatively low over the poles, with a fairly uniform average horizontal pressure gradient between the two.

Hence the average upper-tropospheric winds of the northern hemisphere, directly driven by this pressure gradient and swayed by the earth's deflective force, are from southerly to southwesterly to westerly (drifting, that is, generally towards the northeast, or spirally towards and around the pole). In the southern hemisphere, driven by similar but oppositely directed forces, the average upper-tropospheric winds are mostly northwesterlies, drifting generally southeastward.

Either of these average drifts can be temporarily upset by regional vagaries, such as a southward surge of cold upper air in the north hemisphere. But as a norm or average condition, the two great upper drifts perennially move from the equator spirally eastward and poleward in each hemisphere. The planetary pattern of lower winds (below the two-and-ahalf-mile level) is more complex, even in idealized form. For the lower winds are influenced by surface friction, and the earth's deflective force acts on them as a more drastic limitation on poleward or equatorward progress.

Thus the planetary lowerwind pattern includes, not one wind belt in each hemisphere, but three. Moreover, these wind belts, and the pressure belts which separate them and regulate them, are in fact much modified and distorted (at least in the northern hemisphere) by the alternating surfaces of great continents and great oceans. Perhaps the easiest and best introduction to them would be a voyage through ideally typical weather at just 5° of latitude (three hundred sea miles) each day down the whole 'length' of the world, along the ocean-parting meridian of long. 30° W. from arctic to antarctic.

Any such imaginary meridianal voyage would cut across the following idealized or average wind-andweather belts: a zone of arctic easterlies above lat. 60° N.; a stormy 'polar frontal' zone (actually variable in both width and location) around lat. 60° N. or south of this; prevailing southwesterlies-to-westerlies (the familiar prevailing winds of temperate latitudes) between lat.60° N. and lat. 30° N.; a dry, clear calm belt called the 'horse latitudes' (probably because horses often died here on becalmed sailing ships) around lat.

30° N.; northeast 'trades' (so called by reason of their steady, perennial character and consequent importance to commercial sailing ships) from south of lat. 30° N. down towards the equator: a humid, oppressive, and showery calm belt called the 'doldrums' (because of their dispiriting or dulling or 'dolting' effect on human sailors unwillingly becalmed in them) around the equator; southeast trades from near the equator down to somewhere near lat. 30° S.; the southern horse latitudes around lat. 30° S.; the southern prevailing westerlies between lat. 30° S. and lat. 60° S.; the southern polar-frontal zone around lat. 60° S. or north of here; and the antarctic easterlies south of lat. 60° S.

The steady poleward air transfer of the prevailing westerlies and southwesterlies in our latitudes is compensated by occasional large and violent southward outbreaks of polar air (cold waves), especially in winter, when the polar anti-cyclone extends down over Alaska and northern Canada. Southward outbreaks sometimes also occur far aloft, with 100-m.p.h. north winds in the stratosphere.

Similar equatorward (northward) outbreaks occur in the southern hemisphere. Considering first the northern hemisphere: There is an average high-pressure area, or anti-cyclone, around the 'cold pole' (despite the fact that many arctic observers have noted low pressures). This polar high-pressure area drives the cold, arctic lower air spirally outward (or southward and westward) as the arctic easterlies (mostly northeasterlies) down towards lat. 60° N.

Along lat. 30° N. (the horse latitudes) there is a subtropical high-pressure belt, the northern slope of which drives the prevailing westerlies (mostly southwesterlies) up towards lat. 60° N. Around lat. 60° N., or south of there, these two great and conflicting wind systems meet along the low-pressure belt of the polar-frontal zone

— a stormy region which may actually swing far north or south, and may extend as separate and vari-formed fronts, between various unlike air masses, through a wide sweep of latitude. The subtropical high-pressure belt in lat. 30° N. (or its southern slope) also drives the northeasterly trades, as a general southwestward-to-westward drift around the earth, down into the equatorial low-pressure belt of the doldrums.

In the southern hemisphere, the subtropical high-pressure belt around lat. 30° S. drives the southeast trades up into the doldrums, and also drives the southern prevailing westerlies (mostly northwesterlies) down into the low-pressure polar-frontal zone around lat. 60° S. or north of there, where they meet and battle with the antarctic easterlies (mostly southeasterlies) from the south-polar high-pressure area.

In temperate latitudes, both north and south, lower winds have the same general or average direction as upper winds, and the prevailing westerlies form a deep average pattern (seldom fully duplicated in actual existing winds) that extends with increasing force up through the troposphere, moderating again in the lower stratosphere, and then mostly reversing, in light to moderate intensity, above twelve miles or so.

Within the tropics, on the 'winter side' of the equator, the easterly trades give way, at about two and a half miles altitude, to the planetary upper westerlies. Called the counter-trades or anti-trades, these tropical upper winds blow mostly from southwesterly in the northern hemisphere and northwesterly in the southern. The same planetary upper westerlies are present, to some extent, in arctic regions above the arctic easterlies.

MAJOR SCIENTIFIC ISSUES AND MEASUREMENT NEEDS

Changes in the ozone layer can be divided into two categories: natural changes and man-made changes. Separating these components is the goal of much ozone and trace gas research. Since ozone can be transported by stratospheric winds, there is significant interannual variability in column ozone amounts. Ozone is likewise influenced by aerosol amounts, the formation of nitrogen radicals associated with high-energy particles, and variations in the ultraviolet radiation from the sun.

Manmade changes generally include increased chlorine and hydrogen amounts from industrial gases and increased aerosols and nitrogen radicals from airplane exhaust. Many of our current scientific issues and future measurement needs centre around the interaction of the ozone layer with these pollutants and separating natural changes in the ozone layer from man-made processes.

Natural Changes

Interannual and Long-term Variability of the Stratospheric Circulation

Because the stratospheric circulation is strongly dependent on the dissipation of large-scale waves in the stratosphere, interannual variability of the wave amplitudes has an important impact on ozone transport. Winds and temperatures derived from 3-D GCMs and assimilation models include such interannual variability and can be used to assess the impact on ozone transport. 2-D models can incorporate prescribed variability to simulate interannual ozone transport.

Accurate assessment of the large-scale waves and the transport circulation is necessary for understanding the variability of ozone trends. One of the failures of the 3-D models is inadequate simulation of the QBO. The QBO is a 24-30-month oscillation of the zonal wind in the tropical lower stratosphere that is driven by tropical waves.

The QBO affects the stratospheric temperature distribution and produces a secondary circulation which transports trace gases and aerosols. For ozone, the QBO can generate variations from the climatological mean of 5-10 DU in the tropics. There is also a QBO-ozone signal outside the tropics of 10-20 DU.

The QBO provides one of the largest components of the interannual variability of the column ozone values. Because the geostrophic relationship breaks down in the tropics, direct tropical wind measurements are critical to precisely measuring the QBO and for understanding the effects of the QBO on the circulation. Data-sparse regions and infrequent sampling of wind fields all preclude good quantitative studies of the tropical circulation and its effect on ozone.

