UNIVERSITY OF THE ARCTIC

Module 2
Northern Climates

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Key Terms and Concepts

- potential insolation
- sensible heat
- latent heat
- net short-wave radiation
- albedo
- net long-wave radiation
- net all-wave radiation
- continental climates
- maritime climates
- air mass
- humidity
- cyclonic precipitation
- orographic precipitation
- polar front
- climograph

Learning Objectives/Outcomes

This module should help you to

1. develop an understanding of the components of radiation and energy balances and the influence of these energy fluxes on seasonal variations in air temperatures.

2. distinguish maritime climates from continental climates.

3. develop an understanding of atmospheric circulation and the processes that generate precipitation in northern environments.

4. practise interpreting climographs.

5. provide concise definitions for the key words of this module.

Reading Assignments

Rouse (1993), chapter 3: "Northern Climates," in *Canada's Cold Environments*, 65–92.

Young (1989), chapter 2: "Polar Weather and Climate," in *To the Arctic: An Introduction to the Far Northern World*, 25–43.

Overview

As we saw in Module 1, northern climates are generally categorized into Subarctic, Low Arctic, and High Arctic climate zones. These climate zones are largely defined by annual variations in temperature and precipitation, soil water balance, the nature of the vegetation cover, and the nature of the underlying permafrost. Low Arctic and High Arctic climates are characterized by average summer temperatures less than 10^{o}C, small soil moisture deficits, tundra vegetation, and continuous permafrost. Subarctic climates, on the other hand, are characterized by average summer temperatures greater than 10^{o}C, moderate soil moisture deficits, boreal forest vegetation, and discontinuous or sporadic permafrost.

This module examines the influence of the flux of solar radiation and sensible and latent heat on northern climates. The nature and magnitude of the processes that affect temperature and precipitation are examined using examples from northern Canada.

Lecture

Solar (Short-Wave) Radiation

The most important factor controlling the climate of any region is *insolation*, that is, short-wave radiation from the Sun and its absorption and dispersion in the atmosphere and at the Earth's surface. As noted in Module 1, northern environments share these important characteristics:

- They experience large seasonal variations in insolation.
- Only small quantities of energy are available to perform the work associated with the transformation of materials via the rock cycle and hydrologic cycle at the Earth's surface.

These characteristics are related to the tilt of Earth's rotational axis, known formally as the *obliquity of the ecliptic*. Obliquity influences insolation in several ways:

- It affects the travel path of solar radiation through the atmosphere; hence the degree of absorption and dispersion of this radiation within the atmosphere (see fig. 2.1).
- It affects the degree of dispersion of solar radiation at the Earth's surface (see fig. 2.1).
- It affects day length (see fig. 2.2); hence the total quantity of insolation reaching the Earth's surface.

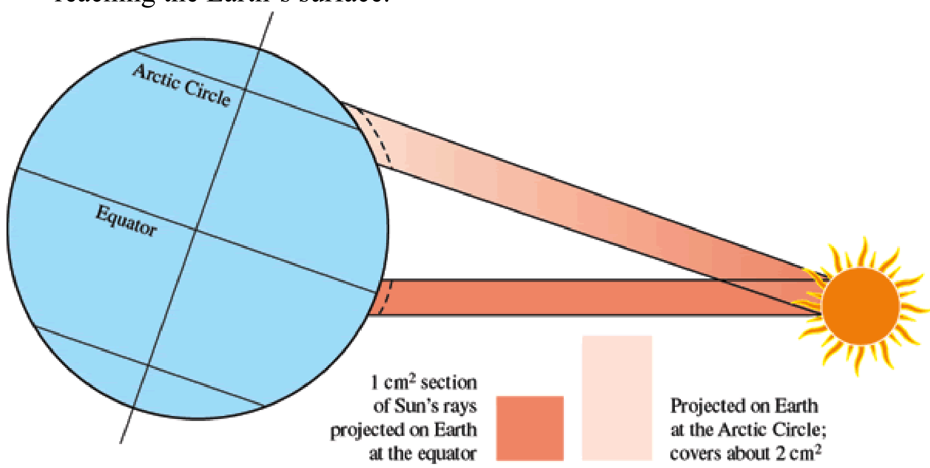

Fig. 2.1 The influence of obliquity on the travel path of solar radiation through space and the intensity of solar radiation received at the equator and at the Arctic Circle

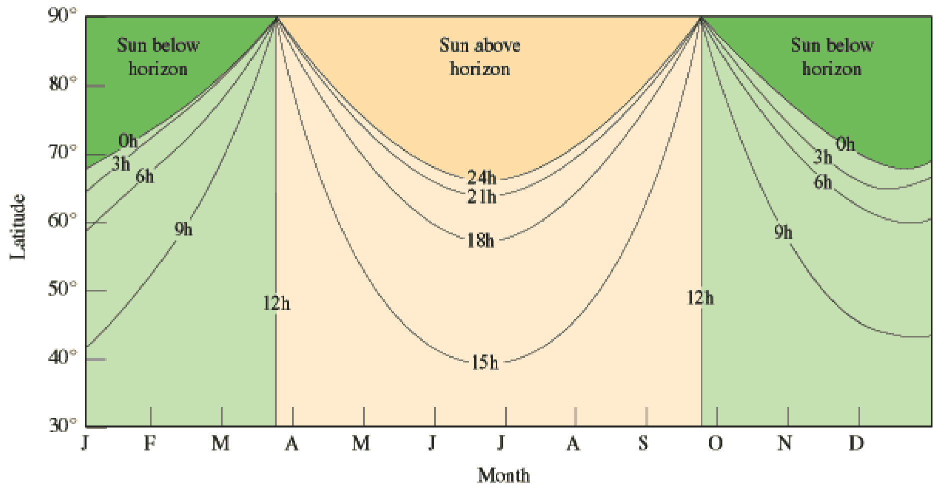

Fig. 2.2 The influence of obliquity on the seasonal variation in day length, measured in hours, at various latitudes in the northern hemisphere

The maximum receipt of solar radiation in Subarctic, Arctic, and north Polar zones occurs during the northern hemisphere summer. In contrast, the minimum receipt of solar radiation in Subarctic, Arctic, and north Polar zones occurs during the northern hemisphere winter (see Module 1).

