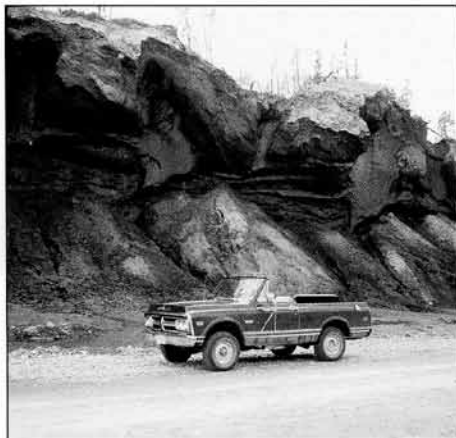


# PERMAFROST

## A Guide to Frozen Ground in Transition



by Neil Davis

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## When the Ground Freezes

### ■ Introduction

A person need not live where there is permafrost, but only where the temperature falls below freezing at least a few nights each year, to observe first-hand some of the intriguing and important phenomena that can make permafrost such a treacherous material. The key here is the movement of water molecules through the ground during and after freezing. When water freezes, its volume expands by 9%, but when water-saturated soil freezes, its expansion might exceed 50%, 100%, or even more. Such radical change could not happen if the soil were isolated (say, if placed in a steel or plastic bucket). However, extreme expansion can and does take place in natural settings where fine-grained soil is in contact with underlying material that provides a water supply. The water migrates upward into the freezing soil and turns to ice, thereby increasing the volume of the soil, the phenomenon known as frost heave.

### ■ Pipkrakes, the Crunchy Product of Ground Freezing

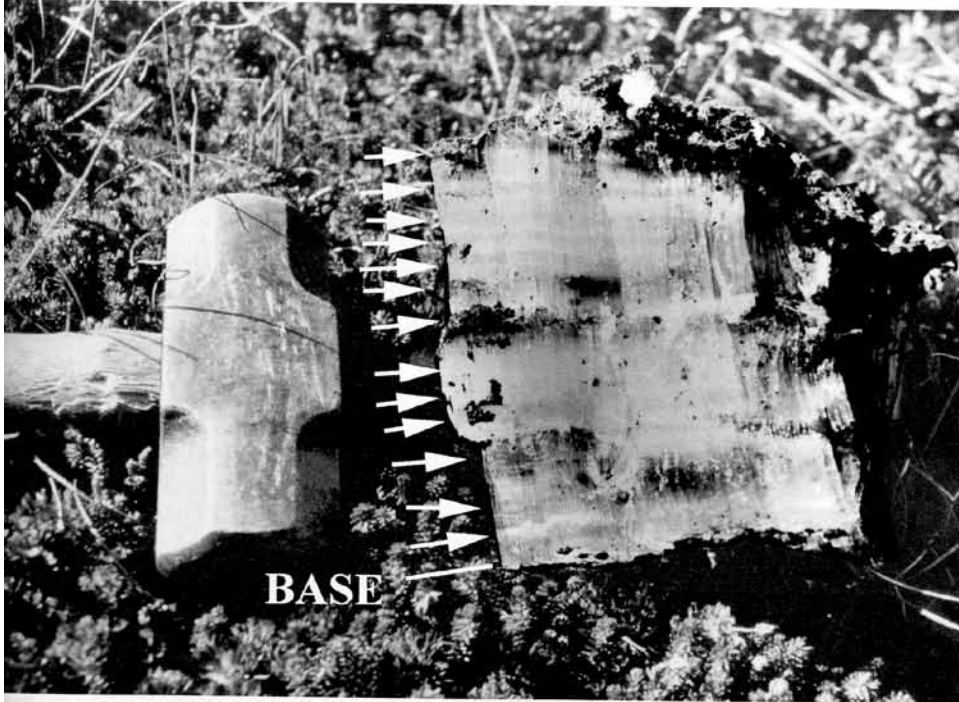
A vivid and easily seen example of the consequences of water's ability to migrate to where the soil is freezing is the formation of *pipkrakes*. At one time or another most everyone has walked out on a frosty morn and experienced the pleasure of padding through the pipkrakes. These little pillars of ice crunch nicely beneath the feet. Also called *needle ice* or mushfrost,

as well as Kammeise in German, and shimobashira in Japanese, pipkrakes are vertically elongated crystals of ice that grow just beneath the ground surface. One cold night ( $-2^{\circ}\text{C}$  or lower) might produce a layer of pipkrakes one or more centimeters thick. Several cold nights in succession interlaced by daytime warming will produce multiple layers of pipkrakes, creating an array 10 cm or more in thickness. A prolonged spell of temperature below  $-2^{\circ}\text{C}$  can cause the individual needle-like piprake crystals to grow to lengths exceeding 40 cm (15 in).<sup>1</sup> **Figure 3.1A** contains examples found at sea level in Washington state, an area better known for its apples than its frozen ground, but pipkrakes grow there, too. **Figures 3.1B** and **3.1C** show Alaska pipkrakes.



**Figure 3.1A** Pipkrakes formed during several cold nights at sea level on San Juan Island, Washington. The arrows mark the lower edge of the piprake layers. The two photos at bottom show pipkrakes removed from the location shown in the photograph at upper right. See also Plate 3.

1. Schmid (1955), cited by Washburn (1980); Tricart (1970).



**Figure 3.1B** Pipkrakes photographed by Richard Reger in Alaska. Notice the needle-like character of those in the upper photo and the more massive character of those below. Arrows mark the nightly layers formed.



**Figure 3.1C** Is this the granddaddy of all pipkrakes? Billy Conner, a researcher with the Alaska Department of Public Facilities and Transportation is shown with a substantial specimen. Perhaps formed by another mechanism, this large ice mass protruding approximately 1 meter above its surroundings was photographed by Richard Veazey near the trans-Alaska pipeline on Alaska's North Slope. See also Plate 4.

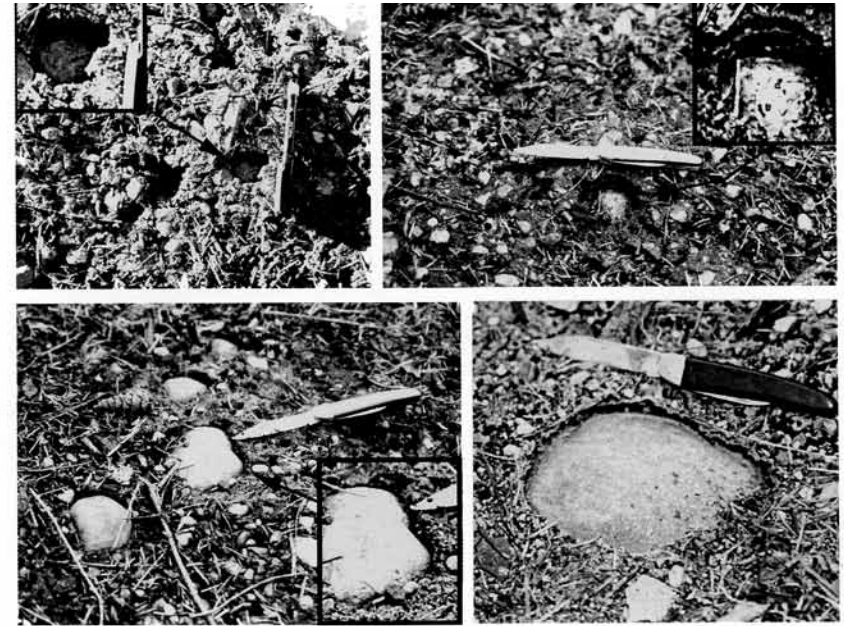
Pipkrakes grow from the bottom up, initially starting out one to several centimeters below the ground surface. The growth then lifts that surface layer upward. Three primary factors intertwine to determine the growth of piprake crystals: the temperature regime, the soil characteristics, and the water supply. For the crystals to continue their growth the temperature of the air in contact with the ground must be low enough to carry away the energy released by the forming ice (the latent heat of fusion, 80 cal/gm). Plenty of water must be available in the soil below where the pipkrakes are forming, and the soil there needs to have characteristics that promote the movement of the water up to the base of the piprake layer. Minor differences in the amount of organic matter on the soil surface can have major effects by altering the thermal conductivity of the material, so piprake fields may contain many irregular pillars of ice

that lift some parts of the ground surface well above others. The ground then becomes karst-like on a microscale.

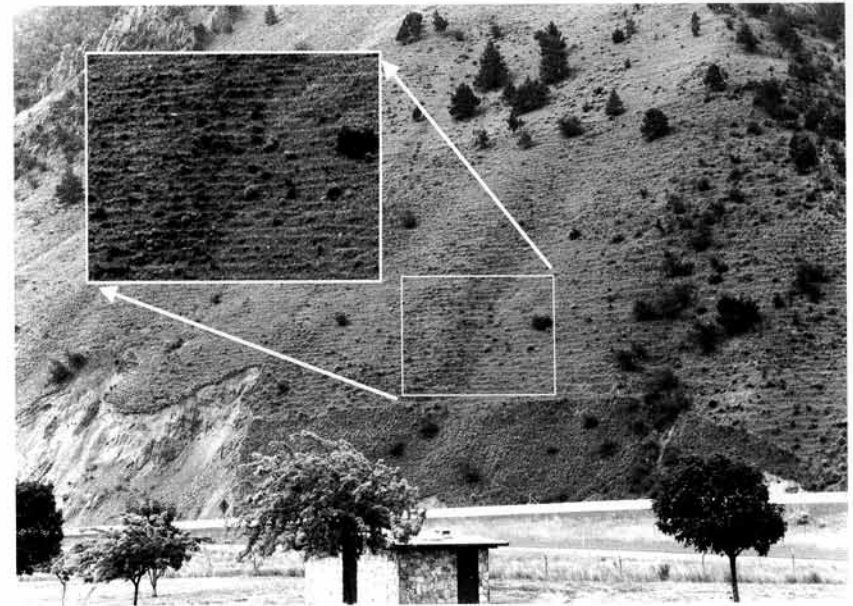
Pipkrakes tend to grow in loose soil, and they tend to loosen the soil, so their presence is a self-perpetuating phenomenon that can have major effect on the soil surface. One consequence in areas of high wind is that the formation of pipkrakes, by lifting up and loosening fine-grained soil, can lead to the phenomenon known as soil *deflation*, the removal of the smaller grains (clay, silt, and sand) by wind action. The effect is sufficiently deleterious on some agricultural lands that farmers sometimes try to compact pipkrake-prone soils.

Of general geomorphological significance is that pipkrake formation can contribute to *cryoturbation*, a churning of soil that can cause sorting by particle size and differential motions, especially on sloping ground. The pipkrake needles grow perpendicular to the ground surface, so if the surface slopes, the growth and subsequent thaw of the needles will tend to transport any lifted material downslope. Also, the needles often develop curvature under the load of surface material carried, so even on level ground material lifted by the needles may fall back to a new position. On sloping ground, pipkrakes sometimes cause displacements of several to several tens of centimeters during the course of a year.<sup>2</sup>

Pipkrakes can help sort near-surface material in two ways. As seen in **Figure 3.2**, the formation of pipkrakes tends to lift the small-sized soil material up away from pebbles or rocks. Also, if rocks lie atop the ground the formation of pipkrakes can lift the rocks above the surrounding surface. Canadian geocryologists have reported the lifting of stones weighing as much as 10 kg.<sup>3</sup> Measurements of downslope movement caused by pipkrake formation show a variety of results, but in most instances the measurements indicate that the needle ice growth moves fine-grained material much faster than coarser material and therefore helps to sort it. Geocryologists think this action contributes to the formation of a variety of *patterned ground* features, among them stripes which extend up and down gradual alpine slopes and *cryoturbation steps* which form staircase-like on steeper slopes (25° to 38°) in relatively dry midlatitude areas that experience freeze and thaw.<sup>4</sup> The cryoturbation steps shown in **Figures 3.3A** and **3.3B** are in northeastern Oregon. Such steps are easy to observe in many



**Figure 3.2** Soil uplifted by pipkrakes away from pebbles in a wet driveway on San Juan Island, Washington. See also Plate 5.

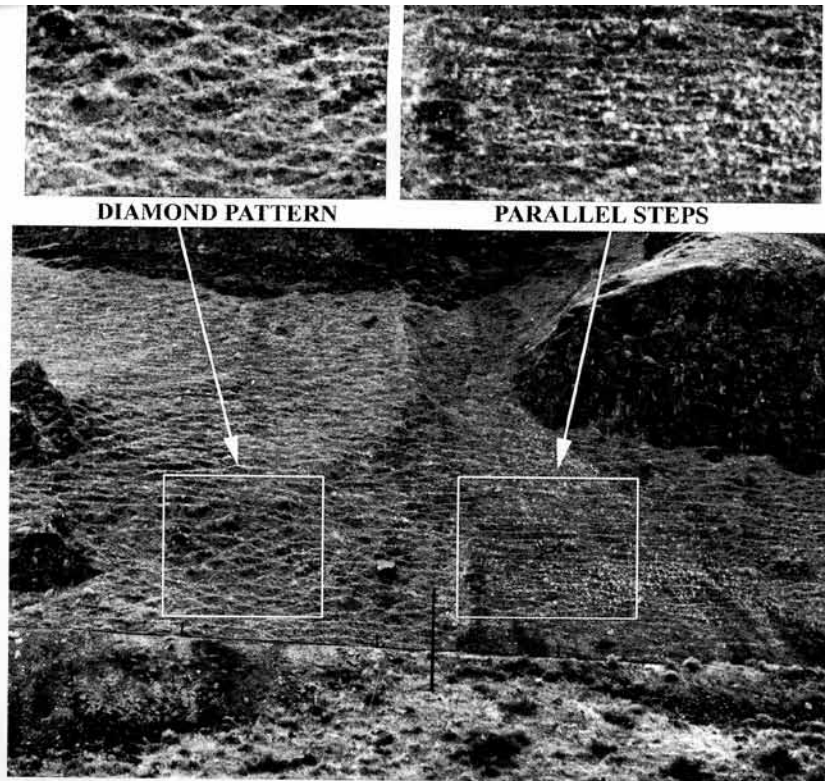


**Figure 3.3A** Cryoturbation steps (terraces) on a 35-degree north slope overlooking the Weatherby Rest Stop at Mile 335 on Interstate 84 in northeastern Oregon. The estimated spacing of these parallel steps is 0.5 to 1 m. Photographed in summer during the late 1990s.

2. MacKay and Mathews (1974a;1974b); Lewkowicz (1988); Washburn (1980).

3. MacKay and Mathews (1974a).

4. Sharp (1942); Washburn (1980); Lewkowicz (1988).



**Figure 3.3B** At left, slightly sloping cryoturbation steps crossing over each other to form a diamond pattern, and, at right, a pattern of parallel horizontal steps. The parallel pattern is far more common than the diamond pattern. Limited observations suggest that, when seen, the diamond pattern typically appears near the bottoms of slopes that exhibit the horizontal steps higher up. Photographed March 1999 looking south from Interstate 84 in northeastern Oregon.

lightly vegetated hilly regions of southern Canada and the northern United States.<sup>5</sup> Generally the steps lie parallel to the slope but some slopes display a diamond pattern of inclined steps crossing over each other, as in Figure 3.3B.

5. History buffs take note that on June 25, 1876, General George Custer led his troops through intermittent arrays of cryoturbation steps as he approached the location of his last stand on the hill above the Little Bighorn River in southern Montana. Surrounded by more pressing matters, it is doubtful if the doomed general or any of his men noticed the steps. The upper slopes of Custer's hill are too gentle to support the development of the steps, and the same is true of the nearby hill where Major Marcus Reno's command managed to hold off the Sioux Indians. However, between those hilltops stands the third important prominence of the battle, Weir Point, with one upper slope steep enough to display a few poorly developed cryoturbation steps. Other more well-developed steps appear on the steepest slopes in the battlefield and also north of it on the bluffs facing eastward onto the Little Bighorn.

Pipkrakes grow best in loose soil containing a substantial proportion (greater than 20%) fine-grained particles, but not too many clayey fines because they can inhibit the movement of water up through the soil. This movement of water is crucial to the growth of the piprake crystals, and also to the growth of other forms of annual and perennial ground ice found in cold regions. So an important factor in what happens when the ground freezes is the nature of the soil, especially the size of the soil particles involved.