External Influences

Solar ultraviolet radiation and precipitating energetic particles can strongly influence ozone amounts. In order to understand the anthropogenic changes in ozone, we must maintain reliable measurements of the solar ultraviolet input to the middle atmosphere. Solar variations in the UV produce ozone changes on the same order of magnitude as the current observed midlatitude changes. Proxies for the UV changes have been historically used to estimate the response of ozone to solar ultraviolet changes. With direct measurements from UARS, these proxies have been shown to inadequately represent changes in ultraviolet flux. Particle events generate NOx compounds which catalytically destroy ozone, but these events tend to be confined to the upper stratosphere.

Large events, which tend to be more episodic, may affect polar ozone at lower levels. The impact of NOx generation through particle precipitation on the natural ozone layer is a major scientific question.

Natural Aerosols and PSCs

Aerosols and PSCs are believed to play a major role in ozone loss. Irregular volcanic inputs of SO_2 with the subsequent formation of sulfate aerosols have an impact on the ozone layer. There is some evidence suggesting that increasing amounts of background aerosols are a result of subsonic aircraft emissions in the lower stratosphere.

A major scientific question is whether the background amounts of these aerosols are increasing and, if so, determining their origin. Monitoring the aerosol amounts within the stratosphere and determining their trend is a primary measurement requirement to understand ozone loss. During the 1980s it became apparent that aerosols play an important role in the chemistry of the stratosphere.

Observations of large decreases in ozone over Antarctica during the Southern Hemisphere spring were not accounted for by theory, until several researchers hypothesized that heterogeneous reactions on PSCs might be converting inactive chlorine compounds into reactive forms.

In a similar fashion to PSCs, heterogeneous reactions upon sulfuric acid droplets at midlatitudes convert N_2O5 into HNO3 and shift the ratio of HNO3 to NO_2 normally present in the stratosphere. Throughout the stratosphere, reactions on and inside aerosol particles are therefore important.

To understand the effectiveness of the heterogeneous reactions, it is important to know:

- The temperatures of the aerosol particles;
- The surface and volume densities of the aerosol particles, which are derived from the aerosol extinction, and a knowledge of the size distribution;
- The composition and phase of the aerosol particles;
- The concentration of the reactants in the aerosol;
- The duration of time over which the heterogeneous reactions occur.

A theoretical framework, by which heterogeneous rates of reaction are quantified, is given in Hanson. An important research goal is the ability to observe the yearly episodes of ozone loss in the polar regions, to measure this loss as reservoir chlorine levels change with time, and to be able to relate the changes in observed ozone to a quantitative understanding of heterogeneous processes.

In principle, one should be able to identify the composition of stratospheric aerosol from multi-wavelength extinction data. Multi-wavelength observations of midlatitude sulfuric acid droplets have an extensive history. Observations of El Chichón aerosol, post El Chichón aerosol, and of Mt. Pinatubo aerosol yield spectral data consistent with theoretical expectation. Analysis of multi-wavelength observations of PSCs is a developing research topic.

Recent attempts to use spectra to determine PSC composition are showed by Toon and Tolbert. Several years ago, ice and nitric acid trihydrate particles were thought to be the primary composition of PSCs. Recent studies have shown that some PSC particles are liquid, and not that of crystalline NAT. As additional laboratory coldtemperature measurements of the indices of refraction of PSC composition candidates become available, the ability to classify PSC composition from spectra will improve.

Although PSCs are now known to be instrumental in polar ozone loss, their amounts and types must be monitored. The major difference between the Antarctic ozone depletion and the less-severe Arctic depletion appears to be the result of a lack of denitrification in the Arctic. Fundamentally, denitrifi-cation is a function of temperature and the size of PSCs. Above frost point the PSC size is generally too small to precipitate nitric acid from the stratosphere. If temperatures reach frost point, larger PSCs form, which are able to remove nitrogen acid from the lower stratosphere. The temperature history of the air parcel may play an important role in the PSC size distribution as well. Photolysis of the nitric acid is key to halting the ozone depletion during winter.

With the increase of greenhouse gases, the stratosphere is expected to cool and thus increase the probability of PSC formation as well as increase the surface area and heterogeneous reaction rates on sulfate aerosols. Preliminary studies suggest greenhouse gas increase could have a major role in polar ozone depletion through increased probability of PSC formation. Monitoring stratospheric aerosol loading and PSC amounts is critical for understanding ozone loss.

Man-made Changes

Man-made changes in ozone mostly arise from the manufacture of unreactive chlorine-containing compounds such as the CFCs. These compounds reach stratospheric altitudes where photolysis by ultraviolet radiation releases chlorine with subsequent destruction of ozone through catalytic cycles. Aviation also has an impact on ozone through the release of nitrogen radicals in aircraft exhaust.

Trends in Chlorine Source Gases

Chlorine source gases and their respective trends are the major drivers behind decreases in stratospheric ozone. A comprehensive discussion of chlorine source gases is contained in WMO.

Historical Trends in Chlorine Source Gases

All chlorine in the stratosphere comes from tropospheric sources, predominantly the man-made CFCs and chloro-carbons. The man-made sources account for about 7/8th of the total stratospheric chlorine. CFCs are currently being phased out in favour of the hydrochlorofluorocarbons. Extensive measurements of the chlorofluorocarbons CFC-11, CFC-12, and CFC-113 have indicated a steady increase in their tropospheric mixing ratios for more than a decade. Most recent data suggest that the growth rate for these species has begun to decrease. Measurements taken from Tasmania suggest that levels of the important chlorocarbon CCl4 in the troposphere are also decreasing.

As HCFCs are introduced as substitutes for CFCs, it may be expected that their mixing ratios in the troposphere will increase well into the next century. HCFC-22 data show a near-linear growth rate in recent years. HCFC-141b and HCFC-142b have been available only recently as CFC replacements. These species are clearly increasing in the troposphere, but further data is required to get reliable growth rates for long-term studies.

CH3CCl3 data also indicate a reduced growth rate that is a result of recently-reduced emissions, but also possibly due in part to increasing hydroxyl levels. Data for dichloromethane, methyl chloride, and chloroform currently exhibit no long-term trends. Continued tropospheric measurements of these gases are required to estimate ozone depletion potential.

Stratospheric Chlorine

An extensive compilation of measurements of chlorine source gases in the stratosphere can be found in Fraser. The most comprehensive suites of simultaneous measurements of chlorine constituents in the stratosphere include the Atmospheric Trace Molecule Spectroscopy experiments of 1985, 1992, and 1993, and the Airborne Arctic Stratospheric Expedition II measurements of 1991, 1992.

The data from these missions have provided invaluable information on the stratospheric chlorine burden and the partitioning among the various chlorine species. Based upon the 1985 ATMOS data, Zander *et al.* determined a total stratospheric chlorine level of 2.55 ± 0.28 ppbv. Further, they concluded that above 50 km most of the inorganic chlorine was in the form of hydrogen chloride, and that the partitioning of the chlorine among sources, sinks, and reservoir species was consistent with that level of total chlorine. From the 1992 ATMOS flights, total stratospheric chlorine was estimated to be 3.4 ± 0.3 ppbv, an increase of approximately 35% in seven years. This increase is consistent with that predicted by models. Schauffler *et al.* inferred total chlorine levels of 3.50 ± 0.06 ppbv from the AASE II data near the tropopause, a value which is in excellent agreement with the 1992 ATMOS values.