For most of the year, high-latitude regions situated between lat 55^{o} N and lat 90^{o} N receive much less insolation than do midlatitude and low-latitude regions (see fig. 2.3). However, in the summer, high-latitude environments receive more insolation than do midlatitude and low-latitude environments (see fig. 2.3). The combination of a shorter travel path for radiation through the atmosphere, fewer and/or thinner clouds in the atmosphere, and longer day lengths contribute to increasing insolation in this region during the summer.

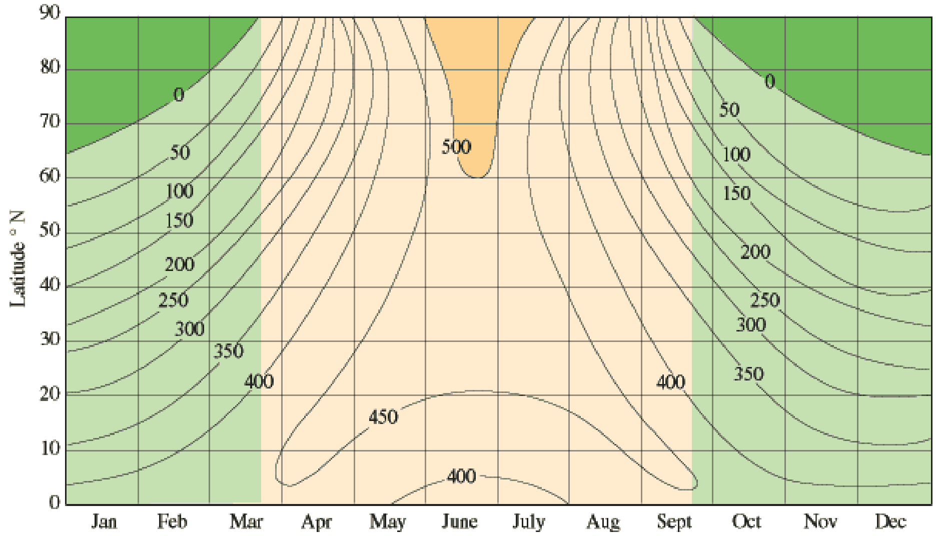

Fig. 2.3 The influence of obliquity on the seasonal variation in the receipt of solar radiation, measured in watts per square metre (W/m^2), at various latitudes in the northern hemisphere

Water droplets, ice crystals, various gases, and particulate matter present in the atmosphere diminish the quantity of insolation by absorbing and reflecting the short-wave radiation transmitted through the atmosphere. This diminution of insolation in the atmosphere is reflected in the ratio of *incoming solar (short-wave) radiation (K↓)* to *potential insolation (I_o)*, and the incoming solar radiation is expressed as a percentage of potential insolation. See table 2.1.

Table 2.1 Incoming solar radiation expressed as a percentage of potential insolation, for Canadian localities (after Rouse 1993)

High Arctic: Alert, Nunavut (82^{o}30′ N):

K↓= 90 W/m^2, I_o = 175 W/m^2, so that K↓ /I_o x 100% = 51.4%

Low Arctic: Iqaluit, Nunavut (64^{o}45′ N):

K↓= 119 W/m^2, I_o = 215 W/m^2, so that K↓ /I_o X 100% = 55.3%

Subarctic: Churchill, Manitoba (59^{o}45′ N):

K↓= 132 W/m^2, I_o = 242 W/m^2, so that K↓ /I_o x 100% = 54.5%

Midlatitude: Toronto, Ontario (43^{o}40′ N):

K↓= 146 W/m^2, I_o = 312 W/m^2, so that K↓ /I_o x 100%= 46.8%

These data illustrate several trends. First, the quantity of potential insolation and the quantity of incoming short-wave radiation decreases polewards. These trends directly reflect the influence of obliquity on insolation received at the Earth’s surface. Second, significant quantity of solar radiation is dispersed within the atmosphere via absorption and reflection; approximately 50 per cent of the potential solar radiation arriving at the outer surface of the atmosphere is dispersed prior to being transmitted as short-wave radiation from the atmosphere to the Earth's surface.

What happens to incoming solar radiation at the Earth's surface further reduces the quantity of available energy. A portion of the *incoming solar radiation (K↓)* is absorbed at the surface, and the remainder is reflected and returns to the atmosphere as *outgoing short-wave radiation (K↑)*. The quantity of radiation leaving the surface is expressed as a ratio to the quantity on incoming radiation, *K*↑:*K*↓. The percentage of the radiation reflected by the Earth is referred to as the *albedo* of the surface (see table 2.2). In general, light-coloured surfaces exhibit high albedos; hence, they reflect much of the incoming short-wave radiation back to the atmosphere. Conversely, dark-coloured surfaces exhibit low albedos; hence, they absorb much of the incoming short-wave radiation that they receive.