### ■ Soil, the Mineral and Organic Matrix

The word *soil*, like so many others, has multiple meanings. In this book I use "soil" in a broad sense to mean the uppermost layer of material at the earth's surface that is perhaps frozen or wherein freeze and thaw processes are taking place. In a narrow sense, "soil" is only the topmost part of this layer, the part that supports plant life. The nature of a soil at any location depends much on climate, its original source material and length of time it has been exposed to weathering processes. A typical soil is a complex of irregularly fragmented particles, both crystalline and amorphous, which are arranged in intricate geometric patterns of uncounted variety. Near-surface soil is always in a perpetual state of change; it swells and shrinks as it wets and dries or freezes and thaws, and it interacts mechanically, chemically, and electrically and with whatever materials fill its pores, be they gases, liquids, or transient solids.

With regard to what happens when the ground freezes and thaws, the single most important characteristic of a soil is the size of its particles. Soil particle size fixes both the size of the pores in the soil and the surface area at the interfaces between the soil particles and the liquids and gases in the pores. If the soil particles are large, the pore spaces are large and the interfacial surface area is small, whereas if the particles in the soil are small, the pore spaces are small and the surface area is large.

*Soil texture* is the term used to describe the size range of particles in soil. Traditionally, soil scientists divide particles into size ranges called separates or fractions. Going from the smallest to the largest grain sizes, the fraction diameters are: clay, < 0.002 mm; silt, 0.002 to 0.05 mm; sand, 0.05 to 1 mm; very coarse sand, 1 mm to 2 mm; and gravel, 2 mm to 5 cm.<sup>6</sup> Upwards of that are pebbles < cobbles < boulders, with no sharp

6. Several different schemes exist, but they all are very similar [Hillel (1980) p 57].

**Table 3.1** Soil Texture by Grain Size<sup>7</sup>

Texture Name	Diameter Limits
Boulders > Cobbles > Pebbles	Larger than 2 cm
Gravel	2 cm – 5 mm
Very Coarse Sand	1 mm – 2 mm
Sand	0.05 mm – 1 mm
Silt	0.002 mm – 0.05 mm
Clay	Less than 0.002 mm

distinction between them. See **Table 3.1**. Soils are often described by their predominant fraction; i.e. sand or silt. If no one texture size predominates, the soil is called **loam** if it contains more than 35% clay and less than 27% silt. If the proportion of silt to clay is reversed, the soil is called **silt loam**. Loess, a soil type common in interior Alaska and other periglacial areas, is wind-transported silt and sand (plus minor clay) material mostly in the size range 0.005 to 0.5 mm, with particle size usually decreasing with distance from the original source location. Brown, tan, or gray, loess's color reflects its depositional and postdepositional conditions and processes. Gray loess contains more organic material or ferrous iron than tan loess.

The size distribution and the shapes of the particles in a body of soil are important in determining the soil's behavior, in part because these characteristics affect the packing, i.e., the geometrical relationships between particles. The packing within a box containing marbles of uniform size is different than that if the marbles are of various sizes, and certainly marbles, coins, and wooden matchsticks each pack differently. More important yet is the fact that particle shape and size distribution affect surface area, and the surface area of the particles in a soil profoundly influences how strongly the soil will adsorb water or other substances. Spherical soil particles have the lowest **specific surface** (surface area per unit weight or per unit volume), and platelike particles have the highest. Sand, with its generally spherical particles, typically has specific surface less than 1 m<sup>2</sup>/gm, whereas the specific surface of clay may exceed 100 m<sup>2</sup>/gm because of the smaller particle size and the platy shape.<sup>8</sup>

7. Based primarily on a table by Wimer (1965).

8. Montmorillonite clay has specific surface 800 m<sup>2</sup>/gm because its platelets are extremely thin [Hillel (1980) pp 65–67].

Yet other factors come into play to affect the cohesiveness of a soil. A soil might contain only crystalline particles that are completely unattached to each other so that the soil is very loose—for example, a dry desert sand—or it might contain materials that tend to cement the crystalline particles together. Amorphous substances such as humus matter or iron and aluminum oxides and carbonates can act as glues to make the soil cohesive. Shape and history come in here as well, as is evident to a stone mason. He much prefers flat rocks to round rocks when building walls, and even with the flat rocks he knows that flat rocks dumped on the ground result merely in a pile of flat rocks, but if stacked carefully they will form a strong structure, especially if set with cement in the joints.

Adding further to the intricacy of processes that take place in a soil and affect its characteristics is the electrical nature of soil particles, particularly those with irregular shapes. Clay particles, thin flat crystalline entities, typically carry an irregular distribution of charge, and additionally, an excess of negative charge. Both the irregular distribution and the excess of negative charge cause the clay particles to interact electrically with other particles and any substances within a soil's pore spaces. An explanation for the typically negative excess charge on a clay particle is the occasional appearance in the platy crystal structure of a triply charged aluminum ion (Al<sup>3+</sup>) where a quadruply charged silicon ion (Si<sup>4+</sup>) normally should reside, and sometimes doubly charged magnesium (Mg<sup>2+</sup>) ions substitute for the triply charged aluminum ions in the crystal structure. Each substitution creates a dearth of positive charge, with the overall result that the crystal lattice has too many negative charges. Also, unbalanced charges on atoms sometimes appear at the edges of the flat crystal structures, there creating collections of excess positive charge.

Collections of unbalanced charge do not persist in nature because their electrostatic fields quickly attract balancing charges of the opposite sign. Both air and water contain adequate supplies of ions, so the negatively charged flat surfaces of clay particles quickly build a balancing layer of positive ions just outside the surfaces.<sup>9</sup> If a clay particle is in contact with air, the balancing positive layer is thin and tight against the clay sur-

9. Air at the earth's surface contains hundreds to thousands of ions per cubic centimeter, and even pure water contains both positive and negative ions because a tiny fraction (one in ten million) of the molecules are disassociated.

face. If the clay is immersed in water, the balancing layer is more diffuse and extends out into the liquid in the pore spaces in a distribution that minimizes free energy.<sup>10</sup> Like clay particles, humus particles in a soil body are typically negatively charged, and therefore they too attract balancing positive charge layers. As in the case of the water molecule, such a particle then has a dipolar charge distribution, because the centers of positive and negative charge are not colocated. Humus particles also act as cementing agents.<sup>11</sup>

Included in the materials attracted to clay and humus particles are the positive ends of water molecules' dipoles. This electrical attraction, along with hydrogen bonding of the water molecules to the soil, explains why clay and humus matter adsorb water so strongly.<sup>12</sup> Another consequence of the electrical nature of clay and humus particles is the development of electrostatic attractive forces between the particles that help draw the particles together. Acting in opposition are electrostatic repulsive forces between soil particle surfaces of like sign. These attractive and repulsive forces and also Van der Waals forces operate with different intensities and vary differently with distance; the way they balance out will help determine if soil particles tend to coagulate or disperse.<sup>13</sup> Affected are both the degree to which soil particles coalesce to form larger aggregate assemblies and the strength of a soil.

In sum, soil is a highly complicated medium composed of mineral and organic materials of various sizes and shapes. The sizes and shapes involved greatly affect the strength of a soil and how it interacts with liquids and gases within the pore spaces. These ongoing interactions are mechanical, chemical, and electrical, and so soil is a complex substance prone to change with time as the soil's environment changes. Typically, the character of a soil depends strongly on climate, and it varies with depth beneath the surface. Additional discussion of these variations is in Appendix B.

10. The configuration—negative charge at or just inside the clay surface, and a balancing layer of positive charge just outside—is called an electrostatic double layer. Ions in the double layer can be exchanged for others in a surrounding solution. This ion exchange process is crucial to plant growth because it is a mechanism for supplying the food plants need, the exchange taking place between root hairs and the surrounding liquids or directly between root hairs and soil particles [Hillel (1980) pp 77–83].

11. Hillel (1980) p 77.

12. Hillel (1980) p 85.

13. Hillel (1980) pp 87–90.

## ■ The Movement of Water through the Ground under the Influence of Suction

The Universal System Happiness Rule's insistence that every system strive to hold minimum free energy provides the fundamental explanation for the formation of pipkrakes. Water transported to the bottom of the pipkrake layer freezes into pipkrake ice because the water molecules possess less free energy when locked into ice crystals than when in liquid or gaseous form. The water molecules move horizontally or upward through the ground toward the freezing front because by doing so they decrease their free energy.

Differences in free energy make things happen: the differences may cause molecules to combine chemically with others, to change state (solid, liquid, or gas), or to move from one place to another. Here we are concerned with change of state of water molecules and the motions they undergo through soils. For this purpose it is useful to think of free energy simply as potential energy. When water falls on the surface of the ground in the form of rain or snow, some of the molecules evaporate back into the atmosphere, some sink into the soil, and others run off along the surface. In each instance, differences in free energy determine the fate of the molecules (even those evaporating because they are supplied with additional free energy from elsewhere).

The simplest fate to describe and understand is that of the water molecules that run off along the surface on the road to lowered free energy. They are moving downhill under the influence of gravity, and are impeded in that motion mainly by the water molecules' high viscosity, large surface tension, and proclivity to cling to the upper surface of the ground—all products of water's ability to form hydrogen bonds.

More complicated is the motion of water molecules that enter the ground. These molecules feel the pull of gravity, but other important forces come into play, and they are very powerful because of the abnormally graspy nature of water molecules, both for each other and for the soil particles they contact. Forces involving *adsorption*, *capillary* action, and *osmosis* literally suck water down into the ground and later may lift it up again to the surface or higher, into plants and trees. In dealing with movement of fluids and vapors through substances, scientists use various ways to describe quantitatively how these forces operate, and one of the easiest involves the concept of suction.

Before delving into that, it is worthwhile to note that in soils and plants the most important movements of water molecules involve those in liquid

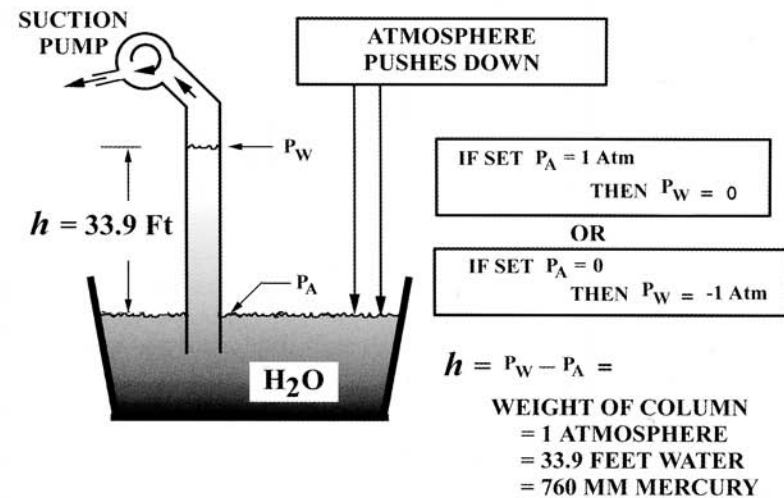
form. Water vapors, i.e., collections of  $H_2O$  molecules in gaseous form, play a role, especially in the movement of water through unsaturated coarse-grained soils, but the primary transport through fine-grained soils and through plants involves water in liquid form.<sup>14</sup> Hereinafter, unless the context suggests otherwise, the word “water” is used to imply the liquid state of this substance.

### ■ Suction and Several of Its Causes

Since all healthy mammals are born with an innate ability to apply suction it is at the outset a familiar idea. Furthermore, suction is measurable and therefore has useful quantitative meaning. The ordinary everyday kind of suction is what gets soda from a straw and gasoline out of an automobile’s tank. It operates to lift water when the handle of a shallow-well pump descends. This kind of suction has a very definite limit because the operating force is atmospheric air pressure. The weight of the atmosphere pressing against the surfaces of the soda, the gasoline, and the well water causes these liquids to rise in the tube when suction is applied. The limit to the suction obtained in this fashion is the weight of the atmosphere, at sea level  $1.033 \text{ kg/cm}^2$  ( $10.33 \text{ tonnes/m}^2$ )<sup>15</sup> or  $14.7 \text{ lbs/in}^2$ . The same weight comes from a column of water  $10.3 \text{ m}$  ( $33.9 \text{ ft}$ ) high, and a column of the much heavier liquid mercury  $76 \text{ cm}$  ( $29.9 \text{ in}$ ) high.

If as shown in **Figure 3.4** we place a tube in an open vessel of water—like the straw in the soda bottle—and suck on the tube we are withdrawing air molecules from it and thereby are reducing the pressure inside the tube. The weight of the atmosphere pressing on the open surface outside the tube pushes water up into the tube, and if we create a perfect vacuum in the tube by sucking out all the air the water column rises to a height of  $10.3 \text{ m} = 33.9 \text{ ft}$ . The pressure at the top of the column is zero, and at the bottom the pressure is equal to the weight of the atmosphere.

A convenient measure of pressure for many purposes is the *atmosphere (Atm)*, and it is meaningful as well since everyone can readily



**Figure 3.4** When all the air is sucked out of a tube immersed in water, the pressure of the atmosphere will push water  $10.3 \text{ m}$  ( $33.9 \text{ feet}$ ) up the tube.

comprehend its magnitude—imagine, for example, balancing on your hand a stick or other object that weighs  $14.7 \text{ lb}$  and has an area of  $1 \text{ in}^2$ . That’s how a pressure of  $1 \text{ Atm}$  feels. Within the earth, and in the same fashion, the weight of overlying material causes increasing pressure downward; the pressure is near  $10,000 \text{ Atm}$  at a depth of  $30 \text{ km}$ , and at the center of the earth it is near  $3 \text{ million Atm}$ . (In dealing with pressure, scientists may use other units such as the bar, the pascal and newtons/ $\text{m}^2$ ;  $1 \text{ atmosphere}$  equals  $1.013 \text{ bar}$ , and  $1 \text{ bar}$  equals  $100 \text{ kilopascal}$  or  $10^5 \text{ newtons/m}^2$ .)

It is obvious that the pressure at the level of the base of the column shown in **Figure 3.4** is everywhere  $1 \text{ Atm}$ , that the pressure declines with distance above the surface, and exactly at the top of the column it is zero. Describing the situation in another way, we can say that a pressure gradient exists within the column, and it is such that the difference in pressure from top to bottom of the column is  $1 \text{ Atm}$ . That pressure difference (the gradient times the distance) does not depend on what we choose as a reference pressure. For many purposes it is convenient to set the reference level at atmospheric pressure; that is, to choose sea level pressure as zero. Ordinary pressure gauges used to measure air pressure in automobile tires read this way, and if they did not, a person using one would have to mentally subtract  $1.033 \text{ kg/cm}^2$  or  $14.7 \text{ lbs/in}^2$  from the gauge readings, a bothersome chore. Adopting the same idea, and again referring to

14. I stress this point here because it is possible to find literature dealing with permafrost and related topics that places too much emphasis on the role of water vapor and also may misidentify the relevant transport processes. For example, a valuable engineering guidebook by McFadden and Bennett (1991) contains statements about the role of water vapor and osmosis on pages 89–91, 167, and 185–86 that are at variance with other references cited here.

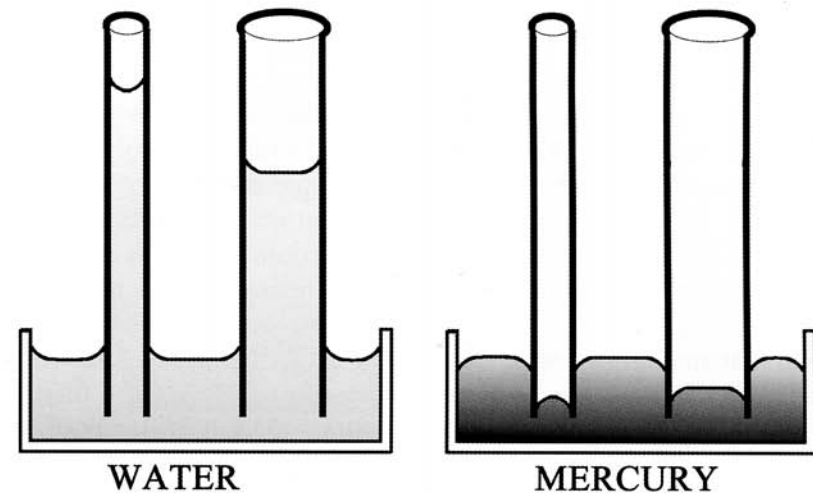
15.  $1 \text{ tonne} = 1000 \text{ kg} = 1 \text{ metric ton}$ .



Figure 3.4, we can set the pressure at the exposed liquid surface equal to zero. Then  $P_a = 0$ , and  $P_w = -1$ . The difference in pressure is still the same:  $P_a - P_w = 0 - (-1) = 1$ , but we have forced ourselves to take on the concept of *negative pressure*. Suction creates negative pressure.