Recent HCl data from HALOE on UARS reveal a trend in HCl versus time at 55 km compared with the estimated total Cl trend based on tropospheric emissions. Of the total stratospheric burden, only about 0.5 ppbv is estimated to arise from natural sources in the troposphere but these estimates have yet to be confirmed by direct or remote observations. HCl emissions from major volcanic eruptions provided negligible perturbations to the levels of HCl in the stratosphere.

Depletion of Ozone by Stratospheric Chlorine

Estimates of the severity of ozone depletion in the future can only be determined by atmospheric model simulations. The level of confidence in these models is based upon their ability to simulate present atmospheric distributions and their ability to simulate recent trends. A discussion of the strengths and weaknesses of current assessment models is contained in WMO.

Model simulations of ozone change spanning the period 1980 to 2050 were conducted as part of the WMO assessment process. Two scenarios were adopted for the assessment studies: 1) the emissions of halocarbons follow the guidelines in the Amendments to the Montreal Protocol, Scenario I; and 2) partial compliance with the guidelines, The results of the model calculations for

Scenario I. The per cent change in the ozone column at 50° N in March for each of the models participating in the assessment. Decreases of up to approximately 6.5% are seen to occur just prior to 2000. The recovery time to 1980 levels varies widely for the different models, from as early as 2020 to well past 2050. The individual models all showed reasonable agreement among themselves for the present-day ozone distributions, but begin to differ substantially as the atmosphere is perturbed away from its existing state by increasing levels of nitrous oxide, methane, halocarbons, and other influences.

Uncertainties in the absolute levels of depletion predicted by the models are difficult to evaluate for these long-term scenario calculations. The trends in the source gases are changing, and the trends in the stratospheric reservoir gases, which are dependent on transport into the stratosphere, will respond.

Thus, measurements of the chlorine source and stratospheric reservoir gases must be made to test models against observations. Critical gases in the suite of required measurements are the reservoirs HCl and ClONO2. The predictive capability of these assessment models directly rests on additional measurements of chlorine source gases, reservoir gases, and gases which are sensitive to transport processes.

Effects of Aircraft Exhaust

Long-lived source gases are unreactive in the troposphere and hence can enter the stratosphere at the ambient tropospheric concentrations. In the stratosphere, these gases undergo photolysis or react with radicals to release their potential ozone-destroying catalytic agents. In contrast, aircraft flying in the stratosphere will directly inject catalytic agents into the stratosphere.

The primary agents for potential ozone change which have been considered in studies of aircraft exhaust are the nitrogen oxides and water vapour. Now that heterogeneous reactions on background aerosols and PSCs are known to play an important role in the ozone balance of the stratosphere, the evaluation of the effects on ozone of NOx from supersonic aircraft flying in the stratosphere has changed significantly.

The impact on column ozone of a fleet of supersonic transports is now calculated to be of the order of 1% or less. An important possibility is that the sulfur in the exhaust will lead to the generation of numerous small particles which will add to the aerosol surface area. An increase in surface area will enhance the conversion of chlorine from its reservoirs to ClOx and thus could lead to an increased loss rate for ozone. Another possibility is that the other condensibles in the exhaust, water vapour, and nitric acid could impact the formation or duration of PSCs. Initial calculations show this effect to be small and transport studies show that injection into the polar vortex is unlikely but there is still uncertainty about what will happen as the stratosphere cools with increasing CO_2 concentrations.

All of the chemical effects of HSCT exhaust depend on how much of the exhaust products accumulate in the stratosphere and where they accumulate. The same is true for the exhaust of the subsonic fleet, which is released in the upper troposphere and lower stratosphere. The three major potential effects of the subsonic fleet of aircraft are ozone increase due to the smog-like photochemistry of NOx, CO_2 increase due to fuel consumption, and cirrus cloud formation from the water vapour. The importance of aircraft NOx to ozone generation in the upper troposphere and lower stratosphere is not completely understood. Aircraft NOx sources have to be compared to the NOx sources due to lightning, stratospheric intrusions, and the lofting of ground-level pollution in cumulus clouds. Thus the role of aircraft as a source of upper tropospheric NOx and its impact on lower stratospheric ozone is uncertain. Also uncertain is whether heterogeneous chemistry on ice crystals plays a significant role in the NOx budget. Understanding the impact of supersonic and subsonic aircraft exhaust on the stratospheric chemical balance is a complex problem.

Knowledge of meteorological conditions is required to compute exhaust dispersion. Knowledge of aerosol chemistry is required to understand the aerosol formation process and its impact on the background conditions. Finally, a good understanding of the lower stratosphere chemistry is required to understand the direct impact of the NOx pollutants.

Summary of Science Issues

The investment by the scientific community in instrument and model development has produced a significant increase in the understanding of stratospheric chemical and dynamical processes. Although some fundamental questions of ozone loss have been answered, new questions have arisen. For example, the long-term response of the ozone layer to natural fluctuations is still not well understood.

The secular decrease in ozone following the eruption of Mt. Pinatubo was clearly associated with aerosol loading of the stratosphere—but the nearly one-year delay in the appearance of maximum ozone loss is still not explained. More fundamentally, the midlatitude trend in column ozone loss reported by Stolarski is still not explained. Our understanding of the more-subtle chemical processes is still quite incomplete, which increases our uncertainty in the forecast predictions. Under the Atmospheric Effects of Aviation Programme the impact of stratospheric and tropospheric aircraft pollution on stratospheric ozone is now being investigated.

The research studies have reemphasized that our understanding of stratospheric transport is not complete with regard to transport, especially the containment of the pollutants within the midlatitude release regions and the distribution and magnitude of stratosphere-troposphere exchange, especially exchange of ozone. Many of the issues associated with the stratospheric circulation are above the observing range of current stratospheric aircraft. The analysis of UARS measurements has also revealed the tremendous advantages of global chemical data sets.

Finally, the most extensive observations of solar UV and energetic particle impact on ozone have been made recently by UARS. Unfortunately, these observations have been made during the declining phase of the solar cycle, and we have not developed a longenough baseline of measurements to quantify the impacts of changing solar conditions. Long-term measurements of solar UV and total solar irradiance are needed during the waxing phase of the solar cycle.

REQUIRED MEASUREMENTS AND DATA SETS

The minimum measurements, their accuracies, and the instruments which will make the measurements. Often, key measurements will be made by more than one instrument, which gives the whole measurement suite a level of robustness in case of instrument failure.

Meteorological Requirements

An understanding of the photochemistry of the stratosphere is clearly contingent on high-quality observations of temperature. The temperature field affects stratospheric physical processes in a number of ways. First, temperature fields are used to calculate geopotential heights and winds via the hydrostatic and geostrophic approximations.

Second, temperatures affect the radiation field, particularly in relation to the longwave cooling in the stratosphere. Third, temperatures affect the chemistry via temperaturedependent reaction rates, and via the formation of PSCs. Hence, accurate and precise temperatures provide a basic foundation for stratospheric chemistry, radiation, dynamics, and transport. Low-quality tropical meteorological observations are an impediment to our understanding of the interaction of the tropics and the middle latitudes. The National Plan for Stratospheric Monitoring 1988-1997 set down the minimum requirements for meteorological variables between 1000 and 0.1 hPa.