Table 2.2 The albedo (K↑/ K↓ x 100%) for various surface covers (after Sellers 1965)

Fresh snow:	75%–95%
Old snow:	40%–70%
Sea ice:	30%–40%
Water (60^{o} N):	21% in winter; 7% in summer
Dry sand:	35%–45%
Moist, dark soil:	10%–20%
Dry, dark soil:	20%–35%
Dry, light soil:	25%–45%
Deciduous forest:	50% in winter; 15% in summer
Coniferous forest:	67% in winter; 16% in summer
Tundra:	83% in winter; 16% in summer

Seasonal variations in the quantity of energy available at the Earth's surface also reflect variations in surface albedo. The presence or absence of snow and ice at the surface strongly influences the surface albedo. The seasonal variation in surface albedos for land surfaces covered by forest and tundra vegetation reflects the presence of snow and ice in the winter season. Snow and ice transform a dark-coloured, vegetated land surface into a light-coloured, highly reflective surface. Open water surfaces also exhibit seasonal variations in albedo. These variations in albedo are related to Sun angle, that is, the angle of the Sun above the horizon). In the winter season, the Sun angle is low and the incoming short-wave radiation approaches the water surface at an oblique angle; this results in a higher albedo. In the summer season, the Sun angle is high and the incoming short-wave radiation approaches the water surface at a greater angle; this results in a lower albedo.

With respect to short-wave radiation, the balance between incoming and outgoing radiation (K↓ – K↑) can be expressed as a quantity K*, *net short-wave radiation*. This quantity varies with latitude, decreasing in value towards the North Pole (see table 2.3).

Table 2.3 Annual net short-wave radiation for Canadian localities (after Rouse 1993)

High Arctic: Alert, Nunavut (82°30′ N):	
	K* = 46 W/m^2
Low Arctic: Iqaluit, Nunavut (64°45′ N):	
	K* = 62 W/m^2
Subarctic: Churchill, Manitoba (59°45′ N):	
	K* = 90 W/m^2
Midlatitude: Toronto, Ontario (43°40′ N):	
	K* = 122 W/m^2

Terrestrial (Long-Wave) Radiation

We must also consider the energy exchanged from the Earth’s surface to the atmosphere. Radiation emitted from the Earth’s surface is referred to as terrestrial (long-wave) radiation (L↑). The quantity of energy emitted from the surface is a function of surface temperature: the warmer the surface, the greater the quantity of radiation emitted; and the cooler the surface, the lesser the quantity of radiation emitted. A portion of the radiation emitted from the Earth’s surface is absorbed by gases in the atmosphere and is subsequently re-emitted towards the Earth’s surface (L↓).

With respect to long-wave radiation, the balance between incoming and outgoing radiation (L↓ – L↑) can be expressed as a quantity L*, *net long-wave radiation*. This quantity is invariably negative. (See table 2.4.)

Table 2.4 Annual net long-wave radiation for Canadian localities (after Rouse 1993)

High Arctic: Alert, Nunavut (82°30′ N):	
	L* = –43 W/m^2
Low Arctic: Iqaluit, Nunavut (64°45′ N):	
	L* = –50 W/m^2
Subarctic: Churchill, Manitoba (59°45′ N):	
	L* = –53 W/m^2
Midlatitude: Toronto, Ontario (43°40′ N):	
	L* = –66 W/m^2

Radiation Balance

The sum of net short-wave radiation (K*) and net long-wave radiation (L*) is known as *net all-wave radiation* (Q*) and is expressed as follows:

$$Q^* = K^* + L^*$$

$$= (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow)$$

where Q^* represents net all-wave radiation

K^* represents net short-wave radiation

$K\downarrow$ represents incoming short-wave radiation

$K\uparrow$ represents outgoing short-wave radiation

L^* represents net long-wave radiation

$L\downarrow$ represents incoming long-wave radiation

$L\uparrow$ represents outgoing long-wave radiation

Net all-wave radiation represents the quantity of energy that is available to do work at the Earth's surface. This quantity of energy varies with latitude, decreasing in value towards the North Pole (see table 2.5).

Table 2.5 Annual net all-wave radiation for Canadian localities (after Rouse 1993)

High Arctic: Alert, Nunavut (82°30′ N):	
	$Q^* = 3$ W/m^2
Low Arctic: Iqaluit, Nunavut (64°45′ N):	
	$Q^* = 12$ W/m^2
Subarctic: Churchill, Manitoba (59°45′ N):	
	$Q^* = 37$ W/m^2
Midlatitude: Toronto, Ontario (43°40′ N):	
	$Q^* = 56$ W/m^2

Energy Balance and Air Temperatures

What happens to the energy that is available at the Earth's surface? It is transformed into *sensible heat* and *latent heat*. Sensible heat contributes to heating the ground surface, heating the air in contact with the ground, and heat stored in the ground. Latent heat contributes to the melting of snow and ice, the thawing of permafrost, and the evaporation of water. The relationship between net all-wave radiation (Q*) and the work that can be performed by this energy is expressed by the *energy balance*:

$$Q^* = Q_H + Q_E + Q_S$$

where Q_H represents the sensible heat flux

Q_E represents the latent heat flux

Q_S represents the heat flux into the ground

Net all-wave radiation varies seasonally. The factors that contribute to the seasonal variation in net all-wave radiation and the various energy balance components are as follows.

Winter Season

The Sun is below the horizon for much, if not all, of the winter season. During those periods when the Sun is above the horizon, the combination of low Sun angle, short day length, and high surface albedos ensure that net all-wave radiation is low. The small quantity of available energy is used to melt or sublimate snow and ice and/or to evaporate water from open water surfaces. Little or no energy is available to warm the ground surface; therefore, cold air temperatures characterize the winter.

