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Temporal Weather Impacts Upon Exterior Intrusion Detection Systems

Charles C. Ryerson and Lindamae Peck

December 1995

Abstract

Fundamentally, an electronic exterior intrusion detection system (IDS) cannot directly detect intruders; it detects a variation in the condition being monitored, extracts characteristics of that variation, and assesses whether such a variation probably is caused by an intruder. However, exterior IDSs do not operate in a benign natural environment. Their environment is constantly changing as a result of solar-driven energy and moisture fluxes that create the weather. These weather changes often cause variations in the conditions being monitored by IDSs. The challenge, therefore, is to recognize how and when IDSs are responding to some change in their natural environment, rather than to intruders. This report is a technical analysis of causes of weather-driven temporal changes in the environment that impact the operational efficiency of IDSs. The report is intended to assist security designers in selecting suitable IDSs for a site and to assist security managers in operating IDSs at the required level of reliability. This is accomplished by identifying temporal variations in weather that are sufficiently general to be identified as patterns, and by identifying how different IDS phenomenologies respond to these patterns. The result is an understanding of how weather conditions influence the ability of types of IDSs to detect reliably activities representative of an intruder while successfully discriminating against weather-created conditions within a detection zone. The main body of the report is organized by temporal scale: diurnal, quasi-periodic, and seasonal. Within each temporal scale, weather processes common at that scale are explained Topics covered include air and soil temperature, soil moisture, precipitation, snow cover, winds, fog, storms, urban and topographic effects, vegetation effects, and solar radiation. Appendixes include a summary of weather impacts on IDSs, extensive indexes by IDS and weather topics, cloud and fog illustrations, explanations of the relationship of weather patterns to forecasts, suggestions about how to perform environmental site surveys, and sources of additional weather information.

Cover: Drifting snow burying a 2-m-high security fence.

For conversion of SI units to non-SI units of measurement consult ASTM Standard E380-93, *Standard Practice for Use of the International System of Units,* published by the American Society for Testing and Materials, 1916 Race St., Philadelphia, Pa. 19103.

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OFFICE OF THE CHIEF OF ENGINEERS
and
U.S. AIR FORCE ELECTRONIC SECURITY AND COMMUNICATIONS CENTER

PREFACE

This report was prepared by Dr. Charles C. Ryerson, Research Physical Scientist, Snow and Ice Division, and Dr. Lindamae Peck, Geophysicist, Geophysical Sciences Division, Research and Engineering Directorate, of the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL).

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CHARLES C. RYERSON AND LINDAMAE PECK

INTRODUCTION

Ambient weather at any given site is the net result of regional energy exchanges modified by local topography. When assessing whether a change in operating environment has influenced an electronic sensor system, both short-term and long-term temporal (time) variations in weather must be considered. This report is a technical analysis of the causes of diurnal (daily), quasi-periodic, and seasonal weather changes that affect the operational efficiency of exterior intrusion detection systems (IDSs). Temporal variations in weather that are sufficiently general and repeatable to be considered as weather patterns are identified and explained. The report is intended to assist security designers in selecting IDSs suitable to a site and to assist security managers in operating those IDSs at the required level of reliability.

This report may be used in several ways. Readers who want a full understanding of temporal weather patterns and processes and their impact upon IDS operations should read the full text of the report. Those who wish a brief synopsis of weather effects upon IDSs should read Appendix A, Environmental Influences on Exterior IDSs. Security designers and IDS managers who wish to find all weather impacts upon a specific type of IDS should refer to Appendix B, Indexing from an IDS Perspective, to locate the appropriate explanations in the main report. Readers wishing to determine the weather conditions associated with a specific change in site environment, such as soil freezing, should refer to Appendix C, Indexing from a Weather Perspective, to locate the proper explanations. Those who wish to determine how to translate everyday weather forecasts to expected local weather conditions explained in the report should refer to Appendix E, Relationship of Weather Forecasts to Weather Patterns. Appendix F, Environmental Site Surveys of IDS Zones, explains how to map locations of weather-related impacts at an IDS installation. Finally, additional sources of weather information, such as forecast information and consultants, may be found in Appendix G, Weather and Climate Information Sources.

Weather-related damage to IDSs is considered only marginally in this report. Equipment failure is distinct from environmentally dependent but recoverable changes in detection capability. For example, temperature-related changes in the stiffness of a security fence are relevant, because associated differences in the motion of the fence under dynamic loading by a climbing intruder may cause the detection capability of a fence-mounted IDS to vary. However, low-temperature embrittlement, such as may lead to structural failure of IDS hardware or fences, is not.

Statements regarding changes in an IDS's detection capability implicitly relate to a given, constant alarm threshold. Some or all of the instances of reduced probability of detection caused by a change in site conditions might be compensated for by lowering the alarm threshold of the IDS. The question, then, is whether security personnel would be aware that a change in IDS sensitivity was needed to maintain a required probability of detection. For this report, the impact of the environment on IDS reliability is expressed as a change in detection capability (because the report is unclassified, the actual change in probability of detection is unspecified), with the assumption that the alarm threshold of the IDS has not been adjusted in compensation. Cited changes in detection capability are based on controlled intrusions and long-term monitoring of IDSs at SOROIDS (SOuth ROyalton Intrusion Detection Systems), the CRREL IDS research facility in South Royalton, Vermont, and pertain to properly installed and well-maintained IDSs. Weather-related changes in detection capability may well be different for poorly installed or poorly maintained IDSs.

Background

Variations in the detection capability of exterior intrusion detection systems as the result of changes in their operating environment have been investigated through a collaboration between CRREL and the Air Force Electronic Systems and Communications Center. The outcome is an understanding of how certain site conditions influence the ability of types of IDSs to reliably detect activities representative of an intruder while successfully discriminating against other causes of disturbances within a detection zone. Ideally, an IDS would have a high probability of detection and a low nuisance alarm rate under the full range of operating conditions encountered at a site. This is probably an unrealistic expectation except at sites with extremely low variation in weather. Generally, it is necessary to maintain an awareness of changing site conditions, to assess their impact on IDS detection capability, and to take compensatory measures as required.

Depending on the design of an IDS, an intruder-caused disturbance that occurs slowly enough or that does not repeat frequently enough (in a specified time period) will not result in an alarm no matter how strong or extensive that disturbance is. It may be that the disturbance is not sensed by the IDS because of a mismatch between the temporal variability of the disturbance and the response characteristics of the IDS's sensor (e.g., thermal detector, microwave radar, vibration transducer, etc.). Or it may be that, in an attempt to exclude sensor response to non-intruder activity, the IDS only acts upon sensor output that repeats a certain number of times within a certain time period. In either case, by design, time is a factor in whether an IDS responds to a change in its operating environment. Relatively slow changes of large significance, such as the freezing of soil, may not be too detrimental to IDS performance while they are occurring, but the contrast in conditions (unfrozen vs. frozen soil) may cause great differences in detection capability. Conversely, it may be the variability of a condition, rather than its severity, that determines its impact on an IDS. For example, a steady, strong wind or constant heavy rainfall may be less troublesome to a particular IDS than either wind gusts or rapid changes in precipitation rate.

The most direct way of determining when site conditions have adversely affected the performance of an IDS is to monitor that IDS's proximity-to-alarm status. This is a method of knowing exactly what margin exists between the current excitation of the IDS and its alarm threshold, and therefore if the IDS is presently prone to environmentally caused nuisance alarms or if it is too insensitive to the level of disturbances typically caused by an intruder.

A less absolute method of determining IDS–environment interaction is to monitor the IDS's operating conditions. As discussed in this report, it may be difficult to characterize adequately the conditions pertaining to a particular IDS's perimeter zone if the perimeter is so large and the variation in terrain and exposure so significant that environmental conditions differ substantially along the perimeter.

The least specific approach to determining when the detection capability of an IDS has been affected by changes in its operating conditions is to extrapolate from a general knowledge of weather patterns and resultant changes in near-surface conditions. Although the accuracy of this approach is less than that obtained through real-time monitoring of IDS proximity-to-alarm or site conditions, it often is the only information available, especially at a temporary site or during IDS design and installation. It is also relevant when employing the particular brands of IDSs that have been designed to self-adapt to changes in their operating environment; automatic adjustments to the alarm criteria of such IDSs occur independently of action by security personnel. For these reasons, weather patterns—their causes and geographic variability—in daily, intermediate, and seasonal time scales, and weather-related changes in near-surface conditions—the operating environment of IDSs—are compiled and explained in this report. These changes are related to their impact on the detection capability of various classes of IDSs. Therefore, the specific goals of this report are to identify the IDS meteorological environment and to explain how it fluctuates and may affect the operation of IDSs over diurnal, quasi-periodic, and seasonal time scales.

Overview

Electronic intrusion detection systems operate in a wide diversity of spatial and temporal weather scales. They are located, at the macro-spatial scale, in meteorological regimes ranging from frigid polar and alpine locations to tropical wet or arid regimes. Within these macrometeorological (continental-scale) regimes, they operate in diverse environments that are dependent on local terrain and elevation. Individual IDSs also experience a multitude of mesometeorological (covering thousands of square kilometers), as well as micrometeorological, conditions that, in composite, comprise the local meteorology of the IDS site. The micrometeorological regimes, usually a few hundred square meters to less than a meter in area, may be caused by security systems themselves, such as shadows cast by perimeter fences or walls, or by minor differences in soil compaction and ground cover due to a recent excavation for the installation of sensor cable, or even by the changing color of grass along a sidewalk as it wears due to foot traffic by guard patrols.

Weather experienced by IDSs also varies widely over several temporal scales. On a diurnal scale, most meteorological variables change as solar energy fluctuates with the sun's position in the sky. Weather conditions continue to change through the night, without the sun's presence, as a result of nonsolar energy exchanges.

Some IDSs have a diurnal variation in reliability. Both passive infrared and video-motion IDSs tend to have fewer nighttime nuisance alarms (excluding ones caused by animals) than daytime alarms. For the passive infrared IDS, this is because the surface temperature of the ground cover or pavement in the IDS's detection zone is more uniform at night, when solar heating is not occurring (Peck 1993b). At night, video-motion IDSs require artificial lighting, which provides more uniform and stable illumination than sunlight.

The magnitude of diurnal change observed on any given day is also a function of quasi-periodic weather changes caused by the passage of synoptic-scale (thousands of square kilometers in area) storm systems. These quasi-periodic systems, operating typically on a scale of days, but occasionally hours to weeks, deliver their own weather signatures and also modify or enhance diurnal patterns depending upon the type, size, and strength of the synoptic system in place. Therefore, quasi-periodic synoptic weather systems, with their own unique patterns, superimpose themselves upon diurnal patterns to either enhance or suppress diurnal changes.

Seasonal weather changes are the longest period fluctuations affecting IDSs. Caused by the revolution of the earth in its solar orbit, change occurs on the scale of months, repeating annually. Seasonal change is superimposed upon the quasi-periodic and diurnal fluctuations that occur at all IDS sites, further complicating the weather conditions and the magnitude of their changes at any time. Whereas diurnal and seasonal changes have a relatively high degree of repeatability, quasi-periodic fluctuations introduce both temporal and magnitude variabilities that are more difficult to anticipate and adequately compensate for in advance.

UTILITY OF STANDARD WEATHER REPORTS

In general, "weather is the condition, or state, of the atmosphere at a given place and time" (Ahrens 1982). This means, effectively, that the state of the atmosphere, as indicated by the magnitude of variables such as temperature, wind speed, pressure, and precipitation, is constantly changing with place and with time. Weather observers concentrate on changes in weather with time and location to provide current conditions for aviators, for example, but also to aid weather forecasting. As a result, observers are usually concerned with a limited number of easily measured weather variables that are ready indicators of large-scale, usually quasi-periodic weather changes. These measurements are typically made with standardized instruments in standard exposures at standard times and are intended to avoid unique local micrometeorological influences. Minimizing local effects allows ready comparison of weather among stations separated by large distances, allowing large patterns to become visible. Accordingly, the primary goal of weather forecasters is often not to maximize observation and prediction of highly localized and diurnal effects, though these are often incorporated into public and military forecasts, but to maximize their ability to recognize regional quasi-periodic weather events. This limits the usefulness of standard weather observations for anticipating weather-related changes in IDS performance.

IDSs do not experience weather conditions identical to those observed by instruments for weather forecasting, even when the instruments may be located on a facility protected by the IDSs. First, weather instruments are intentionally exposed in a standardized manner to avoid representing micrometeorological conditions that may not be representative of the region. Second, weather instruments may not report factors that affect IDSs. For example, rain gauges report only a net effect, such as rainfall quantity (mm) or rate (mm hr^{-1}), although the size of the drops is relevant to IDSs affected by raindrop impact or by transmission loss in rain. Third, the distribution of weather instruments will be much more limited than the exposure of IDSs that circumscribe a perimeter of many kilometers. Even if weather instruments were fortuitously collocated with an IDS zone, the information obtained would not necessarily assist in determining environmentally related changes in IDS detection capability. Weather variables measured for aviators and forecasters are often not those that most affect an IDS. In fact, variables affecting an IDS may often not be considered meteorological, but more properly climatological, because they are a result of interaction of the weather with the earth's surface. This is consistent with the general definition of weather given earlier, as the state of the "atmosphere." Weather forecasters are primarily concerned with the atmosphere rather than the interactions of the atmosphere and the earth's surface.

IDSs experience weather "at the earth's surface." As a result, they often experience weather conditions that are different from those observed by meteorologists, and the variables that affect IDSs are not usually measured by weather observers.

DIURNAL WEATHER EFFECTS

Diurnal weather is driven primarily by the daily solar cycle, which is determined by the path of the sun through the sky. Though modified by periodic changes in regional weather and by the seasons, daily weather at every location exhibits generally repeatable patterns. The following discussions identify those daily-changing weather conditions that most affect IDS operations and explain how and why they change and their relationship to other conditions.

Energy budget

Without the sun there would be little varability in weather. Weather is driven by the differential availability of heat energy throughout the atmosphere. The atmosphere is essentially a heat engine, responding to heat excesses in some locations and deficits in others. Since weather is the state of the atmosphere at a given place and time, it expresses the condition of the gases that comprise the atmosphere. The density of atmospheric gases is controlled by energy through the universal gas laws. Warming of unconfined gases makes them expand in volume and decrease in density. Cooling causes them to contract volumetrically and increase in density. Cool, dense air sinks, whereas warmer, less dense air in the same vicinity rises. This vertical motion, or convection, eventually creates advection, or wind, as air masses flow laterally to replace the risen air. On the smallest scale, this is responsible for local winds, but on the large scale it is the cause of general global circulation.

Energy is transported throughout the atmosphere by four mechanisms; radiation, conduction, convection, and phase changes (principally evaporation and condensation). Among the four transport methods, radiation is solely responsible for providing all energy from the sun. And since the earth must continually lose as much energy to space as it gains from the sun to maintain a relatively constant terrestrial temperature, radiation is the mechanism releasing energy back to space.

Some radiative energy is converted to heat energy and transported throughout the atmosphere. Heat is used to change the temperature of objects and to change the phase of water. Together, radiation and heat comprise the total energy budget of the earth–atmosphere system.

To explain the energy budget, we first explain its most important component, solar energy. Then we trace the operation of the energy budget to illustrate how energy is exchanged throughout a typical day.

The energy budget has the following processes operating (Fig. 1). Shortwave solar radiation propagates through the atmosphere to the surface, with losses due to reflection by clouds and scattering or absorption in the atmosphere. Absorbed shortwave radiation heats the atmosphere and the earth's surface. Warm objects deliver their heat to cooler objects in the vicinity and to the atmosphere through longwave radiation, conduction, convection, and phase changes. Air overlying warm soil is heated by longwave radiation and conduction from the soil, and in turn rises. This convection eventually creates wind, which, on the large scale, is a mechanism for distributing heat energy globally.

Shortwave solar radiation

The largest component of the energy budget is solar energy. It provides all of the incoming en-

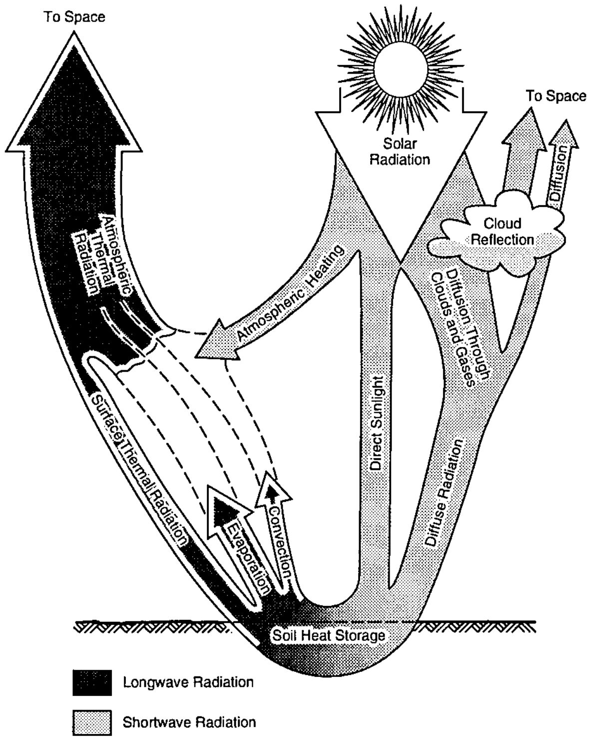


Figure 1. Mean global energy budget. The arrow width represents the relative magnitude of energy flow. Though a simplification of reality, the diagram includes all major mechanisms involved in transporting heat at any given geographic location. Not included here are processes that transport heat from one place to another, including advection of sensible and latent heat by winds and ocean currents, freezing and melting phase changes, and the Greenhouse Effect.

ergy that drives the weather and heats the earth's surface.

The amount of solar radiation absorbed at a site, and therefore how warm exposed surfaces become, depends on several factors, including reflection, day length, sun angle, and the surface area of objects exposed to the sun. The percentage of solar radiation reflected to space by clouds and thus denied to the earth's surface varies with the amount of cloud cover, cloud thickness, and cloud type.

The most significant cause of solar attenuation by clouds is reflection (Fig. 1). Approximately 50% of the planet is covered with cloud at any time, and the average reflectivity, or albedo, of clouds is 50%. Therefore, about 25% of solar radiation entering the atmosphere is reflected to space. On a clear day at any given location little radiation is reflected to space by the atmosphere, whereas about 50% is reflected on an overcast day. In general, fog and thick stratus clouds reflect the most sunlight, followed by nimbostratus, thinner stratus, and stratocumulus, with the least reflected by thin stratus and high, frozen cirrus (Table 1) (Houghton 1985) (Appendix D). This is why temperatures drop so rapidly when scattered, low-level, thick clouds pass over as well as why temperatures do not warm rapidly until morning fogs "burn off."

Clouds also scatter light (Fig. 1). On a completely overcast day only scattered sunlight reaches the surface as generally shadowless diffuse light. Since video-motion IDSs respond to changes in visual contrast in the scene being monitored, diffuse illumination, which reduces visual contrast, generally results in fewer nuisance alarms that are due to wind-induced motion of objects or of the camera itself (Peck 1989a). Adversely, lack of direct sunlight impedes visual detection of a camouflaged intruder who casts no shadow by which to be located. Intermittent cloud cover is a primary cause of nuisance alarms by video-motion IDSs as portions of the scene alternate between being shadowed and sunlit. Under diffuse illumination, aboveground IDSs, security fences, and their surroundings experience less significant solar warming because little direct solar radiation is available to heat objects at the earth's surface.

Occasionally the opposite effect, a temporary increase in the amount of solar radiation received at the ground above what can be delivered directly from the sun, occurs due to cloud cover. This probably would not affect temperatures at a site, but could momentarily affect visual systems because of its transient nature. Very high solar irradiances can occur just before or just after occlusion of the sun by a cloud. This is especially effective if the ground surface is covered with snow or other highly reflective materials. The process occurs when sunlight strikes the surface and is reflected to the cloud. The sunlight

Table 1. Measured cloud albedos of various cloud types (from Houghton 1985).

Cloud form	*Albedo (%)*
Stratus (152 m thick)	5–85 (average 30)
Stratus (152–305 m thick	9–77 (average 60)
Stratus (305–610 m thick)	50–73 (average 70)
Stratocumulus	35–80
Subtropical cumulus	38–45
Nimbostratus	64–70
Cirrostratus	44–59
Cirrus	15–20

Table 2. Albedos of various surfaces (from Shapiro 1982).

	Dry	*Wet*	*Wetness unspecified or indifferent*
Soil			
Dark	0.13	0.08	
Light	0.18	0.10	
Dark ploughed	0.08	0.06	
Light ploughed	0.16	0.08	
Clay	0.23	0.16	
Sandy	0.25	0.18	
Sand	0.40	0.20	
White sand	0.55		
Surface			
Asphalt			0.10
Lava			0.10
Tundra			0.20
Steppe			0.20
Concrete			0.30
Stone			0.30
Desert			0.30
Rock	0.35	0.20	
Dirt road	0.25	0.18	
Clay road	0.30	0.20	

	Growing	*Dormant*	*Verdure unspecified or indifferent*
Fields			
Tall grass	0.18	0.13	0.16
Mowed grass	0.26	0.19	0.22
Deciduous trees	0.18	0.12	0.15
Coniferous trees	0.14	0.12	0.13
Rice	0.12		
Beet, wheat	0.18		
Potato	0.19		
Rye	0.20		
Cotton	0.21		
Lettuce	0.22		
Snow			
Fresh	0.85		
Dense	0.75		
Moist	0.65		
Old	0.55		
Melting	0.35		
Ice			
White	0.75		
Grey	0.60		
Snow and ice	0.65		

	Water surface Solar elevation above horizon (degrees)									
	72	*58*	*49*	*41*	*33*	*27*	*20*	*15*	*9*	*3*
Direct radiation albedo	0.03	0.03	0.03	0.04	0.06	0.08	0.13	0.23	0.41	0.76

is then reflected a second time, back to the surface, by the cloud base to a location that is receiving direct solar radiation between the clouds. The amount of radiation is momentarily increased above cloudless values. Visual systems that react slowly to changes in light intensity could momentarily lose contrast and become ineffective. This would impair the use of surveillance cameras both for assessing intruder activity and for video input to video-motion detection systems.

Upon reaching the surface, most sunlight is either absorbed or reflected by objects at the surface, though some scattering occurs within the leaf canopy of forests and dense plant covers (Chang 1968). The ratio of reflected to incident solar radiation, a surface's albedo, is a function of surface color, texture, transparency, and orientation to the sun. The albedos of clouds are listed in Table 1, and albedos of common substances found at the earth's surface are found in Table 2 (Shapiro 1982). In general, darker materials with rough textures reflect less solar radiation than do smooth, light-colored materials. In addition, albedo typically increases as the angle of the sun above the horizon decreases. This is especially true for water, where albedo increases rapidly as the sun drops below 20° above the horizon. At high sun angles the albedo of water drops to about 3% (Shapiro 1982). Reflected energy may be lost to space, or it may be lost to other objects that either reflect it themselves or absorb it.

Nonreflected solar radiation is absorbed and used for heating. This is one reason for large differentiations in surface temperatures on sunny days, as materials with low albedos absorb more solar radiation that is converted to heat.

A second factor affecting the solar warming of an object is the length of time that sunlight strikes it. An IDS or security fence that is affected by solar loading, either of itself or its background, has a period of potential vulnerability that is dependent on the local day length. Day length is dependent on the latitude of the site, the season, and the existence of solar shadows created by surrounding landforms, structures, and vegetation. Day length increases with latitude in the summer hemisphere and decreases with latitude in the winter hemisphere. A location with no relief, such as a body of water that reaches to the horizon in all directions, experiences a day length that is completely a function of the season and the latitude. As examples of day-length changes with season and latitude, at 25°N latitude, near Key West, Florida, day length is about 10.5 hours on 22 December at the winter solstice, 12 hours on 21 March and 23 September at the equinoxes, and 13.3 hours on 22 June at the summer solstice. Farther north, at 45°N in Minneapolis, Minnesota, day length is about 1.8 hours shorter than Key West on 22 December, and about 2.4 hours longer on 22 June (Brown 1973). On the equinoxes, day length everywhere is 12 hours.