Their requirements were: 2.7-km vertical resolution; 12- hour time resolution; 1-K precision for temperature, and 5 m s-1 precision for winds. Current radiosonde and rawinsonde measurements have precisions of a few tenths of a kelvin and 1-4 m s-1 wind speed in precision. Unfortunately, the balloon-borne rawinsonde system is limited to altitudes below 30 km. For higher altitudes, the meteorological rocket network provided some data, but the network has been effectively discontinued. Satellite systems are now relied upon to provide all of the meteorological information above 30 km. The National Oceanographic and Atmospheric Administration TIROS Operational Vertical Sounder SSU instrument has an error of 2 K at 10 hPa rising to 4 K at 1 hPa. The TOVS weighting functions are about 10-12-km deep. Later NOAA sounders use the Advanced Microwave Sounding Unit instead of MSU/SSU. The AMSU weighting functions are about half the depth of the TOVS functions. AMSU temperature measurements are limited to the atmosphere below 50 hPa. The UARS Microwave Limb Sounder has a vertical resolution of a few km, although its horizontal coverage is inferior to nadir-sounding TOVS and AMSU instruments. Improved understanding of stratospheric chemistry and heterogeneous processing suggests that improvement of temperature measurements will have an impact on our ability to predict where the heterogeneous reactions will take place.

Lower stratospheric temperature measurements made during the numerous polar aircraft missions suggest that the meteorological analyses in the Southern Hemisphere are warm biased by about 2 K. This suggests that the Earth EOS stratospheric temperature accuracy requirements should be less than 0.5 K. It is also important that good temperature measurements be made near the tropical tropopause, especially in cloudy regions where air is entering the stratosphere through the tropopause. Direct stratospheric wind data are needed where the divergence fields are significant. The current wind requirement for assimilation models is an unbiased horizontal wind field accuracy of 2-5 m s^{-1}. These should be global measurements with a vertical resolution of a few kilometres. Presently, the UARS High Resolution Doppler Imager satellite wind instrument makes these measurements at the upper end of the limit. A joint effort between NOAA, the Department of Defence, and NASA will produce the National Polar Orbiting Environmental Satellite System.

This system will take needed operational data and certain long term observations for climate studies. This system will be in place after 2008. The requirements for meteorological variables between 1000 and 1 mb include the following: temperatures accuracys of better than 1.5 K with vertical resolutions of 1 to 5 km from the ground to the mesosphere. The local revisit time is 6 hours and horizontal cell size of about 50 km. The NPOESS also has operational requirements for total column and profile ozone, aerosols and winds.

Chemical Measurement Requirements

Atmospheric composition measurements form a cornerstone of any global change strategy. Chemical and dynamical measurements must be made in both the stratosphere and the troposphere. Indeed, chemical measurements around the upper troposphere and lower stratosphere should be among those with the highest priority.

Science Questions

The science questions for stratospheric processes are mostly focused on the changes in the stratosphere expected to take place as anthropogenic pollutants accumulate in the middle atmosphere. Greenhouse gases are expected to substantially increase during the EOS period. Stratospheric halogens are expected to increase until 1999, then level off and slowly decline as a result of international regulations.

The increases in these gases should produce chemical and dynamical changes. The magnitude of the stratospheric cooling in response to increasing greenhouse gases should far exceed the tropospheric warming because there are fewer feedback mechanisms which buffer the radiative interaction. Increases in chlorine and bromine will cause decreases in stratospheric ozone.

The ozone decrease could be exacerbated by colder lower stratospheric temperatures caused by increasing greenhouse gas concentrations. For example, the colder stratospheric temperatures may lead to an expansion of the extent of PSCs and, hence, polar ozone depletion. The complex chemistry of the stratosphere can only be understood in detail by measuring a broad range of species over varying conditions with global coverage and over at least an annual cycle.

The first area which merits further observational and theoretical study is polar chemistry processes. Direct, simultaneous measurements of HOCl, HNO3, and N_2O5 are critical since these gases are believed to be involved in PSC surface chemistry. Also, polar night observations, above 20 km, of the chemically-active species, along with PSC measurements, are needed in understanding polar ozone depletion. These regions are not presently accessible with balloons and aircraft. In order to understand the large ozone depletion at midlatitudes, simultaneous measurements of N_2O5, HOCl, HNO3, and HCl are needed to assess the role of heterogeneous chemistry on background aerosols. Since OH and HO_2 drive the chemistry of the lower stratosphere, global measurements of these gases are required to evaluate ozone losses, especially any zonal asymmetries.

lso, lower mesosphere observations of OH and HO_2, along with O3 and temperature, are likely to be key links in understanding the large O3 decrease expected to occur near 40 km as chlorine levels continue to rise. It is clear that full understanding of these changes requires not just O3 and ClO measurements, but HOx and NOx measurements as well. Measurements in the lower mesosphere, where the chemistry is more simple, may provide the best data set for this analysis.

Key Trace Gas Measurements

There are several scientific requirements to address middle-atmosphere chemistry issues:

- The self-consistency between the source gases and the resulting active reservoir gases needs to be tested for the four major families that are important to ozone chemistry. The four families and most important species measurements required are: oxygen family, hydrogen family, nitrogen family, and chlorine family. Stratospheric chlorine is predicted by atmospheric models to increase by 20% in the next five years; thus, our understanding of the production and partitioning among the individual family constituents needs to be verified.
- The changes in the Antarctic/Arctic lower stratosphere constituents during the ozone hole period in the winter and spring need to be monitored. Since significant changes have been detected during the 1980s and 1990s in the polar

regions, these geographical areas require special attention and monitoring.

- There are a few chemical process studies which require investigation as indicated below.
  - — The HOx family is fundamentally important in stratospheric chemistry, but the database for that group remains one of the poorest in the atmosphere. Global measurements of the latitudinal, seasonal, and diurnal variation in the HOx family and related species, H_2O and O3, are needed to address this deficiency.
  - — Models for the past decade have predicted less ozone in the upper stratosphere than is measured. Several species need to be measured in the upper stratosphere to help resolve this difficulty.
  - — Models, in general, predict less odd nitrogen in the lower stratosphere than observed. Measurements of odd nitrogen species, NO_2, HNO3, N_2O5, and $ClONO_2$ in the lower stratosphere will help to deal with this problem.
  - — Another odd nitrogen species, HNO3, is not modeled accurately in the wintertime in the mid-to-high latitudes. A measurement of HNO3, N_2O5, H_2O, and aerosols should help confront this problem.
- Global observations of ozone in the lowermost stratosphere with high horizontal, vertical, and temporal resolution are needed in order to quantify the ozone budget in that region of the atmosphere and, in particular, to determine the spatial and temporal distribution of ozone fluxes from the lowermost stratosphere to the troposphere. These fluxes, which are presently known only to within about a factor of two, are important for the ozone budget of the lowermost stratosphere and are crucial for understanding the ozone budget of the upper troposphere.

Generally, the accuracies needed are 5-10% of the ambient concentrations found in the lower stratosphere.

Stratospheric Aerosols and PSCs

The remote sensing of the composition of aerosol at midlatitudes is fairly straightforward. However, remote sensing of the composition and phase of the aerosol particles, for the case of the PSCs, is a developing topic of research. One research goal is to see to what extent it is possible to estimate the composition, phase, area, and volume densities from orbital observations.