Spring Season

Both Sun angle and day length increase during the spring season. This combination of factors serves to increase net all-wave radiation. Much of the available energy is used initially to melt and sublimate snow and ice on the surface. Once snow and ice cover is removed, the surface albedo is reduced by the presence of dark-coloured surfaces. Increased absorption of radiation provides energy to heat the ground surface, raise air temperatures, evaporate water, and start to thaw the permafrost. Air temperatures gradually increase throughout the spring.

Summer Season

The Sun is above the horizon for much, if not all, of the summer season. The combination of high Sun angle, long day length, and low surface albedos ensure that net all-wave radiation is high during this season. A small portion of the available energy is used to remove any remaining snow and ice on the ground surface. The remainder of the available energy heats the ground surface, evaporates water, raises air temperatures, and continues thawing the permafrost, which leads to the development of the active layer of permafrost.

Fall Season

Both Sun angle and day length decrease during the fall season. This combination of factors decreases net all-wave radiation, thus reducing the available energy available to heat the ground surface, raise air temperatures, evaporate water, and sustain the thawing of permafrost. Long-wave radiative cooling of the surface commences, leading to the progressive cooling of air temperatures and refreezing of the active layer of permafrost.

A comparison of the seasonal variations in the energy balance of tundra surfaces is provided in table 2.6. Note that the pre-melt season is winter; the melt season is spring; and the snow-free season is summer.

Table 2.6 Seasonal variations of energy balance components in High Arctic and Low Arctic tundra environments in Canada (after Rouse 1993)

1. **High Arctic: upland tundra, Axel Heiberg Island**

   Pre-melt:

   Q_H = 10 W/m^2: sensible heat; raises air temperature (56%)
   Q_E = 6 W/m^2: latent heat; melts snow and ice (33%)
   Q_S = 2 W/m^2: heats ground; raises snow and soil temperatures (11%)

   Q^* = 18 W/m^2

   Melt:

   Q_H = 27 W/m^2: sensible heat; raises air temperature (32%)
   Q_E = 28 W/m^2: latent heat; melts snow and ice (34%)
   Q_S = 28 W/m^2: heats ground; raises snow and soil temperatures (34%)

   Q^* = 83 W/m^2

Snow-free:

Q_H = 31 W/m^2: sensible heat; raises air temperature (35%)

Q_E = 46 W/m^2: latent heat; evaporates water and thaws permafrost (51%)

Q_S = 13 W/m^2: heats ground; raises soil temperatures (14%)

Q* = 90 W/m^2

2. **Low Arctic: upland tundra, Churchill, Manitoba**

Melt:

Q_H = 8 W/m^2: sensible heat; raises air temperature (30%)

Q_E = 9 W/m^2: latent heat; melts snow and ice (33%)

Q_S = 10 W/m^2: heats ground; raises snow and soil temperatures (37%)

Q* = 27 W/m^2

Snow-free:

Q_H = 49 W/m^2: sensible heat; raises air temperature (41%)

Q_E = 49 W/m^2: latent heat; evaporates water and thaws permafrost (41%)

Q_S = 20 W/m^2: heats ground; raises soil temperatures (18%)

Q* = 118 W/m^2

The data presented in table 6 reveal several interesting patterns. First, net all-wave radiation is greater in the snow-free period than in the melt period at both sites, reflecting the transition to high Sun angles, longer day length, and lower surface albedos during the summer season. Second, the partitioning of net all-wave radiation into sensible and latent heat fluxes varies with the seasons. In the pre-melt (i.e., winter) season, Q* is low and the bulk of this energy is used in heating the air (Q_H) above the snow-covered ground surface. In the melt, or spring, season, Q* has increased and this energy is used largely to raise the temperature of the surface snow pack (Q_S) and initiate melting of snow and ice (Q_E). Surplus energy is available to heat the air above the ground surface (Q_H). In the snow-free (i.e., summer) season, Q* is at its highest. This energy is used to warm the air temperature (Q_H), to thaw permafrost and evaporate water from the surface (Q_E), and to warm the ground (Q_S). Together, the latter two processes account for the bulk of the energy expenditure at the ground surface.

Maritime and Continental Climates

Annual variations in air temperature in response to insolation differ over ocean and land surfaces. The difference in the thermal response of continental and ocean surfaces can be attributed to these factors: (1) differences in the specific heat of water, soil, and rock; and (2) differences in heat transport.

Specific heat refers to the quantity of energy required to change the temperature of one gram of a substance by one degree Celsius. When land and ocean surfaces are exposed to equal quantities of solar radiation, soil and rock, which have low specific heats, will warm up more than water, which has a high specific heat. In addition, solar radiation can penetrate to depths in water ranging from approximately 10 cm (for red light) to more than 100 m (for violet-blue light), but only to depths less than 20 mm in soil (Sellers 1965). Ocean currents serve to circulate heat through large volumes of water, resulting in slow heating of the ocean surface. In contrast, heat is conducted slowly through soil resulting in rapid heating of the ground surface.