Day length at a site is often shorter than that calculated on the basis of site latitude and the date because surrounding relief, structures, and vegetation all modify day length, usually shortening it by blocking the sun (Fig. 2). IDSs at SOROIDS are located within a deep valley at about 43°N latitude. Surrounded by hills reaching about 10° above the horizon in all directions, on 22 June day length is shortened by 1.7 hours, and on 22 December by 2.7 hours. Sunrise does not occur until the sun rises above the hilltops, and sunset occurs when the sun disappears behind the hills.

An IDS or security fence on a south-facing slope could be in full, intense radiation, whereas those on north slopes could be immersed in deep shadow even in the middle of the day. Since the sun's altitude is always changing from day to day at any given solar azimuth, the solar shadow pattern will change each day and so present a different set of conditions to an IDS (Mateer and Godson 1959). Depending on the orientation of a perimeter fence, the solar shadow pattern may progress along it, moving from one IDS's zone to the next (Fig. 3). This is obviously a more serious problem on clear days than on overcast days when scattering by clouds reduces or eliminates shadows. Security designers and IDS managers can map expected solar shadow patterns along fence lines for any date using a sun chart and simple instructions (List 1958, Ryerson 1984). Shadow mapping can also be accomplished with most computerized geographic information systems (GISs). Structures and trees cause shadow problems similar to those caused by relief, though they are generally not as severe as relief-caused shadows.

The angle of incidence of solar radiation, or sun angle, is the third factor that impacts the solar warming of IDS hardware, security fences, and terrain through its effect upon radiation intensity. As with a flashlight aimed at a wall in a darkened room, the more directly, or closer to 90°, the solar beam intercepts a surface, the more intense the radiation (Fig. 4). At low sun angles, vertical objects such as security fences receive far

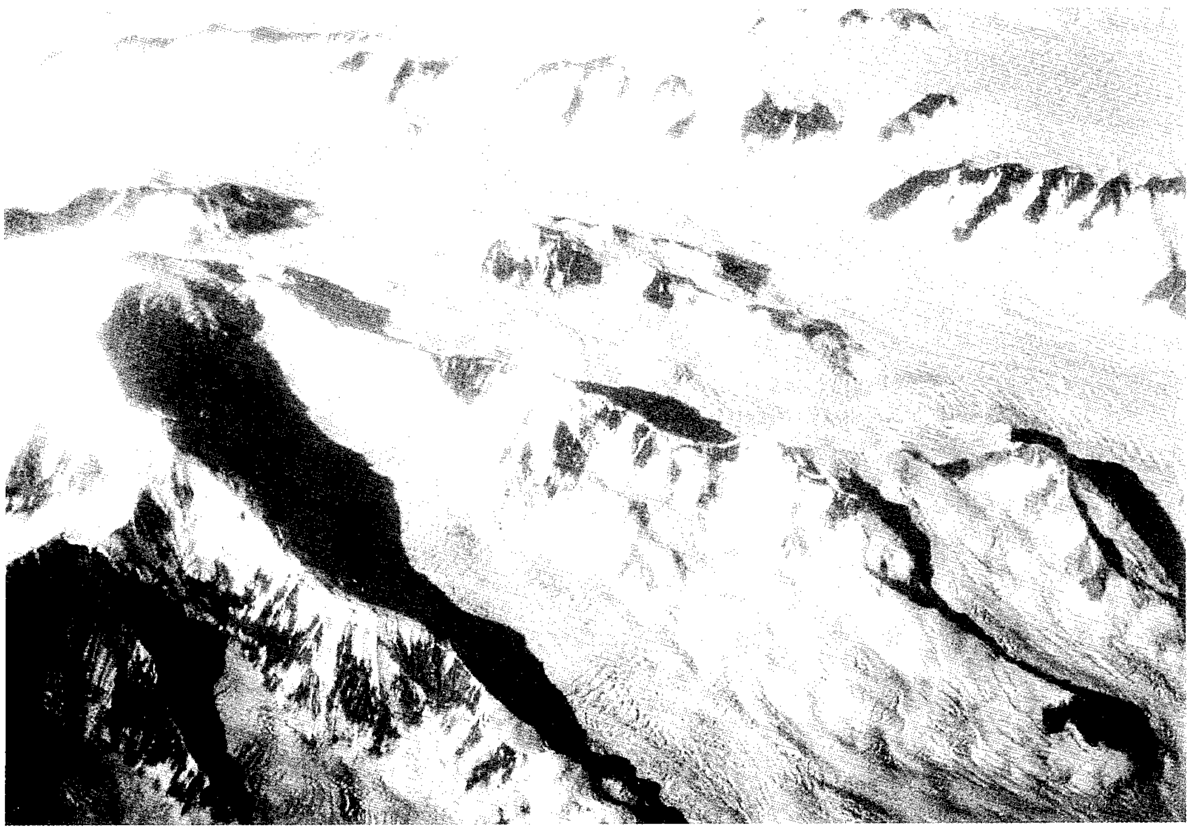


Figure 2. Topographic shadows cast in late afternoon (St. Elias Mountains, Alaska).

more radiation than horizontal surfaces because the fence is nearly 90° to the solar beam, the situation for greatest intensity of radiation. On a midwinter day, for example, the intensity of radiation received by a security fence panel facing south can be 100% greater than the amount of radiation received by the horizontal surface upon which the fence is mounted (Fig. 5). Since the stiffness* of a security fence can vary significantly with the magnitude of solar heating through thermal expansion and contraction, there may be substantial differences in a fence's resistance to wind-induced motion from sector to sector of an IDS depending on the time of day. The magnitude of a temperature-related change in fence stiffness may lessen with time, however, as the fence loosens under the effects of thermal cycling and overall relaxation (Peck 1993a). All types of security

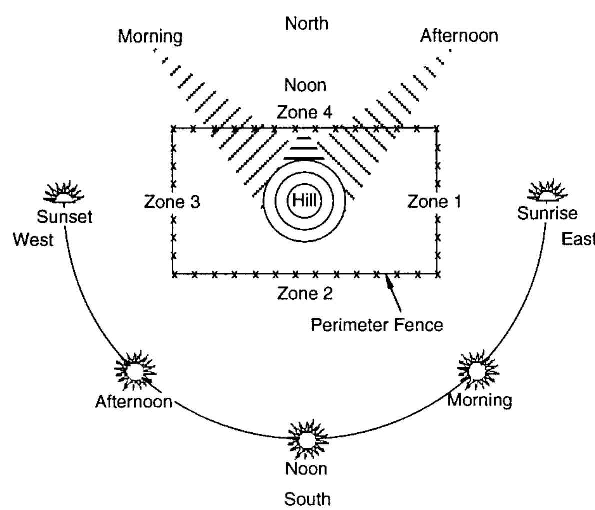


Figure 3. Shadow cast by a symmetrical hill onto a perimeter fence by morning, noon, and afternoon sun. As the day progresses, the shadow moves continuously from west to east, changing shape and size as the sun's altitude and direction change. This diagram applies only to the Northern Hemisphere.

* See section on fence-mounted IDSs in Appendix A for an explanation of stiffness.

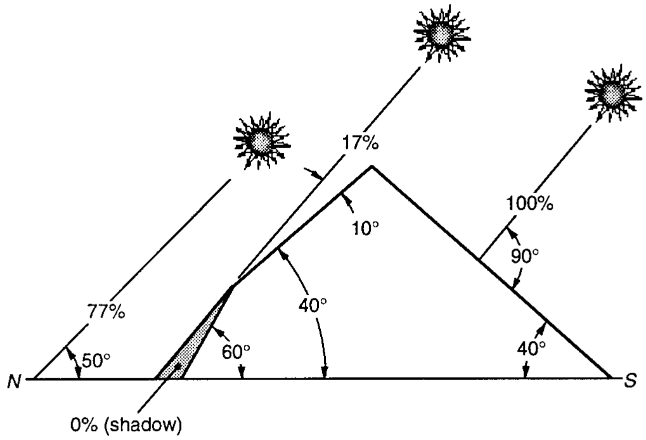


Figure 4. Relative intensity of solar radiation on inclined surfaces. Solar elevation is 50°.

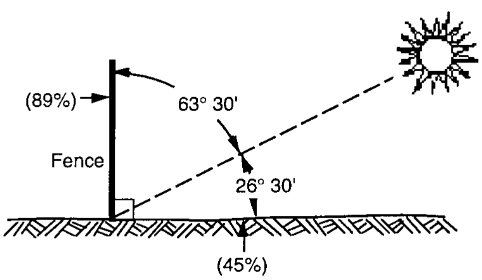


Figure 5. Solar angles to horizontal and vertical surfaces on 22 December at 40° north latitude at solar noon. The fence fabric receives approximately twice as much direct beam solar radiation as the soil surface.

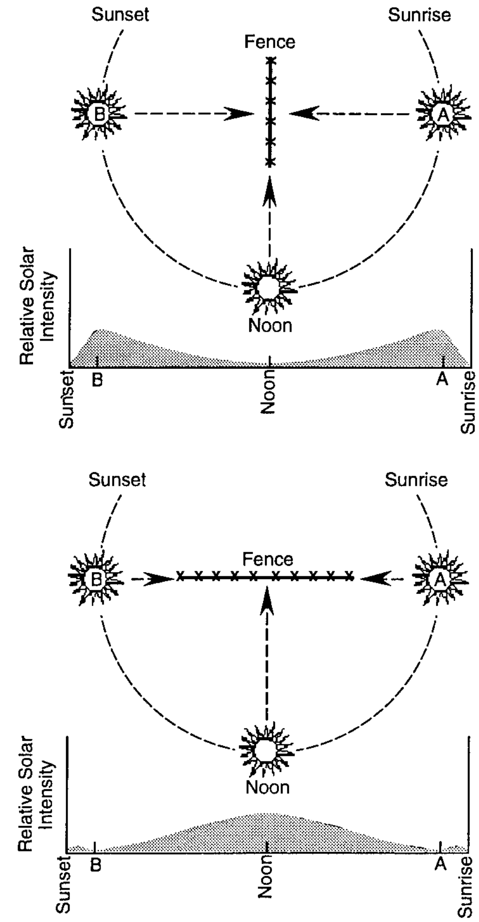


Figure 6. Relative diurnal solar intensity changes on fence fabric as a function of solar azimuth and elevation. (a) North–south-oriented fence and morning and afternoon solar peaks, with a solar noon minimum. (b) East–west-oriented fence with major radiation peak at solar noon and minor peaks immediately after sunrise and before sunset.

fences experience heat-induced expansion and contraction. The resulting displacement of structural elements, once frictional resistance is overcome, may be more abrupt with, for example, weld mesh fences bolted to fence posts than it is with chain-link fences, for which the tension of the panel is distributed over a large area of continuous fence fabric.

The area of an object exposed to solar radiation is a fourth factor determining the amount of radiation that is received. Because security fences typically completely surround compounds, on a sunny day at least part of that fence is probably receiving maximum intensity radiation over its surface, while simultaneously another part is receiving minimal radiation. The portions of fence being highly radiated or minimally radiated at any one time change throughout the day as solar altitude and solar direction change. If a security fence is oriented in a north–south direction, at solar noon the sun is aligned with the fence's longitudinal axis. Only the top of the fence is exposed to the sun, and minimal radiation is striking its fabric (Fig. 6). Solar heating of the fence fabric is therefore least at this time when the sun is highest in the sky and the solar beam is theoretically strongest. Although the direct solar radiation on a security fence may be least at midday, that is usually the warmest portion of the day, so the fence is warmed primarily by contact with hot air rising from the heated ground surface. North–south-oriented portions of the fence receive two solar radiation peaks each day, one when the fence is exposed to the morning sun incident from the east, and again in the afternoon when the western side of the fence is exposed to the setting sun (Fig. 6). Fences oriented in an east–west direction may receive only one peak of radiation, near solar noon (although they could receive three peaks: near sunrise, at solar noon, and near sunset) (Fig. 6).

The complexity of the surface geometries, coupled with the changing position of the sun through the day, causes a myriad of solar microclimates along security fences that affect the temperature-related tension of the fence and any attached IDS sensor cables. This complex pattern of solar microclimates may render some portions of the fence more resistant to motion imparted by an intruder's activities (cutting, climbing) at any one time. In addition, these complex solar exposure patterns affect the reliability of IDS electronics because of the potential for repeated thermal shock due to solar loading patterns through the day. Mapping and understanding fence solar energy patterns could explain patterns of nuisance alarm occurrences and identify when the detection capability of fence-mounted IDSs is diminished.

Longwave radiation

Heated materials at the earth's surface radiate more strongly at the far-infrared wavelengths than at the visible and near-infrared wavelengths of the much hotter sun. Longwave radiation (3–50 micrometers) interacts differently with the atmosphere and with objects at the earth's surface than does shortwave (0.3–3 μm) solar radiation. The atmosphere is nearly transparent to shortwave radiation, and thus it is not heated very effectively by direct absorption of shortwave solar radiation. The atmosphere is, however, an effective absorber and reradiator of longwave radiation. In addition, most objects in the environment are strong absorbers of longwave radiation. That is, there is almost no reflection of longwave radiation. For example, snow reflects 40% to 90% of incident shortwave radiation, but it absorbs nearly all longwave radiation. This means that snow warms and melts slowly in the presence of strong visible sunlight (provided it is not in contact with warm air). However, snow responds quickly to longwave energy radiated from tree trunks, for example, contributing to the snow-free ring often found at the base of tree trunks.

A classic example of longwave radiative exchange occurs on cold, clear nights. Automobile windshields are canted considerably from the vertical, whereas the side windows are often nearly vertical. Automobile windows have little mass and thus change temperature rapidly. On cold, clear nights, the windshield is exposed to the night sky, which often has a temperature well below freezing. The side windows are exposed to warmer surrounding objects such as vegetation or structures. As a result, the windshield loses more energy radiatively to the night sky than it receives in return, cools rapidly, and may drop below the dew point or frost point and become coated with dew or with hoarfrost. The side windows are radiating to and receiving radiation from warmer objects. As a result, they do not cool as deeply as the windshield and usually, therefore, do not dew or frost as heavily.

IDSs experience these same radiative environments. Objects exposed directly to solar radiation and the warm daytime sky receive an abundance of both shortwave solar radiation and longwave sky radiation, and therefore heat. Fence panels, for example, often absorb large quantities of radiation and heat rapidly, particularly those made of thin wire or welded mesh with relatively small heat capacities. They radiate, convect, and conduct heat away strongly as they heat during the day, but more energy is often received radiatively than is lost. By early afternoon, under the joint effects of solar warming and contact with warm air, the fence panel has heated to its warmest temperature of the day and may be loosest at this time.

After sunset the sky cools rapidly, especially if it is cloudless. The fence continues to lose heat to the air by radiation, convection, and conduction without any offsetting heat gain from solar radiation. An exposed fence on a hilltop, for example, may radiate so much energy to the very cold sky that its temperature drops below the temperature of the local air and surrounding objects. The fence then becomes a heat sink and receives energy by conduction, convection, and radiation from the soil, vegetation, and nearby buildings or other objects.

A fence located in a valley, especially if surrounded by vegetation and buildings, is oriented broadside to objects warmer than the sky. As a result, its radiative cooling is not as rapid and, were no other cooling mechanisms occurring, it would remain warmer through the night than would the fence on the hilltop. As a result, the fence in the valley experiences a smaller radiative temperature change through the diurnal cycle than does the hilltop fence.

Cloudy skies and cloudless, hazy, humid skies are warmer at night than are clear skies because the abundant water vapor and liquid water cloud droplets absorb longwave radiation readily and heat and reradiate the energy. In addition, growing clouds are a source of heat as water vapor condenses to water and releases latent heat of condensation. Clear skies found in the winter are relatively devoid of water vapor, so longwave radiation from the surface escapes readily to space. The amount of heat loss from exposed IDSs such as microwave antennas, passive infrared units, or exterior-mounted processors can affect IDS performance, depending on the temperature-sensitivity of electronic components and their response to thermal shock conditions.

Convection

Air warms through direct contact (conduction) with heated surfaces and through radiational heating. Surfaces heated by exposure to shortwave solar radiation emit longwave radiation that in turn heats the air and adjacent objects. The vertical motion, or convection, of air that results from warm air rising and cool air descending eventually creates advection, or wind, as air masses flow laterally to replace the risen air. The occurrence of these local, convective winds is dependent on the heat available at the ground surface. They rarely form over snow-covered ground because the maximum potential surface temperature (0°C) cannot adequately heat the overlying air (Dewey 1977). This is a beneficial circumstance for IDSs that are adversely affected by wind-induced motion, such as fence-mounted IDSs or line-of-sight IDSs with low tolerance for misalignment. Bare surfaces, particularly highly absorptive asphalt or freshly plowed soil, that have no natural constraint on heating will more than likely be the source of convectively generated wind activity.

Phase change

Phase changes are important for transporting heat to, from, and within the atmosphere. The change of water molecules from solid to liquid to gas and reverse requires heat to be consumed or released, respectively. Heat consumed, for example, is used either to raise the temperature of the material (sensible heating) or to change the phase of the material through evaporation or melting (latent heating). Damp or wet surfaces do not heat under solar loading until the available moisture has been evaporated. If moisture is present as either water or ice, energy is always consumed first to change its phase before energy is used for sensible heating of the atmosphere or soil. This is one reason why IDSs sensitive to thermal radiance may be less troubled by nuisance alarms until after evaporation of morning dew or melting of frost is completed and the background, particularly if it is spatially heterogeneous, begins to heat. Adversely, a dampened intruder may escape detection because his thermal contrast with a wet background may remain below the alarm threshold of a passive infrared IDS until after his clothing has dried.

A snow cover in contact with warm air increases in temperature at the surface from below freezing to 0°C through sensible heating, but then remains at that temperature while undergoing melting, a phase change. The stability of a snow cover's surface temperature under conditions of melting is the reason it is a favorable background for passive infrared IDSs (Peck 1993b). Similarly, sensible heating of ice-coated or wet surfaces is inhibited until the ice is melted and any water (dew, rain, meltwater) is evaporated. Wet surfaces are a favorable background for passive infrared IDSs because backgrounds that are thermally heterogeneous when dry become thermally uniform backgrounds when wet (Shushan et al. 1991).

Temperature

Diurnal temperature changes—of the air, soil, and all objects in the environment—have potentially large impacts upon the effectiveness of IDSs. Temperature affects the tension of fences by causing the materials to expand or contract and may also cause movement of fence posts through heaving of the soil as it freezes and settling as it thaws. The relative rate and magnitude of temperature changes that are distributed across a thermally heterogeneous background determine whether passive infrared IDSs will be subject to nuisance alarms (Peck 1993b). The difference in temperature between an intruder and the detection zone background is the basis of passive infrared IDSs, which respond to the magnitude and rate of change of thermal radiance. Lacombe (1993, 1994) investigated how the detection capability of passive infrared IDSs would vary over 24 hours because of differences in the thermal contrast of an intruder with a variety of backgrounds (concrete, asphalt, gravel, sand, grass) under the same

Table 3. Thermal properties of assorted materials (from Geiger 1965).

Type of material	*Density (g cm^{-3})*	*Heat capacity (cal cm^{-3} °C^{-1})*	*Thermal conductivity (cal cm^{-1} s^{-1} °C^{-1})*	*Thermal diffusivity (cm^{2} s^{-1})*
Iron	—	0.82	210	260
Concrete	—	0.5	11	20
Rock	2.5–2.9	0.43–0.58	4–10	6–23
Ice	1.7–2.3	0.46	5–7	11–15
Wet sand	—	0.2–0.6	2–6	4–10
Wet clay	1.7–2.2	0.3–0.4	2–5	6–16
Old snow (density 0.8)	0.8	0.37	3–5	8–14
Still water	—	1.0	1.3–1.5	1.3–1.5
Wet moorland	0.8–1.0	0.6–0.8	0.7–1.0	0.9–1.5
Dry clay	—	0.1–0.4	0.2–1.5	0.5–2.0
Dry sand	1.4–1.7	0.1–0.4	0.4–0.7	2–5
New snow (density 0.2)	0.2	0.09	0.2–0.3	2–4
Dry wood	0.4–0.8	0.1–0.2	0.2–0.5	1–5
Dry moorland	0.3–0.6	0.1–0.2	0.1–0.3	1–3
Still air	—	0.00024–0.00034	0.05–0.06	150–250

weather conditions. His study made use of model predictions of passive infrared sensor response (Lacombe 1992, Lacombe and Peck 1992) and intruder surface temperatures (Lacombe 1993). Temperature also affects the operation of buried systems through related changes in the elastic properties and moisture content of the soil. Buried ground-motion IDSs have diminished detection capability in frozen or dry, hard-packed soil compared with less firm thawed (e.g., damp) soil (Peck 1994b, d). Conversely, buried electromagnetic IDSs respond to the presence of an intruder with a larger margin of reliability when the soil's electrical conductivity is less (e.g., dry or frozen soil), because the electromagnetic field that defines the detection zone is stronger (Peck 1992b, 1994b). The transmission loss experienced by bistatic microwave IDSs is greater for microwaves propagating over wet soil or vegetation, so periods of high temperature that promote surface drying are favorable to these systems, but if contact with warm air causes melting of a snow cover, the increased wetness of the snow is detrimental (Peck 1991, 1992a).

The rate and amount of temperature change a material experiences is a function of its specific physical characteristics and its ability to store and gain or lose heat via the various heat transport mechanisms. The physical properties of a material affect how rapidly it changes temperature (thermal lag or thermal diffusivity), how much it changes temperature (heat capacity), and the volume of the material that is heated or cooled (thermal conductivity). These physical characteristics are most effectively demonstrated and contrasted by water and dry soil.

Heat capacity is a measure of a material's ability to change temperature as a result of a given change in caloric heat content, with low heat capacity materials experiencing a larger temperature change for a given heat change. The volumetric heat capacity of water is 2.5 to 10 times larger than the heat capacity of soil minerals (Table 3) (Geiger 1965). Therefore, for a calorie of energy gained, a cubic centimeter of dry clay or sand soil minerals will warm 2.5 to 10°C, whereas water will heat only 1°C. Furthermore, thermal conductivity, or the ability to transport heat conductively, is 2 to 6 times greater in motionless water than in dry soil, so energy is conducted through water more readily. Accordingly, water bodies heat and cool slowly, demonstrate large thermal lags, and undergo small temperature ranges (also because of the ease of convective overturn of the water), and heating or cooling generally involves a large volume of water. In comparison with water, dry soil heats and cools rapidly, has small thermal lag, experiences a large range of temperature, and heating and cooling generally involves only a small volume near the soil surface.

Two dramatic examples of the effect of land vs. water on local climate are illustrated by two California cities located only 140 km apart: San Francisco and Sacramento. San Francisco is located within a highly marine climate, sandwiched between the Pacific Ocean and San Francisco Bay.

Table 4. Daily temperature ranges (°C) in marine and arid climates (from Ruffner and Bair 1979).

	San Francisco	*Sacramento*
Daily minimum—Jan	7.7	2.9
Daily maximum—Jan	13.4	11.8
January daily range	5.7	8.9
Daily minimum—July	11.9	14.3
Daily maximum—July	17.8	34.1
July daily range	5.9	19.8

The airport weather station is located 3 m above sea level. Sacramento is in the arid Napa Valley directly east of San Francisco. Its airport weather station is 7 m above sea level. However, despite their closeness and similarity in elevation, there are large differences in their daily temperature regimes (Table 4). January daily temperature ranges in Sacramento average about 1.6 times those of San Francisco, and July temperature ranges in Sacramento are 3.4 times larger. In addition, the minimum daily January temperature in Sacramento is lower than San Francisco's, and the Sacramento July daily maximum is higher. The larger range and higher and lower extreme temperatures in Sacramento are due to its distance from water and its location within an arid climate.

Most wet surfaces cannot warm to a temperature above about 26°C because, at the maximum natural rates of short- and longwave radiation and convective transport possible to a water surface, water rapidly converts the absorbed radiation to latent heat and evaporates rather than sensibly warming (Deacon 1969). On the other hand, temperatures can reach as high as 82°C at the surface of soils in warm deserts (Griffiths and Driscoll 1982). The impact of this on intruder detections by passive infared IDSs is that a wet intruder may have a radiative temperature similar to that of a damp background and lack sufficient thermal contrast for detection. Alternatively if the intruder is cool because of evaporation from his clothing, then the thermal contrast with a dry, heated background may be large. A second consequence is that backgrounds that are not uniformly wetted—such as soil that remains damp due to localized pooling of water or perhaps cinder block retaining walls that retain moisture in the block cores—will heat differentially under thermal loading, leading to a background that may be troublesome for passive infrared IDSs because of its nonuniform temperature distribution and the consequent spatial variation in temperature changes under identical solar loading.