It is known that the volume densities of NAT and ternary particles are different. Carslaw and Drdla have shown that ternary solutions best describe some of the NASA High-Altitude Research Aircraft data. Beginning with a sulfuric acid droplet core, the volume density of the aerosol increases as temperatures become colder. These equilibrium curves were calculated using different amounts of ambient HNO3. Remote-sensing observations of temperature versus volume density will likely help classify the composition and phase of the PSC particles. It is also known that HNO3 is incorporated in ternary, nitric acid dihydrate and NAT particles as a function of temperature. Therefore, the simultaneous observation of aerosol extinction and HNO3 gas mixing ratios should help one to classify regions of PSCs as to composition and phase.

Since the microphysics of PSC particles is very temperature sensitive, absolute temperatures need to be measured to plus-or-minus 2 K, since curves of temperature versus volume density for NAT, ternary, and NAD particles differ by only a few kelvins. Remote-sensing observations also average over many kilometres along a horizontal ray path. Vertical coverage is usually on the order of several km.

Thus, the fine-scale structure of PSCs, as sampled by ER-2 instruments, cannot be resolved by the remote sounder. Another complication is due to present limitations in the theoretical understanding of how PSCs form, which compositions are formed, and the need for additional laboratory work to quantify at cold stratospheric temperatures the rates at which realistic PSC particles convert inactive to active chlorine compounds, and the need for additional laboratory measurements of the refractive indices of

PSC and sulfuric droplets. Current research will see to what extent it is possible to refine present capability to quantify the mechanisms of PSC chemistry, as observed from orbit.

Solar Ultraviolet Flux

Solar radiation at wavelengths below about 300 nm is completely absorbed by the Earth's atmosphere and becomes the dominant direct energy input, establishing the composition and temperature through photodissociation, and driving much of the dynamics as well. Even small changes in this ultraviolet irradiance will have important and demonstrable effects on atmospheric ozone.

Radiation between roughly 200 and 300 nm is absorbed by ozone and becomes the major loss mechanism for ozone in the middle atmosphere. Likewise, solar radiation < 200 nm is absorbed predominantly by molecular oxygen and becomes a dominant source of ozone in the middle atmosphere, so changes in these ultraviolet wavelengths will have, to first order, an inverse influence on ozone.

These two atmospheric processes, driven by solar radiation, become the major natural control for ozone in the Earth's stratosphere and lower thermosphere. To fully understand the ozone distribution will require many coordinated observations and, in particular, a precise measurement of the solar ultraviolet flux. The visible portion of solar radiation originates in the solar photosphere and has been accurately measured for about fifteen years.

Apparently, this radiation varies by only small fractions of one per cent over the 11-year activity cycle of the sun, with comparable variation over time scales of a few days. The ultraviolet portion of the solar spectrum comprises only about 1% and originates from higher layers of the photosphere. As we move to shorter and shorter wavelengths, the emission comes from higher and higher layers of the solar atmosphere.

Unlike the solar photosphere, these higher levels are much more under the influence of solar activity, as manifested, for example, by increasing magnetic field strength. As the magnetic activity increases or disappears, the solar radiation, especially the ultraviolet, undergoes dramatic variations modulated by the 27-day rotation period of the sun. Near 120 nm the variation over time periods of days to weeks can be as large as 50%, and over the longer 11- year solar cycle the variation can be as large as a factor of two. Towards longer wavelengths, the solar variability decreases to levels of about 10% at 200 nm and finally to only about 1% at 300 nm. Longward of 300 nm, the intrinsic solar variability is probably only on the order 0.1%, roughly commensurate with measurements of total solar radiation.

The challenge during the EOS time period is to provide measurements of the solar ultraviolet with a precision and accuracy capable of tracking the changes in the solar output. Ideally, the instrument will be capable of measuring changes as small as one per cent throughout the EOS mission. This requirement is extremely challenging for solar instruments, especially those making observations at the ultraviolet wavelengths, which are notoriously variable.

The harsh environment of space, coupled with the energetic solar radiation, rapidly degrades optical surfaces and usually makes the observations suspect. Some manner of in-flight calibration is required to unambiguously separate changes in the instrument response from true solar changes.

Validation of Satellite Measurements

The role of validation of satellite-based chemical measurements cannot be over stressed. Validation measurements, especially measurements of the same species using two different techniques, have proved to be invaluable for understanding satellite trace species measurements. The very successful UARS validation campaign has contributed a great deal to understanding the individual UARS measure-ments. The validation campaigns perform two major functions. First, they test the ability of a satellite instrument to make a measurement by giving an independent data point to compare against. Second, if the validation measurements are performed as part of a larger, coordinated campaign, the validation measurements done using aircraft and ground-based measurements can be used to link the small-scale geophysical features that they can observe with the large-scale geophysical features observable from space.

WEATHER MEASUREMENTS

As we all know without being told, it is our perceptions of the weather that both enable and compel every one to take a personal interest in it. And yet we can not go very far with perceptions alone in comparing the weather of one day with that of another, or the weather of one place with that of any other place. With respect to temperature, for instance, we may use such comparative expressions as cold, cool, comfortable, warm, hot; and that is about all. And even these few gradations are uncertain, for when the air is cool to one person it may be cold another; nor can either say with any assurance when it ceases to be cool and begins to be cold. And the situation is worse for most of the other weather elements.

In fact, for some of them, such as atmospheric pressure and electrical state, we have no distinct perceptions at all. However, every weather element affects inanimate objects and in such manner that measurements of almost any desired degree of refinement are easily made and recorded, measurements, or instrumental readings, that are of great and increasing importance in the industries, that are the data from which weather forecasts are deduced, and that at the end of a few years' accumulation give us reliable knowledge of the climate of the place at which they were obtained.

Early beginnings of weather measurement

No measurements of this kind were made in ancient times, except crude estimates here and there of the direction of the wind, and equally crude measurements of the amount of rainfall as indicated by the catch in certain exposed vessels. Both those very obvious and simple measurements have been made from time to time and at one place or another during the last 2000 years, and perhaps much longer.

But no other weather element was measured—there was no instrument to measure it with—until less than 300 years ago; not, indeed, until, grown-ups, retaining and respecting to the inquisitiveness of the child, began to put questions direct to Nature instead of turning to Aristotle or supporting wild fancies by convenient interpretations of passages from the Bible. Instrumental meteorology was born at Padua, Italy, in the first half of the 17th century, the child of the thermometer, invented by Galileo, and of the barometer, invented by Torricelli. Since that time it has grown mightily, with many new features added and refinements acquired, but always its chief dependence has been on its parents, the thermometer and the barometer.

Temperature

Every one has a temperature sense, at least enough to distinguish between hot and cold, and he will readily agree that a hot object has a different temperature from that of a cold one, but even so he can not, without special instruction, know just what temperature is, or make any distinction between temperature and heat. Indeed the wisest in Nature's ways were themselves a long time in finding this out. All the Solons and Solomons down to near the middle of the 18th century appear to have accepted the existence of heat without attempting to determine its nature, or even to speculate about it.

Then, with the advent of the private laboratory and the eagerness to understand natural phenomena, came philosophizing and speculation, and hear came to be regarded as a sort of material substance, but one without weight, an imponderable, since by experiment objects were found to weigh just as much when cold as when hot. Near the close of the century (18th) it was found that much heat can be added to ice at its melting point without changing its temperature, from which it was inferred that heat and temperature are radically different things. At about the same time it was found that an indefinite amount of heat can be gotten from a couple of objects just by rubbing them together; and finally, near the middle of the 19th century, it was found that there is a definite relation between of work and quantity of heat.