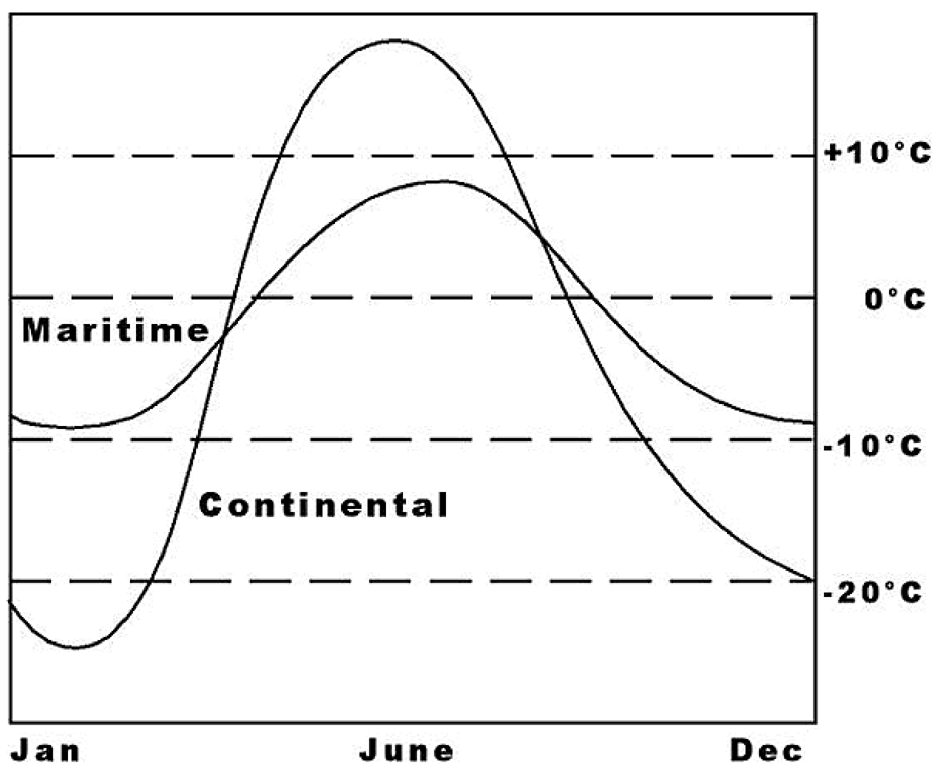

Fig. 2.4 Comparison of annual variations in surface air temperatures for regions experiencing maritime and continental climates. A small annual amplitude in air temperatures characterizes maritime climates. In contrast, a large annual amplitude in air temperatures characterizes continental climates.

The contrast in the thermal response of land and ocean surfaces is reflected in the variations in air temperatures over these surfaces. (See fig. 2.4.) *Continental climates* exhibit a large annual range in air temperatures, reflecting the fact that in the absence of a cover of snow and ice, land surfaces heat and cool rapidly. Cool summers and intensely cold winters are characteristic of continental climates in northern Canada. *Maritime climates*, on the other hand, exhibit a much smaller annual range in air temperatures. The cool summers and mild winters that are characteristic of maritime-influenced climates (e.g., Cfb, Cfc) disappear in northern Canada. The persistence of snow and ice on land and sea ice on the ocean diminishes the moderating influence of the maritime environment. High albedos associated with persistent covers of snow and ice result in a decrease in net all-wave radiation over these surfaces, and the energy available at the surface is consumed largely by melting snow and ice rather than heating the land and water. The result is that coastal communities such as Churchill, Manitoba, and Kuujjuarapik, Quebec, exhibit continental climates. (See fig. 2.8.)

Atmospheric Circulation and Precipitation

Annual variations of insolation across the surface of the Earth create regions of energy surplus (generally between lat 0^{o} and 30^{o} N) and energy deficits (generally between lat 30^{o} and 90^{o} N). (See table 2.7.) The radiation deficit at midlatitudes and high latitudes represents the quantity of energy in the form of sensible and latent heat that must be transported polewards in order to maintain a global energy balance. This transfer of energy from the tropics to the poles is performed by atmospheric and oceanic circulation.

Table 2.7 Annual radiation balance of the Earth-atmosphere system in the northern hemisphere (after Rouse 1993)

Latitude (^{o} N)	**W/m^2**
0–10	44
10–20	41
20–30	8
30–40	–16
40–50	–38
50–60	–66
60–70	–95
70–80	–116
80–90	–132

The imbalance in the distribution of energy creates temperature gradients at the Earth's surface and in the atmosphere above the surface. The temperature gradient in the atmosphere generates pressure gradients in the atmosphere:

- A region of low atmospheric pressure develops in the lower troposphere over the region of energy surplus; a corresponding region of high atmospheric pressure develops in the upper troposphere over this region.
- A region of high atmospheric pressure develops in the lower troposphere over the region of energy deficit; a corresponding region of low atmospheric pressure develops in the upper troposphere over this region.

Atmospheric circulation occurs in response to these pressure gradients: air flows from a region of high atmospheric pressure to a region of low atmospheric pressure. In the winter, regions of high atmospheric pressure develop over Siberia and the Canadian Arctic Archipelago, while regions of low atmospheric pressure develop over the North Atlantic and North Pacific oceans. Atmospheric circulation is meridional in character in this season: the flow of air tends to be along a north-south atmospheric pressure gradient.

In contrast, during summer, a region of high atmospheric pressure develops over the central Arctic Ocean, while regions of low atmospheric pressure develop over the Labrador Sea and Gulf of Alaska. Atmospheric circulation is zonal in character in this season: the flow of air tends to be along a west-east atmospheric pressure gradient.

The surface atmospheric pressure patterns reflect the thermal influence of the cold, sea ice–covered Arctic Ocean in summer and the cold land surfaces in winter.

In general, warm, moist air originating in Equatorial and Subtropical regions flows northwards; and cold, dry air originating in the Polar and Subarctic regions flows southwards. The exchange of sensible heat and latent heat between the Earth's surface and the atmosphere influences the temperature and humidity of air masses in the atmosphere.

Three principal air masses influence the nature of precipitation in northern environments: the *maritime polar* (mP), the *continental polar* (cP), and the *continental arctic* (cA). The physical properties of these air masses are summarized in table 2.8, and the source regions for these air masses are illustrated in figure 2.5.