Now suppose that the suction applied to the tube were only enough to raise the column to height  $h = \frac{1}{4}h_0$ . The pressure at the top of the column would then be  $-\frac{1}{4}$  Atm, and we could say that we have applied a suction  $S = \frac{1}{4}$  Atm to the column. Doubling the suction to  $\frac{1}{2}$  Atm doubles the column length to  $h = \frac{1}{2}$ . Doubling it again brings us back to where we started  $h = h_0$ , and suction  $S = 1$  Atm, but in the process we have reaffirmed that suction is a force that creates reduced pressure, and we have seen how it can have precise quantitative meaning. Notice also that since  $1 \text{ Atm} = 76 \text{ cm mercury} = 29.9 \text{ in mercury} = 33.9 \text{ ft H}_2\text{O}$ , any of these other lengths also can be used to indicate the magnitude of suction. Vacuum-suction (the suction produced, as Webster says in defining suction, by “the act or process of exerting a force on a solid, liquid or gaseous body by reason of reduced air pressure over part of its surface”) cannot exceed  $1 \text{ Atm} = 33.9 \text{ ft H}_2\text{O}$ , but the concept of suction can be extended to describe pressure reductions created by other, more powerful processes such as capillary action, osmosis, and temperature differentials. These processes can create larger suction (negative pressures) amounting to tens or even hundreds of atmospheres. The important consequence of suction, regardless of their cause, is that they frequently cause the movement of material from one place to another.

A powerful creator of suction is capillary action, and it is a very important mover of water through soil and plant material. The usual way to illustrate the way capillary action works is to place small-diameter tubes in containers of water and mercury, as shown in **Figure 3.5**. The water rises up the tubes, but the mercury moves *down* the tubes. This difference in behavior is because water wets glass, but mercury does not. With its graspy hydrogen bonds water reaches out to the oxygen atoms in the capillary tube and so not only clings to the walls, it tries to crawl up them so that the surface of the water becomes dished in what is called a concave *meniscus*. Mercury atoms have high attraction for each other (mercury’s extremely high surface tension expresses this) but no attraction for glass. Therefore mercury retreats from the walls of a glass capillary tube, sinking in on itself to form a convex meniscus. In establishing these shapes and positions within the capillaries both liquids are merely behaving according to the Universal System Happiness Rule that requires them to arrange themselves to have minimum free energy. Mercury’s free energy is minimized by the



**Figure 3.5** Water rises and mercury falls in a capillary, the distances increasing as the capillary becomes smaller.

element’s becoming as spherical as possible because of its high surface tension and its complete lack of attraction for glass, but the free energy of water is minimum when the hydrogen bonds (like Velcro hooks) are enmeshed with oxygen atoms in the capillary walls, and the more hydrogen bonds formed with the oxygen in the walls the better.

In a capillary made of a material that water wets, the water will reach up the wall of a capillary as far as it can, but only to a height that depends on the water’s surface tension and on the diameter of the capillary. In crawling up the wall the water must lift its entire mass against the force of gravity using a force that is equal to the water’s surface tension  $\sigma$  times the circumference of the capillary  $2\pi r$ , where  $r$  is the radius of the capillary. The opposing downward force is the weight of the water in the column and that is equal to the volume of water multiplied by its density  $\rho$  and the acceleration of gravity  $g$ . Since the volume is the height of the column  $h$  times the area  $\pi r^2$ , the downward force is  $\pi r^2 h \rho g$ . Balancing the upward and downward forces ( $2\pi r \sigma = \pi r^2 h \rho g$ ) requires that the height of the column be  $h = 2\sigma / r \rho g$ , and so we see that the height of the column is directly proportional to the surface tension and inversely proportional to the radius.

The pressure in the water just at the meniscus ( $P_{\text{liquid}}$ ) is reduced below the pressure in the air just above ( $P_{\text{air}}$ ) by an amount equal to the

downward force per unit area generated by the column, and that is  $h\rho g = 2\sigma/r$ . Thus  $P_{\text{air}} - P_{\text{liquid}} = 2\sigma/r$  is the suction produced by the capillary action. Notice that as the radius  $r$  of the capillary is made smaller both the suction and the height of water in the capillary increase. In principle, both would approach infinity as the radius of the capillary approaches zero. Even if other limiting factors did not interfere, the water column should eventually break despite its great tensile strength. Remarkable as it seems, water in thin columns does have great tensile strength, as great as that of some steel. Laboratory tests indicate a practical limit near 130 Atm, more than twice the tensile strength water needs to pull itself all the way up through the capillaries of even the tallest trees.

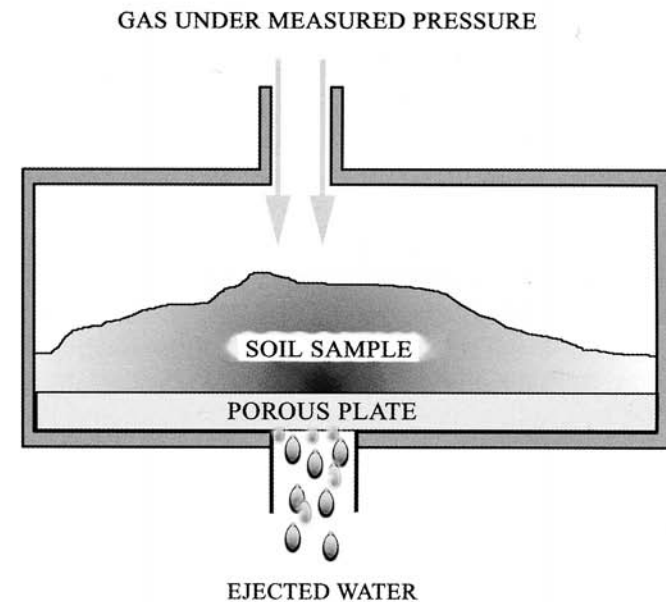
Water's ability to move so far up glass capillaries or others such as those in plants that contain ample oxygen in the capillary walls is made possible by the clinging of the water to the wall surfaces, the process called adsorption, and which in the case of water is fostered by its proclivity for hydrogen bonding. (Adsorption is very different from absorption, the process of one substance taking another into its internal structure.) Van der Waals forces play a key role in adsorption, and water adsorbs well on certain substances for yet another reason: in addition to making hydrogen bonds, the water molecule is highly dipolar (that is, it has a large dipole moment, as is indicated by water's large dielectric constant), and that causes water to adsorb to materials like clay and humus particles that also are dipolar. It is an electrostatic attraction between the poles of opposite sign in the adjacent dipolar materials. Silica gel is another strongly dipolar material, and that is why packets of silica gel are included in shipping containers surrounding items that need to be kept dry.

Crucial to the amount of water adsorbed onto another substance is the available surface area. Since the ratio of surface area to volume increases the more finely a material is divided, the smaller its particles the more it adsorbs. A single rock weighing one pound has a surface area far less than 1 ft<sup>2</sup>, but if that rock is ground up into fine clay its surface area increases to several tens of acres. Finely divided charcoal, much used to purify air of odors and poisons, has a surface area exceeding 100 acres per pound (90 hectares/kg).

Adsorption can be onto a smooth surface but it is especially effective on pitted and rough surfaces like those of soil particles because their nooks and crannies increase surface area and also take up the role of capillaries. In fact, it is thought that adsorption typically depends much on capillary action, but the fundamental cause of adsorption is the collection of all the intermolecular forces that are in operation at the interface between the adsorber and

the adsorbate. When molecules of water or other adsorbed substances stick to surfaces most tightly they are, in essence, experiencing a suction created by hydrogen bonding, the electrical attraction between dipoles of the involved substances, Van der Waals attractions involving the molecules of the water and the adsorber, and electrostatic forces arising from distributions of surface charge. This suction is typically even larger than the purely capillary suction. (Additional discussion of how water's high viscosity, high surface [interfacial] tension, and wetting ability affect its behavior in capillaries and at interfaces appears in Appendix A.)

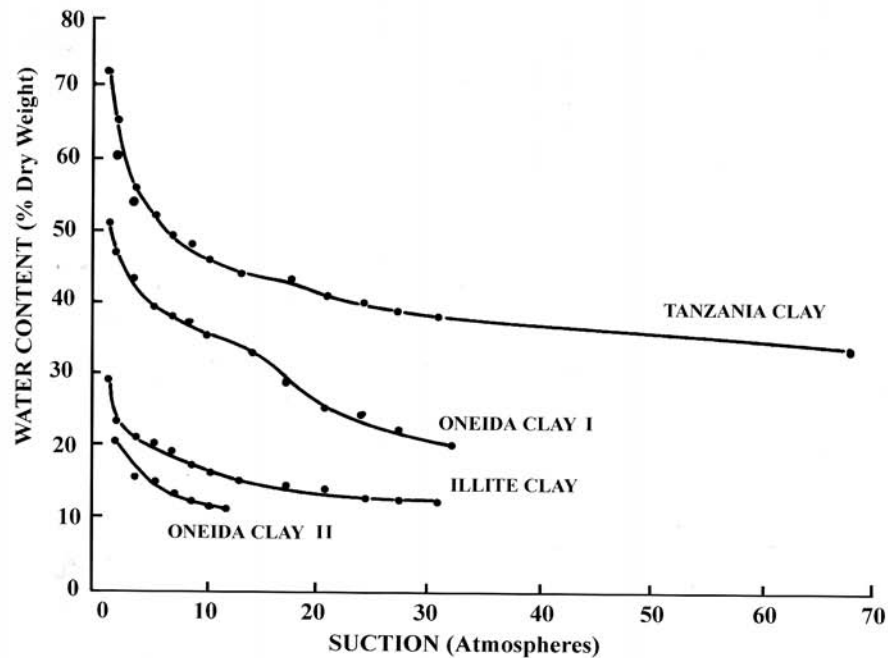
Since the tightness with which molecular forces at interfaces and capillary action hold water to a soil is of major importance to plant life, soil scientists have devised ways to measure how easily a soil gives up adsorbed and capillary water. One direct way is place the soil on a fine porous plate in a container that can be pressurized to force water from the soil into the atmosphere, as illustrated in **Figure 3.6**. The pressure required to force water out is equal to the suction force holding the water in, so at any pressure the gauge reading directly indicates the suction holding whatever water remains in the soil. A plot of the data obtained as the pressure is



**Figure 3.6** Apparatus for measuring the soil moisture characteristic curve of a soil. The gas pressure directly measures the suction.

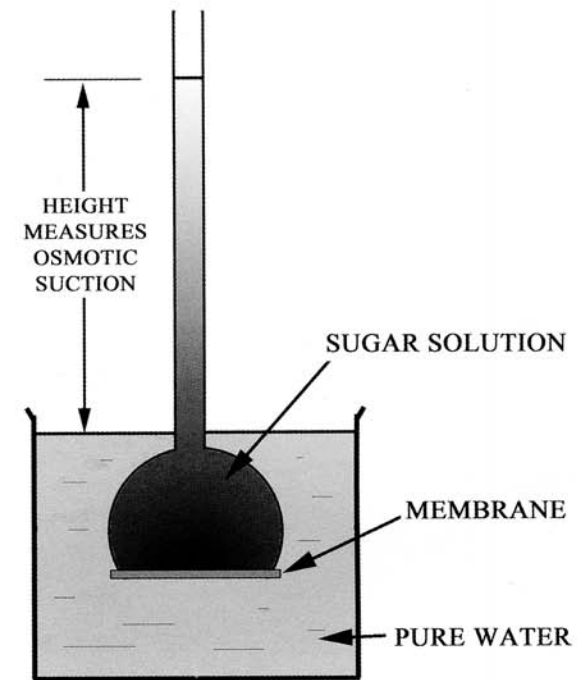
increased gives for a soil sample a curve of water content versus suction. Several examples of these *soil moisture characteristic curves* appear in **Figure 3.7**, which shows that suctions of 10 or more atmospheres are required to remove the first half of the water contained in various clay soils. The most loosely held water, that held in the soil by capillary action in the larger pores, comes out first, and an increasing proportion of that remaining is the tightly held adsorbed water. In general, the soil moisture characteristic curves show no obvious sharp transition between capillaries and oxygen-rich or dipolar adsorptive surfaces, a signal suggesting that the underlying physical causes of capillary action and adsorption are one and the same: forces acting between molecules lying near each other at boundaries between different substances.

Plants also have another tool for collecting water, osmosis. Osmosis is the one-way movement (diffusion) of a substance through a membrane that is tight enough to exclude some molecules but porous enough to allow the passage of others. Plant and animal cell walls are examples of such semipermeable membranes. An easy experiment displaying the effect of osmosis and how



**Figure 3.7** Soil moisture characteristic curves for several clay samples. Redrawn from Figure 7.6 of Williams and Smith (1989).

to measure the suction it generates is shown in **Figure 3.8**. A membrane made of skin, parchment, or cellophane separates sugar water from pure water. Sugar does not pass through the membrane, but water does, so the solution rises up the tube to some distance  $h$ . The distance  $h$ , after correction for capillary effects, is a measure of the osmotic suction. It is found to be directly proportional to the concentration of sugar above the membrane. Familiar examples of osmosis are the passage of nutrients through the walls of the stomach and intestine to the bloodstream, and both the inward diffusion of oxygen and the outward diffusion of carbon dioxide through the linings of the lungs. Osmotic suctions in plants can be very large; the largest observed are in saltbush leaves, in excess of 200 Atm.<sup>16</sup> Clay soils may also in some instances behave in osmotic fashion, causing minor and usually insignificant differential movements of solutions of differing character.<sup>17</sup>



**Figure 3.8** Water passes through the membrane, but sugar does not.

16. Meyer (1973). Even higher suctions (lower negative pressures, down to -330 Atm) have been observed in laboratory situations.

17. Hillel (1980) pp 245-46.

### ■ Effects of Adsorption, Capillary Suction, and Osmotic Suction in Plants and Soils

In situations such as depicted in Figures 3.4, 3.6, and 3.8, a movement of liquid or gas takes place until equilibrium occurs. It is a transient movement, lasting only long enough to bring the overall system to minimum free energy by transporting a sufficient number of molecules to locations of least free energy. However, if other processes act to remove molecules from the low free-energy end of the system, and to supply them to the other end, flow will continue. In fact, a higher extraction than supply rate generally will increase the suction within the system, until either the supply matches the extraction or the supply limits the transfer through the system.

Such effects are thought to explain how water can rise up through even the tallest of all trees, the giant redwoods exceeding 110 m (360 ft). **Transpiration** from leaves rapidly removes water so that osmosis suctions in the range 5 to 40 Atm typically occur. These large suctions act in conjunction with those produced by adsorption on the walls of plant tissues and capillary action in plant structures. The resulting huge pressure deficits (suctions up to 50 Atm) essentially suck water up through the tree; lifting water to such heights is possible because water clings to itself so well (has such high surface tension), behaving like thin steel wires in small capillary passages.

As plants lose water to the atmosphere by transpiration most of them try to replenish it by sucking water in through their roots, and that effort is strongly affected by soil **capillarity** and other processes that dictate the rates at which water moves through the soil. Water falling on dry soil rapidly diffuses downward until the water molecules adhere to all the surfaces of the soil particles and partially fill the intervening pore spaces. The pore space is about 30% in sandy soils, and about 50% in clay soils. After several days of rain, the rate of water movement downward slows, and the water content becomes fairly uniform in the wetted soil layer. This soil typically also contains much air in the pore spaces unless the soil lies below the water table, and in that case the pore spaces fill with water and the soil is said to be saturated. Most plants grow best in partially aerated soil that is able to release water to the plant roots, but a certain portion of the water adheres so tightly to the soil particles that it is unavailable. Adsorption and perhaps some capillary action locks up about the first 20% (by volume) in clay soils but only the first 5% in sandy soils. One consequence is that sandy soils easily and quickly make water available for

plant growth, whereas clay soils release water more slowly. In their pore spaces clay soils are capable of containing more water than sandy soils, so even though they hold back a greater proportion, they may supply more in the long run.<sup>18</sup>

In saturated soils the free energy is fairly uniform throughout, hence pressure deficits between locations are generally less than 1 Atm. The deficit may be near 15 Atm in soils that have released all the water they can to plants (all but the first 5% in sandy soil and 20% in clay soil), and if the water content falls below those levels the pressure deficits (the suctions) get very large, in the range of hundreds of atmospheres. The free energy of the remaining water molecules is so low that plants are unable to suck them away from the soil particles. Even with such large pressure deficits, water usually cannot move through soil fast enough to satisfy the needs of plant growth. When the water cannot come to the plant, if it is to avoid wilting the plant has to go to the water, and that is what happens. It sounds unbelievable, but reported measurements on rye grass indicate that the average rye plant increases the length of its root system by about 5 km (3 mi) each day, and by doing so gains enough water to replace that lost through transpiration.<sup>19</sup>

To recapitulate: although osmotic suction is relatively minor, capillary suction and the suction produced by adsorption are important forces affecting the motion of water through soil. Another powerful suction, **cryosuction**, also comes into play when the temperature drops below freezing. Like the other kinds of suction already discussed, cryosuction can be very large, ranging to up above 200 Atm, so it too will move water through the ground, permitting the formation of pipkrakes and other kinds of ice segregations.