Thus, tediously and laboriously, we learned that heat is not a substance at all, but a something interchangeable with work, or as we now call it, a form of energy—energy being capacity for doing work. But this doesn't yet tell us what temperature is. Perhaps we might regard temperature as a measure or index of the concentration of heat inasmuch as the same amount of heat, as near as we can judge it, makes a cup of water much hotter than a pot of water, provided neither is brought to the boiling point.

This is easily understood, but like lots of other simple definitions and explanations it is all wrong, as Nature proves by citing an endless number of facts like these: The same amount of heat that would just melt a pound of ice, making no change whatever in the temperature, would heat a pound of ice water to a scalding temperature, 176° F., roughly, or seven and three-quarter pounds of ice-cold lead up to its melting point. Evidently, then temperature is not a measure of the concentration of heat.

But there is another experiment that helps us to a definition of temperature that appears to be satisfactory. The experiment is this: When an object that is decidedly cold to the touch is put in intimate contact with one that feels distinctly warm, and both are protected from contact with other objects, the cold one always gets warmer and the warmer one cooler. From this we conclude that temperature is that thermal state of an object that enables it to communicate heat to other objects. We may assume that there is a flow of heat from each object to the other, whatever their temperatures, as there certainly is when this exchange is by means of radiation, but in all cases the net result is a gain of heat by the colder body and a loss by the warmer.

Thermometers

It has been found by experiment that all objects change in volume with change in temperature, and this has furnished our most convenient and common method of comparing and labeling temperatures. A small bore glass tube, for instance, with a bulb at one end is filled to the desired point on the stem with mercury, or other fluid which will expand faster than glass, and sealed off, and the whole made into a thermometer is follows: First, the bulb and stem are placed in a mixture of pure ice and water, which always has the same temperature, and the end of the mercury column marked on the glass tube.

This point is labeled 0 (zero) if we are making a centigrade thermometer, or 32 if it is to be a Fahreheit thermometer. Next, the bulb and stem are immersed in a bath of live steam over boiling water and the new position of the end of the mercury column marked. The atmospheric pressure is noted at the same time because the temperature of the steam varies with this pressure. If the pressure is just right (if not right the proper correction can be made) the new mark on the stem is labeled 100 or 212 according as the thermometer is to be centigrade or Fahrenheit. Finally, if the bore of the stem is uniform, the distance between these two marks is subdivided as finely as desired and as many as necessary of the division marks appropriately labeled. Also similar divisions may extend beyond each of the initial end marks.

This construction of a thermometer is based on the fact that the volumes of the glass vessel and the mercury both increase almost uniformly with increase of temperature, but unequally—the mercury fastest. There are, of course, many refinements not indicated here in the manufacture of good thermometers, but the principle remains the same. Of course some other fluid than mercury may be used, and is used, for measuring temperatures below -40° F., the temperature at which mercury freezes.

Maximum Thermometers

When the tube of the thermometer is sharply constricted a short range, generally near the bulb, as in the case of the well-known clinical thermometer, the mercury passes by that place on increase of temperature due to the force of expansion, but does not pass it, the column promptly pulling in two, on decrease of temperature. It can, however, be thrown back by a proper swing of the instrument. This thermometer therefore can be, and is, used to mark the highest temperature to which it was exposed since it was last adjusted. It is the standard instrument used for obtaining the day-to-day maximum temperature of the air.

Minimum Thermometers

If alcohol is used as the thermometer fluid, and a miniature glass dumb-bell, colored so as to be more easily seen, is placed within the fluid in the bore of the stem, and the whole mounted horizontally, then with increase of temperature the alcohol flows along the tube past the dumb-bell, which fits loosely, and without disturbance to it.

On the other hand, when the column contracts the surface film at its outer end drags the dumb-bell along with it. Hence this thermometer, much used in meteorological work, gives the lowest temperature, as indicated by the outer end of the sliding index, since its last adjustment—an operation that consists only in holding the thermometer sufficiently inclined to let the index drop into contact with the surface of the fluid column.

The "empty" portion of the stem should contain air so as to reduce the trouble due to a shift of a portion of the alcohol to the outer end of the stem by evaporation and condensation.

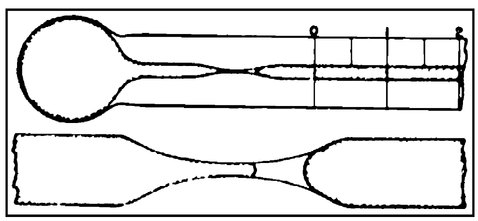

Fig. ***Constriction in the Stem of a Maximum Thermometer.***

Telethermoscope

Often while indoors we need to know what the temperature is outdoors, and therefore have devised instruments that give us that information without the trouble of going to the roof, or wherever the outdoor thermometer may be. An excellent instrument of this kind is based on the fact that the electrical resistance of a metal wire (pure nickel is excellent) varies with temperature. We therefore can expose such a wire, properly protected from injury, to the free air, and connect it with a small electric battery and a current indicator on our desk, if we like.

We then can vary an adjustable resistance in the circuit until it is equal the current indicator, to the resistance at that moment of the thermometer wire. If also the action that adjusts the balancing resistance correspondingly moves a pointer, a result easily secured, we can, of course, mark each position of that pointer with the then temperature of the exposed wire.

After this we have only to close the circuit (open except when an observation is being made) and turn the pointer to the position of balance, a matter of only a few seconds, to read on our desk, and at any time, the temperature of the air outdoors, or of any thing else to which the thermometer is exposed.

This instrument is extensively used in meteorological offices. It enables the observer to answer frequent requests for the current temperature without leaving his desk to the neglect of other duties.

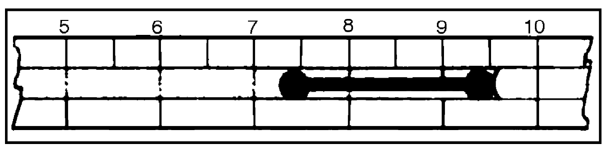

Fig. ***Dumb-bell in the Stem of a Minimum Thermometer.***

Thermograph

Reading a thermometer every few minutes day and night, month in and month out, is wholly impracticable. Yet a continuous record of the temperature is desirable, for many significant changes occur at brief intervals. Hence thermographs have been devised that keep an absolutely continuous record of the temperature of the air. The basic element of the most extensively used instrument of this kind. Consists of a thin-walled, flattened, metal tube; curved along the flat sides, completely filled with alcohol and tightly closed; in short, a Bourdon tube of the kind that is widely used in measuring gas and liquid pressures.

With increase of temperature the enclosed liquid, expanding faster than the metal, tends to straighten the tube; and with decrease of temperature curves it more. Hence by rigidly fastening one end of it to a fixed object, and a pen to the other or free end, we have directly, or through a multiplying linkage, a means of obtaining a continuous record of the temperature on a properly graduated moving sheet of paper.

Telethermographs

Obviously the recording device of a thermograph can be at one place and the thermal element at another, with either electrical or mechanical connection. Both kinds of instruments are in use, but they are not numerous since the need for them is relatively slight.