Soil temperatures

The thermal response of soil varies with moisture content: heat capacity increases as pores fill with water, but thermal conductivity increases only until adjacent soil particles are connected by a continuous water film (Lowry 1967). At very low soil moisture, soil particles are in poor contact with one another and thermal contact resistance is high (Carson 1969). As soil moisture increases and bridges the contact between soil particles, thermal conductivity increases. For example, in a silt loam soil in central Ohio (Ryerson 1979), soil thermal conductivity nearly doubled when soil moisture content increased from about 10 to 43%. Increasing soil moisture beyond that of a moderate moisture content does not increase soil's ability to conduct heat because soil mineral particles are better conductors of heat than is water. Additional water only increases heat capacity and thus slows the progress of thermal waves through soil. Although moist soils do have a lower albedo, averaging about 10% less than that of bare, dry soils (Chang 1968), the larger amount of solar radiation absorbed by moist soil is consumed as latent rather than sensible heat. Therefore, moist soils never change temperature as rapidly as dry soils, despite their larger thermal conductivity and smaller albedo, nor do they reach the extremes of temperature that dry soils do. Under freezing conditions, however, frost penetration is more rapid and the depth of the freezing front is greater in moderately moist soil (Ryerson 1979). For example, an increase in moisture content of a silty soil from 10 to 17% by weight results in the soil's latent heat, heat capacity (unfrozen soil), and thermal conductivity (frozen soil) increasing by ~66%, 22%, and 57%, respectively. The net result is that, although more heat must be released in cooling the soil to ~0°C (heat capacity) and then freezing the soil (latent heat), that heat is transported away more readily (thermal conductivity) in frozen soil having the higher moisture content, so freezing proceeds more rapidly in the moister soil (Peck and O'Neill 1995).

Frost penetration along a buried IDS's detection zone will be nonuniform if the moisture content of the soil varies laterally, all other factors (weather, exposure, soil type, etc.) being the same. Soil freezing promotes reliable detection capability of buried electromagnetic IDSs, but reduces the detection capability of buried ground-motion IDSs (Peck 1994b, d).

Soil density also has a major impact on soil temperature. Low-density soils, such as those rich in organic material, sand, and freshly plowed soils, conduct heat poorly if dry (Chang 1968). Lower-density soils change temperature rapidly at the surface and conduct heat only slowly into deep soil horizons, causing them to experience large temperature extremes at the surface. Higher-density soils, such as clay, or compacted soils subjected to heavy foot or vehicular traffic conduct heat readily, have smaller temperature extremes at the surface, and transport heat into and out of greater depths (Chang 1968). A foot path along a fence, or replacing clay soils with sand when trenching along a fence, will increase and decrease the rate of heat transport into and out of the soil, respectively.

Other characteristics affecting thermal conductivity are soil particle size, mineral type (Sartz 1970), and bulk density, which is closely related to particle size and pore space. In general, soil thermal conductivity increases with increase of soil moisture, temperature, density, and particle size (Kersten 1966).

Temperature changes in soils are damped by vegetation cover and litter (duff) layers at the soil surface. Forest humus has a low thermal conductivity, similar to that of snow, and shades the soil surface from direct solar radiation (Satterlund 1972). As a result, duff-covered soils are cooler during the day because they are not exposed to direct solar radiation. They are also insulated from cooling at night because the low thermal conductivity of the duff layer keeps the soil surface warmer, preventing loss of energy to the atmosphere by longwave radiation. Overall, the diurnal temperature amplitude of a litter-covered soil is smaller, or damped, and it responds more slowly to changes in external factors than does a bare mineral soil (Bouyoucos 1913, MacKinney 1929). However, since radiant energy is not being used to heat the soil, it instead heats the duff layer, which, if dry, can warm significantly (Reifsnyder and Lull 1965).

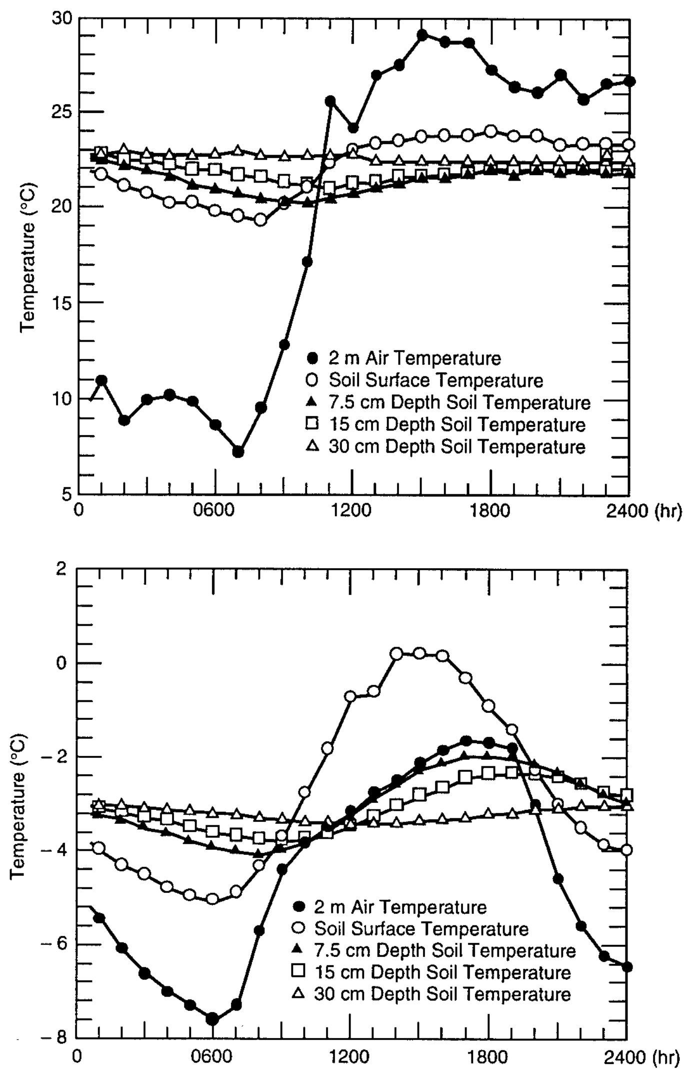


Figure 7. Typical summer and winter diurnal temperature changes in the air at 2 m, at the soil surface, and at depth within the soil. (a) Midsummer diurnal soil temperature amplitude at 15 cm depth is about 25% of the 2-m air temperature amplitude, and it is progressively time-lagged with depth within the soil. The thermal lag at 15-cm depth is about 6 hours. (b) Winter diurnal soil temperature amplitude at 15-cm depth, and under a snow cover, is only about 17% of the 2-m air temperature change, and it is lagged about 5 hours.

Differential damping of soil temperature changes at the soil surface can be particularly troublesome for passive infrared IDSs. During the transition period between being snow-covered in winter and the establishment of a complete vegetation cover in summer, the ground cover may consist of thatch (duff), new-growth grass, and exposed soil (where foot or vehicle traffic earlier churned the surface to mud). This is a difficult background for passive infrared IDSs because the spatial heterogeneity of the soil, thatch, and grass, together with their different rates of response to solar heating and radiative cooling, cause spatial thermal heterogeneity at the soil surface and numerous nuisance alarms (Peck 1993b). Since a passive infrared IDS is designed to alarm at differential changes in thermal radiance, such as would be caused by an intruder moving through its detection zone, nonuniform heating and cooling of the ground cover within its detection zone can readily satisfy its alarm criteria.

Damping of soil temperature changes by snow, vegetation, or duff layers is beneficial if freezing of the soil is to be delayed or avoided, but it is detrimental if drying or warming of the soil is desirable. For example, the detection capability of a buried electromagnetic IDS is highly dependent on the liquid moisture content of the soil (Peck 1992b, 1994d). It is better that the soil remain either unfrozen or frozen (i.e., relatively high or low liquid moisture content) instead of fluctuating between the two states. Similarly, if the system's detection capability were set for dry, low moisture content soil, a rain storm can lead to poor detection capability that persists beyond the storm, i.e., until the wet soil dries. Buried IDSs that detect intruders on the basis of ground motion are more reliable in wet soil, which is less resistant to displacement under contact loading, than in soil that is firm because it is frozen or dry (Peck 1994b, d). Variation in the retention of soil moisture along a perimeter because of different degrees of vegetation cover can cause location-dependent differences in buried IDS performance. This is true also in winter, when frost penetration depends on the moisture content of the soil and so may be variable within an IDS's detection zone (Ryerson 1979, Peck and O'Neill 1995). Security personnel trying to sort out why a buried IDS is more responsive in one zone than another might understandably overlook subtle differences in ground cover.

Soil temperature is also damped with depth within the soil because of soil's generally poor ability to conduct heat and its general high heat capacity. Examples of soil temperature damping and lag with depth are illustrated by profiles at SOROIDS. Figure 7a is a summer situation under a grass cover, and Figure 7b demonstrates a winter situation under a 2–4 cm snow cover. Air and soil temperatures lag and diminish in amplitude in the air and with depth in the soil compared with temperature changes at the soil surface.

Generally, soil temperature fluctuations are larger for the cases of higher daytime insolation on cloudless days, low soil surface albedo, orientation to the equator in the midlatitudes, and bare, moderately moist soils with a high thermal diffusivity. Thermal fluctuations are minimized by very high or very low soil moisture, cloud cover, insulating snow or organic material at the surface, full vegetation cover, and orientation poleward in the midlatitudes.

Freeze–thaw cycles. Soil freezing is not generally considered a diurnal event because penetration of frost into the soil, and subsequent thawing, is a multiday, if not seasonal, event. However, diurnal freeze–thaw cycles do occur that affect surface soil temperatures and change the mechanical characteristics of the soil surface. The middle latitudes and mountains experience more frequent freeze–thaw cycles than warmer and colder climates (Fig. 8). A freeze–thaw cycle is defined as a drop in temperature of the soil sur-

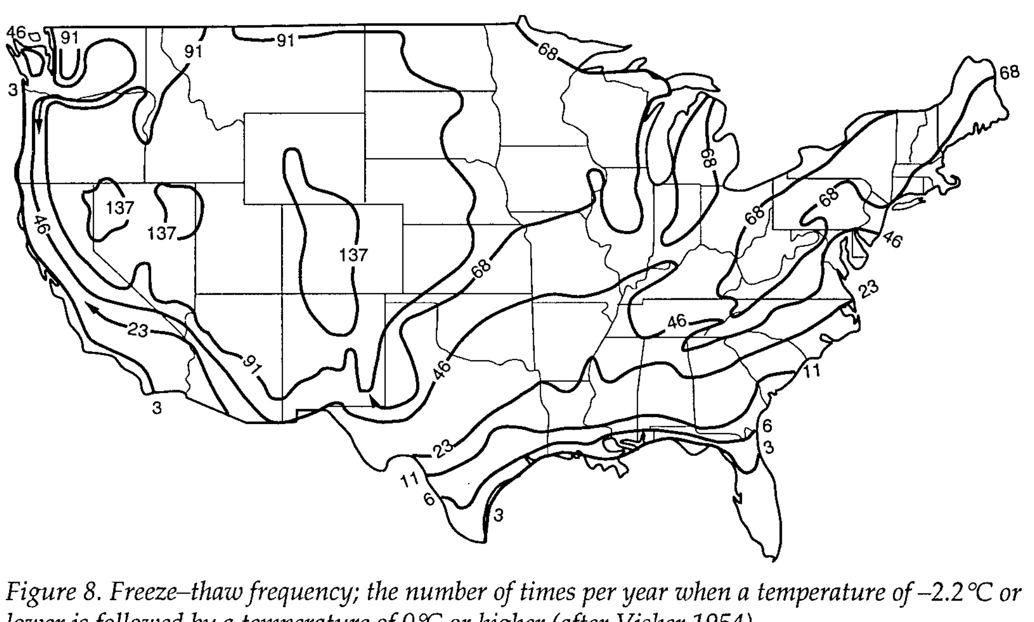


Figure 8. Freeze–thaw frequency; the number of times per year when a temperature of –2.2°C or lower is followed by a temperature of 0°C or higher (after Visher 1954).

Figure 9. Needle ice, about 4 cm long, growing from a moist, unfrozen soil surface. Needle ice growth occurs from the soil surface, therefore ice atop the ice columns, immediately below the lifted soil particles, is oldest. The soil particle caps were lifted from the soil surface.

face to below freezing, with a subsequent rise above freezing, at least once per day.

Diurnal freeze–thaw cycles affect air and soil thermal regimes in unique ways. Diurnal freezing usually results from the invasion of a cold air mass that allows extreme chilling due primarily to loss of surface longwave radiation to space on calm nights. As soil begins to freeze and frost begins to nucleate at the soil surface, 80 calories of latent heat are released for each cubic centimeter of water frozen. Though frost may only penetrate a few centimeters into the soil in one night, the large release of latent heat tends to cancel deeper cooling of the air and soil surface, preventing temperatures from dropping significantly below 0°C. The influx of strong solar radiation the following morning begins the thawing process. Again, however, air and soil temperatures generally do not warm significantly above 0°C until after thawing has occurred. These thermal plateaus at 0°C, "zero curtains," though most obvious in the fall and spring in seasonal frost regions, are observable in the daily cycle.

The bearing capacity and shear strength of bulk soil increase as soil freezes, which improves the traction of vehicles and personnel traversing the surface. However, the soil surface can become more delicate as a result of diurnal freeze–thaw cycles and can collapse with considerable noise under certain conditions from loading by a vehicle or person. This occurs when columns of ice have grown vertically into the atmosphere or immediately below the soil surface during the night (Washburn 1973). This needle ice, columnar frost, or piprake (Fig. 9) grows as a result of radiational cooling at the surface of moist, bare, and sparsely vegetated loam and silt soils (Price 1972). Slow moisture transport in clay soils, because of their smaller pore structure, and in sandy soils, because of the large pore spaces, inhibits needle ice growth in these soils because insufficient water is transported to the surface. Needle ice grows vertically to produce slender columns 0.5 to 3 cm long most commonly, though extreme lengths of 30 to 40 cm have been observed (Washburn 1973). Needle diameters are typically less than a few millimeters. Several tiers of needles can develop if growth from the previous evening does not completely melt or if the freezing process halts and resumes several times during the night. Small soil particles and rocks are often lifted from the surface on top of the needles, in part because these mineral materials cool rapidly and have high thermal conductivities, thus causing freezing at their bases. The soil surface is heaved as a result of a night of needle ice formation, and the surface often appears rough and pocked. Under the weight of a person, needles usually collapse with a deep crushing sound, producing a potentially noisy surface for intruders to traverse.

Frost cracks are another soil phenomenon at low temperatures in seasonal frost regions. Contraction of the soil may lead to the development of a network of visible ground fractures as the soil fails in tension under thermal stress. The cracks persist through the winter, possibly reestablishing themselves in the same locations during succeeding winters. The abrupt motion as a crack incrementally deepens may be a source of nuisance alarms for buried IDSs that respond to ground motion. More certainly, buried sensor cables that are exposed by frost cracks are more vulnerable to damage. If water infiltrates the cracks and then freezes, the associated IDSs may have inconsistent detection capability.

Snow cover and rainfall. Snow is one of the most effective natural insulators, with thermal conductivity tied directly to its density. Newly fallen, low-density snow may have a thermal conductivity similar to that of still air, whereas snow that is warm, dense, and old may attain a thermal conductivity equivalent to that of wet soil (Table 3) (U.S. Army 1956). The generally low thermal conductivity of snow makes it an effective insulator. Snow-covered soil is thermally isolated from diurnal temperature changes occuring in air above the snow surface. In general, the thermal isolation of soil from air temperature fluctuations increases when snow depth is greatest and snow density is least (Anderson 1947). Because of its insulative characteristics, snow delays cooling of the soil if the cover is in place before temperatures drop significantly. Snow also delays warming of the soil surface in the spring because of its albedo, insulative capabilities, and the latent heat required to melt it.

Since snow density is the primary control of snow thermal conductivity, compaction by traffic can have a large effect on the temperature of the snow cover and the temperature of the soil below the snow cover. When an initially undisturbed, 0.25-m-deep snow cover was compacted by driving an Army tank across it, the surface temperature increased by up to 10°C at air temperatures of –5 to –35°C (Jordan et al. 1989). The surface of the undisturbed snow is colder because its lower thermal conductivity prevents soil heat from reaching the snow surface. Compacted snow allows soil heat to be transported more readily to the snow surface, heating the snow surface and cooling the soil below. Snow compaction can increase the complexity of thermal scenes observed by passive infrared IDSs. Even after a new snowfall leaves a visually uniform snow surface, it may still be a thermally nonuniform background, with higher localized surface temperatures above compacted snow. Snow compaction encourages more rapid soil cooling than is experienced under an undisturbed snow cover, while soil beneath uncompacted snow gradually warms as heat flowing from greater depths is retained at the base of the snow cover due to the snow's low thermal conductivity. Buried IDSs may experience changes in detection capability as the soil warms in this manner because of increased unfrozen water content (electromagnetic IDSs) or decreased soil rigidity (ground-motion IDSs).

Snowmelt, which is usually greatest during the day and smallest in volume at night, and rainfall are surges of uniform-temperature water that can produce rapid and dramatic changes in soil temperature, with the largest and most rapid changes occurring at the soil surface (Crawford 1952, Chang 1968). Decreases in soil temperature as large as 4–5°C have been recorded in the upper centimeters of the soil surface within 2 minutes of the beginning of a heavy rainfall, with decreases of about 3°C at a depth of 2 cm within 20 min, and a decrease of 1°C at 10 cm within 1 hour of the start of rainfall. These changes are extreme because 20.8 mm of rainfall fell within 1.4 hours (Geiger 1965). Van Wijk (1963) observed similar changes after an 8-mm rainfall lasting about an hour. Bare soil and pavement surfaces could readily experience similar, or perhaps larger, magnitude decreases in temperature during an afternoon thunderstorm following a clear day of intense solar radiation in an arid environment. The effect of infiltrating rainfall or snowmelt water is to homogenize the surface soil thermally and reduce thermal contrast at the surface and in the near-surface atmosphere above the soil. This is because water has a very high heat capacity; therefore, if rain or snowmelt infiltrate soil, the soil temperature will trend toward that of the infiltrating moisture. This change occurs most rapidly in dry, low-density soils with high infiltration capacities.

Vegetation effects. Vegetation, like litter, decreases the rate and range of diurnal temperature changes in soils. It shades the soil surface, allowing only diffuse solar radiation or patches of direct solar radiation, known as sunflects, to reach the soil surface. The amount of solar radiation available to heat the soil surface beneath a canopy is primarily a function of the inclination and the arrangement of leaves, with vertical-leaved crops blocking about 30 to 50% of insolation and crops with horizontal leaves blocking 70 to 100% (Chang 1968). Taller crops also intercept more sunlight than shorter crops and, as solar elevation decreases, less light reaches the soil surface below.

Even grass with a height of 0.20 m allows only 10% of incident solar radiation to reach the soil surface (Oke 1978). Most green vegetation either absorbs solar radiation for evapotranspiration and photosynthesis, or reflects it, with some transmission of radiation through the leaves.

Leaf motion and changes in the sun's position cause sunflects to move. Their rate of movement and the solar intensity are a function of the height of the canopy opening. Solar intensity within flecks is seldom more than 20% of the above-canopy solar intensity (Reifsnyder and Lull 1965). Despite their lower intensity, sunflects momentarily warm the soil and can produce wide temperature variations in the litter, as much as 16°C in 20 min (Reifsnyder and Lull 1965).

The canopy also depletes longwave radiation from the sky by interception, but it causes more longwave terrestrial radiation to be returned to the soil because of the reduced sky view factor (Oke 1978). Both also tend to reduce fluctuation in temperature below the canopy.

Cloud cover effects. The most significant fluctuations of temperatures at the surface typically occur under clear sky conditions because of the intensity of incident shortwave solar radiation during the day and the extreme loss of longwave radiation from the surface to space. Clouds are moderators of surface temperature. Overcast skies reduce insolation and scatter insolation transmitted through the clouds, thus causing diffused, shadowless light at the surface. Overcast skies are also stronger radiators of longwave radiation than clear skies. Clouds are warmer than the clear

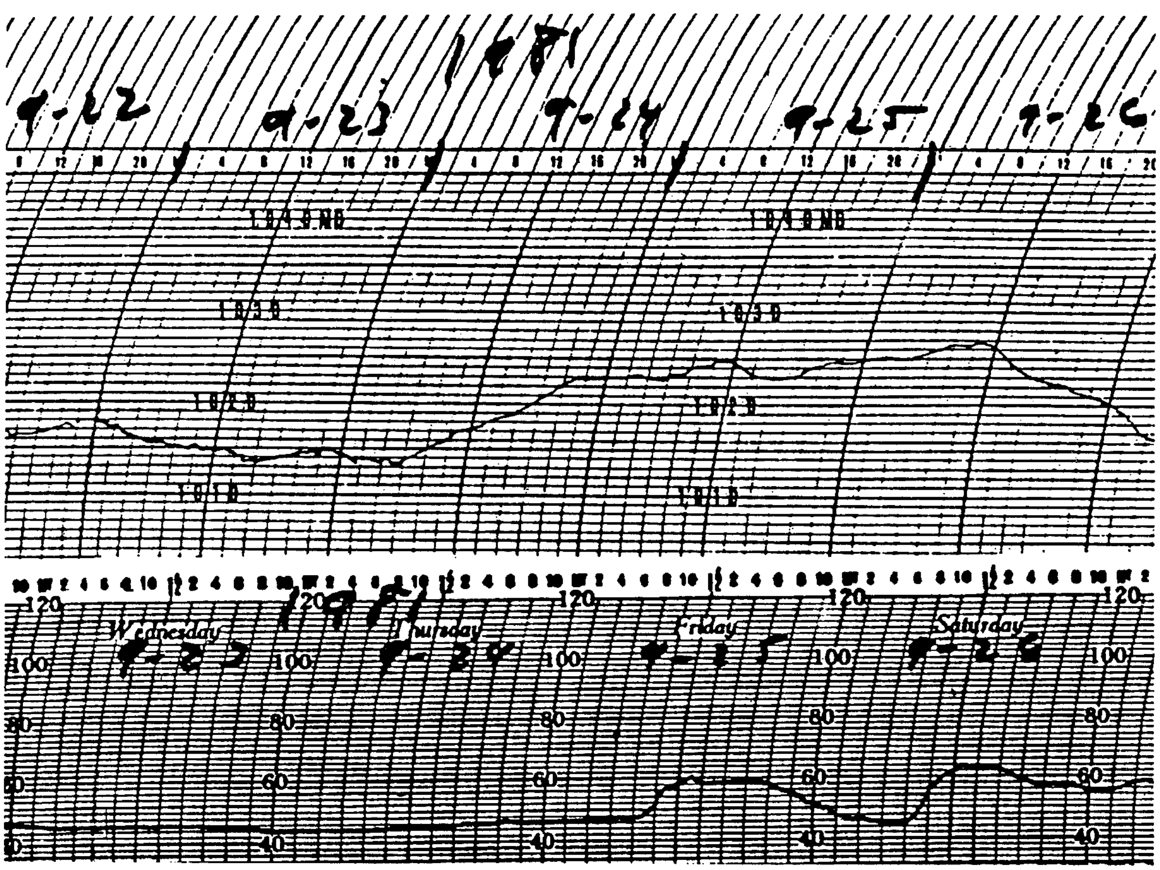


Figure 10. The effects of cloud cover upon daily temperature range, illustrated by a sequence of days recorded at Burlington, Vermont, in September 1981. A low pressure cell arrived over Burlington on the 23rd and departed late on the 24th. The sky was completely overcast, and temperatures changed by only a few degrees Fahrenheit during both days. High pressure built over the area on 25 and 26 September, skies cleared, and air temperature reached its maximum near and after 1200 hr, and minimum temperatures occurred near sunrise, producing diurnal temperature ranges of about 12–18°F.

sky, and they contain liquid water and water vapor, which are both strong longwave absorbers and radiators. Generally, lower, thicker, and warmer clouds intercept and return the most longwave radiation. High cirrus clouds, for example, return only about 4% more atmospheric longwave radiation than does a clear sky, whereas altostratus return 20% more than a clear sky and fog returns 25% more (Oke 1978) (Appendix D). Clouds, therefore, keep the surface cooler during the day by reducing shortwave radiation intensity, and they keep it warmer at night by reducing loss of longwave radiation to space. Four days of clear and overcast weather at Burlington, Vermont, demonstrate that the daily temperature range is highly damped on overcast days when compared with clear days (Fig. 10).