As we move forward it is useful to keep in mind that since suction—however it is caused—is a reduction in pressure from some base level, it can be called a pressure deficit, or a negative pressure. Both terms appear in the literature on soil science, along with others like “pore water pressure,” “potential,” “matric potential,” and “Gibbs free energy.” The plethora of terms used to describe what is really only one thing is confusing. The key thing to remember is the direct inverse relation between the free energy of a substance and the suction it experiences. A quick look at **Figure 3.9** may be most helpful of all.

18. Hillel (1980) pp 164–65.

19. Meyer (1973).

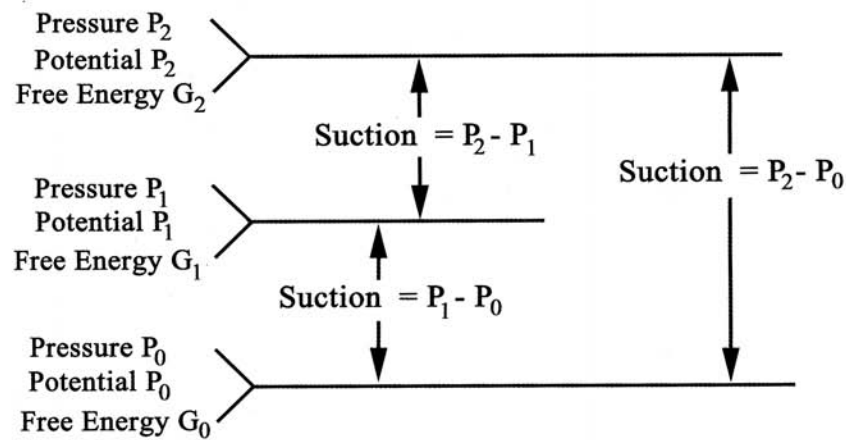


Figure 3.9 A diagram showing the relationship between suction and its related quantities, pressure, potential and free energy.

## ■ Cryosuction in Freezing Soil and the Formation of Segregation Ice

### Introductory comments

**Soil temperatures.** Changes in air temperature produce changes in ground temperature, but the ground responds relatively slowly. The situation is the same as that when a person holds one end of a steel bar in a fire.<sup>20</sup> The end of the bar in the fire starts getting hot right away, but it takes some time for the heat energy going into the end of the bar to travel down its length and burn the holder's hand. The temperature within the bar varies with position along the length of the bar, and thus a temperature gradient exists within the bar. And so it is with soil as the temperature in the air changes; the top layer of soil responds the quickest, and the deeper layers follow suit, but more slowly. If a step change in temperature is imposed at the top of the ground, the change will propagate down at a rate that varies with the square root of the time elapsed. Thus if change induced at the top of the ground can penetrate to a depth of 10 cm in one day, the change will penetrate to 20 cm in 4 days, to 50 cm in 25 days, and to 1 meter in 100 days. That is why cold-area residents sometimes are

20. John S. Wettlaufer, in a private communication with the author in 1999, suggested using this analogy.

chagrined to have their sewer lines freeze up in early summer, long after the coldest winter weather.

**Water migration.** Because of the slowness of the ground at depth to respond to changes in air temperature, soil temperature in cold climates is rarely uniform throughout. To put it another way, the soil almost always has a temperature gradient. Seeking to minimize its free energy, any water in the soil will try to migrate through the soil toward the coldest part—that is, down the temperature gradient. Visual proof that water molecules in vapor or liquid form seek the cold is the condensation—or even ice—that forms on the inside surfaces of single-pane windows on cold days. The cold window panes act as sinks that can literally suck water out of the air because the water molecules lower their free energy by collecting on the cold window surface. Much the same thing takes place in a freezing soil in that water molecules travel to and into the soil and collect there in the form of ice.

**Freezing point depression.** The normal freezing point of water,  $T_0$ , is the temperature at which **bulk water** freezes, i.e., water not in capillaries or otherwise confined.  $T_0 = 0^\circ\text{C}$  when the water is at a pressure of 1 Atm. Increasing the pressure on most substances causes their freezing points to increase, but water is an exception. Its freezing point gets lower with increasing pressure, but only by a slight  $0.0074^\circ\text{C}$  per Atm. Consequently the freezing temperature of any bulk water lying several meters below the ground surface is only a tiny fraction of a degree lower than that at the ground surface, although if it were buried beneath several thousand meters of overburden the freezing point would be depressed by 1 to  $2^\circ\text{C}$ . Salts or other impurities added to water also lower its freezing temperature. Adding 2% (by weight) sodium chloride lowers water's freezing point about  $1^\circ\text{C}$ , and a 22% salt solution has freezing temperature just below  $-20^\circ\text{C}$ . Water in soil contains some salt but so little that its freezing temperature is not significantly lower than  $0^\circ\text{C}$ . Thus in dealing with the freezing of the ground near the earth's surface, depressions of the freezing point  $T_0$  caused by water's salt content or pressures from overburden are of little consequence, and we can think of  $T_0$  as essentially equal to  $0^\circ\text{C}$ . Soil begins to freeze when the temperature falls only slightly below  $0^\circ\text{C}$  and, as further cooling occurs, more and more water freezes out.

### The freezing of soil

If an open container of water is cooled while being stirred so that it everywhere maintains the same temperature, then each gram of water gives up

1 calorie per degree of cooling (its heat capacity) until it reaches the freezing point, 0°C, where the sudden change of state to ice releases 80 cal/gm, the latent heat of fusion. But if that same water is contained within the pore spaces of a soil and adhered to the surfaces of the soil particles, the freezing behavior is quite different. Observations show that the water begins to turn to ice at 0°C, but that some unfrozen water remains in the soil down to quite low temperatures.

Plots of the unfrozen water content of a soil versus temperature such as those shown in Figure 3.10 are called *soil freezing characteristic curves*. As the water freezes, it gives off a known amount of heat (the latent heat of fusion); thus, by measuring the heat given off from a soil sample as its temperature is changed, it is possible to construct its soil freezing characteristic curve. Notice that the general shapes of the soil

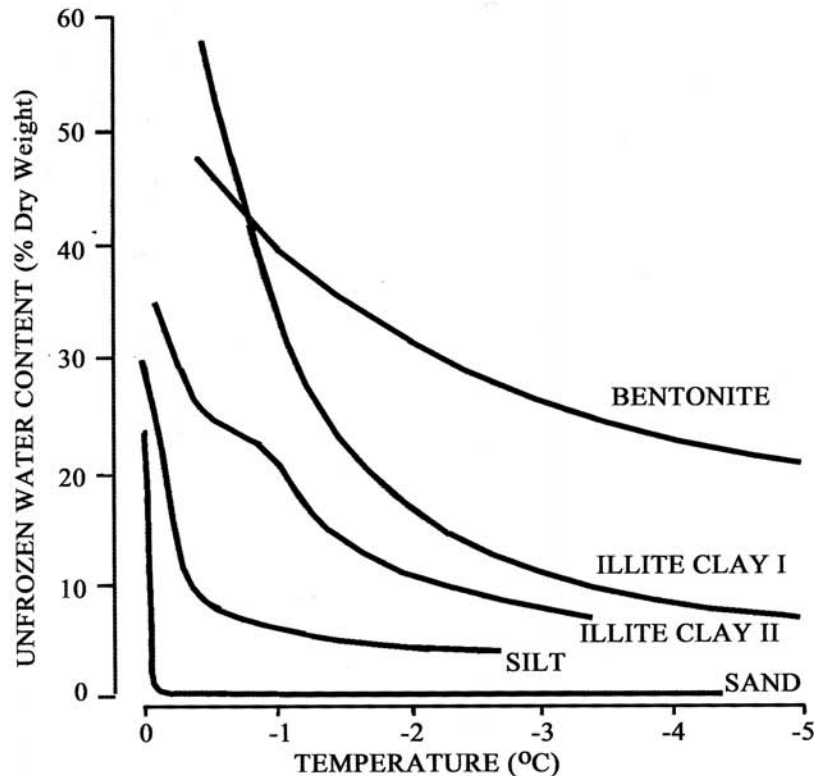


Figure 3.10 Soil freezing characteristic curves for a few clays, a silt and a sand. Redrawn after Figure 1.4 of Williams and Smith (1989).

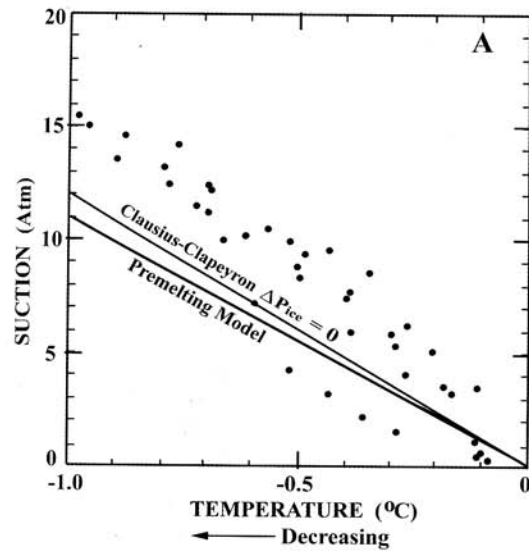
freezing characteristic curves in Figure 3.10 are very similar to the shapes of the soil moisture characteristic curves shown in Figure 3.7. The similarity is not accidental; in fact, the shape of a soil's freezing characteristic curve can be predicted from its soil moisture characteristic curve. Both curves depend directly on the size of the pores in a soil and the total surface area available for adsorption of water. An extreme example of a soil that resists having its unfrozen water driven off by imposed pressure or lowered temperature is bentonite. This fine clay, formed from weathered volcanic ash, has such small pores and high surface area per unit weight that it retains 20% (by weight) unfrozen water even at -5°C.

By comparing a soil's moisture and freezing characteristic curves, it is possible to determine how much water is retained in a freezing soil at a particular temperature and hence to make an experimental plot relating cryosuction to temperature of the freezing soil, as in Figures 3.11A and B. (Another way is to measure the water content directly by one of several different methods, including one known as pulsed magnetic nuclear resonance.)<sup>21</sup> The results are rather stunning because it is evident that cryosuction changes in direct proportion to the temperature as it falls below 0°C, and that the rate of change in cryosuction is very high, approximately 11 to 12 Atm/°C. Thus if a parcel of fine-grained soil has a temperature of -1°C, the cryosuction within that parcel is approximately 11 to 12 Atm, and if another parcel has temperature -10°C, the suction there is 110 to 120 Atm.

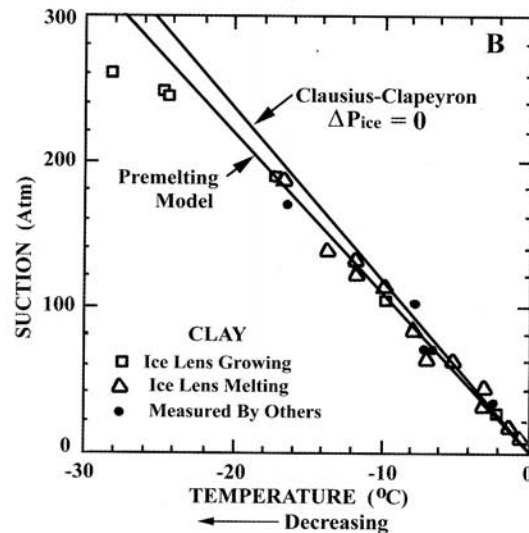
The task now is to explain why the variation of cryosuction with temperature is so great, why water and ice can coexist in frozen soil, and why frozen soil is found to contain layers of pure ice called *segregation ice* that typically increase in thickness with depth, as shown in Figure 3.12. Toward this end, geocryologists have taken several theoretical approaches, all really based on evaluation of the free energy of a soil-water system to see how it varies when changes are made and certain conditions imposed.

Useful at the outset—and relatively easy to come by—is a general insight into why such high cryosuctions develop in a freezing soil. The starting point is the Universal System Happiness Rule—the *stable, equilibrium state of a system is the state of minimum free energy*. To that rule we apply the principle of Le Chatelier which states that if a system is in a state of stable equilibrium and if one of the conditions is changed then the equilibrium will shift in such a way as to tend to restore the original condition. By itself, Le Chatelier's principle does not help much, but thanks to

21. Williams and Smith (1989) p 189.

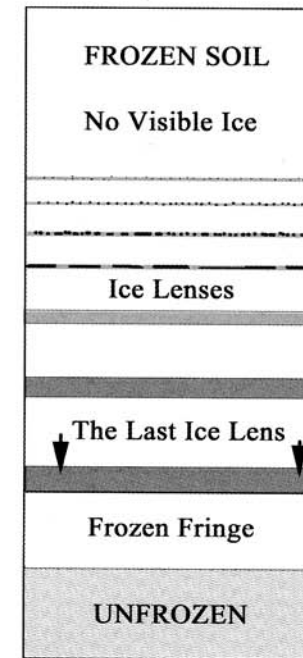


(A)



(B)

**Figure 3.11** (A) Measured suctions in the temperature range  $-1^{\circ}\text{C}$  to  $0^{\circ}\text{C}$  and the suctions predicted by the Clausius-Clapeyron equation (with  $P_{\text{ice}}$  held fixed) and premelting theory. The data points represent measurements on various fine-grained soils, as given in Figure 7.7 of Williams and Smith (1989). (B) Measured suctions in the temperature range  $-30^{\circ}\text{C}$  to  $0^{\circ}\text{C}$  together with the suctions predicted by the Clausius-Clapeyron equation and premelting theory. Based on Figure 9 of Dash et al. (1995).



**Figure 3.12** Segregation ice often forms in repeating layers that tend to thicken with depth. Based on a diagram (Figure 8.8) reproduced by Williams and Smith (1989).

the work of two gentlemen named Clausius and Clapeyron, we have the Clausius-Clapeyron equation that gives the principle specificity. If a system is in stable equilibrium, and if a change is made to one of the conditions that determines the system's free energy—temperature ( $T$ ), volume ( $V$ ) or pressure ( $P$ )—the Clausius-Clapeyron equation describes how the other conditions must vary as the system tries to restore itself to stable equilibrium. Applying this equation (derived in Appendix A) to the situation depicted in Figures 3.10A and B—the coexistence of ice and water in a soil of subzero temperature—yields the relationship:

$$\Delta T = (V_{\text{water}} \Delta P_{\text{water}} - V_{\text{ice}} \Delta P_{\text{ice}}) T_0 / L. \quad (1)$$

where ( $T$  is change in absolute temperature  $T$ ,  $T_0$  is the absolute temperature at the melting point of ice ( $273 \text{ K} = 0^{\circ}\text{C}$ ),  $V_{\text{water}}$  and  $V_{\text{ice}}$  are the specific volumes of water and ice,  $\Delta P_{\text{water}}$  and  $\Delta P_{\text{ice}}$  are changes in pressure on the water and ice, and  $L$  is the latent heat of fusion of water. Suppose that

we apply Equation 1 to a system where the pressure on the water  $P_{\text{water}}$  varies with temperature while the pressure on the ice  $P_{\text{ice}}$  is fixed by giving it a set value, perhaps that on the soil. Then  $\Delta P_{\text{ice}} = 0$ , and by rearranging some of the terms, Equation 1 becomes

$$\Delta P_{\text{water}} = L \Delta T / T_0 V_{\text{water}}$$

Then setting  $\Delta P_{\text{water}} = P_0 - P_{\text{water}}$  and  $\Delta T = T_0 - T$ , where  $P_0$  is the pressure on the water at  $T = T_0$ , the normal freezing point of bulk water,  $0^\circ\text{C} = 273 \text{ K}$ , yields an expression describing how the suction on the water varies with temperature:

$$\text{Suction} = P_0 - P_{\text{water}} = L(T_0 - T) / T_0 V_{\text{water}} \quad (2)$$

Since  $V_{\text{water}}$  is the specific volume of water it equals unity, so the suction depends only of the heat of fusion and the temperature. To see the effect of this expression we can let  $T = -1^\circ\text{C} = 272 \text{ K}$ , then the suction  $P_0 - P_{\text{water}} = L(1)/273$ . The heat of fusion of water is 334 million Joules per cubic meter (or 80 calories per gram), so the suction amounts to 12.2 bars which equals approximately 12 Atm. Thus the difference in pressure on the water and the ice changes by approximately 12 Atm per degree change in temperature, a direct consequence of the combination of the Universal System Happiness Rule and the principle of Le Chatelier. The lowered pressure on the water creates a suction that can draw water into the freezing soil, but again notice that the suction does not depend on the characteristics of the soil; instead it depends only on the heat of fusion of water and the temperature. The straight-line curve labeled "Clausius-Clapeyron" on Figures 3.11A and B represents Equation 2.