Thermometer Shelter

To obtain the correct temperature of the air outdoors it is necessary that the thermometer of whatever kind be protected from sunshine, both direct and reflected, since either, and especially the former, would make it read much too high; also that it be protected from rain and snow as they would make it register too low. At the same time it must be exposed to free and easy circulation of the air.

These ends are sufficiently well obtained by housing the thermometer in a double roofed shelter with louvred sides, all painted white. This shelter does not secure entire freedom from errors, especially when the sun is shining on it, but they seldom are greater than a small fraction of a degree and there- fore may be neglected.

A much more important matter is the location of the shelter and its height above the surface since, in extreme cases, a difference of a few feet in height above the surface means a difference of as many degrees in the temperature of the air. On still, clear days the temperature often decreases very rapidly through the first 10 to 20 feet above a flat barren surface, and increases quite as rapidly during calm clear nights. Hence it is highly desirable to have the thermometer shelter on a grass plot, since that does not heat up greatly, and on a slight elevation to avoid undue accumulation of cold air, and, finally, at a standard height above ground. In fact the shelter and its installation are quite as important in obtaining the true temperature of the air as is the accuracy of the thermometer.

Katathermometer

A good deal has been said and written about this kind of thermometer which, on exposure to the air, indoors or out, gives readings more or less indicative of the degree of comfort or discomfort a normal healthy person would experience in that air at that particular time. Essentially it is only a pair of thermometers, one of which is kept dry and the other (bulb and a portion of the stem) supplied with a drippingly (but not dripping wet jacket of muslin or other suitable material. If both are allowed to cool simultaneously from some fixed temperature, a few degrees higher than that of the body, to some other fixed temperature, a few degrees below that of the body, it will be found that their rates of cooling, and the ratio of these rates, one to the other, depend on the temperature, humidity and movement of the air, and exposure to or protection from sunshine, that is, on the very things that, so far as the weather is concerned, contribute most to our comfort or discomfort. Hence these readings can be, and they have been, coordinated roughly with our feelings.

However, this device is not much used. Perhaps no one needs an instrument to tell him how he feels; while in the process of air conditioning, whether for manufacturing requirements or physical comfort, it is much easier, and generally much better, to measure and control the actual temperature and humidity of the air supplied.

GLOBAL LIMB TEMPERATURE MEASUREMENTS

The tropopause, the boundary between the upper troposphere and lower stratosphere is critical for understanding many important processes in the atmosphere. The tropopause is defined by a sharp change in the vertical temperature gradient, taking place over a few hundred metres at most. Below the tropopause, the troposphere is a region of active vertical mixing. Above the tropopause, the stratosphere is very stable with little vertical mixing.

The match between these dissimilar regions, troposphere and stratosphere, modulates the processes that permit the exchange of mass, trace gases, momentum, potential vorticity, and energy between the two regions. Unfortunately, present observing systems do not observe the UT-LS region with sufficient detail. The NOAA operational temperature sensors are characterized by vertical resolution of the retrievals of the order of 10- 12 km.

The detailed structure of the tropopause is much too thin to be seen by operational systems. However, their cross-track scanning capability gives them the ability to observe horizontal scales of about 100 km. Temperature profiles with much higher vertical resolution can be obtained by observing the atmospheric limb, or horizon. The improvement results from the geometry, since most of the ray path through the atmosphere is within 1-2 km of the lowest, or tangent, point. In addition, the atmospheric signal is seen against the cold background of space. These factors can reduce the height of the vertical weighting functions to 3-4 km, and the effective resolution to ~ 5 km. EOS limb sounders will greatly improve the accuracy, precision, and resolution of temperature measurements in the tropopause region. HIRDLS will determine temperatures with a resolution of 1-1.5 km, through a combination of a narrow vertical field of view, low noise, and oversampling. MLS will make limb temperature measurements with a resolution of 2-3 km.

Higher Horizontal Resolution Temperature Profiles

Previous limb scanners have retrieved temperatures with higher vertical resolution, but, because the vertical scans are made at a single azimuth relative to the orbital plane, the horizontal resolution was limited to the orbital spacing or about 25°. This is sufficient to resolve only about 6 longitudinal waves. However, the UT-LS is a region in which smaller-scale waves from the troposphere are present. The horizontal resolution of previous limb sounders did not allow these smaller-scale systems to be measured.

The consistent scaling between vertical and horizontal scales suggests that higher horizontal resolution is required to sample geostrophic motions. The HIRDLS has the ability to observe and retrieve with a 1-km vertical resolution and 4° horizontal resolution by scanning from side to side.

This resolution allows all horizontal waves, up to ~ wavenumber 45, to be observed with the appropriate vertical resolution. Furthermore, the high vertical resolution of HIRDLS together with the high horizontal resolution and daily observations will for the first time provide accurate global mapping of the distribution of ozone in the lowermost stratosphere, a measurement that is crucial for determining the distribution of the flux of ozone to the troposphere, which in turn is crucial for understanding the budget of tropospheric ozone.

Recently a technological innovation has been proposed for the MLS instrument. Instead of a single heterodyne receiver, Microwave Monolithic Integrated Circuit arrays have been proposed at two frequencies. The array system would allow 100-km x 100-km horizontal resolution temperature, ozone, N2O, and water with lower power and weight. This proposed system is currently being studied by NASA.

Improvements in Chemical Measurements in the Stratosphere

EOS instruments will give significantly improved stratospheric chemical measurements by having better measurement precision, particularly in the lower stratosphere, and a more-complete suite of collocated measurements, especially chemical radicals. The major improvements in these chemical measurements are from MLS and HIRDLS. These two instruments are very synergistic in that HIRDLS will have high resolution in longitude as well as latitude while MLS will be able to make measurements in high-aerosol or cloudy regions.

In addition, high-latitude coverage will be obtained on each orbit from both HIRDLS and MLS, a significant improvement from UARS, which had monthly gaps in high-latitude coverage and did not sample important periods. The high-vertical and horizontal-resolution coverage in the upper troposphere and lower stratosphere by HIRDLS is extremely important because atmospheric waves with smaller horizontal scales can penetrate to these altitudes, creating variations on small scales that are critical to our understanding of wave breaking, mixing, stratosphere-troposphere exchange of mass constituents, and perhaps chemical processing.

Compared with UARS, EOS MLS has tremendous improvements in precision of lower stratospheric measurements due to its increased spectral bandwidth and choice of stronger spectral lines. Whereas UARS MLS was designed primarily for the middle and upper stratosphere, EOS MLS emphasizes the lower stratosphere. Inclusion of the OH measurement in MLS is a major qualitative improvement in the suite of global stratospheric measurements which EOS will provide.

This measurement is possible because of recent submillimetre-wavelength technology advances, which were not available for UARS. The OH measurement will extend to the lowest and the highest regions of the stratosphere, where HOx chemistry is thought to be the dominant ozone loss mechanism on a global scale. It will cover regions where OH is thought to control the conversion of CH4 to H_2O, to control the rates of SO_2 and OCS oxidation to sulfate aerosol, and to be an essential player in controlling the balance between radical and reservoir species in the nitrogen and chlorine families. The global OH measurements over the complete stratosphere by EOS will give unprecedented new information on stratospheric chemistry, and are especially valuable in being made simultaneously with that of the many other EOS chemistry measurements. MLS will also likely be able to measure HO_2 in the upper stratosphere, further testing and improving our understanding of stratospheric hydrogen chemistry.