Table 2.8 Average physical properties of air masses (after Henderson-Sellers and Robinson 1986, 247, table 5.2)

Air mass	**Physical Properties**	**Source Region**	**Average Temperature (^{o}C)**	**Specific Humidity (g kg–1)**
Maritime Polar (mP)	cool, moist	Oceans; polewards of lat 50º N	4	4.4
Continental Polar (cP)	cold, dry	Continents in the vicinity of the Arctic Circle	–10	1.4
Continental Arctic (cA)	very cold, dry	Polar regions	–40	0.1

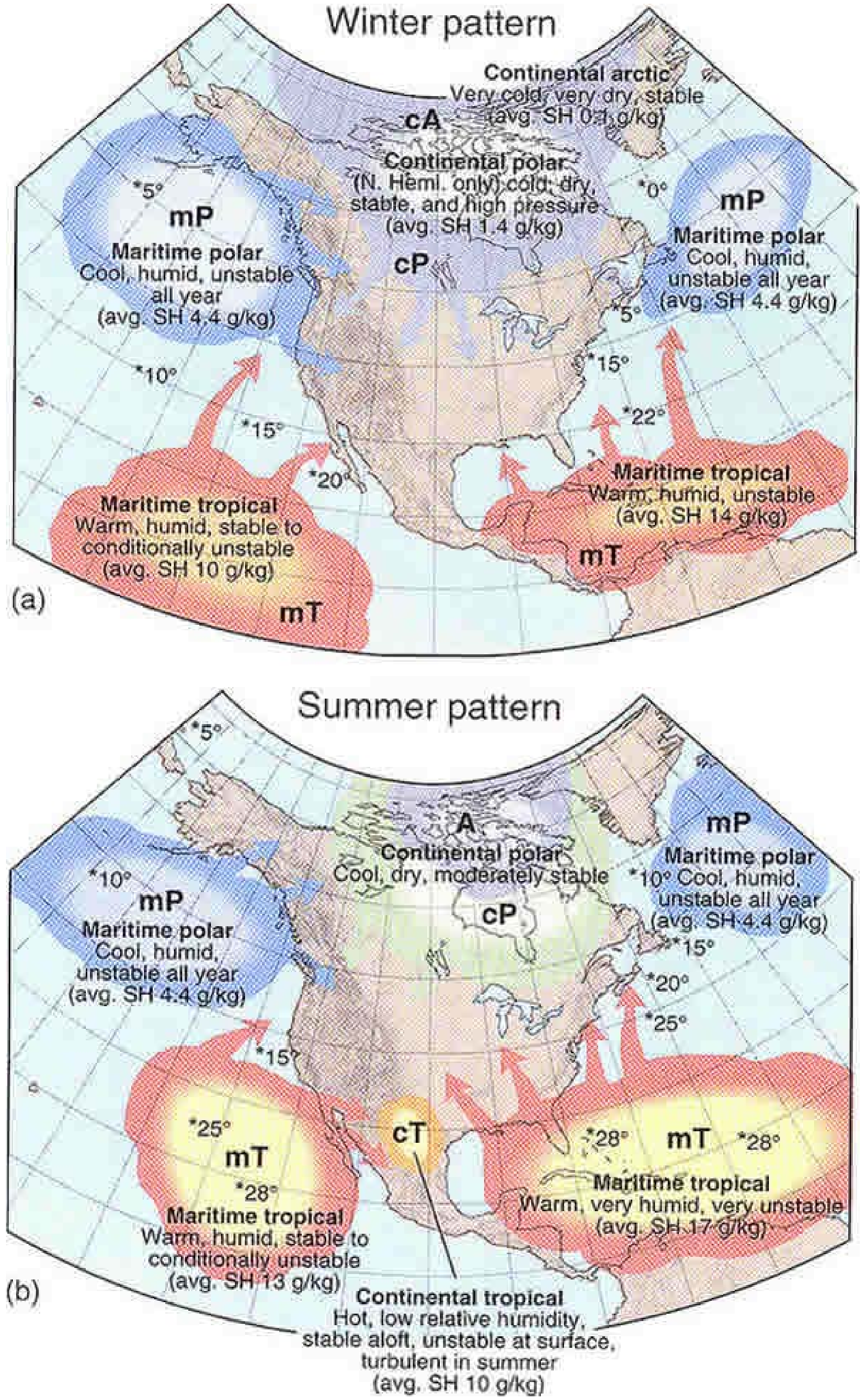

Fig. 2.5 Source regions for the maritime polar (mP), continental polar (cP), and continental arctic (cA) air masses in North America.

Atmospheric circulation moves these air masses out of their source regions so that they interact with one another. The air masses interact across a surface within the atmosphere known as the *polar front*. The position of the polar front varies with the seasons, moving towards the equator in the winter and towards the pole in the summer (see fig. 2.6). Cold air, being denser than warm air, tends to move along the ground surface, displacing warm air upwards and away from the surface along the polar front. This upward motion of warm, moist air is accompanied by cooling, resulting in the condensation of the moisture within the warm air mass. When the rising warm air becomes saturated with water, precipitation is initiated. Precipitation developed in this manner is referred to as *cyclonic precipitation*.