If instead of being perfectly constrained, the pressure on the ice is allowed to vary along with the pressure on the water, the pressure difference  $P_{\text{ice}} - P_{\text{water}}$  also varies somewhat. For example, if both  $P_{\text{ice}}$  and  $P_{\text{water}}$  drift away from the initial pressure  $P_0$  by equal amounts, then  $P_{\text{ice}} - P_{\text{water}} = 11.4 \text{ Atm}/^\circ\text{C}$ , and if  $P_{\text{ice}}$  changes five times as much as  $P_{\text{water}}$  when the temperature falls below  $0^\circ\text{C}$ , then  $P_{\text{ice}} - P_{\text{water}} = 11.1 \text{ Atm}/^\circ\text{C}$ . The more nearly the pressure on the water  $P_{\text{water}}$  is constrained the more nearly the pressure difference  $P_{\text{ice}} - P_{\text{water}}$  approaches the limiting value  $11.0 \text{ Atm}/^\circ\text{C}$ . Thus, according to the Clausius-Clapeyron equation,  $P_{\text{ice}} - P_{\text{water}}$  must always have a value ranging between  $11 \text{ Atm}/^\circ\text{C}$  and  $12 \text{ Atm}/^\circ\text{C}$ , regardless of how the pressures  $P_{\text{ice}}$  and  $P_{\text{water}}$  drift farther apart with decreasing temperature below  $0^\circ\text{C}$ .

### The capillary model of segregation ice formation

Segregation ice typically forms in distinct layers of microscopic to multi-meter thickness that can more than double the volume of a soil, causing it to push even heavy objects upward, the phenomenon known as frost heave.<sup>22</sup> The formation of segregation ice also is responsible for the creation of certain landforms and ground patterns found in areas where freezing occurs. One of the first approaches to the explanation of how segregation ice forms and why it is observed in layers within frozen soil is the capillary (or curvature) model of segregation ice formation.

The main idea behind the capillary model is that the relationship between cryosuction and water content in a frozen soil reflects the distribution of pore sizes in the soil. As the temperature of a water-saturated soil falls below  $0^\circ\text{C}$  the first water to convert to ice is that least attached to the soil by capillary action and adsorption to the soil particles. It is the water with the highest free energy; that is, it is the water experiencing the least suction, the water in the largest of the pore spaces. If the temperature continues to fall, ice invades successively smaller pore spaces as the free energy of the water molecules contained there exceeds the free energy they would have in the form of ice. The capillary model allows prediction of the size of the soil pores that can be invaded by ice at a given temperature.

Recall from the previous section the relationship derived to find the suction produced by capillary action,  $P_{\text{air}} - P_{\text{liquid}} = 2\sigma/r$ , where  $\sigma$  is the liquid's surface tension (interfacial tension  $\sigma_{\text{water-air}}$ ) and  $r$  is the radius of the capillary. In the capillary model of segregation ice formation the assumption is made that ice-water interfaces behave the same as water-air interfaces, and that if a soil contains ice the pressure on the ice  $P_{\text{ice}}$  is also the pressure on the soil  $P_{\text{soil}}$ . In an unburied soil the pressure on the soil is just that of the atmosphere, generally taken as equal to zero. In the capillary model, the suction is the difference between the pressures on the ice  $P_{\text{ice}}$  (or  $P_{\text{soil}}$ ) and the water  $P_{\text{water}}$ , the appropriate *interfacial tension* is that between ice and water,  $\sigma_{\text{ice-water}}$ , and the relevant radius is that of capillaries at the ice-water interface  $r_{\text{ice-water}}$ . Hence

$$\text{Suction} = P_{\text{ice}} - P_{\text{water}} = 2\sigma_{\text{ice-water}} / r_{\text{ice-water}} \quad (3)$$

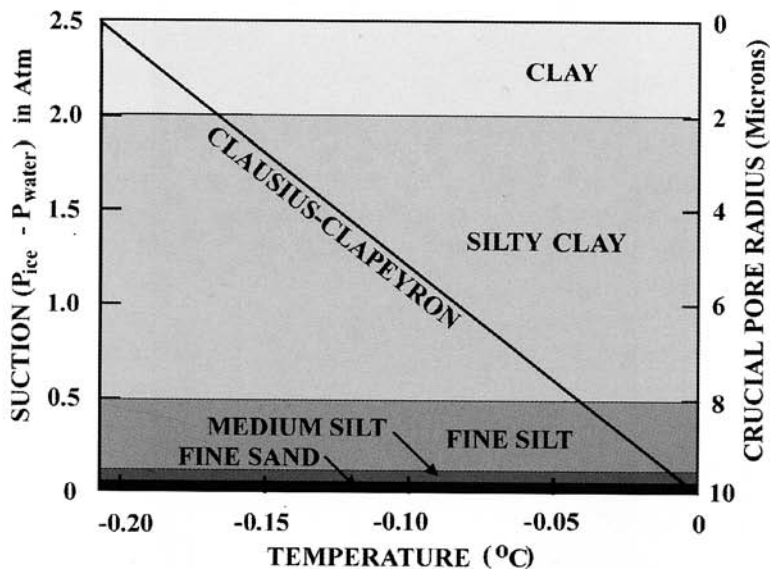
22. The increase in volume in a parcel of freezing soil wherein segregation ice is forming is due to the increase in number of molecules (soil particles plus water molecules), not the expansion by 9% of water as it turns to ice. Although engineers ignored his work for many years, as early as 1918 Taber (1929) performed experiments showing that frost heave created substantial volume changes and that the amount of heave depended on soil grain size.



A tenet of the capillary model also is that the suction  $P_{ice} - P_{water}$  is given by the Clausius-Clapeyron equation when the pressure on the ice is held fixed, that is, by Equation 2. Combining Equations 2 and 3 yields an expression for the ice-water radius

$$r_{ice-water} = 2\sigma_{ice-water} T_o V_w / L(T_o - T) \quad (4)$$

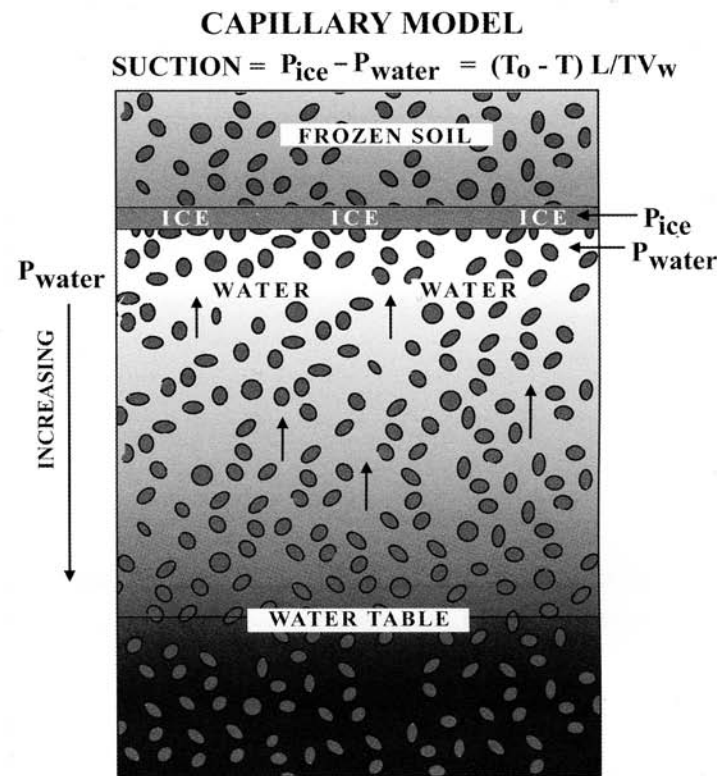
which specifies that the radius of curvature at the ice-water interface must decrease with decreasing temperature below the freezing point of bulk water, a radius assumed to be determined by pore size. Thus ice invades smaller pore spaces as the temperature falls, but the water in those pores with radii smaller than the radius  $r_{ice-water}$  remains unfrozen. **Figure 3.313** illustrates the relationships among pore size, suction and temperature, according to the capillary model.<sup>23</sup>



**Figure 3.13** A diagram showing, according to the Clausius-Clapeyron equation, how suction varies with temperature, and the minimum size of the soil pores that ice can invade (right-hand scale), according to the capillary model of segregation ice formation. Also shown is the relationship between pore size and soil texture. In soil, the suction increases with declining temperature below 0°C. At a temperature of -0.1°C, the suction is about 1.2 atmospheres and the crucial pore radius is approximately 5 microns, within the domain of silty clay. Based on data given in Tables 6.1 and 6.3 of Smith (1985a).

23. Williams (1999).

As the temperature drops below 0°C and the soil begins to freeze, the capillary model calls for the water in the largest pore spaces to freeze first. At a given temperature, if the pore radius  $r_{ice-water}$  is smaller than the minimum pore size then ice can enter the pores and so the freezing front advances into the soil. However, if  $r_{ice-water}$  is larger than the minimum pore size, the freezing front stalls. In that case, the water at the freezing front is at lower pressure than the water in the soil below, and therefore that water moves toward the freezing front in order to lower its free energy, the situation depicted in **Figure 3.14**. The movement of the water to the freezing front brings to it water of higher pressure than  $P_{water}$  in Equation 2 above, and so that water with its higher free energy converts to



**Figure 3.14** A schematic diagram illustrating the capillary model of segregation ice formation. According to this model, when the freezing front stalls, segregation ice forms, the suction (pressure difference between that of the ice and the adjacent water) increases, and water moves upward toward the freezing front. Pressure here is represented by shading; the higher the pressure, the darker the shading.

ice, forming a layer of segregation ice. The heat of fusion given off by the freezing water fights against further drop in temperature, helping to stall the freezing front.

An explanation of why segregation ice forms in multiple, regularly spaced layers, as shown in Figure 3.12, is that the movement of water from the unfrozen soil to the freezing front desiccates the unfrozen soil, thereby lowering the gradient in suction and slowing the movement of water to the freezing front. Less latent heat of fusion is given off, and that may allow the temperature at the forming ice lens to become lower, thereby permitting the freezing front to advance until conditions again become ripe for the formation of another layer of segregation ice. As the freezing front moves down through the soil to greater depth, the changes in temperature become slower, allowing more time for layers to thicken.

The formation of segregation ice in a soil causes its volume to increase, and that volume increase causes the surface of the soil to heave upwards. If the soil in which the ice is forming is essentially at the surface of the ground, then only the pressure of the atmosphere resists the heaving, the situation usually described by setting  $P_{ice} = 0$ . The pressure on the water  $P_{water}$  is the same when the temperature is  $0^{\circ}\text{C}$ , but it becomes more negative as the temperature falls, or alternately, we could say that the suction increases. As long as  $P_{water}$  is negative, the suction draws water to the freezing front, segregation ice forms, and the ground surface heaves upward. However, if the freezing soil is buried deep enough, or by other means a heavy enough load is placed on it, the pressure on the soil and the ice  $P_{ice}$  increases. The pressure on the water  $P_{water}$  must do likewise, always having a value determined by  $P_{ice}$  and the temperature. If  $P_{ice}$  is high enough at a given temperature to bring  $P_{water}$  up to zero, no longer is there a difference in pressure (a suction) between the water at the freezing front and the water in the unfrozen soil. Thus, the movement of water to the freezing front ceases, shutting off further frost heaving. Examination of Figure 3.13 shows that if the capillary model fully describes the formation of segregation ice and its consequent frost heaving, then ice formation could not take place at  $P_{ice}$  greater than about 2.5 Atm. This **shut-off pressure** occurs when the temperature at the freezing front is approximately  $-0.2^{\circ}\text{C}$ .

A soil layer approximately 3 m thick produces an overburden pressure sufficient to shut off the capillary model's formation of segregation ice.<sup>24, 25</sup>

24. Smith (1985a).

25. Although individual mineral soil particles can have densities up to 2.7, bulk soil usually has density only 1.0 to 1.7 times that of water.

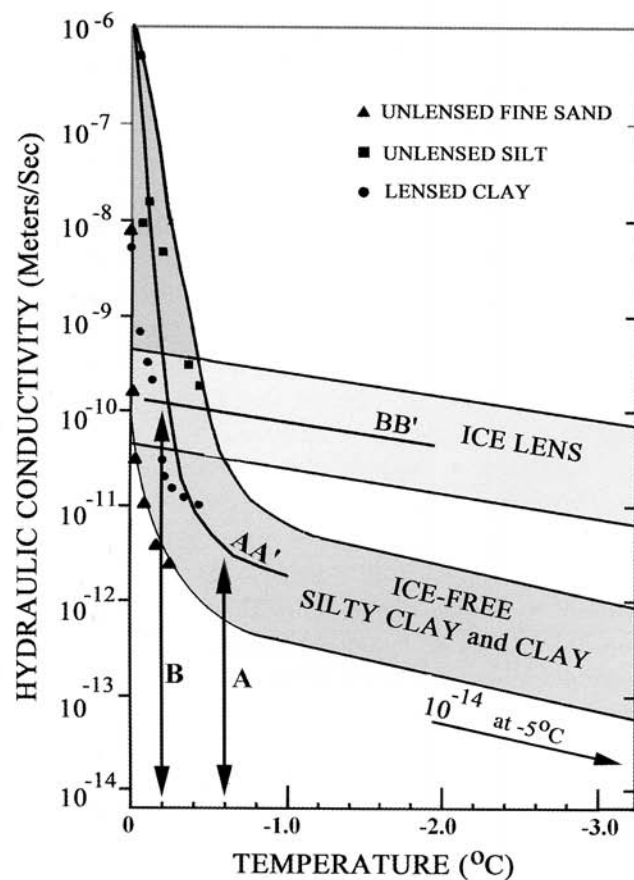
However, field observations have shown that segregation ice can form in soil buried more than 30 m, and one observation has been made of the ice forming at a depth exceeding 100 m.<sup>26</sup> Also, laboratory experiments show the growth of segregation ice at suctions far greater than allowed by the capillary model (see Figure 3.11B).

### The hydrodynamic model of segregation ice formation

The inability of the capillary model to explain the formation of segregation ice at temperatures below about  $-0.2^{\circ}\text{C}$  or heaving pressures greater than approximately 2.5 Atm led to development of the hydrodynamic model. Another consideration was that observations were showing that segregation ice was growing at locations behind the freezing front, whereas the capillary model called for formation of the ice only at the front. If segregation ice is to form behind the freezing front it is obvious that water must be moving across the freezing front through the partially frozen region shown as the "**frozen fringe**" on Figure 3.12. That flow of water presumably is through the unfrozen pores and through unfrozen water lying at soil particle surfaces or perhaps boundaries between ice crystals, a deduction forced by observations that frozen soil and ice layers do transmit water when suction is applied.

The approach in the hydrodynamic model is to examine the flow of heat and mass in a freezing soil, recognizing that these are interrelated, and assuming that the only mass transport is that of water. The procedure is to develop an expression for the transport of heat through the soil and another to describe the flow of water, then to link these together by means of the Clausius-Clapeyron equation and the assumption that the temperature dictates the amount of unfrozen water in the soil, that is, by the soil's freezing characteristic curve. The result is that segregation ice should form at some location behind the freezing front where the effective **hydraulic conductivity** changes most rapidly with temperature. At that point, water flowing through the frozen fringe piles up and turns to ice. At temperatures above  $0^{\circ}\text{C}$  the hydraulic conductivity of a soil varies slowly with temperature, but **Figure 3.15**, a somewhat schematic plot of hydraulic conductivity versus temperature, shows that the hydraulic conductivity of a fine-grained soil changes rapidly as the temperature falls a few tenths of

26. Wernecke (1932) reported segregation ice veins in quartzite rock at 100–130 m depth at Keno Hill, Yukon Territory. The overburden pressure was approximately 24 Atm, and even higher pressure may have been required to fracture the quartzite.