On a cloudless night, metallic objects radiate strongly to the clear sky and so often are colder (as viewed by a thermal imager or as sensed by a passive infrared IDS) than the air temperature. This can result in levels of thermal contrast in the background that are greater than those anticipated on the basis of air temperature. On overcast nights, the difference in air temperature between the cloud base and ground is less, and the background may be more thermally uniform.

Partly cloudy days with scattered cumulus or altocumulus clouds may be the most thermally variable of all. Frequent heating and cooling cycles occur as moving clouds produce traveling shadows that can cause rapid cooling at the surface as a cloud passes overhead, an event that may last only a few minutes. The resulting, but shortened, thermal heating and cooling patterns are similar to those on a clear day, but intensities are subdued. The thermal landscape changes rapidly from one of intense solar radiation between clouds, if the sun is high in the sky, to one of only diffuse solar radiation. This sudden change causes objects with high heat capacities to remain warm when in shadow, but other surfaces begin to cool more rapidly. Such rapid changes may cause a passive infrared IDS to alarm due to localized, abrupt increases or decreases in thermal radiance, but alternatively the IDS's proximity to alarm will be lower while the background is cloud-shadowed (Peck 1993b). Video-motion IDSs also have high rates of nuisance alarms on partly cloudy days because the same portion of the scene will alternate between being sunlit and shadowed (Peck 1989a). Detection capability will be highest during sunlit periods, as it will be most difficult then for an intruder to avoid casting a shadow.

Nonsoil surface temperatures

Except for conditions during and immediately after rainfall, nearly all energy reaching a pavement or rock surface is used to heat the surface; subsequently it is conducted into the material, convected into the atmosphere, or radiated into the atmosphere. No energy is used for latent heat. Pavement and rock surfaces have albedos of 10 to 35% (Geiger 1965, Arya 1988), similar to soil and vegetation surfaces (Table 2). However, pavements and rocks have high thermal conductivities, much higher than water or bulk soils, and heat capacities similar to wet soils. Therefore, their thermal diffusivities are high, giving them the ability to heat or cool quickly (Table 3). The result is that pavement and rock surfaces usually heat to temperatures much higher than the air on sunny days, but cool to only slightly below the air temperature at night because of their ability to conduct heat from the soil below. This effect is most pronounced during the winter when snow is compacted on a pavement or completely removed, thus allowing soils beneath pavements to be rapidly chilled, perhaps freezing and heaving, as heat is conducted to the atmosphere.

On days with discontinuous cloud cover, pavements and rocks heat and cool rapidly with the fluctuations in incident solar radiation. Such a dynamic thermal background scene can cause a high number of nuisance alarms by passive infrared IDSs if the changes in temperature are not spatially uniform on the scale of the IDS's thermal detectors. Lowering the IDS's sensitivity, to reduce the number of nuisance alarms, may allow an intruder to pass through undetected.

Landsberg (1981) recorded temperatures 14°C warmer at the surface of asphalt pavements than in the air above at a height of 2 m, near noon on a summer day in Columbia, Maryland. He also recorded temperatures 3°C warmer than the air over grass, 4°C warmer than the air over bare soil, and 1°C cooler than the air over a lake. At night, the air and the lake surface were the same temperature, but asphalt, grass, and bare soil were all cooler than the air by 2°C, 10°C, and 7°C, respectively (Landsberg 1981). Similar dramatic temperature differences occur for rock surfaces, even at high elevations. Geiger (1965) reports the surface of a 0.25 m^2 gneiss rock reaching temperatures about 19°C warmer than the air on a clear September day at 3050 m in Europe. Rock surface temperatures dropped to about 3°C cooler than the air at night. Lesser extremes were reported over smaller, thin rocks. On overcast days and

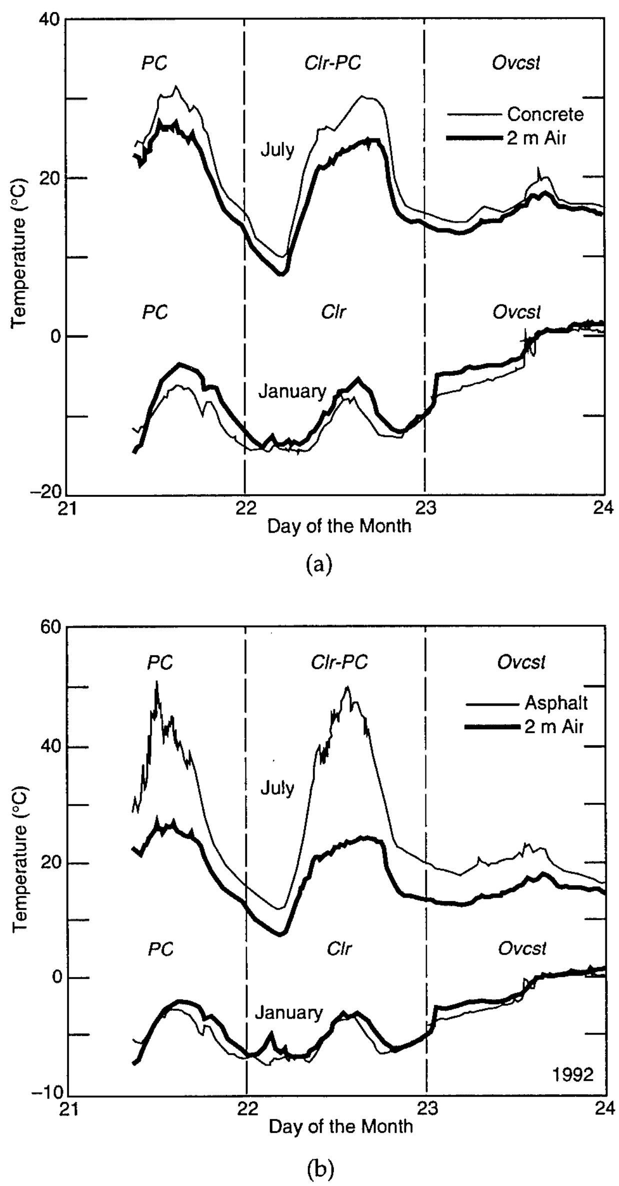


Figure 11. Surface and air temperatures during three days in January and July 1992 in New Hampshire over (a) concrete and (b) asphalt (Lacombe 1993). Air temperature was measured at about 2 m with a thermistor. The temperatures of concrete and asphalt surfaces (2.1 m × 2.1 m in-ground rectangular test cells) were measured with a radiometer. Clr = clear, PC = partly cloudy, Ovcst = overcast.

days with precipitation, rock surface and air temperatures are similar.

Changes in the difference between surface and 2-m air temperatures with daily weather condition and season are illustrated for three days in July 1992 and January 1992 (Fig. 11). Differences between surface and air temperatures are always greater on clear or partly cloudy days than on overcast days. This is best demonstrated over asphalt pavement in July where surface temperatures were about 25°C warmer than the air on partly cloudy days, but less than 5°C warmer than the air on overcast days. During the winter, day-to-day variations in surface and air temperature differences as a function of cloud cover were small because of the lower solar elevation and thus less intense insolation, and perhaps because of the shorter day length. This demonstrates that air temperature measured at about 2 m above the surface is often not a reliable indicator of surface temperatures, especially on sunny days.

Air temperature

Since the lower atmosphere is heated primarily by the conversion of radiation to sensible heat at the earth's surface, the largest temperature fluctuations always occur at the surface. Temperature fluctuations generally decrease with height above the surface because the atmosphere is heated through thermal transport processes only after the surface has been heated radiatively. As a result, air temperatures have a smaller daily range and show increasing thermal lag with height above the ground (Lowry 1967, Griffiths and Driscoll 1982). Thermal lag generally also increases with height above the ground because of the time necessary to heat the air convectively and radiatively by the surface. In relatively dry locations, surface temperature coincides closely with solar noon (Sellers 1965). However, 1.5 m above the surface, maximum temperatures are typically delayed up to 2 hours (Miller 1978).

Air temperature at weather stations is typically measured at a height of about 2 m above the ground surface. Thermometers are shaded by an instrument shelter, and the shelter is usually located over a grass surface. Temperatures within a few centimeters of the grass surface may be 5°C or more warmer than the air temperature at shelter height (Geiger 1965; Fig. 12). Much larger temperature differences occur between the surfaces of objects at the earth's surface and shelter temperature (Fig. 11). Deserts appear to have the largest temperature difference, or lapse, between the surface and shelter height. Oke (1978) reports temperature decreases of 28–29°C between desert surfaces at midday and air temperature at shelter height in Southern California, and temperature decreases of up to 27°C in the lowest 50 mm of the atmosphere over a sandy desert in Saudi Arabia. Deacon (1969) has reported midsummer shelter air temperatures 25–32°C cooler than tar-

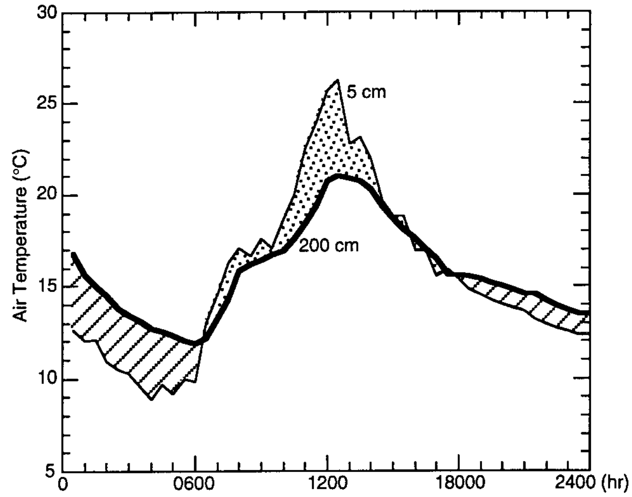


Figure 12. Temperature changes with height as measured at SOROIDS on 5 June 1992. The day was overcast and morning winds were less than 5 m s^{-1}. (a) Early morning air temperature at 200 cm was 2 to 3°C warmer than at 5 cm because radiative cooling occurs most strongly from the surface at night. At 1250 hr, air temperature at 5 cm is about 5°C warmer than at 200 cm because of surface heating by insolation. (b) The rate of change of temperature with height is the thermal lapse rate. Cooling temperatures with height, negative lapse rates as observed between approximately 0600 and 1530 at SOROIDS on 5 June 1992, indicate that air temperature at 5 cm is warmer than at 200 cm. Cooling at this rate indicates that there is strong surface heating by solar radiation and that the lower atmosphere is extremely unstable. During this period wind speeds might increase at the surface due to convection coupling surface air with faster moving air aloft. Before 0600 and after 1530 hr the lapse rate is positive, indicating that air at 5 cm is cooler than air at 200 cm. This indicates that an inversion exists and the air is stable—warmer air overlies cooler air. Decoupling occurs at this time because convection has slowed or stopped, causing wind speeds at the surface to diminish. Lapse rates of ±2°C m^{-1} are very large and demonstrate large temperature changes with height.

mac at the surface, and 20–26°C cooler than the surfaces of railroad rails. Even on a warm December day in New Jersey, during otherwise unspecified weather conditions, a temperature decrease of 4°C was found 22.5 cm above the surface at midday when compared with the temperature of the snowless surface (Mather 1974). These measurements all indicate that at midday, especially on sunny days, temperature measurements at shelter height often strongly underestimate air temperatures near the surface.

Over frozen or very wet surfaces, the vertical change of temperature will be only a few degrees because energy absorbed at the surface is consumed by latent heat for thawing or evaporation. Since the thermal gradient between the surface and shelter height is small over exposed frozen or wet surfaces, the shelter temperature provides a reasonable approximation of the surface temperature.

Temperature changes with height are also noticeable at night, but air temperatures typically increase rather than decrease with height above the surface, and differences are typically less than 50% of those observed at midday (Fig. 12). Because of this it is possible that the air temperature measured at shelter height may overestimate the near-surface temperature and consequently may not be a reliable indicator of when the soil surface is freezing, for example. On clear, calm nights, snow surface temperatures are often 5 to 20°C colder than the standard shelter air temperature. These differences are all smaller under overcast skies and may disappear completely during precipitation. The differences also disappear at least twice a day as the gradient changes from cooling with height during the day to warming with height at night (Fig. 12).

Air temperature, and particularly its rate of change, potentially influence IDS reliability through any temperature-dependent variation in the functioning of sensor or processor electronics. One example of this from SOROIDS was a severe reduction in detection capability of a fence-mounted, optical-fiber IDS on a day that was characterized by rapidly increasing air temperature (–18°C to 5°C in 8 hours). This IDS previously had been reliable (good detection capability for fence taps—a nondestructive substitute for cutting the fence) in the range of air temperatures it experienced on that March day, but it had not been subjected to such a high rate of temperature change during prior controlled intrusions (Peck 1994b).

Relief effects on air temperature. The shape of the land surface, or relief, can have a large impact upon daily air temperature. Several mechanisms responsible for the effect of relief upon diurnal temperatures include lapse rates, cold air drainage, inversions, and fogs.

The atmosphere typically cools with elevation above the general regional land surface because most atmospheric heating occurs at the surface. The "normal" rate of cooling as one rises through the atmosphere in a balloon or climbs a mountain is about –0.65°C per 100 m because of the dominance of surface heating. Therefore, assuming no effects from winds advecting warmer or cooler air, IDSs and security fences distributed about a large area with relief of several hundred meters will experience higher or lower temperatures simply because of the effects of elevation. This is an average, or "normal," decrease of temperature with elevation, referred to as the normal thermal lapse rate. Many relief-related factors are more important controls of diurnal temperature and dominate the lapse rate effects.

Midday temperatures tend to be higher on west-facing slopes in the midlatitudes than on slopes facing other directions. On all slopes the morning sun must often evaporate dew, frost, fog, and clouds before radiant energy can be used for sensible heating. As a result, by the time easterly slopes have begun to experience significant sensible heating, it is often midday, and the intensity of direct radiation has diminished. Westerly slopes, on the contrary, receive their strongest radiation in the afternoon and benefit from the warmer air temperatures at this time and the absence of fog or dew in the afternoon. One exception is that regions with frequent cumulus cloud development in the afternoon, especially from upslope flow of air, may occasionally have cooler westerly slopes than easterly slopes (Geiger 1965). Another is that warm air advected from southerly latitudes, such as air following a warm front, may completely obliterate diurnal heating effects, enabling the warmest temperature to occur at any time of the day.

As a result of these processes, in the midlatitudes southwesterly facing slopes are generally the warmest and northeasterly facing slopes the coldest (Geiger 1965). In the midlatitudes, south-facing slopes are also generally warmer than north-facing slopes because of the greater intensity of radiation on southerly slopes. North-facing slopes usually are more moist, frequently have more lush vegetation in arid regions, generally retain snow covers longer, and freeze deeper in the winter (though they do not necessarily have more numerous freeze–thaw cycles). Fences crossing from southerly and westerly facing slopes to northerly and easterly facing slopes may experience significant differences in the magnitude and timing of daily thermal regimes.

Cold air drainage is probably the most dramatic relief-related temperature effect. According to Charles' gas law, cooling air by 10°C increases its density by about 4% (Geiger 1965). Air masses with low atmospheric water vapor, such as a cold air mass that has invaded from polar continental areas, allow longwave atmospheric and terrestrial radiation to readily escape to space with little return radiation. The greatest cooling from these radiative losses occurs in the atmosphere high above valleys and from the surfaces of hilltops and mountaintops. The most intense cooling near the land surface occurs within a layer of air only a few tens of meters thick. This air layer increases in density as it cools and slowly moves downslope at speeds of a few meters per second (Barry 1992). Cold air over valleys also sinks slowly. This draining air accumulates in valley bottoms and along inflections and basins on the sides of hillslopes (Hogan and Ferrick 1993).

Temperatures often reach their lowest value at night in small hollows and valley bottoms, which accumulate cold air drainage, rather than on mountaintops. After intrusion of continental polar air masses into central Vermont, for example, early morning air temperatures of –35°C have been observed near sea level, but 1000 m higher, temperatures were only –18°C on a mountaintop. Hogan* has observed changes of temperature, on clear, calm mornings in New Hampshire, of 11°C with 60 m change in elevation and 400 m horizontal distance. He also observed cold air layers only 20 m thick. Hogan and Ferrick (1993) found the most dramatic temperature changes over continuous and deep (>30 cm) snow covers. Strong topographic effects have even been observed on relatively low relief. Mather (1974) cites observations made across Toronto, Ontario, on a winter night with temperature changes of 17°C over vertical distances of 50 m and a horizontal distance of 8 km. Temperature decreased in valley bottoms and increased on ridges.

Cold air drainage temperature effects are diminished by cloud cover, winds more than a few meters per second, or shallow (rather than deep) snow cover. They are also diminished by cold air dams along hillsides, such as buildings and wind breaks, that prevent air from freely flowing downslope. If the cold air drainage is sufficiently vigorous, however, air can pile up behind barriers, drain as a pulse, and then build again. Drainage typically is oriented by gravity directly down slopes, but it is occasionally directed by small valleys, buildings, and vegetation (Geiger 1965, Yoshino 1975). Active cold air drainage occasionally begins as early as about one hour before sunset in clear, calm conditions and typically ends 20–30 min after sunrise (Yoshino 1975). In large valleys, cold air drainage can become sufficiently organized to create a general down-valley wind, called a mountain wind, that can reach speeds of 3–4 m s^{-1} (Barry 1992). In some locations, these valley winds can be much stronger. They are documented, for example, to have been strong enough on Whiteface Mountain in the Adirondacks of New York that sufficient snow cover was stripped from ski slopes to force relocation of a ski area.*

Pooling of cold air in valley bottoms forces warmer, less dense air to rise. Air temperature thus increases with elevation in these situations rather than decreases. This reversed, or inverted, temperature change with elevation from the normal is an inversion. A fence line located on a hillside and thus passing through a valley inversion would experience increased temperatures with elevation until the top of the inversion was reached, then decreasing temperatures with elevation. Since warm air is less dense than cold air, and naturally wants to remain above cold air, an inversion is stable. That is, the air does not try to rise or sink, but remains stationary in the vertical. Therefore the coldest air tends to remain at the surface and does not mix with the warmer air above until they are mixed by some outside factor. Mixing can occur as a result of winds from aloft causing turbulence that mixes and warms the surface air, or from solar heating of the earth's surface, which heats the cold air from below, producing mixing as parcels of warm air rise through the cooler air.

Since inversions typically form in valley bottoms and there is usually considerable moisture at the ground surface in the form of streams, moist soil, and water vapor from exhaling animals and combustion, often the air humidifies sufficiently, and cooling is great enough that fog forms. Air temperatures within fog often stabilize near the dew point because of the release of latent heat as the fog condenses, and cooling ceases or continues at a decreasing rate.

Daytime temperatures are usually warmer in valley bottoms than on mountain tops simply because of normal thermal lapse rates (Anthes 1976). However, during clear days with intense solar radiation, valley bottoms often become considerably warmer than mountaintops and hilltops. This is most true if the valley is deep and its sides are steep. Solar radiation enters the valley and heats the valley floor. Longwave radiation is emitted by the surface but often does not escape readily because it is intercepted by the valley sides due to the restricted sky view factor. In addition, the valley has a large land surface area with a relatively small volume of air above the valley to heat. These processes create the fourneaux, or furnace, effect (Peattie 1936) because the valley walls and floor heat the air volume within the valley much as occurs within a furnace. This often leads to the formation of valley winds. The fourneaux effect during the day, and cold air drainage at night, cause valley bottoms to have larger diurnal temperature ranges than hilltops or mountaintops.

Vegetation effects on air temperature. The overall result of vegetation blocking solar radiation from reaching the surface and the near-surface air layer during the hottest part of the day, as well as preventing thermal loss from the soil surface and near-surface air layer at night, is that the largest

* A. Hogan, 1994, CRREL, Hanover, N.H., personal communication.

diurnal temperature changes occur near the top of the leaf canopy (whether it is grass, a corn crop, or a mature forest canopy) and the smallest changes occur near the bottom (Geiger 1965, Oke 1978). Geiger (1965) observed a 13°C diurnal change of temperature near the ground surface in a young fir plantation near Munich, whereas temperatures at the top of the canopy ranged over 19°C. Most vegetation surfaces respond similarly, and it is acceptable in most respects to view forest cover as larger versions of crop or low-growth vegetation cover (Oke 1978). The temperature profile within a vegetation cover can lead to a thermally heterogeneous background that a passive infrared IDS or a thermal imager responds to, but that is not visually apparent to an observer.

The above descriptions refer largely to closed-canopy, healthy vegetation covers. Open vegetation covers allow considerably more shortwave radiation and longwave sky radiation to reach deeper within the canopy, and they allow longwave radiation from the surface to exit the canopy because of the increased sky view factor of an open canopy (Oke 1987). Long droughts change the color of leaves, reduce evapotranspiration, and open the canopy as leaves curl, wilt, or drop, all of which change the vegetation cover's thermal characteristics.

Seasonal changes in the vegetation cover also affect its thermal regime. Leaf-out of tree canopies and the growth of annuals in the spring produce a vegetation cover that increases in density and/or height as the season progresses. Similar changes occur in the fall as leaves change color and are shed from deciduous trees and annuals die, dry, and eventually fall and decompose (Frodigh 1967). During these periods diurnal thermal regimes transition from a bare soil, or bare soil with a debris layer, to a mature vegetated canopy, and finally revert again to a bare soil with an open canopy. Each creates a different diurnal thermal environment.

An IDS located within a canopy will experience different thermal conditions at different locations within the canopy. A passive infrared IDS directed at the edge of a forest canopy or at the top of a low vegetation canopy, even one as low as mowed grass, will observe the greatest diurnal changes. Much more rapid changes occur if the vegetation blows in the wind so that lower, cooler portions of the vegetation are momentarily exposed to the thermal device; this is a likely cause of nuisance alarms by a passive infrared IDS viewing a grass cover on sunny summer days (Peck 1993b). A passive infrared IDS within a partially open forest canopy will observe small flecks of direct solar radiation moving within the canopy and along the forest floor as leaves blow in the wind and as the sun's position changes. The flecks move, change shape, and appear and disappear. For example, mature red pine and red beech stands typically allow less than 25% of incident solar radiation to reach the surface, with the largest amounts reaching the surface when the sun is directly overhead (Reifsnyder and Lull 1965). Finally, an even larger diversity of temperatures will be observed if drought dries the canopy, which curls leaves and allows even more sunlight through.

Urban areas. Temperatures are warmer in urban areas than in rural areas, especially at night. This urban "heat island" effect is a result of the preponderance of dry, rock-like surfaces in the urban landscape, the rapid removal of water by storm drains that could cool the area by evaporation, and the abundant release of heat by industry, automobiles, heating and air conditioning systems, and people (Landsberg 1981). Urban areas need not be large for this thermal effect to be noticed. Even small groupings of buildings have measurable heat islands that can elevate evening temperatures above those of surrounding rural areas. Clusters of structures on most IDS-protected installations will create some heat island effect. Walls and building surfaces, especially those of stone or brick, show the largest contrast in temperature, which is usually most noticeable 2–3 hours after sunset (Landsberg 1981). For small clusters of buildings the effect often disappears after midnight, but for larger clusters of buildings, and structures where heating systems and industrial activities operate all night, the effect persists.