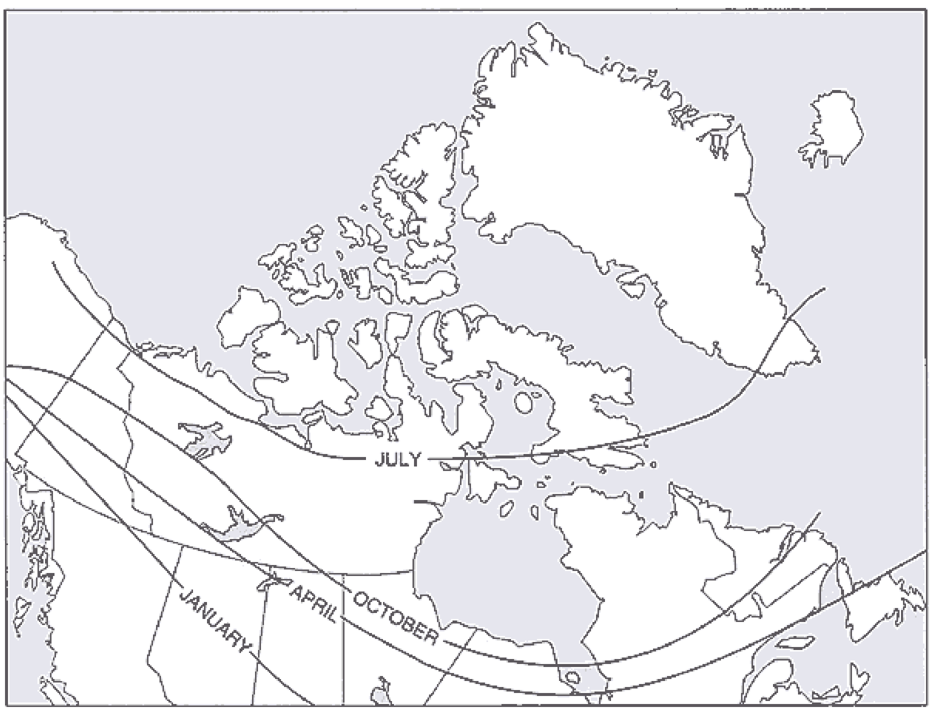

Fig. 2.6 Seasonal positions of the polar front across North America

In the High Arctic, the cA and cP air masses dominate throughout the year; the result is that this region receives less than 100 mm of precipitation annually. In the Low Arctic, the cA and cP air masses dominate in the winter, while the mP air masses become more important in the summer. Cyclonic activity in this region is less frequent than in the Subarctic and is largely confined to the summer season; hence, winters tend to be relatively dry and the bulk of the annual precipitation is received in the summer months. Mean annual precipitation varies from 100 to 300 mm in this region. In the Subarctic, air masses of diverse origin influence the climate; mP air masses influence the climate in this region throughout the year. The greater frequency of cyclonic activity in this region contributes to an increase in annual precipitation. Mean annual precipitation exceeds 300 mm throughout most of this region. See figure 2.7.

Fig. 2.7 Variations in mean annual precipitation within Canada.

The greatest annual precipitation in northern Canada is received in regions where mountain areas lie adjacent to seasonally ice-free water surfaces. The flow of moist mP air masses over these significant topographic barriers generates *orographic precipitation*. As is the case for cyclonic precipitation, the upward motion of moist air over a mountain range is accompanied by cooling, resulting in condensation of the moisture within the air mass. When the rising air becomes saturated with water, precipitation is initiated. The combination of high annual precipitation and high altitudes facilitates the development of alpine glaciers in these coastal regions.

Regional Climate Variations in Northern Canada

For the purposes of this course, you should be familiar with the regional climates of the Subarctic, Arctic, and Polar zones as defined in Module 1. These regional climates are referred to as Boreal Forest climates (Dfc, Dfd), Tundra climates (ET), and Ice Sheet climates (EF). The distribution of these climate zones is illustrated in figure 2.8. A graphic tool referred to as a *climograph* is used to simultaneously display variations in monthly air temperatures and precipitation at a station. Climographs for various Canadian localities can be viewed by using your computer mouse to click on a highlighted place name on figure 2.8.