**Figure 3.15** A somewhat schematic drawing illustrating that the effective hydraulic conductivity of silty clay and clay is higher than that of an ice layer at temperatures just below 0°C, but at temperatures below about -0.3°C water passes more freely through the ice layer than the silty clay and clay. The hydraulic conductivity changes drastically at the temperatures marked A and B, as described further in the text. Data points are taken from Figure 7.10 of Williams and Smith (1989).

a degree below 0°C, and then declines more slowly at lower temperature. On Figure 3.15 the dark band shows the range in hydraulic conductivity of ice-free silty clay and clay, and the line AA' represents a hypothetical average silty soil. The data points are the results of measurements made on a fine sand and also on a clay-silt, both ice-free and containing ice. The rapid change in hydraulic conductivity with decreasing temperature occurs to the left of the double-angled line marking the temperature

where the slope of the line AA' changes drastically, and so the hydrodynamic model calls for the formation of segregation ice in this hypothetical soil somewhere in the temperature range 0°C to -0.6°C. For actual soils the range is typically -0.2°C to -0.3°C.<sup>27</sup> The band marked "ice lens" on Figure 3.15 represents the range of effective hydraulic conductivities of ice lenses, its width reflecting the estimated range, and the line BB' a hypothetical average. At temperatures near 0°C the effective hydraulic conductivity of fine-grained soils exceeds that of ice, but at lower temperature the reverse is true.<sup>28</sup> The hydraulic conductivity of the silt is more than 10 times that of pure ice at -0.1°C, but then as the temperature goes lower, the effective hydraulic conductivity of the silt falls dramatically until at temperatures below -0.5°C it is below that of ice by a factor of 10 or more.<sup>29</sup> Hence if ice is present, water likely will pile up where the temperature is approximately -0.2°C, at least if the line marked BB' on Figure 3.15 represents the hydraulic conductivity of ice reasonably well.

Whereas the hydrodynamic model can explain the development of segregation ice behind the freezing front, it still fails to account for the formation of the ice at temperatures more than a few tenths of a degree below 0°C, and also the high heaving pressures that have been observed. At best, the model allows for heaving pressures of not more than a few atmospheres.

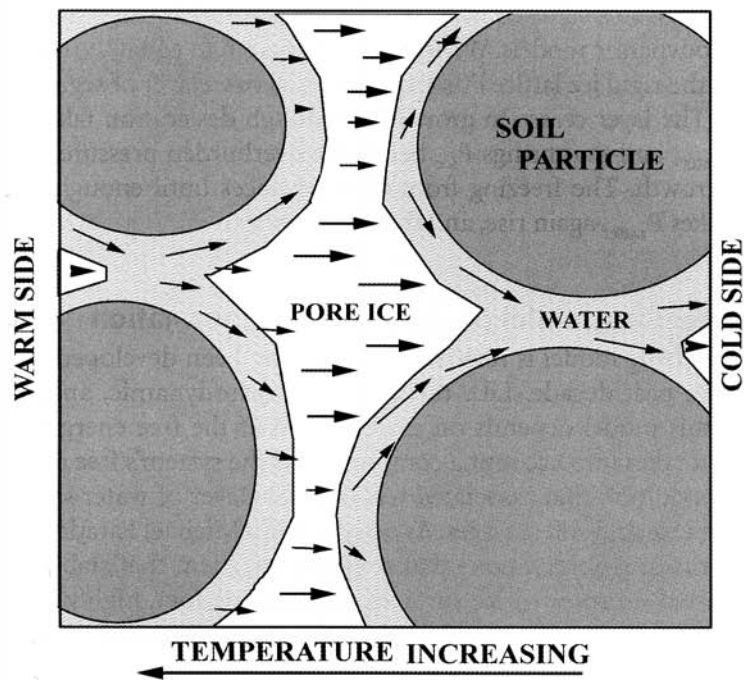
### The rigid-ice model of segregation ice formation

Like the hydrodynamic model, the rigid-ice model seeks to explain the formation of segregation ice behind the freezing front by examining the coupled flow of heat and mass. Ice transport is envisioned to take place by the process of thermally induced regelation. Driven by a temperature gradient, the process is as depicted in Figure 3.16. In this model the ice forms a rigid lattice that is separated from the soil particles by a thin layer of adsorbed water. In the thermal regelation process water molecules move from warmer to colder locations within the system by continuously transforming between liquid and solid phases. Water continuously transforms to ice at the upstream ice boundaries where the free energy of the water slightly exceeds

27. Smith (1985a).

28. Loch and Kay (1978); Horiguchi and Miller (1983) both cited by Williams and Smith (1989) p 215.

29. Oliphant et al. (1983); Ohrai and Yamamoto (1985); Smith (1985) all cited by Williams and Smith (1989) p 213.



**Figure 3.16** According to the rigid-ice model, the flow of water through soil is enhanced at temperatures just below  $0^{\circ}\text{C}$  by thermal regelation, the process involving melting of ice on a warm boundary and refreezing on the opposite cold boundary.

that of the ice at the temperature there, and the reverse occurs at the downstream ends where the ice melts to water. At the warm boundary of an ice body, water traveling in liquid form through the pore spaces transforms to ice, and at the cold boundary the ice transforms back to water. The Clausius-Clapeyron equation is assumed to assure that the pressure difference  $P_{\text{ice}} - P_{\text{water}}$  at both boundaries is determined by the temperature at the boundary location. The thermal regelation process is proposed to increase the effective hydraulic conductivity greatly in the frozen fringe, and if the hydraulic conductivity is high, relatively little suction is needed to move water. Hence, the pressure on the water in the frozen fringe need not go much below the pressure on the water in the unfrozen soil for continued inflow. In essence, the unfrozen soil water acts like a reservoir to help maintain  $P_{\text{water}}$ . But  $P_{\text{ice}} - P_{\text{water}}$  is determined by the temperature, so maintenance of  $P_{\text{water}}$  drives  $P_{\text{ice}}$  upward as the temperature falls. Thus an important characteristic of the rigid-ice model is that  $P_{\text{ice}}$  is allowed to rise,

rather than given a value equal to the overburden pressure as in the capillary and hydrodynamic models. When  $P_{\text{ice}}$  rises enough to equal the overburden pressure, the rigid ice lattice lifts the soil and allows a layer of segregation ice to form. The layer ceases to grow when enough desiccation takes place to lower  $P_{\text{water}}$ , and that brings  $P_{\text{ice}}$  below the overburden pressure, terminating the growth. The freezing front then advances until enough inflow of water makes  $P_{\text{water}}$  again rise, and the cycle repeats.

### The premelting model of segregation ice formation

The *premelting* model is relatively new, having been developed primarily during the past decade. Like the capillary, hydrodynamic, and rigid-ice models, this model depends on examination of the free energy of a system, but it takes into account a component of the system's free energy that the others do not: that associated with a stable layer of water sandwiched in between ice and soil surfaces. As early as 1850 Michael Faraday and others were making observations that suggested to them that stable layers of water existed on snow or ice surfaces. A contemporary, highly influential Lord Kelvin, incorrectly thought that Faraday's observations were due simply to the melting of ice under pressure, so the existence and characteristics of the stable water layers did not receive the attention deserved until much later. Now, the existence of the stable layers is proven, and much is being learned about them.<sup>30</sup>

If a solid substance such as ice is in contact with its vapor phase when the absolute temperature of the solid is below approximately 0.9 its absolute melting temperature  $T_o$  (for ice,  $T_o = 273 \text{ K}$ ), then the interface remains dry; that is, no liquid phase is present. However when the temperature rises to approximately  $0.9T_o$ , then the ice system's free energy is lowered if some of the ice melts to form a layer of water between the ice and the vapor. At first, this so-called premelted layer is only one or two molecules thick, but the layer thickens as the temperature rises, moving toward macroscopically large thickness at the melting temperature. This change from a two-phase to three-phase system is of course a consequence of the Universal System Happiness Rule, and the rule also requires that the thickness of the premelted layer be that which minimizes the overall free energy of the system. Note that the premelted layer is a stable layer, whereas a layer of supercooled bulk water is only metastable—tickle it slightly by introducing a large enough solid surface and it will freeze. Thus

30. Dash et al. (1995).

supercooled bulk water depends on the absence of a solid surface, whereas premelted water requires one.

In the premelting model of segregation ice formation, a first step in the approach is to predict how the thickness of a premelted layer and the pressure within it vary with temperature of a system. The system considered is one consisting of a premelted layer separating an ice wall and a wall of another substance such as soil. As with other models discussed above, the approach is to examine the free energy of the system to ascertain the conditions required for the system to be in equilibrium; that is, to possess minimum free energy. However, the premelting examination goes farther by including in the accounting the free energies of all the interfaces along with the bulk free energies of the ice and the water plus the free energy associated with intermolecular forces acting between all the molecules in the three layers: ice, water, and soil.

The intermolecular forces of concern are Van der Waals forces and electrostatic forces that might involve distributions of surface charge at interfaces and ions within the premelted layer if it contains impurities. These various forces decay with distance in different ways, so the predicted thickness of the premelted layer depends on which type of force is incorporated in the computation. The procedure is to fix the temperature and pressure and then see what thickness of the premelted layer allows the system to have the least free energy.<sup>31, 32, 33</sup> The process leads to an expression for the premelted layer's thickness  $d$ :

$$d = \lambda_v T_o^v / (T_o - T)^v \quad (5)$$

where  $T$  is the temperature,  $T_o$  is the freezing temperature of bulk water ( $0^\circ\text{C}$ ), and  $\lambda_v$  is a fixed coefficient. Equation 5 looks simple, but much is bound up in the coefficient  $\lambda_v$ . It depends on the heat of fusion of water ( $L$ ), on the density of ice  $\rho_{ice}$ , and on  $v$ , and  $v$ , in turn, is a number related to the fashion in which the intermolecular Van der Waals or electrostatic forces decay with distance. For Van der Waals forces,  $v = 3$  or  $4$ , and for electrostatic forces  $v = 3/2$  or  $2$ .

Also coming out of the examination of the free energy of the system is an expression relating the hydrodynamic pressure on water in the premelted

31. Wettlaufer et al. (1996).

32. Worster and Wettlaufer (1999).

33. Note to the mathematically inclined: The expression for free energy is expanded into a power series and then the first-order term is taken as the result.

layer ( $P_{water}$ ) to the pressure on the ice ( $P_{ice}$ ) and the temperature. The pressure in each of these phases is uniform, but the interactions between water and ice molecules create a difference between  $P_{ice}$  and  $P_{water}$ , and these interactions depend on the thickness of the premelted layer. The expression for the hydrodynamic pressure in the premelted layer ( $P_{water}$ ) is independent of the nature of the intermolecular interactions; that is, it does not depend on the numerical value of  $v$ . The pressure in the water is dependent only on the specific density of ice  $\rho_{ice}$ , the heat of fusion  $L$ , and the temperature  $T$ .<sup>34</sup>

$$P_{water} = P_{ice} - \rho_{ice}L(T_o - T)/T_o \quad (6)$$

Notice that this result differs from that given by Equation 2, derived from the Clausius-Clapeyron equation with  $P_{ice}$  held fixed. The appearance of the density of ice  $\rho_{ice}$  in Equation 6 but not in Equation 2 is the cause of the difference, and the consequence is that the predicted change of pressure on the water changes by approximately 11 Atm/ $^\circ\text{C}$  instead of the 12Atm/ $^\circ\text{C}$  predicted by fixing the pressure on the ice and solving for the pressure in the water using the Clausius-Clapeyron equation. Representing Equation 6 on Figures 3.10A and B are the lines labeled "Premelting Model." Curiously enough, the result is identical to that derived from the Clausius-Clapeyron equation with  $P_{water}$  instead of  $P_{ice}$  held fixed.

Notice from Equation 6 that the pressure on the ice  $P_{ice}$  is equal to the pressure on the water plus another pressure equal to  $\rho_{ice}L(T_o - T)/T_o$ . It is convenient to call this pressure the **thermomolecular pressure**  $P_T$ . In the premelting model it is assumed that the pressure on the ice in a freezing soil is initially fixed and equal to the overburden pressure. Fixing  $P_{ice}$  requires that  $P_{water}$  decrease and the thermomolecular pressure increase with decreasing temperature. Hence if there is a temperature gradient in the freezing soil,  $P_{water}$  and  $P_T$  vary with position, and the suction  $P_{ice} - P_{water}$  draws water toward the coldest part. Not only is water pulled to the freezing front from the unfrozen soil, it also travels via the premelted layer into the frozen soil. However, the thickness of the premelted layer declines rapidly with temperature, less rapidly if electrostatic intermolecular forces dominate rather than Van der Waals forces, but in both cases the change with temperature is profound.<sup>35, 36</sup> The thickness at  $-1^\circ\text{C}$  is 1/100 to 1/1000 that at  $-0.1^\circ\text{C}$ , and at  $-10^\circ\text{C}$  the thickness is 1/1000 to 1/1,000,000 that at  $-0.1^\circ\text{C}$ .

34. Dash et al. (1995); Wettlaufer et al. (1996); Worster and Wettlaufer (1999a).

35. Wettlaufer (1999a).

36. Wettlaufer (1999b).

Thus pulling water through the premelted layer is like pulling water through a funnel with extreme taper. Near the relatively warm freezing front the channel is wide and water flows relatively easily, but where the temperature is lower, the channel narrows and so greater suction is required to maintain a given flow. Because the suction changes much more slowly than the thickness of the premelted layer with falling temperature, water will tend to pile up somewhere behind the freezing front. This incoming water increases  $P_{\text{water}}$  locally, and that in turn drives  $P_{\text{ice}}$  upward locally, causing it to exceed the overburden pressure locally, rupturing the soil. The increase in  $P_{\text{water}}$  concurrently raises the free energy of the water above that of ice at the temperature existing and so the water converts to segregation ice. Ice will continue to form as long as enough water flows to the forming layer to maintain  $P_{\text{water}}$  high enough to force  $P_{\text{ice}}$  to exceed the overburden pressure. If enough desiccation occurs then  $P_{\text{water}}$  falls, dragging  $P_{\text{ice}}$  down below the overburden pressure to terminate the growth of the layer of segregation ice. The freezing front advances, and when conditions again become favorable a new layer of segregation ice forms at a different location.

The premelting model allows very high suctions and heaving pressures to develop since the premelted layer is stable down to temperatures near  $-30^{\circ}\text{C}$ , where the heaving pressure exceeds 300 Atm. That is equal to the pressure difference between the bottom and top of a water column 3400 m (11,200 ft) tall. That premelting can generate segregation ice has been demonstrated in the laboratory, in situations not involving curved interfaces so that capillary effects could be ruled out.<sup>37</sup>

### Summary discussion of the models of segregation ice formation

Common to all the models put forth to explain the formation of segregation ice is the idea that temperature-dependent suction (cryosuction) develops in a porous freezing soil and that this suction draws water into the soil. The transported water freezes somewhere within the soil, typically in the form of layers perpendicular to the imposed gradient in temperature, and these increase the volume of the soil. Coexistence of water and ice in the freezing soil requires that the pressure in the water be different from that in the ice, and that these two quantities  $P_{\text{water}}$  and  $P_{\text{ice}}$  be related in a way that depends on temperature, namely growing farther

apart as the temperature of the soil falls below  $0^{\circ}\text{C}$ , by 11 Atm/ $^{\circ}\text{C}$  in the premelting model, and by 12 Atm/ $^{\circ}\text{C}$  in the others that adopt a version of the Clausius-Clapeyron equation wherein the pressure on the ice is held fixed. Both predicted pressure differences (suctions)  $P_{\text{ice}} - P_{\text{water}}$  fit the data at near- $0^{\circ}\text{C}$  temperatures, but the suction predicted by the premelting model fits better at lower temperatures. Since in all cases the suction  $P_{\text{ice}} - P_{\text{water}}$  is constrained to a very narrow range, 11 Atm/ $^{\circ}\text{C}$  to 12 Atm/ $^{\circ}\text{C}$ , the exact value of this suction is not important. It probably varies with change in the rate of water inflow from nearby unfrozen soil, and its variation might be a factor helping to determine the thickness of segregation ice's layers. All models are compatible with the idea that the only way to stop the flow of water into the soil, and therefore to avoid increasing the volume of the soil-ice matrix, is for the overburden pressure to exceed  $P_{\text{ice}}$  everywhere in the freezing soil.