Poorly insulated buildings during the winter, and thick masonry walls exposed to the sun for long periods, also show as thermal anomalies, especially to passive infrared IDSs. Buildings with inadequate insulation lose large amounts of heat through walls and roofs. Thermal infrared cameras are usually used to detect "hot spots" in walls and roofs to evaluate a building's thermal efficiency (Tobiasson et al. 1977, Flanders 1987). A passive infrared IDS will also detect heat leaks through building exteriors. Depending on the spatial distribution and the rate of change in the intensity of the heat leaks with changes in air temperature, or change in the structure's interior temperature as usage varies over 24 hours, this could produce IDS nuisance alarms. Thick masonry walls can also produce thermal patterns that could cause a passive infrared IDS to alarm. For example, cinder block walls are composed of mortared joints with high thermal mass and hollow cavities. Material surrounding the cavities will heat and cool more rapidly and will reach higher daytime temperatures than will the mortared and thicker areas of block. However, if the cavities in a block fill with water, they will remain cooler than solid block areas during the day due to higher thermal mass and evaporative cooling as moisture seeps through the porous block. Conversely, the cavities may appear warmer at night because of the large thermal mass of the water and block. A wall that visually appears to be a uniform background may be a complicated background for a passive infrared IDS or a thermal imager.

Wind effects

Wind can have a dramatic impact on IDS operations and can significantly modify thermal regimes by promoting mixing, which reduces temperature contrasts. Wind is a result of atmospheric pressure differences from one location to another. Winds of concern to IDS operations are those that arise from local convective systems on diurnal time scales and macroscale winds that are driven by the passage of large, regional pressure systems.

Diurnal winds form, change in speed and direction, and subside as a result of daily heating and cooling patterns. They usually do not carry over continuously from one day to the next, although they can occur on consecutive days. That is, diurnal winds rarely last longer than the period of daylight or darkness, depending upon the thermal regime driving them. Nuisance alarms by fence-mounted IDSs may duplicate diurnal wind patterns, although this is modified by temperature-related changes in fence stiffness that may not coincide with the wind patterns. Nuisance alarms with ground-motion IDSs caused by wind-induced motion of above-ground objects coupling into the ground also follow diurnal wind patterns. Video-motion IDSs will have wind-related nuisance alarms on sunny days if movement of sunlit objects, such as flags or tree leaves, changes the distribution of shadows in the camera scene. If standing water is nearby, wind-induced ripples will cause the reflection of visible, microwave, or infrared radiation to vary, which may cause nuisance alarms by video-motion, microwave, and passive infrared IDSs, respectively. Moving water above buried electromagnetic IDSs may also cause them to alarm.

Diurnal winds are geographically local and are typically not as strong as regional winds created by synoptic-scale weather patterns. In fact, diurnal winds usually only appear when synoptic-scale winds are weakest. Occasionally, exceptionally strong and destructive local diurnal winds occur, such as katabatic (downslope) winds off glaciers, but they are relatively uncommon. Diurnal winds are thus most evident during calm, clear conditions typical of the center of regional-scale high pressure cells that often produce clear skies and weak pressure gradients.

Local winds are driven by changes in the temporal and spatial heterogeneity of energy budgets, which are evident through surface temperatures. As a result, local winds are strongly dependent upon the local weather, local geographical patterns of vegetation, relief, and land and water bodies, and preceding weather conditions that modify soil moisture and any snow cover.

As discussed in the context of the diurnal energy budget, heating of the earth's surface, due primarily to solar radiation, warms by conduction the layer of air immediately at the surface. That air then rises (convects) if it is warmer and less dense than surrounding air (positively buoyant or unstable). Air that sinks because it is colder, and thus more dense, than surrounding air is negatively buoyant and stable. Parcels of air at the same temperature as surrounding air are neutrally buoyant and also stable, and so do not rise or fall because of density differences. Thus, air that rises is unstable, whereas neutrally or negatively buoyant air is considered stable. Stability and instability are of major importance in driving winds at all scales in the atmosphere, but they are particularly obvious in local winds.

As unstable parcels of air rise from the earth's surface, they become entrained within air that is moving horizontally—the higher-altitude regional winds. The regional wind consumes energy in accelerating the air parcel, slowing the regional wind in the vicinity of the convecting parcel and accelerating further the parcel of rising air. Since the top of the rising parcel of air is immersed within higher-speed regional winds, the top accelerates, effectively dragging the bottom of the parcel along. That is, momentum is transferred from the higher-altitude regional winds down to the surface. Rising parcels of air, through convection, thus accelerate surface air and decelerate

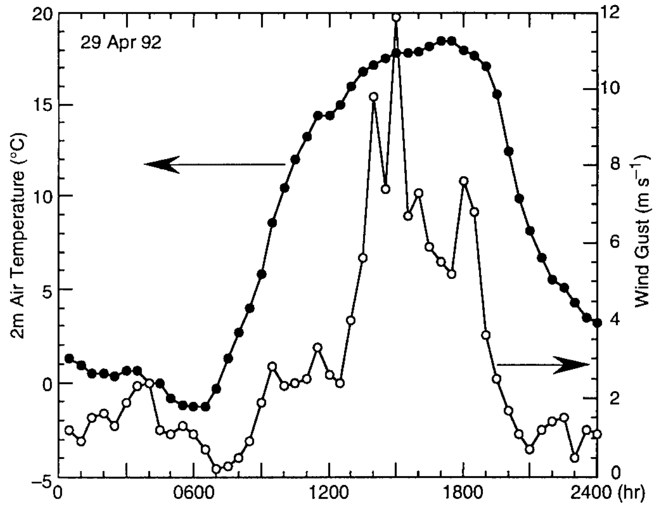


Figure 13. Increase of 2-m wind speed during hours of surface heating due to rising parcels of warm air convecting from the surface and coupling surface winds with winds aloft. Nocturnal decoupling occurs because convection ceases shortly after sunset and surface winds slow. Measurements were made at SOROIDS on 29 April, 1992.

regional winds aloft because the parcels couple the regional winds to surface air. High-speed surface winds created by this process are favored by cloudless skies, which allow strong daytime surface heating by intense solar radiation. These winds begin several hours after sunrise because of the time necessary for the surface to heat and the coupling process to become established.

Wind speeds at night are generally low because heating at the surface, and thus convection, ceases. As surface cooling occurs at night, inversions often form near the surface and enhance stability. Air at the surface becomes either neutrally or negatively buoyant, and surface winds decouple from the regional winds. The surface winds thus decelerate and may even calm when convection ceases. In contrast, nighttime regional winds that are many meters above the surface accelerate because there is no drag on them by rising air parcels.

Mather (1974) cites 13 years of comparisons between wind speeds at 67 m above sea level at San Francisco and wind speeds at 794 m elevation on Mt. Tamalpais 15 km away. At 0600 hr, wind speeds at San Francisco averaged 3 m s^{-1}, whereas on Mt. Tamalpais winds averaged about 8.5 m s^{-1}. After sunrise, wind speeds converged until about 1300 hr, when speeds at both locations became approximately equal, about 5 m s^{-1}.

Arya (1988) cites measurements of 40-day averaged wind speeds at 8 m, 50 m, 100 m, 200 m, and 500 m at a smooth rural site in Australia. Wind speeds at 8 m showed a night-to-day increase from about 4 to 6 m s^{-1}. At 50 m, wind speeds changed from about 7 m s^{-1} at night to 6 m s^{-1} during the day. From 100 to 500 m, wind speeds averaged between 8.5 and 10.5 m s^{-1} at night to between 6.5 and 7.5 m s^{-1} during the day.

Wind speeds at SOROIDS are measured at heights of 2 and 4 m. The diurnal pattern described above is very evident at the site, even in April, a time when soils are typically quite wet (Fig. 13). Surface winds remain less than 2 m s^{-1} throughout the night, but increase to maximum gusts of about 12 m s^{-1} at midday; they decline again to less than 2 m s^{-1} within 2 hours of sunset. Gustiness is often observed midday at SOROIDS because of the discontinuous nature of convective activity. Wind speeds increase and decrease as parcels of air rise one after another. This diurnal cycle of wind activity causes a corresponding variation in the wind loading of the fence (see Appendix A, Fence-mounted IDSs) and consequent wind-induced fence motion. Fence-mounted IDSs are more prone to nuisance alarms during the windy portions of the day, with the frequency as well as speed of wind gusts influencing proximity to alarm, although the temperature-related stiffness of the fence and the rigidity of the fence posts are also factors (Peck 1992b, 1993a).

Urban winds

IDSs located in urban areas may experience a subdued diurnal wind regime, or perhaps even a reversal from the rural patterns discussed above, because of the structure of cities and because of the urban heat island, which is most pronounced at night. As in rural areas, during daylight hours wind speeds often increase at the surface in cities, but to a lesser extent because of friction at the surface. However, at night the thermal contrast between urban and rural areas increases. Less nocturnal cooling occurs in cities than in the country (Landsberg 1981). The urban heat island often produces a thermal plume that rises from the city surface into the regional winds, coupling the surface at night to higher-speed regional winds above the city. As a result, wind speeds in the city often decrease less at night than in rural areas or may even increase. This is especially true of larger urban areas where the heat island is more pronounced and where tall buildings and deep, narrow canyons can cause significant turbulence and acceleration of wind through the streets. Consequently, fence-mounted IDSs close to clusters of buildings or in urban areas may show a different pattern of nuisance alarm activity than do similar IDSs in more open locations.

Topographically driven winds

Land and sea breezes are local winds usually found only along the shores of large bodies of water, such as sea coasts and lakes with more than 250 km^2 of water surface (Griffiths and Driscoll 1982). As with most diurnal effects, land and sea breezes are most prominent when solar radiation is strongest and regional wind speeds are weakest. On many low-latitude coasts where solar radiation is strong, sea and land breezes are regular patterns of the daily weather. In the midlatitudes, they are generally weaker and less frequent. They are also principally a summer phenomenon.

Sea breezes occur only along narrow strips of coastline and usually extend inland a few kilometers along lakes, but perhaps several tens of kilometers along the ocean (Oliver and Hidore 1984). The strongest sea breezes are reported to extend up to 60 km inland (Smith 1975). Sea breezes are simple thermal winds. Due to thermal differences between land and water, convection begins over the land. As heated air rises, a pressure gradient is established between the warm land and the colder water where convection is not occurring. Cool, dense air is drawn from the water to the shore as a sea breeze front that progresses inland at a rate of 1 to 7 m s^{-1} (Barry and Chorley 1970, Smith 1975). The depth of a sea breeze circulation is typically less than 1 km (Barry and Chorley 1970, Smith 1975, Griffiths and Driscoll 1982). Sea breeze fronts generally begin a few hours after sunrise, though they may initiate later in the day, depending upon soil moisture and cloud cover. Though the primary driving mechanism typically ends at sunset, sea breeze fronts have been observed to continue inland well after sunset (Smith 1975). The sea breeze not only produces a change in wind direction and an increase in wind speed, but it also lowers air temperature, increases humidity, and may trigger afternoon thunderstorms at the frontal boundary where warm air is pushed aloft.

The land breeze occurs at night and is much weaker than the sea breeze. Air moves from land to water because land surfaces chill more quickly and to a lower temperature than do water surfaces, thus establishing a pressure gradient reversed from the sea breeze. Wind speeds are typically lower in a land breeze, 0.5 to 1.4 m s^{-1}, because land–water temperature differences are smaller at night and because there is greater friction onshore where the breeze originates (Griffiths and Driscoll 1982).

Sea breezes can cause rapid and dramatic cooling in coastal areas. For example, at Chicago the sea breeze has been observed to develop on 36% of summer days. Temperatures occasionally are 16°C lower than inland values after the sea breeze front arrives, and temperatures can cool by 4°C in one hour (Smith 1975, Oliver and Hidore 1984).

Mountain and valley winds are diurnal reversals of airflow along valley sides and longitudinally along valley axes. Valley breezes occur during daylight hours during clear summer weather, and generally begin within an hour after sunrise (Barry 1992). They result from intense solar heating of the valley sides, which heats air adjacent to the valley walls, lowering the air density and increasing its buoyancy. This air rises up the valley sides as upslope winds. The upslope winds eventually turn and flow up the longitudinal axis of the valley along both the valley walls and the valley floor. Upvalley wind speeds are usually stronger along the valley longitudinal axis than along the slopes, with the highest speeds usually in a jet several hundred meters above the valley floor (Barry 1992). Maximum upvalley wind speeds several hundred meters above the surface are frequently about 3 to 4 m s^{-1}, with some peak gusts of 10 to 11 m s^{-1} reported (Yoshino 1975). The upvalley wind typically ends within an hour after sunset (Barry 1992).

Mountain winds blow at night, typically beginning about an hour after sunset at the end of the valley wind and ending at about sunrise (Barry 1992). They flow from the mountain sides down to the valley and result from rapid radiative cooling of the high-elevation ground surface and chilling of the air above it. The air eventually cools sufficiently to become more dense than air at lower elevations, and so, because it is negatively buoyant, it sinks. Though adiabatic heating occurs as the air sinks, it remains sufficiently cold to continue flowing as a downslope wind. When the sinking air travels the longitudinal axis of the valley, it is called a downvalley or mountain wind. Maximum wind speeds are typically up to 4 m s^{-1} at heights of 100 to 400 m (Barry 1992). Mountain winds frequently flow in pulses downslope as the downward-draining air encounters obstacles and periodically overtops them. These surges are reported to occur with varying frequency, from 25 minutes between pulses to several hours (Barry 1992).

Though the speeds of mountain and valley winds are low, especially near the ground surface, their daily reversal of direction and their changing velocities could shake IDS fences, especially if the fences are on the valley sides.

Leaf whirls and dust devils

Local diurnal heating, especially during clear, dry weather in late summer and fall in the midlatitudes and arid regions, can create leaf whirls and dust devils. Leaf whirls are small, rotating windstorms that are approximately 1 to 2 m in diameter and a few meters high. They rotate at about 0.8 to 1.4 m s^{-1}, move across the ground at speeds up to 0.5 m s^{-1}, and last from a few seconds up to a minute.* They are usually observed in the fall in the midlatitudes as leaves are whirled in the air. Leaf whirls are rotating low-pressure systems, like tornadoes and hurricanes, though much smaller. In addition, they are not necessarily caused by heating at the earth's surface; instead, eddies and turbulence on a gusty day can cause them. As a result, they are most common on windy and gusty days rather than during calm weather. Though not significant storms, they can shake security fences, perhaps causing fence-mounted IDSs to alarm, and may cause loss of alignment of bistatic microwave or beambreak IDSs, making them more prone to nuisance alarms.

Dust devils are larger than leaf whirls and often appear to be small tornadoes. They are small rotating systems that are produced where the ground is heated to a very high temperature, such as over deserts and dry, plowed fields. As a result they occur frequently during cloudless, sunny weather instead of the thunderstorm conditions that spawn tornadoes. Dust devils typically are 10 to 20 m in diameter, rotate at 8 to 14 m s^{-1}, move at speeds of 1.6 to 5.6 m s^{-1} over the ground, and last 5 to 12 minutes* (Sinclair 1964). They usually do not extend more than 2000 m above the ground. Dust devils are sufficiently powerful to pick up large amounts of dust, and often tumbleweeds and other small debris, and roll a person along the ground. Consequently, they are certainly able to severely shake or even damage security fences and IDS hardware.

* D.L. Jones, 1977, Southern Illinois University–Carbondale, personal communication.

Moisture patterns

Though some moisture-related diurnal phenomena have been discussed, there are a number of diurnal condensation patterns that are noteworthy because of their potential deleterious effects on IDS performance. These include patterns of cloud cover, precipitation, and fog. As noted several times earlier, diurnal effects are most pronounced within regional high pressure when the atmosphere is calm and generally clear. This is also true of all of the following phenomena.

Cloud cover effects

Clouds often have distinct diurnal patterns under regional high pressure. This is especially true in the summer. Clear or hazy humid days allow abundant solar radiation to reach the surface, causing parcels of air to rise convectively. These parcels contain large amounts of water vapor if a humid air mass has settled over the region, as is common on the trailing side of a high pressure cell in the midlatitudes. The rising air expands and cools adiabatically. If the parcel of air cools to the dew point, condensation begins and liquid cloud droplets form. The latent heat released warms the air parcel within which the condensation is occurring, thereby increasing its positive buoyancy and allowing it to rise higher than it may have, had condensation not occurred.

Once condensation begins within the rising air parcel, a cumulus cloud forms that will grow until it reaches neutral buoyancy (see Appendix D). If neutral buoyancy is reached before the cloud is about 1000 m thick, precipitation is unlikely and the cloud is what is commonly called a fair weather cumulus cloud. Fair-weather cumulus typically have bases less than 2000 m above the ground and grow in patterns, with the distance between clouds nearly equal to the width of the clouds. Drifting fair weather cumulus reach maximum size in mid- to late afternoon and cast cooling shadows along the ground that can cool the temperature of the air quickly by 3°C or more. Thus IDS systems are often alternately heated and cooled as these drifting clouds pass overhead on warm afternoons. This has an adverse effect on performance if the IDS's electronics cannot rapidly adjust to the prevailing temperature.

Occasionally, if the upper atmosphere is sufficiently cold or if sufficient latent heat is released, these cumulus will continue to grow vertically. This may allow the cloud to grow to thicknesses much greater than 1000 m; in extreme cases, cloud tops may reach 10,000 m above ground level with bases still within 2000 m of the surface. These may become afternoon air mass thunderstorms, which can be severe, with strong, gusty winds and downbursts, heavy downpours, lightning, and hail. Tornadoes are occasionally spawned by these storms, but they are more common within well-organized storm systems created by regional low-pressure cells. Though severe, air mass thunderstorms usually are not long-lived and do not remain active long after sunset. They can survive for hours after sunset, however, if there is sufficient residual heat at the surface following a warm day. These afternoon storms are also common along coasts where the local sea breeze front may trigger the storms. Air mass storms are rare in the morning.

Convective activity creating fair weather cumulus clouds and air mass storms weakens at night and ceases by early morning because there is no longer a heat source at the surface to encourage convection. As a result, the cloud tops cool radiatively to the cold upper atmosphere and to space, becoming dense and negatively buoyant. By morning the previous day's cumulus have often sufficiently contracted vertically and evaporated that they have become stratus clouds, or they may have completely evaporated.

Dew and frost

Another common diurnal moisture pattern is the formation of dew or frost on a surface. Dew and frost nearly always form at night and survive until shortly after sunrise when the surface has been heated by the sun sufficiently to melt and/or evaporate the moisture. Dew is the condensation of water vapor from a gas to the liquid form. Frost, however, is the direct deposition of water vapor to ice, without passing through the intermediate liquid form (Ryerson et al. 1994). Frost forms fine, often delicate, ice crystals rather than clear, frozen drops as would occur if dew had simply frozen.

Dew and frost only form when objects at the earth's surface cool to below the dew point or the frost point. The two conditions that encourage this to occur are moderate levels of atmospheric water vapor and an effective nocturnal cooling mechanism. Water vapor sources can be advected humid air or evaporation of water from local sources. The latter situation is most common in autumn when the soil is still warm from the summer and the air is cold, especially at night. Warm, moist soil and warm water bodies such as lakes and ponds evaporate large quantities of water vapor into the air as the result of the large vapor-pressure gradient between their moist surfaces and the dry air.

The nocturnal cooling mechanism is radiative. Escape of longwave radiation to space allows rapid cooling at the surface. Low to moderate atmospheric water vapor promotes cooling, for water vapor blocks the escape of longwave radiation. Therefore, the requirements of water vapor and loss of thermal energy radiatively are mutually exclusive conditions. Since the cause of cooling is nearly always radiative, dew and frost are favored by clear nights that allow ready loss of surface energy to the atmosphere radiatively.

Frost and dew are spatially heterogeneous, forming with greatest intensity on objects that cool the most. Objects exposed directly to the night sky lose energy most rapidly. Although exposure to the night sky is usually necessary to form dew and frost, it is often not sufficient. At some locations, heat lost through rapid radiation to space may be replaced from another source. For example, bare soil exposed to the atmosphere often conducts heat from depth to the surface nearly as rapidly as it is lost radiatively, producing little cooling at the soil surface. Dew and frost rarely form on pavements in the fall and early in the winter for this reason. On the other hand, leaves and stems of plants, even if lying on the ground, are often separated from the soil surface by a highly insulating litter or duff layer. Since soil heat cannot reach the surface, the vegetable matter exposed to the night sky chills rapidly and becomes thickly covered with large deposits of frost or dew. A spatial pattern of deposited dew and frost eventually forms as a result of the varying exposure of objects to the sky and their ability to receive heat from other sources. Security fences, for example, often become coated with dew before the bare ground or the concrete fill that anchors the fence posts does. Dark-colored or black surfaces usually radiate heat more effectively than smooth, metallic-colored objects. Therefore, a black-painted fence will cool more deeply at night and frost more heavily than will a galvanized fence.

Dew or frost that evaporates in place probably is unlikely to cause nuisance alarms with fence-mounted IDSs because the weight loss would be gradual, not leading to an abrupt rebound of the fence or sensor cables. Water that drains to low points on the fence panel and drips off, however, may cause nuisance alarms when it hits the sensor cables of fence-mounted IDSs. Whether water that collects on a sensor cable itself is a source of nuisance alarms as it drips from the cables should depend on how firmly the cables are attached to the fence panel and on how taut the cables are. Placing the sensor cables of fence-mounted IDSs in conduit would avoid this potential source of nuisance alarms. A disadvantage is that the conduit itself would be a larger collecting surface for snow, which when it dropped from the conduit (perhaps during wind-induced motion of the fence) could in turn cause nuisance alarms upon hitting IDSs installed lower on the fence.

Conditions during which dew forms vary somewhat with season and location. Measurements in the Netherlands in September and October indicate that dew formed when relative humidity was 94% or higher, air temperature and dew point temperatures were separated by a degree or less, and wind velocities were 1 to 1.6 m s^{-1} (Janssen and Römer 1991). Dew did not form when relative humidity dropped to 80–85%, the dew-point temperature spread increased to 2°C or greater, and wind velocities increased to over 2.5 m s^{-1}. In general, too much wind causes mixing of the air and does not allow the surface to cool to the dew point. In addition, higher winds mix dry air from aloft down to the surface, thus reducing humidity. Yet, some air movement is necessary for dew formation because moist air must be carried to cold objects for sufficient condensation to occur. Therefore, dew and frost usually form under only a narrow range of wind conditions.

Once dew or frost forms, air temperatures will not rise significantly in the morning until most of the moisture has evaporated away (Baram and Azar 1992). When moisture is present at the earth's surface, nearly all of the solar radiation received will be consumed evaporating that water until it is gone. Usually evaporation or sublimation takes less than an hour for dew or frost, with a similar delay in daytime heating of the surface. Consequently, the thermal appearance of the background is uniform and remains so until the dew or frost has evaporated. Once that has occurred, the background objects heat differentially according to their different thermal properties.

Fog

Fog by definition is a cloud that forms along the ground; it is identified by its method of formation. There are several types of fog, but only a few recur diurnally and then only during the clear and calm conditions associated with high pressure.

Steam, or convection, fog is a nocturnal fog that is most common in the fall in the midlatitudes, but it can occur whenever cold, dry air moves over a warm water body. The warm water surface heats the layer of cold air contacting it, making the surface air less dense and causing it to rise, carrying water vapor with it. However, the rising air is immediately chilled by contact with the cold air above. The rising air cools below the dew point and produces wisps of rising fog that look like the streamers of steam that rise off a pot of water being heated on a stove (Appendix D). Steam fog does not require wind, but it can form in much higher winds than most fogs. Steam fog has been observed at sea in winds of 5 m s^{-1} and higher. Winds rarely carry it for any distance onshore, and thus it has little effect on IDSs unless they cross or are on the shore of water bodies. Steam fog is strongest at night when the air is coldest. It does occur during the day, but with less intensity than at night. Occasionally the morning evaporation, or sublimation, of dew and frost from solar heating will also produce a thin steam fog that persists until all of the deposit has vanished.

Radiation fog is the most common diurnal fog, forming at night under clear and nearly calm skies but becoming most obvious in the morning after having cooled during the entire previous night. Radiation fog forms as a result of nocturnal radiation cooling. Generally, a cold, dry air mass exists over the location before the fog forms, allowing radiation to freely escape the surface. Moisture from the surface evaporates into the air only to be cooled below the dew point, forming a usually thin, horizontally banded fog near the surface (Appendix D). Radiation fogs can be very shallow, perhaps only a few meters deep. They generally form most intensely over stands of high grass and often over swampy ground where convection fog occasionally contributes. Radiation fog rarely forms intensely over roads or other surfaces where subsurface soil heat can readily replace heat lost radiatively. As a result, radiation fog is geographically dispersed over surfaces that are thermally isolated from soil heat and have moisture available for humidifying the air.