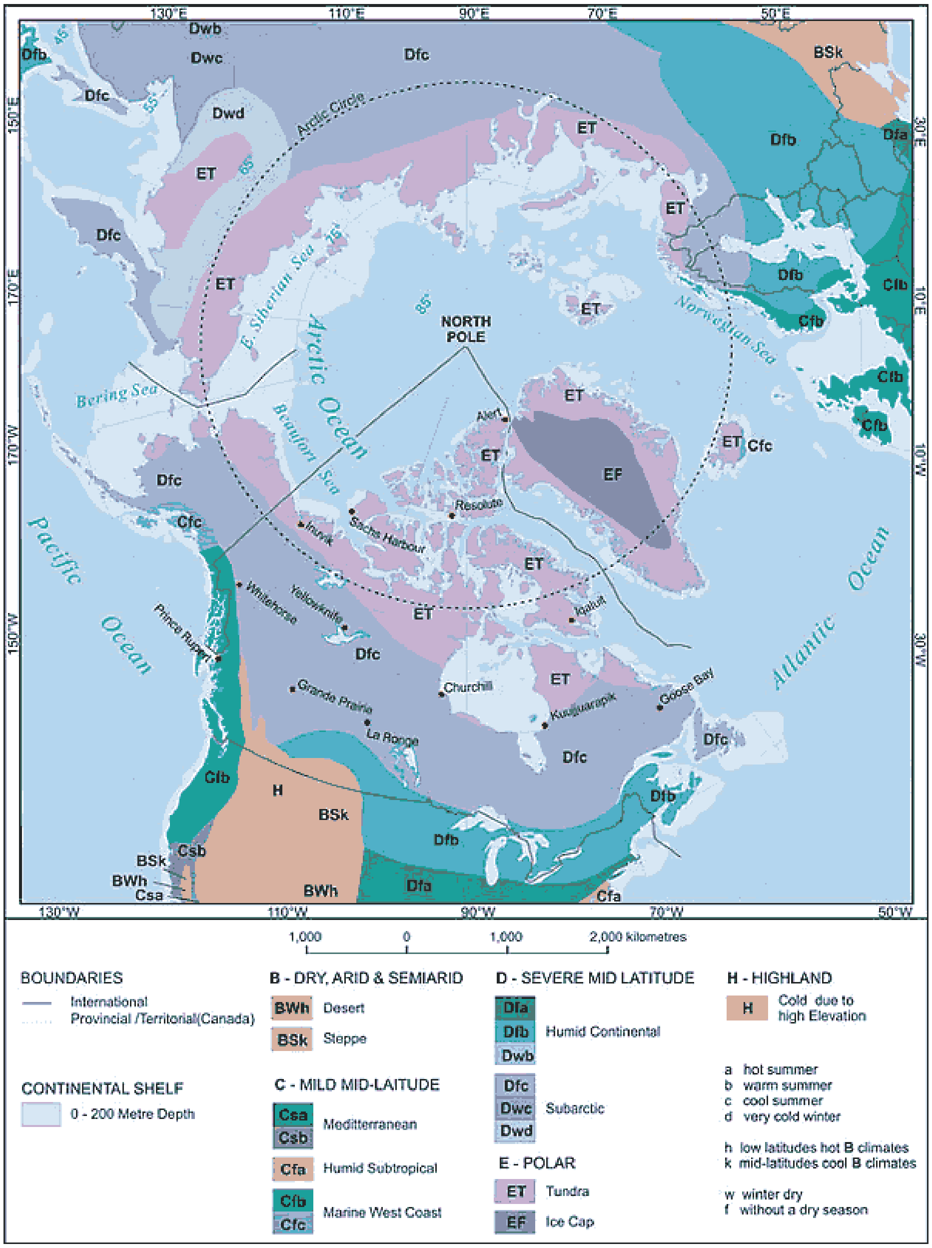

Fig. 2.8 Regional climates in the circumpolar North as defined by the Köppen-Geiger climate classification system.

Student Activity

View the climographs that appear in figure 2.8 and examine the regional variations in rainfall and snowfall in northern Canada.

Summary

The quantity of solar radiation received at the Earth's surface is the single most important factor affecting the climate of a region. Northern environments experience large seasonal variations in insolation, and only small quantities of energy are available to perform the work associated with the transformation of materials via the rock cycle (e.g., the mechanical weathering of rocks) and the hydrologic cycle (e.g., the thawing of snow) at the Earth's surface. The persistence of snow and ice on land and sea ice on the ocean affects both the radiation and energy balances of northern landscapes. High albedos associated with persistent covers of snow and ice result in a decrease in net all-wave radiation. Melting snow and ice rather than heating the land and water largely consumes the energy available at the surface. This situation produces the cold winters and cool summers that characterize northern environments.

Precipitation in northern environments is generated by the interaction between cold, dry air masses (cP, cA) and cool, moist air masses (mP) along the polar front in the atmosphere. Arctic regions are dominated by cold, dry air masses throughout the year; mean annual precipitation varies from 100 to 300 mm, most of which falls as snow in the fall and winter. Cool, moist air masses occur with greater frequency in Subarctic regions; mean annual precipitation exceeds 300 mm, much of which falls as rain in the summer.

The greatest annual precipitation (greater than 400 mm annually) in northern Canada is received in regions where mountain areas lie adjacent to seasonally ice-free water surfaces. Orographic lifting of moist air masses over mountain ranges contributes to the abundant precipitation in these regions.

Study Questions

A. Multiple Choice Questions

1. __________ precipitation develops from the interaction of the _____ and _____ air masses along the polar front.

   a. Orographic; cP; mP

   b. Cyclonic; cP; mP

   c. Cyclonic; cA; mP

   d. Orographic; cA; mP

2. The transition from winter to summer at high latitudes is accompanied by ____________ of the albedo of land surfaces and __________ of the albedo of water surfaces.
   a. an increase; a decrease
   b. an increase; an increase
   c. a decrease; a decrease
   d. a decrease; an increase

3. The albedo of the ground surface relates to ________________________.
   a. the flux of short-wave radiation between the atmosphere and the surface
   b. the flux of latent heat between the atmosphere and the surface
   c. the flux of sensible heat between the atmosphere and the surface
   d. the flux of water vapour between the atmosphere and the surface

4. __________ heat influences the flux of __________ from the Earth's surface and the __________ of air masses.
   a. Sensible; water vapour; temperature
   b. Sensible; short-wave radiation; temperature
   c. Latent; water vapour; humidity
   d. Latent; long-wave radiation; humidity

5. Net all-wave radiation increases in response to _____________________.
   a. a decrease of Sun angle and day length
   b. a decrease in Sun angle and an increase in day length
   c. an increase in Sun angle and a decrease in day length
   d. an increase in Sun angle and day length

Answers to Multiple Choice Questions

1. b
2. c
3. a
4. c
5. d

B. Essay Question

With reference to the radiation and energy balances, discuss the processes that contribute to the monthly variations in air temperature and precipitation illustrated in climographs for Alert, Nunavut; Iqaluit, Nunavut; and Churchill, Manitoba.

Glossary of Terms

air mass	a large volume of air with unique properties of temperature and specific humidity that extend horizontally over thousands of kilometres.
Arctic Archipelago	the group of islands to the north of mainland Canada, including Baffin, Ellesmere, and Victoria Islands.
High Arctic	the part of the Canadian Arctic that lies within the Arctic Circle.
Low Arctic	the part of the Canadian Arctic south of the Arctic Circle.
meridional	of or relating to a meridian.
polar front	a surface within the atmosphere across which air masses interact.
Subarctic	*noun* the region immediately south of the Arctic Circle. *adjective* characteristic of, pertaining to, or inhabiting this region.
troposphere	the lowest region of the atmosphere, extending to a height of between 8 and 18 km and marked by convection and a general decrease of temperature with height.
zonal	of or relating to zones.

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