Recall that the shapes of soil freezing characteristic curves are very similar to, and can be predicted from, soil moisture characteristic curves. That fact suggests that the small-scale intermolecular forces which cause water to cling to soil so tenaciously are much the same whether the soil be frozen or unfrozen. Ground temperatures rarely if ever get down to  $-30^{\circ}\text{C}$  (approximately 0.9 times the absolute melting temperature of ice). Therefore, an interesting consequence of premelting is that the phenomenon guarantees that all or virtually all of the soil particles in the world's frozen ground, be they in the active layer or in permafrost, will have a coating of liquid water surrounding them. This coating provides an avenue along which water can travel whenever it is under suction.

### ■ Frost Heave

Most geocryologists use the term "frost heave" to mean the increase in volume undergone by a soil because of the segregation ice forming within it. By this strict definition, any increase of a soil's volume due to the freezing of its resident pore water is excluded, but that is usually of minor consequence anyway, never exceeding 9% and typically far less. Much greater expansions do occur, and so it is to explain how these expansions come about and to predict how large they will be that geocryologists have developed the various models to explain the formation of segregation ice and its legacy, frost heave.

Such a delicate balance must occur among so many contributing factors that accurate prediction of frost heave is not yet within reach. Reviewing the

37. Wettlaufer (1999a).

key features of the models brings out several important requirements for the formation of segregation ice.

Segregation ice will form if:

1. Subzero temperature gradients and soil pore sizes permit cryosuction to develop, and
2. a source of water is available, and
3. the permeability (effective hydraulic conductivity) allows water transport to the site where the ice can form, and
4. the thermal conductivity and temperature gradient permit a balancing of heat energy in and out (including the latent heat of fusion)<sup>38</sup> sufficient to stall the freezing front or allow advance at a rate no faster than the segregation ice accumulates, and
5. the suction on the water in the soil is such that the conversion of water to ice allows the ice pressure to equal or exceed the pressure from the weight of overlying material.

The ubiquitous growth of piprake ice shows that these five necessary conditions are easily met at the surface of the ground when the air temperature falls only slightly below 0°C. The piprakes grow where the ground is wet and loose, so only minor cryosuction is required to bring water to the freezing front. Here, segregation ice readily accumulates because the lack of overburden minimizes downward pressure and permits easy removal to the air of the freezing water's heat of fusion.

Below the surface of the ground, things become more complicated. The observed development of multiple layers of segregation ice in freezing soil, both in the laboratory and in natural settings, hints strongly that the system cycles through values that generate the five necessary conditions and then cause them to hold no longer. In a given soil the two variables that can change are the temperature and the availability of water. Segregation ice forms in a soil having favorable characteristics (mainly pore size and effective hydraulic conductivity) when the temperature-driven suction delivers the proper amount of water to and beyond the freezing front.

38. The thermal conductivity and the effective heat capacity change abruptly at the freezing point because both depend on the water content, and that in turn depends on temperature. Ice is more conducting than water, so as more water converts to ice in a freezing soil its thermal conductivity increases. The effective heat capacity in the temperature range -3°C to 0°C is very high because in this range much change of phase takes place and the water's latent heat of fusion is so large.

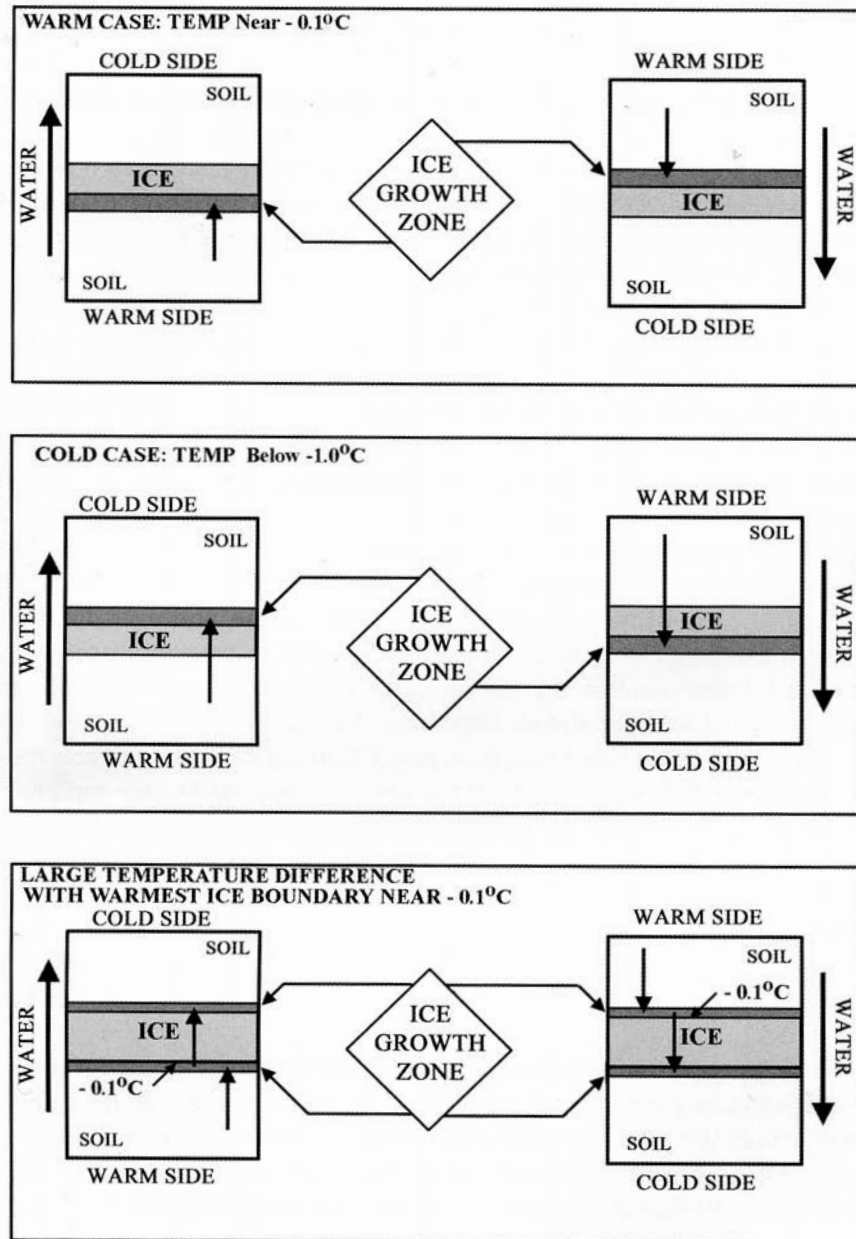
The segregation ice can continue to form as long as the water supply is adequate, but the necessary suction tends to deplete water from the surroundings. If the supply of water becomes inadequate, the growth of segregation ice terminates and the freezing front advances. Later, and at some distance beyond, continuing temperature decline will increase suction and bring enough water to begin forming a new layer of segregation ice.

Multiple layers of segregation ice tend to increase in thickness with depth, as was shown in Figure 3.12. As the active layer freezes from the top down, the temperature changes more slowly with depth, and that gives the existing combination of suction, effective hydraulic conductivity and water supply a longer opportunity to maintain proper conditions for continuing growth of a segregation ice layer.

Gradations and boundaries in the soil help determine the location and shape of bodies of segregation ice. Thick accumulations are often found near boundaries between zones of fine-grained material and more permeable coarse-grained material that favors movement of water to the site where segregation ice is forming. Compositional variation of any sort that affects the effective hydraulic conductivity, thermal conductivity, or heat capacity can influence the location, orientation, and shape of ice layers. And any ice layer that forms tends to foster additional accumulation of segregation ice because the boundaries of the initial ice layer generally are locations of abrupt changes in effective hydraulic conductivity. That segregation ice does tend to build up at these sites is verified by laboratory observations of simultaneous enlargement of multiple ice layers within frozen soil.<sup>39</sup>

Recall from viewing Figure 3.15 that at temperatures barely below 0°C the effective hydraulic conductivity of ice is less than that of saturated frozen soil but that at lower temperature the effective hydraulic conductivity of the ice exceeds that of the soil. Thus, depending on the temperature, water molecules might encounter either increase or decrease in effective hydraulic conductivity as they move through frozen soil into an ice layer. If they encounter a decrease they should tend to pile up at the boundary, causing a growth of segregation ice there. If the temperature at the boundary causes the effective hydraulic conductivity of the ice to be greater than the hydraulic conductivity of the soil, then the water will tend to pile up at the exit boundary of the ice body. **Figure 3.17** indicates schematically on which sides of ice layers segregation ice should tend to accumulate during

39. Williams and Smith (1989) pp 232-33.



**Figure 3.17** Top panel: When the soil and ice temperature is barely below freezing, and the ground surface is colder than at depth, water moves upward to create new ice on the bottom of a pre-existing ice layer (as at left), but if the temperature decreases downward, water moves down and new ice forms at the top of an older ice layer. Center panel: When the soil and ice temperature is well below freezing, just the opposite occurs: new ice accumulates at top during the fall and at the bottom during the spring. Bottom panel: If a large temperature difference appears across a pre-existing ice layer, then ice can form on both sides of it, fall or spring.

freeze and thaw periods when the temperature at their boundaries is just below and well below freezing.

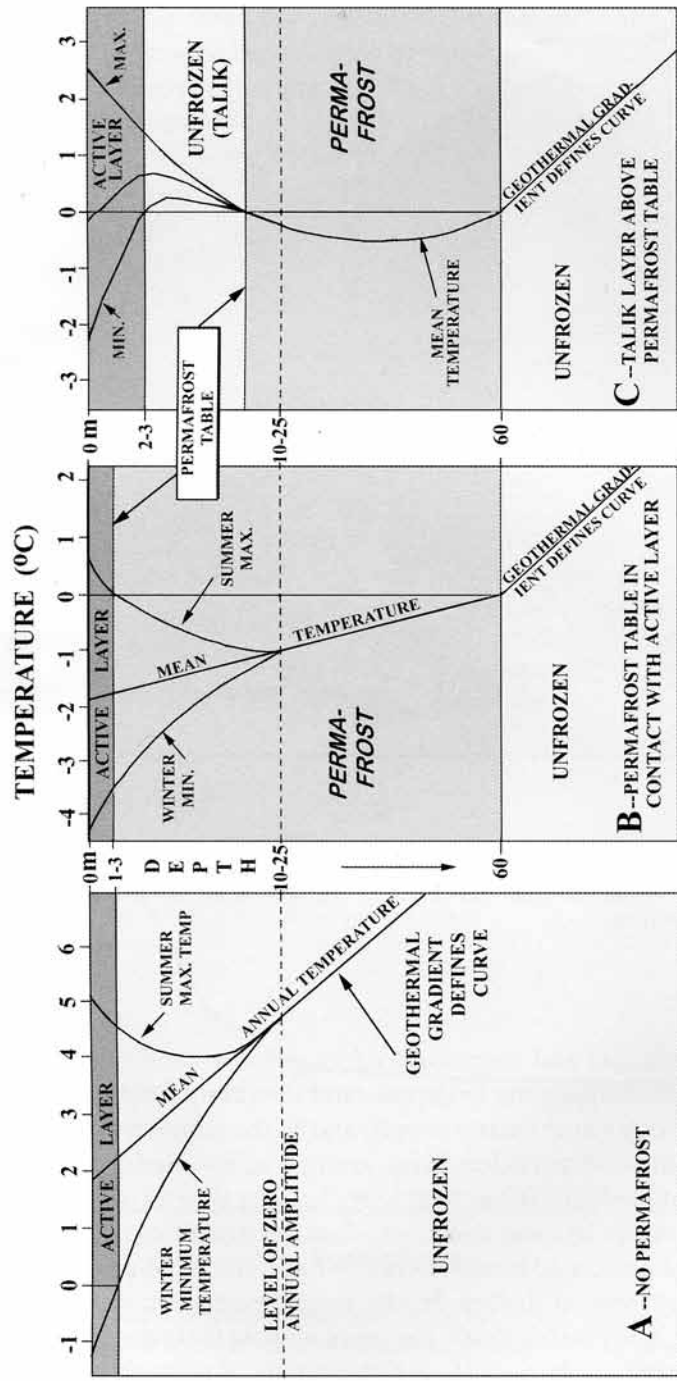
That part of the permafrost layer lying above the level of zero amplitude (see **Figure 3.18**) in the annual temperature variation undergoes temperature fluctuations ranging between  $0^{\circ}\text{C}$  (at the very top of the permafrost) down to the minimum temperature of the permafrost, about  $-10^{\circ}\text{C}$  in northern Alaska and  $-13^{\circ}\text{C}$  in northeastern Siberia. At or somewhere below the level of zero amplitude, typically at depth 10 to 25 m, the normal upward geothermal heat flow causes increasing temperature (the geothermal gradient ranging from  $1^{\circ}\text{C}/20\text{ m}$  to  $1^{\circ}\text{C}/60\text{ m}$  [ $1^{\circ}\text{F}/40\text{ ft}$  to  $1^{\circ}\text{F}/110\text{ ft}$ ]). Thus, wherever the top of the permafrost lies above the zero-amplitude level, a continuing growth of ice layers can occur. In winter cryosuction pulls water upward through the permafrost and in summer it helps drive the water down from the thawing active layer above. So, as in a tidal channel, though much more slowly, water sweeps back and forth through the top of the permafrost with little volume, and over distances perhaps measured only in centimeters. Still, over the course of years the flow can cause—millimeter by millimeter—growing accumulations to any horizontal ice layer already in place.

As long as the ice layers are thin, at any time they are likely to accumulate segregation ice on one side or the other, as indicated in the top panel of **Figure 3.17**, but the accumulation might shift from one side to the other as the temperature changes from near zero to somewhat colder (compare left-hand or right-hand parts of the top and center panels in **Figure 3.17**). Then if the ice layers become thick enough for the temperature gradient to produce substantial temperature differences across the layers, simultaneous accumulation can occur, as in the bottom panel of **Figure 3.17**.

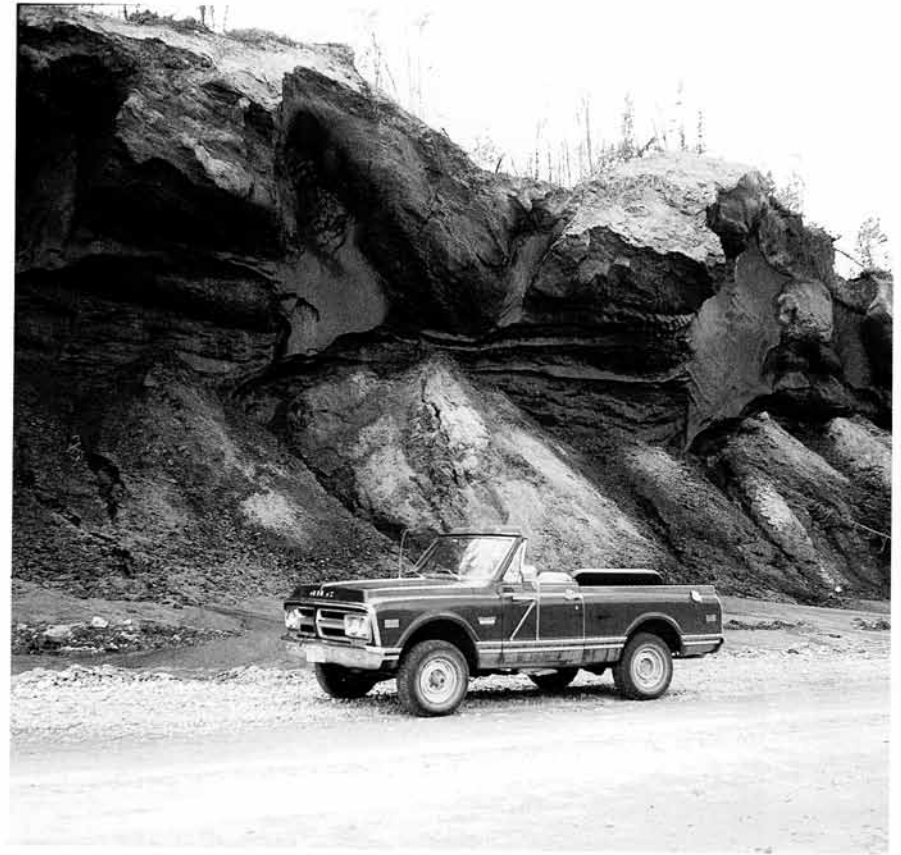
The long-term consequence of the *cryogenic* pumping of water through frozen ground is an enrichment of the ice content of the uppermost part of the permafrost layer and the formation of layers and lenses of pure ice near the top of the permafrost, either singly or in association with other forms of ground ice, such as shown in **Figure 3.19**. Geocryologists sometimes refer to the new segregation ice forming within previously frozen ground or that accumulating to pre-existing ice layers as *aggradational ice*. In some instances it becomes very thick, commonly in excess of 10 m in the Mackenzie delta of northwestern Canada, and sometimes more than 30 m.