Valley fog also forms at night and dissipates within a few hours after sunrise, depending upon its thickness (Appendix D). It forms because of drainage, or katabatic flow, of air that has been cooled on mountainsides or hillsides at night. These mountain breezes gradually fill the valley bottoms with cold air that continues to chill radiatively (see Wind Effects above). This process generally produces a temperature inversion, with warmer air in the valley bottom being forced to rise as it is displaced by cold air. Water bodies and moist soil in the valley then humidify the cold air below the inversion. When the air mass below the inversion reaches the dew point through a combination of cooling and humidification, a fog forms that fills the valley bottom. Valley fogs are often deep and thick and, depending upon the valley's orientation to the rising sun, may not dissipate for several hours after sunrise. Winds are generally light in valley fogs with sufficient air movement only to thoroughly mix air below the inversion.

As a general rule, fogs are expected to form whenever the air temperature and dew point temperature are separated by less than 2°C. This produces a high relative humidity that can spontaneously form a fog within only a few minutes. Once a fog forms, especially if it is thick, the air temperature generally does not drop significantly below the dew point temperature, because the latent heat that is released continuously as the fog forms keeps the air warm but at the dew point. Weather forecasters often use this phenomenon to predict the nighttime low temperature.

Horizontal visibility within a fog is usually 1 km or less. However, visibility within a fog varies considerably depending upon the direction of view. Fogs are thickest when they are viewed parallel to the ground, even though they are often thin at the ground surface. However, they often appear thin, or appear to disappear when viewed from below or, occasionally, from above. This is particularly true of radiation fogs. This is because fogs are typically thin in the vertical, but cover large horizontal distances. Visibility also varies with time, as the size and quantity of water drops change during the formation and dissipation of fog. The evolution of a fog is evident when its liquid water content (total volume or mass of droplets per unit volume of air) is monitored.

Depending on its severity, fog may adversely affect several types of IDSs. Reduced visibility in fog diminishes the detection capability of video-motion IDSs and reduces the effective range of surveillance cameras. Although an intruder's thermal contrast with his background may be high in fog, attenuation of that contrast over the distance separating the intruder and a passive infrared IDS may render the thermal contrast insufficient for detection by the IDS.* Lacombe combined results from models of thermal infrared transmittance (Electro-Optical Systems Atmospheric Effects Library [EOSAEL] [Duncan et al. 1987]) and detection capability of a particular passive infrared IDS (Lacombe 1992, Lacombe and Peck 1992). He determined that fog had a minor impact on intruder detection for ranges less than 50 m, provided visibility was greater than 500 m. If the intruder were to make use of any of several simple countermeasures that are effective in fog, then he would probably pass undetected. For an active system, such as a near-infrared beambreak IDS, transmission loss in fog is a major cause of nuisance alarms and may place a maximum limit on the practical length of the detection zone in fog-prone areas (Peck 1994a). Because fog density can vary with height, the level of obscuration at a person's eye level may be significantly different from that nearer the ground or at the top of a fence. This means that, for example, a near-infrared beambreak system located at the ground surface to detect crawling intruders may not be experiencing the same transmission loss as a second beambreak system mounted on top of a fence to detect climb-over intrusions (Peck 1994a).

QUASI-PERIODIC WEATHER EFFECTS

Quasi-periodic weather changes occur over periods of days and are neither primarily seasonal nor diurnal in nature, though their intensity and certain other characteristics do change diurnally and seasonally. Quasi-periodic weather events are driven by perturbations in the global weather circulation system, which are evident on weather maps as migrating pressure systems. These migrating pressure systems are large storms, typically the lows on weather maps, and large areas of fair weather, typically the highs on weather maps. The frequent perturbations in glo-

* J. Lacombe, 1994, CRREL, Hanover, N.H., personal communication.

bal circulation that give rise to quasi-periodic weather are most common in the midlatitudes. Understanding the reasons for this requires a basic understanding of global atmospheric circulation.

General circulation

The atmospheric circulation system is thermally driven and modified in its structure by the earth's rotation and the distribution of land and water over the planet. As a result, an understanding of general circulation draws upon the discussion of the daily energy budget presented earlier.

Since the earth is a spheroid with the equator on the plane of the ecliptic, the sun's rays intercept the earth's surface more directly at the equator than at the poles on the average over the year. The intensity of the solar flux is a function of the trigonometric sine of the elevation of the sun above the horizon. As a result, considerably more solar flux is intercepted by the earth–atmosphere system in the equatorial latitudes than in the polar regions. Consequently, the earth–atmosphere system gains more energy from shortwave solar radiation equatorward of 35° north and south latitudes than it loses radiatively to space, resulting in a net gain of energy over the year. The reverse is true of latitudes poleward of 35° (Musk 1991). If there were no processes for redistributing thermal energy throughout the atmosphere, the tropics would continually grow warmer and high latitudes would cool to very low temperatures. This does not happen.

The atmosphere moves in response to excesses and deficiencies in thermal energy from one location to another. Heating causes air to expand, decreasing its density and increasing its buoyancy. Cooling causes air to contract, increasing density and decreasing buoyancy.

In the tropics air is heated, expands, and rises vertically because of its buoyancy. As it rises, the total mass of air in the atmospheric column changes little, but the mass is stretched over a great height, from the surface to about 15,000 m. In addition, as the air is rising, more air is leaving the column at the top than is entering at the bottom. There is little friction at high altitudes hindering air from leaving the column, but high friction at the earth's surface slows the entrance of replacement air into the column. Thus, at the earth's surface air pressure decreases as more air exits the top of the column than enters the bottom. In the tropics the center of the atmospheric mass is at about 6000 m, with ~50% of the atmospheric mass below and ~50% above.

In the polar regions, the atmosphere is much colder on the average than in the tropics. Nearly the same total mass of air exists in the polar atmospheric column as in the tropical column (the total mass of air in the atmospheric column rarely varies by more than ±5% over time and from place to place). However, the cold polar air is more dense, so the same mass of air occupies a smaller volume; that is, the height of the air column is less, approximately 12,000 m, and the center of the atmospheric mass is at about 5000 m. The air in the column is trying to sink and diverge at the surface. Air pressure increases at the surface as more air enters the top of the column, where friction is less, than exits the bottom at the earth's surface where friction is greater. Since air pressure is the mass of air pressing down from the atmospheric column above any location, air pressure at 6000 m in the tropics is higher than air pressure at the same altitude in the polar regions (where the overlying mass is less at that altitude). As a result, the pressure surfaces are inclined, and at high altitudes the pressure gradient is from the tropics to the poles, and warmer air at higher pressure at high altitudes in the tropics moves poleward to replace colder air leaving the poles. Conversely, at the earth's surface, air moving down the pressure gradient (high to low) flows from the poles to the tropics, thus advecting cold air to the tropics (Fig. 14).

If there were no other factors, then circulation in the Northern and Southern Hemispheres would each consist of one large convection cell with air moving equatorward at the surface, and poleward aloft. Though this could occur, and does in concept, the overall motion is complicated considerably by the earth's rotation (the Coriolis effect) and the distribution of land and water. The general movement of air poleward and equatorward, and these other complicating factors, are responsible for the quasi-periodic traveling pressure systems characteristic of weather in most parts of the world, but primarily in the midlatitudes.

The surface area of the earth becomes smaller poleward. This is graphically evident on globes because the meridians converge toward the poles and the parallels become shorter. As a result, air moving poleward at higher altitudes from the tropics is compressed into a smaller and smaller area, whereas air moving equatorward spreads out into a larger surface area. Since, when moving poleward, area is decreasing and volume is constant, then the vertical dimension must increase. This effect is most noticeable at 25 to 30°

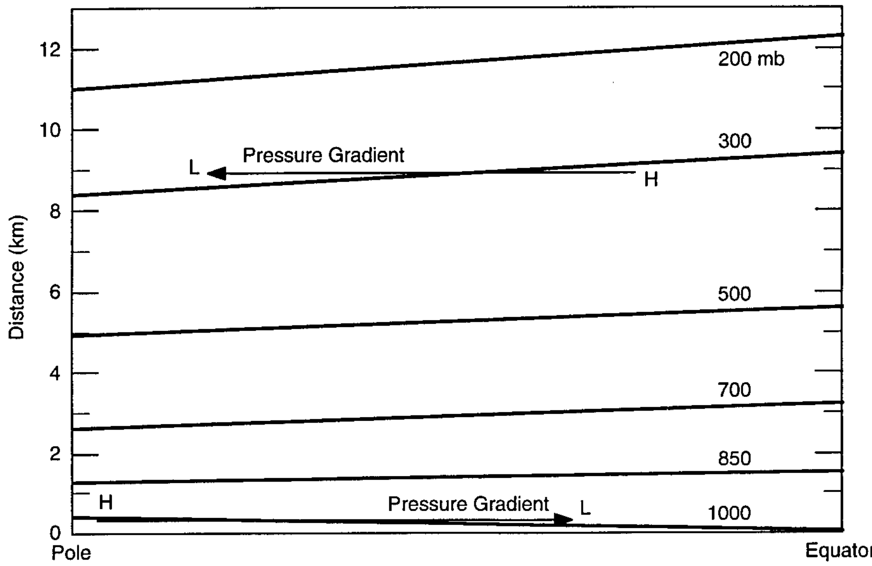


Figure 14. Profile of baroclinic atmosphere demonstrating general cause of poleward flow of winds aloft and equatorial flow at the surface.

north and south latitude, where poleward-moving air converges and "piles up." This is especially true over the oceans, where temperatures at the surface are colder on average than over land masses at the same latitudes. Since this converging air cannot escape to outer space, it is forced to sink. As the air sinks it is compressed and heats adiabatically, the reverse of cooling with expansion. The result is that the sinking air warms adiabatically, becoming positively buoyant, but is forced to continue sinking because of the mass of air converging from above. This increase of air mass in the atmospheric column creates higher pressure at the earth's surface at these locations, called the subtropical highs. Whereas the tropical surface low and polar surface high are primarily thermal in origin, the subtropical surface high is primarily dynamic in origin. That is, the subtropical highs are produced primarily as a result of masses of moving air and not primarily as a result of heating or cooling.

When this slowly sinking air reaches the earth's surface, it spreads, or diverges, with some of the air mass moving equatorward to feed the equatorial low and the remainder moving poleward. The poleward-moving air is warm and dry if it descended over a land mass (the desert regions of the planet are at 25–30° north and south latitude for this reason), but if descent occurred over an ocean, the air is greatly humidified at the ocean surface as it begins movement poleward or equatorward.

Masses of warm air move poleward from the subtropical highs at the earth's surface and encounter cold air moving equatorward from the poles. These warm air/cold air encounters occur generally in the midlatitudes, between approximately 30° and 60° north and south latitude. Since warm and cold air masses are of different densities, they do not readily mix but form boundaries between themselves, called fronts, with warm air typically on the equatorial side and cold air on the poleward side.

Movement of the air masses does not cease after they meet and form a front. The process cannot be likened to two automobiles that collide and stop. Instead, if cold air is moving equatorward more vigorously than warm air is moving poleward, the cold air mass pushes beneath the warm air mass and replaces warm air at the surface. Although some warm air rises over the cold air due to the warm air's lower density, this is a cold front because cold air is displacing warm air at the surface. If the warm air mass is moving poleward more vigorously than the cold air mass is pushing equatorward, then the warm air pushes the cold air poleward and replaces it at the surface. This is a warm front. The warm front is generally less vigorous than the cold front because the warm air cannot efficiently move the denser cold air, which tends to hug the earth's surface, so it overrides the cold air as it slowly pushes it. Another difference is that the warm front also spreads the process, and the weather it creates, over a larger geographic area than that associated with a cold front. Clouds and precipitation typically form in both cold and warm fronts where warm air is forced to rise above cold air.

In areas where fronts are moving rapidly or upper atmospheric conditions are encouraging air to rise, such as when general weather movement is from southwest to northeast, warm air rises over both warm and cold fronts more vigorously. As the warm air rises, one can think of it as rising as a column of air. Rising warm air enters the bottom of the column more slowly than air exits at the top because topography and vegetation create more frictional resistance to air movement at the earth's surface. This results in a decrease in the mass of air in the column and thus less mass pressing on the earth's surface below the column, reducing the air pressure or the weight of the air column resting on the earth's surface. This process, in greatly simplified form, creates the traveling extratropical low pressure cells, identified with Ls on weather maps, and their attendant warm and cold fronts. They are responsible for most of the adverse weather in the midlatitudes.

The weather setting in the midlatitudes is, in synopsis, generally organized by warm and cold air masses that are usually within high pressure cells, interacting and producing warm and cold fronts along their boundaries. Where this interaction is most vigorous, cells of low pressure form along the fronts. These traveling extratropical low pressure cells are the storm systems of the midlatitudes and produce the weather, occasionally violent, that on intermediate time scales determines an IDS's operating conditions. These storms overpower and obliterate the diurnal patterns discussed earlier, which usually only form within the high pressure cells between storms.

High and low pressure areas generally form as localized cells of pressure rather than as broad bands of high or low pressure. The air within these cells rotates around the center of the high or low as the air tries to move from the high to the low. This rotation is imparted by the earth's rotation (decrease of earth's rotational velocity from equator to poles) and is a result of the so-called Coriolis effect. The earth's rotation imparts rotation to all pressure cells, large or small, from tornadoes to hurricanes and the large air masses and extratropical cyclones. However, lows and highs rotate in opposite directions, and are reversed in their rotation in the Northern and Southern Hemispheres because, when viewed from the respective poles, the earth apparently rotates in opposite directions. When viewed from above the North Pole, the earth rotates counterclockwise, but when viewed from above the South Pole, it appears to rotate clockwise. As a result, low-pressure cells rotate counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. Highs are reversed from this, rotating clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere.

The midlatitudes, as a result, are where maximum transfer of heat from the tropics to the poles occurs, since this is where air masses are exchanged. About 70 to 90% of all tropical-to-polar energy exchanges occur in the form of sensible and latent heat during atmospheric circulation (Musk 1991). The remainder of exchange is by ocean currents. This need for energy to be transported poleward and for cold air to be returned to the tropics is the basic reason that the atmosphere moves.

Extratropical cyclone structure

The organization of extratropical cyclones, first recognized by the Norwegians, is generally consistent enough to identify and predict repeatable weather patterns as they traverse a region. Extratropical cyclones are a product of diversity, occurring when cold, dense, and dry air masses meet warm, less dense, and usually moist air masses. This meeting usually spawns an organized storm system with a life cycle and a sequence of weather that is unique to the portion of the storm traversing an area.

Extratropical cyclones form, or strengthen, in readily identifiable areas known as regions of cyclogenesis. Cyclogenesis is most vigorous where contrasts between warm and cold air are greatest. These areas of large contrast are a result of the position of high pressure cells (the subtropical highs), the polar highs, and the distribution of land and water in the middle latitudes.

The subtropical highs are generally strongest over oceans and weakest over land masses, especially in the summer hemisphere, because air that is dynamically forced to sink from aloft is chilled at the ocean surface. As a result, the highs tend to be centered over the ocean basins, with circulation from the highs extending well over adjacent continents. Since highs rotate clockwise in the Northern Hemisphere, cooler and usually drier air is pulled equatorward on their eastern sides. On their western sides, conversely, warmer and usually more moist air is pulled poleward from the tropics. Since the subtropical highs extend over the continents, warm, moist air is advected over the eastern sides of the continents, and colder

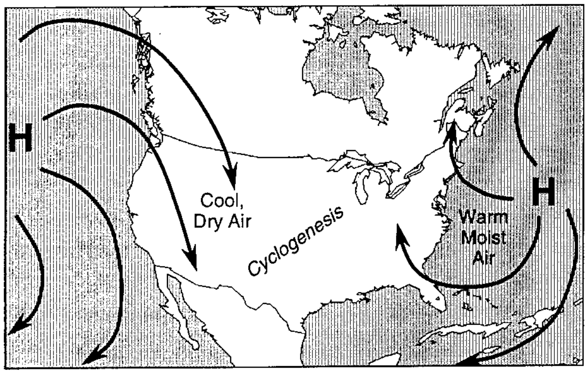


Figure 15. General region where most cyclogenesis occurs in North America, situated between the permanent subtropical highs located in the Atlantic and the Pacific Oceans.

air is advected over the western sides (Fig. 15). Within the continents, therefore, warm air from the southeast often collides with cold air from the northwest. This is the fundamental pattern that occurs in North America and the other continents crossing the midlatitudes, though there is extensive modification by the size of the land masses, such as Asia, and the blocking effects of mountain massifs.

Warm and cold air from adjacent highs meet within the continents. Where this activity is greatest, which in North America is the Great Plains, warm air is forced to rise most vigorously over cold air. This is where low pressure forms at the surface and cyclogenesis occurs (Fig. 15).

The organization of a subtropical cyclone is a further result of the location of the high pressure cells around the low. On the eastern side of the low, where cyclogenesis is occurring in Figure 15, warm air is being vigorously pushed poleward by the high to the east (over the Atlantic Ocean for North America), and the push of cold air from the north is relatively weak, especially in the summer months. Since warm air is pushing poleward more vigorously on the eastern side of the low, a warm front typically forms on the east side. In contrast, on the western side of the low, cold air is being pulled more vigorously from the northwest than warm air is being pulled from the south. On the west side a cold front usually forms as a result of the more vigorous push of cold air from the northwest. The result is that most extratropical cyclones produce warm fronts on their eastern sides and cold fronts on their western sides. The center of the low is usually between the two fronts (Fig. 15 and 16).

The sequence of weather within warm and cold fronts is logical and, within a mature cyclone, very identifiable. Before the passage of a warm front, the air at a site is cool. As the front approaches, warm air overruns the cool air at the surface and occupies the top of the column. As the front approaches, it progressively occupies more of the column as it erodes the cool air at the earth's surface. Once the warm front has passed, all of the cool air has been eroded away, and warm air completely occupies the column. This persists until the first effects of the cold front are felt, when cold air begins to progressively underrun and replace the warm air from the surface upward.

The track of a cyclone and its fronts can be plotted on a weather map, and the location-dependent weather events that an IDS will experience in the course of the storm can be predicted. Weather conditions are typically most changeable and severe on the equatorward, or tropical, side of the storm; they are discussed first.

Weather on equatorward side of cyclones

The idealized, Norwegian model of a typical extratropical cyclone is organized geographically with a warm front on the eastern or leading edge of the storm and a cold front trailing the storm to the west (Fig. 16a). The lowest air pressure is located where the warm and cold fronts meet, and the path of this lowest pressure, traced on the earth's surface, is the storm track. Air pressure at the surface increases radially but asymmetrically from the center of the low, with slight "valleys" of low pressure radiating away from the storm center where the warm front and the cold front contact the earth's surface. Lower pressure along the fronts indicates that these are areas of rapidly rising air and more vigorous storm activity.

Warm fronts. As a storm approaches from the west, the earliest indication of its approach is often provided by slowly falling air pressure and the appearance of thin, wispy cirrus clouds (Appendix D) at distances of up to 1600 km ahead of the storm (U.S. Army 1976). The falling barometer and the cirrus clouds result from warm air rising away from the surface. In a warm front, cold air at the surface is pushed poleward and usually eastward by warm air, an inherently inefficient process because lower-density warm air is pushing higher-density cold air. As a result, the warm air overrides the cold air and produces a gently sloping ramp upon which the warm air rises (Fig. 16b). The slope of a warm front surface is very small, typically ranging from 1:100 to 1:200, or 0.6° to 0.3° of slope (Musk 1991). At a distance of 100 to 200 km ahead of the surface front (the intersection of the frontal surface with the earth's surface), the warm air may be only 1000 m above the earth's surface. Therefore, high mountains and hills often intercept the warm front long before the frontal surface reaches locations in lowland areas. Security fences that traverse large ranges in elevation may experience warm frontal weather at higher elevations before the lower elevations experience the same weather. This may cause location- and time-dependent variations in IDS detectability in the course of the same weather event.

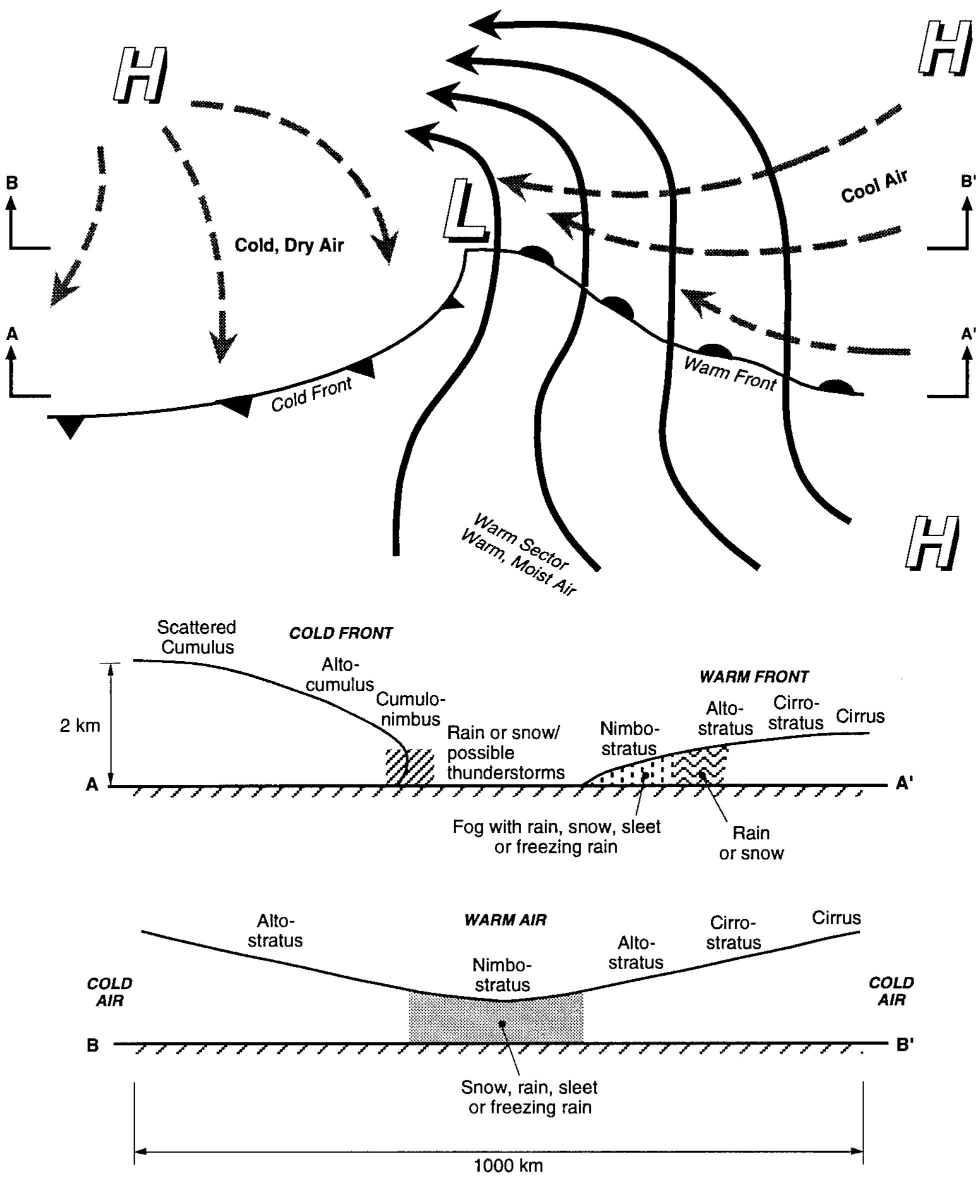


Figure 16. Idealized extratropical cyclone showing fronts, air mass temperature and humidity, wind directions, clouds, and precipitation for the Northern Hemisphere. (Top) Warm fronts are typically located on the eastern side of these storms, and cold fronts on the west. Broken arrows indicate the motion of cold air at the surface, and solid arrows indicate motion of warm-sector air into the storm at the surface, and hence its rising over the cool surface air of the warm front. (Bottom) East–west profiles of an extratropical cyclone equatorward of the storm center (A–A'), showing positions of warm and cold fronts, precipitation, and clouds, and poleward of the storm center (B–B'). The vertical scale is exaggerated approximately 80 times.