HIRDLS NO_2 measurements provide an important component of NOx, and, because of its reactions with ClO, the formation of $ClONO_2$, also measured by HIRDLS. HIRDLS measurements of N_2O5 and HNO3, along with the $ClONO_2$, provide a fairly complete set of measurements of the NOy species. Note that the ratio of NO_2/HNO3 provides an additional way of deriving the OH concentration. MLS will also measure middle- and upper-stratospheric BrO, the dominant radical in the bromine chemical family. No global stratospheric bromine measurements have been made to date, and the BrO measurement will be important to test our understanding of this chemistry. Other important improvements in the suite of chemical measurements include the simultaneous and commonly-calibrated measurement of HCl and ClO by MLS. This allows very accurate monitoring of the ClO/HCl ratio, which provides a sensitive indicator of our understanding of chlorine chemistry and early detection of changes.

The MLS and HIRDLS N_2O measurements will allow much more accurate distinction between chemical and dynamical changes in the atmosphere. Total ozone measurements using TOMS will continue with a flight on a Russian Meteor satellite in 2000. Total ozone measurements will continue with an advanced Ozone Monitoring Instrument contributed to EOS Chem by the Netherlands. OMI is a hyperspectral imaging spectrometre operating in the ultraviolet and visible. Its hyperspectral capabilities will improve the precision and accuracy of the total ozone measurements by better estimates of cloud heights and aerosols effects. OMI will make additional measurements which will be highly complimentary to EOS Chem and SAGE, such as column amounts of NO_2, background SO_2, OCLO, BrO, aerosols, and derived UV-B. After EOS-Chem, total and profile ozone will monitored by the NPOESS using an advanced TOMS and a separate profiling instrument capable of high vertical resolution measurements in the lower stratosphere.

Improvements in Measurements of Aerosols

EOS instruments will improve upon the ability demonstrated by UARS for several reasons. Temperature retrievals will be more refined, and the horizontal and vertical resolution will be better. HIRDLS resolution will be on the order of 1 km in the vertical coordinate, and 4° latitude-by-4° longitude in the horizontal coordinate, and will retrieve temperatures with an accuracy of 1K at altitudes below 50 km.

The HIRDLS experiment has four spectral channels which are specifically dedicated to obtaining aerosol extinction measurements, but will obtain aerosol information from many of the other channels as well. The measurements will observe and distinguish midlatitude sulfuric acid droplets and PSCs. H_2O and HNO3, two gases which are incorporated into stratospheric aerosols, will be retrieved by HIRDLS. In addition, $ClONO_2$ and N_2O5, species important in heterogeneous processing, are retrieved by HIRDLS.

The SAGE III aerosol data will consist of vertical profiles of aerosol extinction at seven wavelengths extending in altitude range from the lower troposphere or cloud top to about 40-km altitude, with vertical resolution of 1 km. From the seven-wavelength aerosol extinction measurements, parametres describing the aerosol physical size distribution, such as volume density and surface area density, can be estimated with good accuracy, with uncertainties of the order of 10%.

The wide spectral extinction behaviour from the SAGE III measurements should also provide sufficient information to distinguish between a single-modal versus a bimodal aerosol size distribution, especially important immediately after volcanic eruptions. The SAGE III instrument will provide better tropospheric aerosol measurements with extinction measurements with multiple wavelengths through the mid-troposphere, and at least at two wave- lengths down to sea level. For the measurements of PSCs, SAGE III will be the first satellite instrument to provide size distribution information. The SAGE III aerosol information will be key to understanding the role played by stratospheric aerosols and PSCs in heterogeneous chemical reactions and ozone depletion and will be indispensable for understanding aerosol radiative forcing.

Improvements in Measurements of the Solar Ultraviolet Flux

The EOS Solar Stellar Irradiance Comparison Experiment has the unique capability of observing bright, blue stars employing the very same optics and detectors used for the solar observations. With the assumption that these stars are extremely stable, the response of SOLSTICE could be accurately monitored by observing a single star. However the calibration plan uses twenty or more of these reference stars, and it is the average flux from the entire ensemble of stars that provides an even more reliable reference for the solar observations. Repeated observation of the stars should yield an average flux level that is unchanging in time. A decrease in the level is an indication of loss in instrument sensitivity and adjustments are made accordingly. Since the same optics and detectors are used for the solar observations, the same corrections apply, and the resulting solar measurement is reliable and free of any instrumental effects. The large dynamic range between the stellar and solar signals is on the order of eight to nine orders of magnitude, and is easily accommodated by changing the entrance apertures, spectral bandpass, and integration times.

Full Meteorological and Chemical Assimilation of EOS Data Sets

The Data Assimilation Office at NASA/Goddard Space Flight Centre already provides routine support to the stratospheric aircraft missions planned to study stratospheric ozone. Daily analyses are produced for the Stratospheric Tracers of Atmospheric Transport mission, and, given the duration of this mission, this will evolve into the operational support of the EOS Morning Platform.

During the aircraft deployment periods, forecasts are produced to aid in-flight planning. These forecasts enhance the ability of the mission planners to target air of specific chemical characteristics. The current DAO meteorological analyses have a large impact on stratospheric chemistry studies. The wind fields are of sufficient quality to remove the dynamical uncertainty from trace observations. This allows a highcaliber examination of chemical processes, a benefit that has been realised for satellite, balloon, and aircraft data. There are active efforts of the DAO to improve the subtropical winds and the deep vertical motions that link the stratosphere and mesosphere. Future plans call for a straightforward extension of the application of winds from the DAO assimilation to more-general problems.

This includes stratospheric-tropospheric exchange, as well as broader issues of tropospheric chemistry. Advanced assimilation systems will directly assimilate constituent observations. First, long-lived tracers will be assimilated. Initial studies of N_2O from UARS show that assimilation can in fact provide verifiable global information from the non-global UARS coverage pattern.

An important goal of the DAO constituent effort is to improve wind estimates, especially in the tropics. Wind inversion techniques are being actively investigated. In addition, at least two university proposals have been recently submitted to attempt to assimilate aerosol observations within the DAO system. These proposals are examples of the long-term efforts to assimilate the complete chemical suite of measurements, with the goal of bringing to bear the quantitative analysis of data assimilation on the internal consistency of the chemical observations of UARS and CHEM.

FOREIGN PARTNERS AND OTHER MEASUREMENT SOURCES

Foreign partners are critical to the scientific success of the global stratospheric measurements programme. The scale of collaboration ranges from individual science teams, to collaborative instruments, to reciprocal flights of instruments, to mission planning. The timing of the EOS CHEM mission has been structured to follow the launch of the European Space Agency's Environmental Satellite I. This mission, to be launched in 2000, follows the successful NASA UARS mission. ENVISAT will make many critical trace species measurements in the time period of maximum stratospheric chlorine. This long-term data set, UARS-ENVISAT-CHEM, will be absolutely critical to our understanding of the role of trace species in controlling ozone in the stratosphere. Other international space platforms will carry a few stratospheric instruments: The French Systeme pour l'Observation de la Terre satellite will carry the Polar Ozone Aerosol Measurements II aerosol and ozone measuring system.

The Russian MIR space station will fly the Fourier Transform Spectrometre, DOPI. In addition to these space instruments, several stratospheric aircraft campaigns are planned for the next few years.

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