Based on field observations, laboratory experiments, and interpretations guided by the models of ice segregation, it is clear that key factors





**Figure 3.18** Schematic plots of ground temperature where : A—no permafrost is present below the active layer, B—the permafrost table is in contact with the bottom of the active layer, and C—the nonequilibrium situation where climatic change or other cause has melted the upper portion of the permafrost to leave an unfrozen layer (talik) between the bottom of the active layer and the permafrost table.

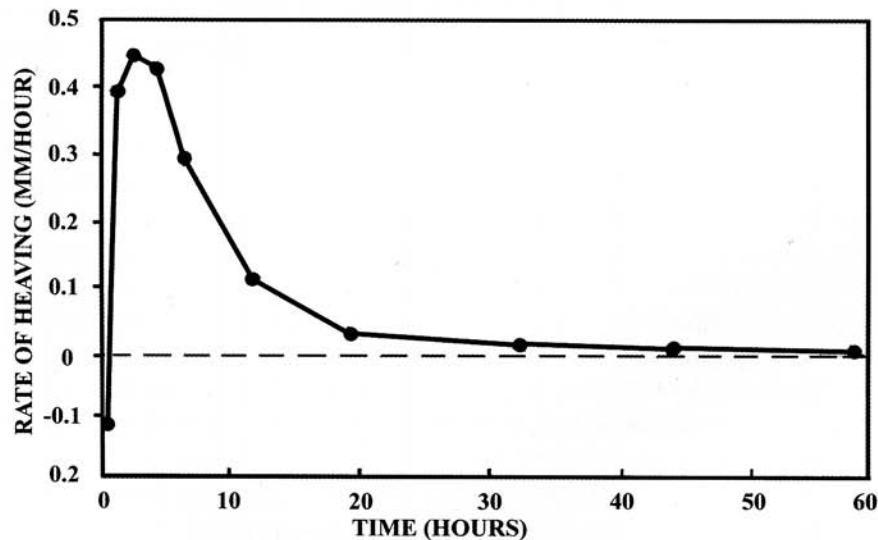


**Figure 3.19** Ice wedges and multiple layers of segregation ice in a roadcut near Fox, Alaska. See also Plate 6.

determining the rate and magnitude of ice growth—and hence of frost heave—are water supply, soil grain size, and the thermal regime. The low permeability of clay limits water supply, and so the largest frost heaving is in soil with silt-sized particles. Large-grained soils—sands, gravels, and also peats—support little if any frost heave because they do not allow large suctions to develop. In most instances when the temperature falls rapidly (and hence the associated heat flow rate is high) the rapid advance of the freezing front, at least at shallow depths, limits segregation ice to thin layers, and so the heave also is small. However, even in that situation of rapid change, the heave can be very large if the supply of water is copious and

the soil permeable. Then enough water may migrate to and freeze near the freezing front to stall the advance of the front by giving up its heat of fusion and thereby forming a very thick ice layer that rapidly heaves the ground surface upward.

When the ground first freezes, the rate of heaving due to forming segregation ice is small, but it increases to a maximum when conditions cause the freezing front to become nearly stationary. Thereafter, as shown in **Figure 3.20**, the rate of primary heaving falls away as the rate of heat removal and the rate of water transport to the front decline. This fast initial ice formation and heaving presumably occur at temperatures only slightly below 0°C since they are restricted to a location at or very near the freezing front. The near-zero temperature there limits the ice pressure  $P_{ice}$  (the heaving pressure) because the difference in ice and water pressure  $P_{ice} - P_{water}$  is also near zero, and the water pressure  $P_{water}$  will always have some finite value, typically one less than the pressure from the weight of overlying soil. If freezing is occurring at temperature  $-0.1^\circ\text{C}$ ,  $P_{ice} - P_{water} = 1.1 - 1.2$  Atm, and if the water pressure is 1 Atm or less then the heaving pressure cannot exceed 2.2–2.4 Atm. That is the pressure exerted by a soil overburden layer approximately 6 to 10 meters thick (approximately 10 metric tons/m<sup>2</sup> or 2000 lbs/ft<sup>2</sup>). Thus, the



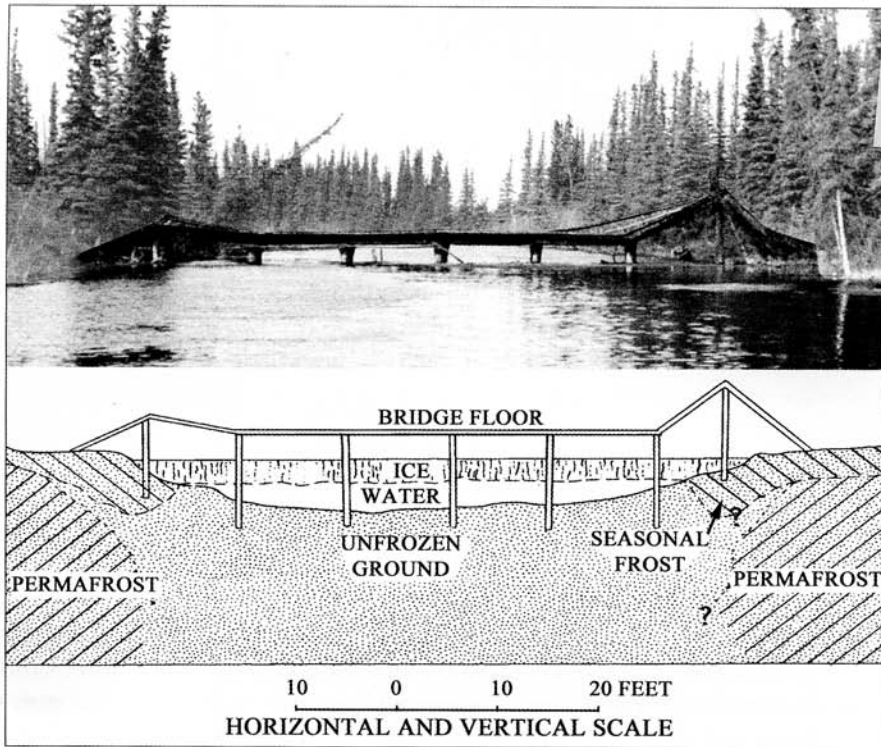
**Figure 3.20** When a wet, fine-grained soil freezes, the rate of heaving climbs quickly to a peak and then declines. Redrawn after a diagram by Garand (1981) reproduced as Figure 6.8 in Smith (1985a).

initial heaving—sometimes called the **primary frost heaving**—can readily lift the uppermost layer of soil and rather substantial objects placed upon it such as houses, roadways, and railroads.

If the overburden load is great enough it can stop the forming of segregation ice, and hence, frost heave. Then any further freezing forces the unfrozen water into the ground below, and if the soil there is very fine-grained the required pressure is very large—hundreds of atmospheres. Pressure-measuring devices placed in freezing soils have shown that such large pressures do occur, and the fact that the formation of segregation ice obviously has taken place at depths where the overburden load is in the vicinity of 20 Atm adds further verification. The initial (primary) frost heaving—that developing at or very near the freezing front where the temperature is near 0°C—is incapable of producing large pressures, and so in explaining these the rigid-ice and premelting models of ice segregation come into their own. They allow for ice formation behind the freezing front where the temperature can be several degrees below freezing and the ice pressure high. In keeping with the identification of the near-0°C ice segregation as causing primary heaving, some geocryologists identify the slower but more powerful expansion as **secondary frost heaving**. Whereas primary heaving probably occurs mainly within the annual freezing of the active layer, secondary heaving can extend down into permafrost to depths of 30 or more meters. The amount of segregation ice (and heaving) tends to decline with depth, and so most of the overall heaving is usually within the upper few meters of the permafrost layer.

**Figure 3.21** presents a classic photograph, well illustrating both how large frost heave can be and how much it depends on local conditions, especially water supply. Although frost heaving of a few centimeters per year is typical, one set of bridge pilings in Siberia heaved 0.6 m (2 ft) in one year, and the piles at the right-hand side of the bridge in **Figure 3.20** heaved 3.3 m (11 ft) in one year.<sup>40</sup> Railroads are particularly susceptible to problems with frost heaving since the rails need to be kept fairly level in order to avoid train cars becoming decoupled or derailed. The Alaska Railroad has found it necessary to expend funds each year for shimming up track, trimming off uplifted pilings, and replacing them. One cold day during the winter of 1950–51 the rear part of a passenger train came uncoupled while moving over a frost-heaved bridge just outside Fairbanks. The passengers sat cooling

40. Péwé (1982).



**Figure 3.21** Frost-heaved bridge spanning the outlet of Clearwater Lake near Big Delta, Alaska. Compiled from a photograph taken by M. F. Meiser in August 1951 and a drawing by Troy L. Péwé, reproduced as Figures 30 and 31 by Oscar J. Ferrians et al. (1969).

their heels and all other parts of their anatomies for several hours while the engineer drove on the 80 km to Nenana, discovered his problem, and back-tracked to the bridge raised by segregation ice.

### ■ Other Kinds of Ice Found in the Ground

Frost heaving is generally considered to be the result of volume change wrought by segregation ice, that resulting from the freezing of water transported by cryogenic processes. Several other kinds of ice form in the soil, and also may cause uplift of the ground surface. One descriptive scheme geocryologists use has five categories: 1) segregation ice, 2) pore

ice, 3) ice-wedge ice or foliated ice, 4) pingo ice, also called intrusive ice or injection ice, and 5) buried ice.<sup>41</sup> These various forms of ground ice may develop in association and even grade into one another, so sometimes it may be difficult to determine which kind is involved in a particular deposit.

### Pore ice

*Pore ice* is that formed from water already in place within the soil. Any expansion of the soil that may result because of ice's 9% greater volume than water is in addition to cryogenic frost heave. Even if the freezing of pore ice causes little or no overall expansion of the soil, it may temporarily or permanently modify the soil by squeezing or otherwise altering the soil particles. Pore ice comprises a substantial part of the upper few meters of permafrost, and the ice in permafrost below depths of 10 m is mostly pore ice.

### Ice-wedge ice or foliated ice

*Ice-wedge* ice is that formed in permafrost from surface water running down into small (1 to 3 cm) cracks opened by thermal contraction of the soil. The process is repetitive, progressing over the course of many summer thaws and winter freezes, so ice-wedge ice typically is highly foliated, each layer representing a freeze-thaw episode.

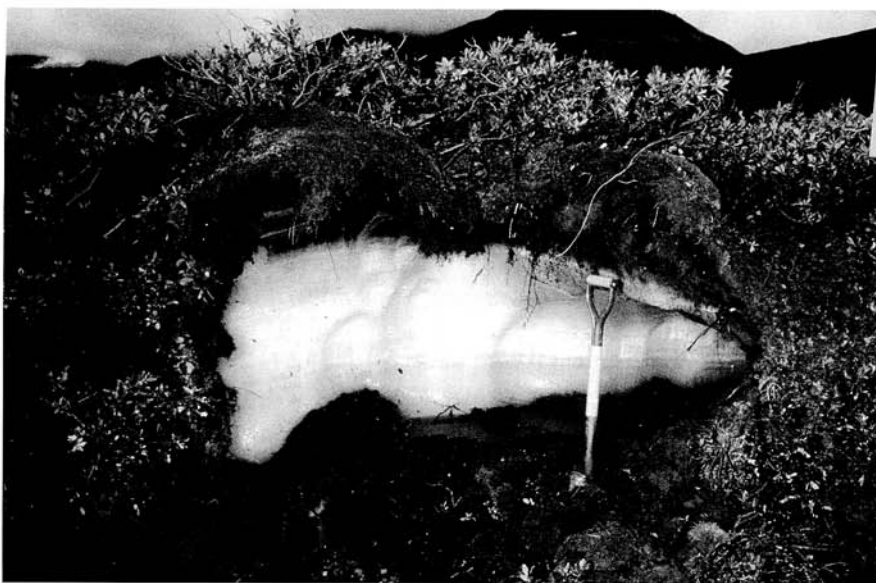
### Pingo ice

*Pingo* ice (also called intrusive ice or injection ice) is that which forms from water under hydrostatic pressure in permafrost areas. It may be milky from entrapped air bubbles but is typically clear or nearly clear and is often found in planoconvex masses that, when forming, have lifted overlying soil upwards to create minor mound structures such as shown in **Figure 3.22** or the much larger mounds called pingos, discussed in Chapter 4.

### Buried ice

Depositional events such as avalanches, mudslides, or windstorms sometimes bury sea ice, lake ice, river ice, overflow ice (aufeis), glacier ice, and

41. Péwé (1975b).



**Figure 3.22** A layer of pure ice pushing up the moss in northern Alaska. Photograph by Richard Veazey. See also Plate 7.

recrystallized snow. This buried ice does not directly lead to frost heaving, but it might contribute to it by affecting where subsequent segregation (or aggradation) ice forms.

### Another icelike substance: gas hydrates (clathrates)

**Gas hydrates** are icelike crystalline solids formed from methane gas and water molecules. In one kind of gas hydrate (Structure I), 46 water molecules join to form a large lattice structure in which some methane molecules ( $\text{CH}_4$ ) participate, and in another (Structure II) 136 water molecules form a lattice in which ethane molecules ( $\text{C}_2\text{H}_6$ ) can join. The maximum participation allows one methane molecule in Structure I for every  $5\frac{3}{4}$  water molecules, and one ethane molecule for every 17 water molecules in Structure II. This packing brings the methane and ethane molecules into a volume 200 times smaller than they would occupy if by themselves under the same conditions of pressure and temperature.<sup>42</sup> It is as if the gas molecules were locked into a water molecule pressure vessel. However, the vessel

42. Ehlig-Economides (1981).

is stable only in a limited range of pressure and temperature conditions. At pressure 20 Atm the gas hydrate is stable only at temperatures below  $-10^\circ\text{C}$ , at 10 Atm it is stable only below  $-30^\circ\text{C}$ , and at 1 Atm stable only below about  $-80^\circ\text{C}$ . The necessary conditions for stability exist in the coldest of permafrost areas at depths below 100 m and perhaps down to a depth of approximately 1000 m where heat flow from the earth's interior raises the temperature above the maximum for stability at the pressures involved. Conditions for stability should also pertain beneath the oceans, at depths near 250 m in cold arctic waters, and at depths below 400 m in warmer waters.<sup>43</sup>

Drilling into gas hydrates can be hazardous because the release of pressure on the hydrates can cause an almost explosive expansion that can blow drilling equipment back up the hole. However, the hydrate deposits in and below permafrost also represent a potentially huge energy resource.<sup>44</sup>

## ■ Chapter Summary

Various kinds of suction—vacuum suction, capillary suction, osmotic suction, and cryosuction—act to move water from one place to another because the movement lowers the overall free energy of the system. Cryosuction, the suction that develops in the pores or on the walls of fine-grained materials at subfreezing temperatures, is a powerful water-moving force. A familiar example of its operation is the freezer burn that occurs when a poorly wrapped slab of meat is placed in a freezer. The cryosuction pulls the water out of the meat and converts it to ice crystals on the meat's surface or elsewhere in the freezer compartment. The same thing happens to the soil near the surface of the earth when the air temperature goes down below  $0^\circ\text{C}$ . Cryosuction in the soil pulls water toward the cold surface where it often forms ice crystals called pipkrakes or needle ice. Cryogenic suction depends on both temperature and soil characteristics, particularly the size of the particles in the soil. Because of water's abnormal characteristics, cryosuction increases dramatically as the temperature falls, by 11 to 12 Atm per centigrade degree below  $0^\circ\text{C}$ , and in a fine-grained soil, it can exceed 100 Atm. An important consequence of cryosuction is the formation of microscopic to thick segregated layers of pure ice within fine-grained soils when adequate water is available and the

43. Williams and Smith (1989) pp 46–48.

44. Ehlig-Economides (1981).

thermal regime is favorable. The formation of this segregation ice can increase the volume of a soil by 50% or more, and in the process heave the ground surface upward. Silty soils are the most prone to frost heave, and in them the heaving is powerful enough to lift the uppermost layer of soil and structures built thereon. Segregation ice is widespread, and it forms in both seasonally and perennially frozen ground. Of the mechanisms responsible for forming the various types of ice found in the ground—segregation ice, pore ice, foliated ice-wedge ice, pingo ice, and buried ice—the one creating segregation ice is the most complex, and also the least understood.