As air rises, pressure decreases, causing the air to expand and cool adiabatically. Warm, moist air rising over the colder surface air along the three-dimensional warm frontal boundary slowly cools to the dew or frost point and produces clouds. Generally, little moisture reaches the leading edge of the storm, but when it does, it is at high altitudes and temperatures are well below the freezing temperature of cloud droplets, so cirrus clouds form—harbingers of the arriving storm.

As the warm front approaches, cloud cover thickens and the cloud base decreases in height. Clouds evolve sequentially from cirrus to cirrostratus, to altostratus and eventually to nimbostratus (stratus with precipitation), and the frontal surface overhead lowers in height (Fig. 16b). Before the front passes, cloud cover is typically completely overcast and frontal fog often forms immediately before passage, usually within precipitation. Temperatures rise slowly, perhaps accompanied by a slow increase in humidity, though changes in both temperature and humidity are most rapid at frontal passage. Winds are usually easterly, but may vary to southerly and increase in speed as the front approaches. Strong, gusty southerly winds are not uncommon several hours before the front passes, though it can be quite calm. Gustiness, rather than wind strength, may be more troublesome to fence-mounted IDSs, to IDSs for which alignment is important (beam-break, bistatic microwave), and to buried IDSs that respond to motion of above-ground objects that couples into the ground.

Precipitation within warm fronts varies considerably with the season and normally begins up to 500 km ahead of the surface front (U.S. Army 1976). Many warm fronts are weak, producing little or no precipitation. Strong fronts during warm weather may produce up to several days of dark, gloomy weather with light, continuous drizzle or rain that increases in intensity as the front approaches. Thunderstorms are occasionally observed within warm fronts, but they are not typical. Warm-front rains often drop considerable volumes of water, but they are usually steady rains that infiltrate and produce little surface runoff from bare soils. These soaking rains allow water to penetrate deeply into the soil, possibly affecting buried IDSs that are sensitive to the moisture content of the soil. Electromagnetic IDSs have reduced detection capability in moist soil (high electrical conductivity), and ground-motion IDSs have improved detection capability in moist soil (low rigidity) (Peck 1994d). The greater the moisture content of the soil, which is often wetted deeply by warm frontal rains, the longer the delay in sensible heating of the soil surface once precipitation has ceased and the ground is absorbing solar radiation. This is conducive to a low rate of nuisance alarms from the passive infrared IDSs, provided that the surface temperature of the soil is stabilized throughout the IDS's detection zone. However, it is an unfavorable situation for bistatic microwave IDSs, which experience greater propagation loss arising from interaction of the microwaves with damp ground and wet vegetation (Peck 1992a).

Warm frontal precipitation during the winter is usually far more complicated. A warm front is an inversion because warm air is overlying colder air. Well ahead of the surface front (where the warm front contacts the earth's surface), warm air has typically risen high enough (as it overrides cold air) that temperatures in the warmer air within the inversion are below freezing. If precipitation occurs, it is in the form of snow because all temperatures in the air column are below freezing. The warm air aloft increases in temperature over time because latent heat is released as condensation occurs to produce clouds and precipitation and because, as the front approaches more closely, overriding air becomes warmer. If the air warms above freezing, the precipitation turns to rain aloft, but the air below the inversion is still quite deep and cold. Raindrops may thus freeze as they fall through the cold air below the inversion and produce sleet (frozen raindrops). Sleet typically does not have a long duration, if it occurs at all, but occasionally it can accumulate to depths of several centimeters.

Eventually the layer of cold air below the inversion becomes thin enough that raindrops passing through it do not freeze before striking the ground. However, they may supercool—cool to below freezing but remain liquid (Bennett 1959). Glaze is then deposited when these supercooled drops strike objects at the surface. Glaze thickness depends upon the orientation and thermal characteristics of surfaces and commonly ranges from a trace to several centimeters. Glaze can reside on objects for days if temperatures do not warm, such as on the polar side of cyclones. However, it may begin to melt shortly after accretion because of the conditions described below.

With the approach of the surface front, the cold air underlying the inversion becomes so thin that raindrops no longer supercool, and freezing ceases at the surface. However, this warm rain occasionally falls onto an impermeable frozen surface if sufficient glaze icing has already occurred. This can cause local pooling of water and even flooding if rainfall is intense. Radiation (visible, thermal infrared, microwave) that reflects from moving water surfaces can cause nuisance alarms by video-motion, passive infrared, and microwave IDSs, respectively. Frontal fog also often occurs at this time because warm rain is falling through cold air. The warm rain evaporates into the cold air and, as with steam fog, causes humidification, producing fog. Near-infrared beambreak IDSs are vulnerable to nuisance alarms in fog, while the detection capability of video-motion and passive infrared IDSs may be diminished because of reduced visual and thermal contrast.

Warm sector. After the warm front passes, all cold air has been removed and the entire column of air from the earth's surface to high altitudes is warm. This is referred to as the warm sector of the storm, a region of warm air that penetrates the storm between the warm front and the cold front (Fig. 16). It is actually the same warm air mass that produced the warm front, but once the warm front has removed all cold air, it becomes the warm sector. Temperatures rise, humidity increases, and winds become generally southerly or southwesterly. Partial clearing of cloud cover may occur, but often the sky remains overcast because of instability of the encroaching warm air. Clouds are typically stratus or stratocumulus. Occasional showers or air mass thunderstorms may occur, though typically in late afternoon in the warmest months because of the greater instability at that time from solar heating.

Cold fronts. Though the sequence of weather conditions through a warm front and into the warm sector can be dramatic, the changes are usually not as great as those observed during a cold frontal traverse. Cold fronts are noted not only for the magnitude of change in conditions that occurs as they pass, but also by the rapidity of change. Cold fronts generally move faster than warm fronts and, because cold, dense air is pushing beneath or underrunning warm air, they create a more effective "ramp" for pushing warm air aloft. The slope of a cold front is typically about 1:30 to 1:140, or about 2.0° to 0.4° of slope (Musk 1991) (Fig. 16b). As a result, warm air rises more rapidly, and storm conditions generated along the frontal surface typically are more violent and cover a smaller geographical area than those within a warm front.

There is usually little obvious warning that a cold front is about to traverse a location. There are no gradual changes in temperature, wind speed, or wind direction. The only indication is air pressure, which continues to drop until the surface front is overhead and rises rapidly thereafter. During the winter, the passage of a cold front is often accompanied by snow squalls with whiteouts, followed by drifting snow that produces blizzard conditions. The primary visual evidence of the imminent passage of a cold front occurs during the summer. If a cold front is vigorous, a squall line (a line of often severe thunderstorms) may precede the surface front by 100 to 300 km or coincide with the surface front passage (Ahrens 1982). These storms are identified by their towering cumulonimbus clouds, which often reach heights of 10,000 m. Strong winds, hail, intense rainfall, lightning, and occasionally tornadoes may be associated with these storms. Severe, destructive downbursts also accompany such storms, producing gust fronts with wind speeds often exceeding 50 m s^{-1}. The weather is most severe if the front passes in the afternoon at the time of maximum heating. Though these storms typically occur in a narrow band less than 80 km wide and are short lived, usually lasting less than a few hours, their destructive winds and lightning could completely disable aboveground IDSs or their supports and destroy portions of a security fence. Intense rainfall may cause fence-mounted IDSs and microwave systems to generate nuisance alarms. Because cold frontal rainfall is often intense, it does not sink into the soil as effectively as long, slow warm frontal rains. Therefore, the background does not become as thermally homogeneous. Accordingly, there may be less reduction in nuisance alarms by passive infrared IDSs; moreover, the detection capability of infrared systems may be decreased if the thermal contrast between a wet intruder and the background is insufficient. Wind-induced motion of water that pools at the surface can cause nuisance alarms from microwave and buried electromagnetic IDSs.

Passage of the surface cold front is characterized by a sharp rise in air pressure, rapid drops in air temperature and humidity, and dramatic shifts in wind direction from southerly to westerly or northwesterly (Fig. 16). Skies usually clear rapidly after the passage of a vigorous cold front with cumulonimbus clouds followed by altocumulus and wind speed increases if the following high pressure is strong (Appendix D). A day or two after frontal passage, skies generally clear completely, winds calm, and cooler, drier weather settles in as high pressure moves in.

Weather on poleward side of cyclones

Weather changes on the poleward sides of traveling extratropical cyclones are typically less dramatic than those occurring on the equatorward side because the warm sector, and thus the surface fronts, are not present (Fig. 16). Though an IDS escapes the direct effects of surface fronts when the storm center passes equatorward, the large weather changes occurring over the time of cyclone passage (24 to 36 hours) have the potential to alter IDS performance.

As the cyclone approaches from the west, the warm front again is marked by a shield of high cirrus clouds (Fig. 16b). The sequence of events through the warm front as it approaches is similar to the situation on the storm's tropical side except that winds change from being easterly or southeasterly to being northeasterly and northerly as the front nears. Cloud cover increases and lowers, as before, but the sequence of winter precipitation may not occur as fully, with all precipitation being snow or perhaps transitioning briefly through freezing rain. In warmer weather, rainfall may be continuous without clearing after the warm front passes. There is no surface warm front passage, and thus no warming, no warm sector, no clearing, and no southerly winds occur.

As the center of the low passes to the east, the weather resembles that of the cold front, but again without rapid changes in temperature and wind direction (Fig. 16). Temperatures cool more slowly, and thunderstorms or snow squalls (depending on the season) may occur, but changes are not sudden. As before, winds eventually change through northerly to perhaps westerly. Throughout this entire sequence, precipitation may occur for a longer period of time because there is no warm sector present (Lydolf 1985).

Quasi-periodic weather features

The weather conditions described here accompany cyclones and may significantly affect IDS performance.

Precipitation is classified by type and by form. Precipitation types refer to the dynamic processes causing the air to cool and produce precipitating clouds, which are cyclonic or frontal, orographic, and convective (Musk 1991). Cyclonic precipitation was discussed under Extratropical Cyclone Structure above. Convective precipitation was addressed in the discussion of diurnal weather, because convection typically reaches its maximum in mid- to late afternoon on days with abundant sunshine. Orographic precipitation is locally generated when air cools as it is lifted over elevated terrain features. Orographic precipitation is usually strongest during cyclone passages and during instability, but it can occur whenever there is sufficient moisture for rising and expanding air to reach the dew point.

The frequency of precipitation types varies by season and location. Convective precipitation dominates in low latitudes in all seasons. Though large low-pressure areas that are often precursors to hurricanes, called easterly waves, do cause some precipitation in the equatorial latitudes, most precipitation is the result of local convection resulting from intense solar heating.

Convective precipitation is extremely localized, typically consisting of storms only a few kilometers to a few tens of kilometers in diameter. Convective storms are usually isolated and spaced apart, with the distance between storms similar to the width of the storms. Convection is most vigorous in the afternoon during the warmest portion of the day. Though often violent, the duration of the strongest portion of convective storms is limited, with activity generally decreasing significantly near and after sunset.

In the midlatitudes, summer precipitation is typically a combination of cyclonic and convective. Convective storms are similar to those found in the tropics. Cyclonic precipitation is usually well organized and covers large areas. However, its duration varies considerably, depending upon location within the storm, from over 24 hours within warm fronts to as little as one hour within a cold front squall line. Cyclonic rainfall can also occur at any time of the day, depending on the timing of the cyclonic passage.

Winter precipitation in the midlatitudes is almost all cyclonic. Cyclonic activity is generally most vigorous during the winter, with stronger and more frequent storms. The primary winter exception to this is lake-effect storms. These are strong, local convective systems that grow as cold, dry air moves over warm unfrozen lakes between cyclonic storms. Lake-effect storms produce high winds, low visibility, heavy snowfall, and drifting and occur on the lee shores of large lakes in the fall and early winter when water temperatures are still warm and cold air is beginning to penetrate from the north. Lake-effect storms are common on the southeastern shores of the Great Lakes and can occur over sea water and impact downwind coastlines, such as western coastal Japan and along the Adriatic Sea.

Orographic precipitation is highly localized and can occur in any season, being a product of surface relief. Precipitation amount and intensity typically increase with elevation, though the elevation of maximum precipitation varies with location. On some mountains, maximum precipitation occurs near the summit, whereas in other locations precipitation maximums may be near the cloud base, with amounts decreasing with elevation thereafter (Barry 1969).

Precipitation forms include, but are not limited to, rain, drizzle, sleet, hail, snow, snow pellets, snow grains, ice pellets, and graupel. Each has its associated weather condition. The most common of the forms potentially disruptive to IDSs are described here.

Rainfall

Rainfall may be classified as rain or drizzle depending upon drop size. Rainfall is usually composed of liquid drops larger than 0.5 mm in diameter, whereas drizzle typically ranges in diameter from 0.02 to 0.5 mm. Rainfall and drizzle are also classified by intensity. Light rainfall deposits less than 2.0 mm hr^{-1}, and light drizzle less than 0.2 mm hr^{-1} (Barry 1969). Heavy rainfall deposits more than 7.0 mm hr^{-1}, whereas heavy drizzle deposits more than 0.5 mm hr^{-1}. The rainfall rate of a typical afternoon convective thunderstorm is about 25.0 mm hr^{-1}.

Mean rainfall intensity is inversely related to storm duration (Barry 1969). Short-duration storms produce the highest mean intensity rainfalls. For example, in Washington, D.C., a 10-min rainfall will typically deposit about 2.5 cm of water, whereas a 60-min rainfall will deposit 5 cm of water on the average (Barry 1969). The 10-min storm thus deposits water at three times the rate of the 60-min storm. Though this ratio changes with location, it does indicate that convective or frontal thunderstorms and squalls, for example, which are generally of short duration, produce the most intense rainfall, whereas warm frontal rainfall, which is of comparatively long duration, is deposited at a low intensity. Since a large portion of summer rainfall is convective, summer rainfall is generally more intense than winter rainfall.

Rainfall intensity affects atmospheric extinction (loss of radiation through scattering and absorption by airborne particles such as water drops; see Appendix A) in both the visible and infrared spectra; the effective range of surveillance cameras is reduced and detection ranges of video-motion and passive infrared IDSs are less. Extinction is primarily a function of the rainfall rate and the drop size distribution, with drop sizes generally increasing as rainfall rate increases; e.g., Laws–Parsons or Marshall–Palmer drop-size distributions for various precipitation rates. Values of extinction coefficients in rain as a function of precipitation rate have been published for visible (0.63 μm) and thermal infrared (10.59 μm) (Rensch and Long 1970), near-infrared (0.94 μm) (Nedvidek et al. 1986), and microwave (10 GHz) radiation (Pendleton and Niles 1994, Ulaby et al. 1981). For rainfall in the range of 8–12.5 mm/hr, the extinction coefficients are: 4.0 (0.63 μm), 5.9 (0.94 μm), 4.6 (10.59 μm) and 0.2 (10.6 GHz). Over a 100-m path length, these equate to transmission losses of 9%, 13%, 10%, and <1%, respectively. For a given rainfall rate, the extinction coefficient in drizzle is approximately three times that in a thunderstorm (Duncan et al. 1987). Rainfall rates are generally much higher during a thunderstorm, however, so the net effect is that extinction is commonly greatest within a thunderstorm. Occasionally, however, widespread rain events can generate higher extinctions than a thunderstorm, because at intermediate rainfall rates the distribution of smaller drops falling in the rainstorm will cause more extinction than the larger drops associated with the thunderstorm. A quantity of small droplets in a volume of air may have a larger cumulative cross section to intercept visible and infrared radiation relative to that associated with fewer, larger droplets in an equivalent volume of air.

High rainfall rates, such as in driving thunderstorms, also shake fences and cause detection systems to vibrate because of the larger drops, which may approach extreme diameters of 6 mm (Rogers 1976). Fence-mounted IDSs at SOROIDS often alarm during the onset of heavy rainfall, although whether this is caused by rain striking the fence or the sensor cables attached to the fence, or both, is not known. Ulaby et al. (1981) conclude that extinction of microwaves by water droplets (fog or rain), which depends on the quantity and size distribution of the drops relative to the wavelength of the radiation, is not significant during most atmospheric weather conditions if the radar frequency is less than 10 GHz. At higher frequencies, extinction of microwaves by rain will cause a bistatic radar to alarm when scattering and absorption sufficiently disrupt the microwave transmission (between transmitter and receiver) that the variation in signal at the receiving antenna is similar to that caused by an intruder moving through the detection zone. Backscattering (the return of radiation toward the transmitting antenna) will cause a monostatic radar to alarm when the return from the raindrops is sufficiently strong and time-varying as to be equivalent to the typical return from a moving intruder. Although microwave (10.525 GHz) IDS alarms during rainfall at SOROIDS were not common, the monostatic radar IDS had more alarms during rainfall than did a bistatic radar IDS operating at the same time, with the former more prone to alarm at the onset of rainfall than throughout a rainstorm. A factor in the microwave alarm occurrences may have been disruptions of the microwave field by movement of nearby chain-link fences under the impact of the rain or accompanying wind. (Manually induced movement of panels of the SOROIDS chain-link fence had been shown to cause alarms by nearby microwave IDSs.) This type of nuisance alarm with microwave IDSs depends in turn on the temperature-dependent stiffness of the fence panels: a colder, stiffer panel moves less under wind loading or rain impact, so for a given wind speed, there were fewer nuisance alarms by the microwave IDSs at lower air temperatures (Peck 1992b).

Rainfall rate also affects the amount of water that puddles on the ground surface, which affects the thermal regime at the soil surface. Soil infiltration capacity is a function of soil type, soil surface conditions, preceding soil moisture amount, and rainfall rate (Morgan 1969). Clays and soils have infiltration capacities of about 2.5 to 15.2 mm hr^{-1}, whereas loamy sands have maximum infiltration capacities of up to 50.8 mm hr^{-1} (Morgan 1969). Soils that have been compacted by foot or vehicular traffic, and soils without vegetation cover, have overall lower infiltration capacities for their respective soil types. Large rainfall amounts and rains that persist for long periods of time also eventually saturate the near-surface soil horizons, causing puddling at the surface. Until the puddles dry, they are a potential cause of nuisance alarms by microwave IDSs and buried electromagnetic IDSs whenever the water surface is disrupted, as by traffic through the puddles or wind-induced motion.

Hail

Hail is associated with strong summer thunderstorms and usually not with subfreezing weather. During the winter, cumulonimbus clouds are often completely below freezing, preventing hail from forming. Hail forms when raindrops falling through cumulonimbus clouds are caught by updrafts and carried back up into the cloud and above the freezing layer. After the droplets freeze, they fall and are coated with liquid water in the warmer air below. They are again carried by updrafts above the freezing level and the coating of water is frozen over the drop much like the layers of an onion. The hailstone falls repeatedly, and the coating–freezing process recurs, until the weight of the hailstone is sufficient to overcome the drag of the updrafts. Eventually the hailstone falls to the ground, sometimes as large as a golf ball, but more typically marble sized. Hail frequently dents automobile sheet metal, causes often severe aircraft damage, and certainly will cause fences to vibrate. If the hailstones are large enough, IDS hardware may be damaged. The occurrence of hail is usually very localized in area and of short duration, typically preceding the storm creating it by a few minutes.

Snowfall

Snow is precipitation resulting from the creation of ice crystals by direct deposition of water from the vapor phase to the solid phase. Snow crystals form when supercooled liquid cloud droplets and either ice crystals or ice nuclei within clouds coexist (Schemenauer et al. 1981). At temperatures below 0°C the vapor pressure of the surface of supercooled water droplets is higher than the vapor pressure at the surface of ice crystals, or ice nuclei, at the same temperature. As a result, as with all gases, water vapor flows down the pressure gradient and evaporates from the droplets and deposits onto existing ice crystals or ice nuclei. Crystals form in a myriad of shapes depending upon air temperature and moisture conditions within the clouds (Schaeffer and Day 1981). As crystals fall, they also intercept supercooled cloud droplets and thus often increase in mass through this process, called riming.

Falling snow is detrimental to many IDS operations. Fence-mounted IDSs alarm when wet snow that has adhered to the fence or sensor cables drops off, shaking the fence and cables (Peck 1993a, 1994b). The visible range of surveillance cameras is reduced. The extinction of visible and infrared wavelengths within snowfall is controlled primarily by snowfall intensity and the area-to-mass ratio of the snowflakes, with thermal infrared and midinfrared transmission being less than that of visible radiation, and near-infrared transmission being comparable (Koh 1989). The aperture diameter of the sensors of a near-infrared beambreak IDS influences how sensitive the IDS is to transmission loss due to scattering by falling snow. A near-infrared system with a sensor lens approximately 2 cm in diameter was very susceptible to nuisance alarms during falling snow at SOROIDS, with the frequency of alarms varying with the snowfall rate. In contrast, a beambreak system with a 9-cm-diameter lens, which was designed to alarm when transmission fell below 1.5% (i.e., when more than 98.5% of the sensor area is blocked), did not alarm during any snowfall events. It did alarm, however, during blowing snow events, which are characterized by high concentrations of small snow particles entrained in the air near the snow surface. In addition, because blowing snow is concentrated near the ground, pre-existing partial obstructions of a close-to-the-surface sensor would increase the probability of nuisance alarms by the beambreak IDS during blowing snow events (Peck 1994b). The transmitter and receiver units of a near-infrared beambreak IDS with 9-cm-diameter sensor lenses were aligned on an east–west line-of-sight at SOROIDS, where the predominant wind directions are northerly and southerly (the range of wind directions is restricted because the site is located in a river valley). Each transmitting and receiving sensor was fitted with a snow cover, a metal tube extending outward ~20 cm from the sensor faceplate. No snow accumulated in front of the lenses (within the snow covers) to block the near-infrared beam and cause nuisance alarms. When the same type of IDS was installed on a north–south line-of-sight at SOROIDS, without snow covers, falling snow usually filled the slight recess in front of the sensor faceplate (for those sensors looking into the wind) enough to cause nuisance alarms that persisted until the snow melted (Peck 1994a). The use of snow covers in this situation would not have eliminated the snow blockage, but only made it more difficult to remove the accumulated snow manually. The problem was corrected by fitting a clear Plexiglas™ disk, provided by the IDS manufacturer, in front of each sensor's faceplate, across the housing of the sensor, to eliminate the recess. Although this type of nuisance alarm results more from a design flaw than from snowfall by itself, it is cited here as being of practical interest.

Snowfall intensity. Airborne snow disrupts visible and infrared transmission to a greater degree than rainfall does, for a given precipitation rate, because of its larger area-to-mass ratio. The area-to-mass ratio of airborne snow depends upon the type of snow crystal and the size distribution of the snowflakes, but these are not commonly measured. There are empirical relationships between precipitation intensity and extinction, with the scatter in the relationships related to differences in snow crystal type, area-to-mass ratio, size, etc. (Hogan et al. 1990, summary in Peck 1989b). Dependence of transmission on wavelength of the radiation when the particles (snowflakes) are so large relative to wavelength that the scattering efficiency is a constant is in part due to the differences in the amount of radiation that is scattered in the forward direction by the snowflakes (see discussion in Appendix A). Lacombe and O'Brien (1982) present the variation in transmission of visible and infrared radiation during a single snowstorm in terms of changes in the type of snow crystal that predominates. When all snow crystal types are included in the regression analysis for a relationship between the extinction coefficient, k_{ext} (km^{-1}), and snow mass concentration, C (g m^{-3}), of the form $k_{ext} = a * C$, the following values of a resulted: 13.41 (0.55 μm radiation), 14.16 (1.06 μm), 19.21 (3.0 μm), and 19.05 (10.37 μm). From these expressions for extinction coefficient, the relative impact of falling snow on visual systems (surveillance cameras, video-motion IDSs), near-infrared beambreak IDSs, and passive infrared IDSs can be approximated. As an example, over a 100-m path through falling snow with a mass concentration of 0.4 g m^{-3} (which Lacombe [1983] relates to a moderate-intensity snowfall), visible contrast is reduced 42%, near-infrared (1.06 μm) transmission is reduced 43%, and thermal-infrared contrast is reduced 53%. Near-infrared and thermal-infrared extinction during a snowstorm also have at times been reported as being less than that of visible radiation. In this case, the cause may be the concurrent existence of fog, which is more detrimental to visible radiation because the sizes of water droplets comprising fogs are closer to the wavelengths of visible radiation. When snowfall is intense, extinction by snowflakes dominates, so the extinction of radiation is independent of wavelength (or apparently greater at longer wavelengths because of forward scattering), but if fog is also present, then extinction by water droplets dominates during periods of low-intensity snowfall (Koh 1986). The effect of airborne snow on microwave IDSs

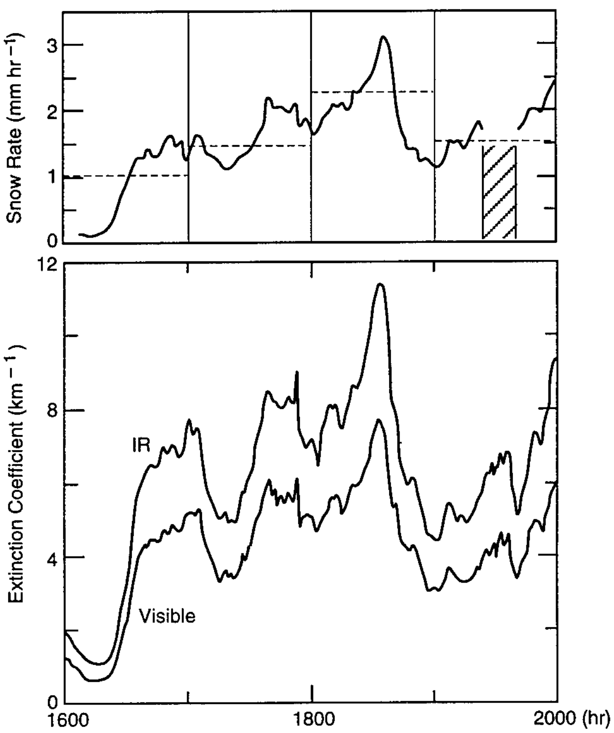


Figure 17. A snowfall that deposited 20 cm of snow at Camp Ethan Allen, Vermont, 1600–2000, 31 January 1982. The upper panel shows the continuous snowfall mass accumulation rate expressed as mm of equivalent melted water. The lower panel provides visible and infrared extinction coefficients during the storm (after Ryerson 1991).

also depends on the wetness of the snow. Scattering and absorption of microwaves by falling, dry snow are slight, and the net effect is that for dry snow the extinction coefficient is approximately 20 to 50 times smaller than that of rain for the same precipitation rate (mm of melted water per hour) (Ulaby et al. 1981). This means that microwave transmission through falling dry snow is less of a problem than through rain, so bistatic microwave IDSs should have fewer nuisance alarms during dry snowfall than during rain. The extinction coefficient of melting airborne snow, however, is much larger than even that of rain, so a heavy fall of wet snow (air temperature ≥ 0°C) is likely to cause nuisance alarms. Ulaby and his colleagues also show for microwaves that the volume backscattering coefficient of falling, dry snow is comparable to that of rain for the same precipitation rate, while that of wet snow is much higher. The implication of this for monostatic microwave IDSs is that, at a given (liquid water equivalent) precipitation rate, wet snow would be more of a problem (higher nuisance alarm rate, perhaps decreased detection range) than rain or falling dry snow. Precipitation rates are generally higher in rain, however, so the net result is that rainfall generally is more troublesome for monostatic microwave IDSs than is falling snow, wet or dry. Precipitation intensity can be determined with a weighing rain gauge.

Comparison of snowfall intensity within one storm measured at Camp Ethan Allen, Vermont, and concurrent extinction coefficient measurements in visible and infrared wavelengths show strong correlations (Fig. 17) (Ryerson 1991). The extinction coefficient equals about 3 km^{-1} at a water-equivalent precipitation rate of 1 mm hr^{-1}, which corresponds to a light to moderate snowfall (Lacombe 1983). For an active visible or infrared IDS with a transmitter/detector spacing of 50 m or a passive infrared system attempting to detect an intruder at the same distance, this translates into a 15 to 20% signal loss.* Several empirical relationships for the extinction coefficient of visible light as a function of precipitation rate are summarized in Peck (1989b), together with a discussion of visual contrast and visibility.

Snowfall intensity varies considerably with time throughout a storm, storm to storm, and by geographic location. The snowfall rate within the Ethan Allen storm varied by 300% over a period of 30 min. Storm-to-storm variation can be as large as within-storm variation, especially if the synoptic situation creating each storm is different.

Maps derived from four years of snowfall data indicate that snowfall intensity varies geographically within the United States (Ryerson 1991). Higher snowfall intensities occur in mountainous regions and in coastal areas of the Northeast (Fig. 18). The lower Mississippi River valley also experiences occasional high-intensity storms, perhaps because it lies in an area of frequent winter cyclogenesis where storms are young and vigorous. Snowfall intensity is generally low in the northern plains, largely because the area is dominated most of the winter by dry, polar air masses and weak, cold lows along the polar front. Strangely, snowfall intensity is also moderate to low in some lake-effect areas south of the Great Lakes. Though lake-effect storms are often intense and visibility is often low, the cause may not be

* J. Lacombe, 1995, CRREL, Hanover, N.H., personal communication.

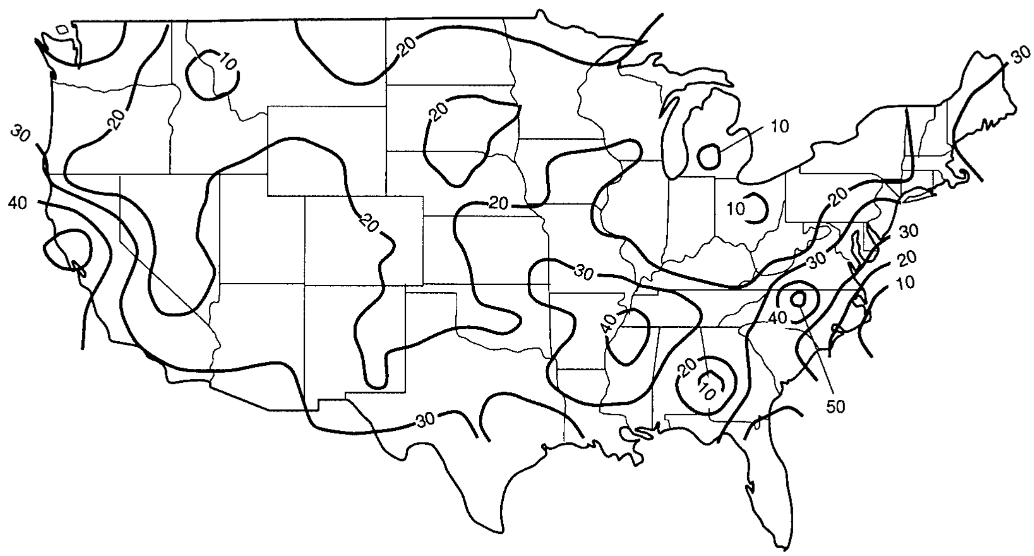


Figure 18. Percent of winter snowfall hours (1984–1987) with snowfall intensity greater than 1 mm hr^{-1} water equivalent, the snowfall rate above which extinction becomes significant.

snowfall intensity alone. Lake-effect storms are produced by steam fog rising from the warm lakes into cold, continental air as it advects to the southeast. As a result, fog can accompany snowfall within these storms near the lake shores. In addition, strong, gusty winds associated with these storms blow considerable snow, further reducing visibility (Eichenlaub 1979, Ryerson 1991).

Snow adhesion. Snow, as a material, is typically considered a dry substance that simply accumulates on the ground surface. In many geographic locations, however, and in some types of storms, snow is sticky and readily adheres to surfaces of any orientation, shape, or surface texture. This is commonly referred to as wet snow, because the condition is most often observed when snow has some liquid content, but dry adhesive snow has also been observed to adhere to wires in Alaska and Japan (Colbeck and Ackley 1983, Peabody 1993). Wet snow causes significant damage to electrical transmission lines in Japan, France, Norway, and Canada and may damage IDSs through the additional weight of the snow mass.

Wet snow occurs when air temperature is at or a few degrees above 0°C (Admirat and Sakamoto 1988). At these temperatures, liquid water is usually present on snow crystals. Upon striking an object, snow metamorphism immediately begins and crystal structure begins to evolve from the sharp, angular crystals of fresh snow to the well-rounded shapes of metamorphosed snow. The mechanism is driven by the relationship of the radius of ice crystals to their melting temperature (Colbeck and Ackley 1983). The lower melting temperature of smaller crystals causes them to melt and refreeze onto larger crystals, causing an overall increase in size of the snow crystals and increased intercrystalline bonding. Wind-packing on fences and wires accelerates the process and densifies the snow. A similar process of snow metamorphism occurs in the snowpack on the ground over months, but the process takes only hours on wires and fences exposed to the wind (Colbeck and Ackley 1983).

The resultant adhering snow becomes quite dense, ranging from 0.5 to 0.9 g cm^{-3}, and adheres strongly, especially if wires, such as trip wires or those used with taut-wire systems, have rotated during the asymmetrical accretion process, causing a round sleeve of snow to form (Colbeck and Ackley 1983). The resulting accretion may be heavy and damage trip wires by causing them to sag or break. The snow fills in fence fabric voids, thus increasing wind load and perhaps causing the fence to be blown over, or it may cause sufficient vibration that fence-mounted IDSs generate nuisance alarms. If rapid melting occurs after accretion of wet snow, especially if sufficient time has elapsed to allow the snow to become very dense, snow falling from fences can cause fence-mounted IDSs to alarm by shaking the fence fabric (Peck 1993a, 1994b).

Wet snow can adhere to the lenses of video cameras, rendering them less effective. Even if the cameras have heating elements, the film of water formed as the snow melts blurs their imagery. During snowstorms, high-threat intruders at SOROIDS have been known to splatter snowballs on the camera faces to assist the snow in disrupting visual surveillance. Wet snow may also adhere to the faceplates of near-infrared sensors (beambreak IDSs), passive infrared IDSs, and microwave antennas. Near-infrared IDSs, and possibly microwave IDSs, would indicate the occurrence of such blockage through nuisance alarms if the accreted snow were extensive and thick enough to severely disrupt transmission. For a passive infrared IDS, however, the establishment of a uniform coating of adhered snow may foreshorten the effective detection zone to the immediate vicinity of the IDS. The initial development of a layer of accreted snow might be accompanied by nuisance alarms until the faceplate filled in completely, but subsequently the likelihood of nuisance alarms and intruder detection would both be less.

Lake-effect snow storms. Locations to the lee of large bodies of water, such as the southeastern shores of the Great Lakes, the eastern shore of Lake Champlain and the Great Salt Lake, the lee side of Hudson Bay, the western coast of Japan, and the shores of the Adriatic and Baltic Seas, all experience lake-effect storms* (Eichenlaub 1979). Lake-effect storms are also the product of quasi-periodic pressure changes, but they are unique because they occur during high pressure, between extratropical cyclones, in normally clear and stable weather. Lake-effect storms only occur where large masses of cold air can advect over large bodies of warm water during the fall and early winter before an ice cover forms.

Lake-effect storms are typically narrow-banded rainshowers in the fall—September and October—and become snowstorms in November and December (Moore and Orville 1990). They form as a result of instability caused by cold air from the west or northwest flowing along the longitudinal axis of the lakes, along the longest fetch. Transfer of heat and moisture from the warm water surfaces to the air, and the subsequent condensation of that moisture in the air, releases large quantities of latent heat that causes vigorous convective activity. This process typically occurs behind cold fronts where colder air is penetrating on the leading side of a high pressure cell. Steam fog rises into the air over the lakes, and a fully developed convective storm develops by the time the heated and humidified air reaches the lee shores. Orographic lift also usually enhances the convective activity on the lee side, making the storms even more vigorous.

Lake-effect storms typically form within westerly winds following cold fronts. As a result, lake-effect areas receive snow much more frequently than non-lake-effect areas. In addition to snow from the usual extratropical cyclone systems, they also receive lake-effect snows between the passage of cyclones when most areas do not receive snow. Lake-effect storms are also often vigorous and are renowned for their heavy snowfalls, high winds, and blowing snow.

Lake-effect storms may affect IDSs in several ways. High winds may shake fences, producing false alarms. Fog, blowing snow, and heavy snowfall may cause high extinction in the visible and infrared wavelengths, reducing the effectiveness of surveillance cameras and video-motion, near-infrared, and passive infrared IDSs. Blowing snow from ground blizzards can completely obscure the snow surface, hiding intruders from view. Heavy snows and drifting can partially or completely cover fences, rendering them ineffective. In addition, the storms often begin and end quickly, are highly localized, and are difficult to predict as to exact location and time. As a result, when the probability of lake-effect storms is high, such as when the air temperature is much colder than the lake water and winds are blowing the length of the lake and over the IDS site, vigilance is necessary to watch for intruders who may take advantage of the chance that IDSs may not be operating reliably.

Snow drifting. Snow drifting is a serious problem in any snowfall area where winds are high and there is little tree cover. Drifting often denudes areas where snow cover is desired, and buries areas that should be clear of snow. Blowing snow may partially or completely bury fences (Fig. 19 and front cover). The movement of snow and subsequent alteration of the snowpack depth alters soil temperatures by exposing soil to air temperatures in areas of snow erosion and insulating soil from the air in areas of snow deposition or drifting. Ground blizzards that accompany drifting also often obscure the lowest meter or two of atmosphere above the snow surface.

Windblown snow particles are generally smaller than snowflakes, so there can be a larger

* A. Hogan, 1994, CRREL, Hanover, N.H., personal communication.

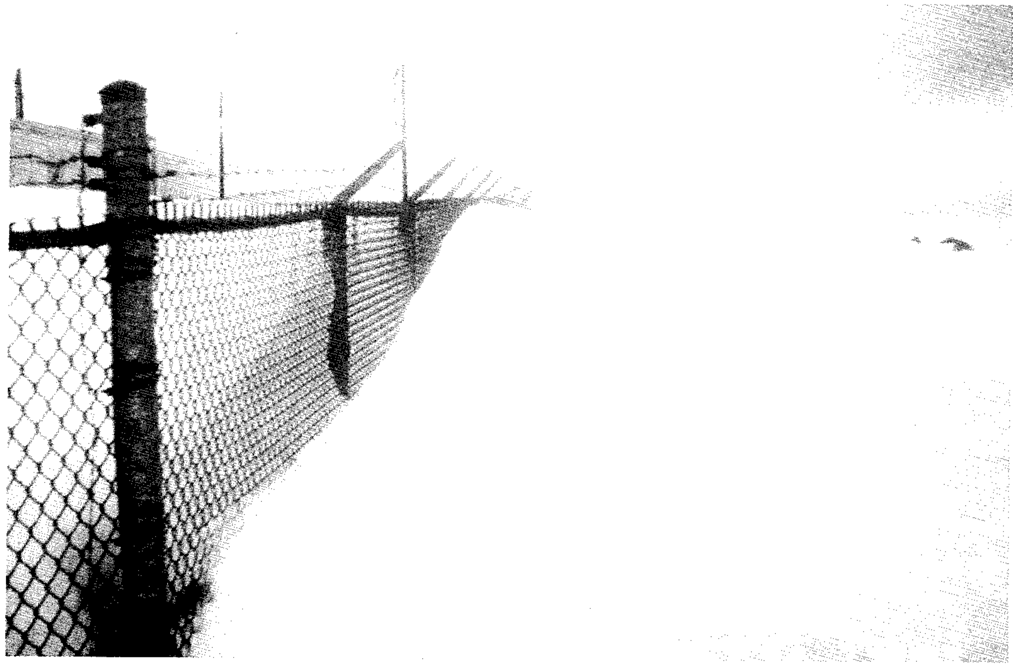


Figure 19. Perimeter fence acting as a snow fence accumulates snow that would normally blow away. The fence in the photograph is 2 m high. In the background the snow cover is almost blown free.

concentration of airborne snow mass near the ground during blowing snow events. For a given mass of airborne snow per volume of air (drift density), the extinction coefficient for visible and near-infrared radiation doubles when the mean particle radius decreases from 90 to 40 μm (Pomeroy and Male 1988); this applies to the suspended layer of blowing snow below a height of 3 m. Extinction in blowing snow is significantly greater than in falling snow (no coexisting fog) of an equivalent mass concentration, e.g., 80% vs. 50% transmission loss over a 100-m path for a drift density of 0.5 g m^{-3} and 40-μm-sized particles. Blowing snow is particularly troublesome for near-infrared beambreak systems if it occurs in conjunction with pre-existing partial beam blockages due to topography or snow on the ground. A near-infrared beambreak IDS (with 9-cm-diameter lenses for transmitter and receiver sensors) at SOROIDS that did not experience nuisance alarms due to falling snow did alarm during blowing snow events (Peck 1994b).

Drifting snow may be acquired from falling snow or it may be lifted from the snowpack surface and transported (Mellor 1965). Loose, unbonded snow is most susceptible to being disturbed by the wind and transported, with snow movement occurring during winds as low as 3 to 8 m s^{-1} (Mellor 1965). Snow aged by freeze–thaw activity and metamorphism may require winds of 30 m s^{-1} for movement to occur. Most snow is transported by saltation, a bounding of snow particles across the surface within about 1.5 m of the snow surface (Verge and Williams 1981). It is thought that additional snow particles are ejected into the airstream by the impact of landing particles (Mellor 1965).

Wind direction and speed, the distribution, size, shape, orientation, and roughness of objects at the earth's surface, and topography all affect the distribution of blowing snow and subsequent drifts (Verge and Williams 1981). Chain-link fences are very effective snow fences. That is, their mesh fabric sufficiently slows the wind speed to cause snow to be deposited near them. Without the fences an area may be blown free. The buried fence shown on the cover of this report protects an area on a flat-topped mountain. The entire mountaintop is blown nearly free of snow except where the building and fence perturb the wind flow sufficiently to cause snow deposition.

Before an IDS is installed, the probable location of pre-existing snow drifts may be determined by inquiring of local observers or by modeling the site and testing within a wind tunnel; the latter approach is effective if the IDS is to be used with new construction that would alter historical patterns of snow drifting. Drifting occurs on both the upwind and downwind sides of fences, which leads to deposition within the clear zone between double fences and alongside the outer or inner fence, depending on wind direction. While fences on high ground may often be blown free of snow, fences in roadway cuts or small open valleys are susceptible to filling in because wind speed is reduced there. After a site has been constructed and operated for several winters, the typical locations of snow drifts can be mapped. From this map a strategy can be developed for snowdrift control by plowing or installation of snow fences. This may reduce the majority of problems with drifting snow by causing it to be deposited before it reaches IDS locations or perimeter fences.

Snow cover. Snow cover is a seasonal phenomenon and could as easily be discussed with seasonal weather effects. However, the snowpack is very dynamic and changes at the quasi-periodic time scale. Therefore, its effect upon IDS operations may change dramatically throughout the winter season as a result of quasi-periodic weather changes. Microwave IDSs that are oriented nearly parallel to the ground experience a propagation loss that is dependent on the wetness and depth of the snow cover (Peck 1991, 1992a). For a bistatic microwave IDS with a 120-m-long detection zone (transmitter and receiver units separated by 120 m) and antenna center height of 60 cm, monitoring the automatic gain control of the receiver, together with snow depth, indicated that a snow depth of ~0.5 m would have saturated the system's automatic gain control so that further propagation loss of the microwave signal would not have been compensated (Peck 1992b). Saturation would occur at a lesser snow depth when the snow cover is wet. Since the detection capability of the microwave IDS would be suspect when the signal being analyzed for evidence of an intruder is below the required minimum for processing, snow cannot be allowed to accumulate in an microwave IDS's detection zone. The depth of snow that can be tolerated depends upon the length of the IDS's detection zone, since the greater the reduction in signal strength due to beam spreading, the less automatic gain control is available to compensate for other sources of signal attenuation, such as the loss of microwaves within a snow layer. Another reason to remove snow from microwave IDS detection zones is that snow that is deep enough to conceal half the body thickness of a crawling intruder may prevent his detection by a bistatic microwave IDS, while snow of any depth reduces the detection capability of a monostatic microwave IDS in a range-dependent manner (Peck 1992b).

As a snow cover deepens, it may extend into the lowest beam of a near-infrared beambreak IDS, eventually causing sufficient blockage that a nuisance alarm is generated. Snow accumulation at the base of a security fence may change the vibration characteristics of the fence panel, thereby affecting the detection capability of fence-mounted IDSs. At SOROIDS, when deep snow covered the lowest 75 cm of the chain-link fence fabric, two optical-fiber (cable) IDSs on the fence detected taps to the fence (a nondestructive substitute for cutting) through the snow cover, although with different degrees of reliability (Peck 1994b).

Conversely, a snow cover is an ideal background for passive infrared IDSs because its spatial and temporal variations in temperature are more limited than with grass or other vegetated surfaces and so the likelihood of nuisance alarms is greatly reduced (Peck 1993b). A snow cover may also be a favorable background for passive infrared IDSs to detect intruders if the intruder cannot sufficiently reduce his thermal contrast with the snow. It is not necessarily the surface temperature of the snow cover that influences whether an intruder has a sufficiently low thermal contrast to escape detection by a passive infrared IDS. If there is a negative temperature gradient from the ground surface to the snow surface, then an intruder crawling through (not on) the snow cover exposes the warmer snow at depth. The intruder's thermal background is therefore this warmer snow, not the cold snow surface, and so the thermal contrast may be less than expected based on the temperature of the snow surface or approximated from air temperature. The intruder may still be detected, provided that the passive infrared IDS alarms not at the intruder, but at the disturbance to the snow cover (thermal contrast between the surface and deeper, exposed snow) as it propagates through the detection zone.

Snow cover is the end product of accumulation and ablation, and it is a composite of the many forms of precipitation that produce it: sleet, rain, freezing rain, and various types of snow. It is also a product of snow ablation processes, including melting at the snow/air interface and at the snow/soil interface, sublimation, and erosion by wind. In addition, the snowpack changes in character with temperature changes, condensation of moisture on the snow surface, surface melting, and rainfall on its surface. Finally, the snowpack experiences continual metamorphism that increases its density and dramatically changes its crystal structure, thus changing physical properties affecting its thermal, acoustic, and mechanical properties. As a snow cover develops, both its increasing depth and the formation of ice layers serve to further isolate the underlying ground and any buried ground-motion IDSs from mechanical waves generated by a moving intruder, particularly if the snow cover is firm enough to support the intruder's weight. The induced ground motion eventually is sufficiently different in its characteristics (amplitude, frequency content) that it does not satisfy the alarm criteria of ground-motion IDSs. If the snow cover is accompanied by frost penetration, it also contributes to the loss of detection capability by buried ground-motion IDSs (Peck 1994d)

Temperature and the temperature gradient within the snowpack affect the hardness, dryness, density, and crystal structure of a snowpack. Warmer regions, such as coastal areas and regions receiving little snowfall because they have near-tropical climates, often have wet, dense snowpacks. Colder, dry, interior regions, such as the northern Great Plains, often have less dense and less cohesive snowpacks. As a result, snow is more likely to be redistributed by wind in regions with dry, less-dense snow (McKay and Gray 1981). If seasonal snowfall amounts are similar, seasonal snow depths may also be greater in colder regions because snow accumulations last longer and less melting occurs during the winter to diminish the pack.

Wind affects the distribution of snow, snowpack density, and crystal structure by breaking crystals into smaller sizes (McKay and Gray 1981). Snow depths are usually greatest in forests and similar areas where wind speeds are lowest, because intercepted snow is also shaken from tree crowns by wind-induced vibration (Miller 1966). In studies of snow distribution by cover type, snow depth was found to be inversely related to canopy density, with a thin, leafless hardwood canopy allowing large amounts of snow to reach the forest floor and conifers allowing only small amounts of snow to reach the forest floor (Eschner and Satterlund 1963). Snow depth was found to decrease by cover type in the following sequence: brush hardwoods, northern hardwoods, red pine, Norway spruce, and open land.

Snowpack densities after individual storms are often directly related to air temperature and wind speed, with higher densities associated with temperatures near freezing and higher wind speeds (McKay and Gray 1981). Fresh snow deposited under calm, cold, and dry conditions has a density of about 0.06 g cm^{-3}, whereas snow deposited under warm, moist gale conditions has a density of about 0.33 g cm^{-3} (Marcus 1969). New-snow density varies with air temperature measured 1.2 m above the snow surface (U.S. Army 1956), increasing by 0.0029 g cm^{-3} $°C^{-1}$ with increases of air temperature from –18 to 4.5°C.

Snow metamorphoses as it ages, changing from small angular crystals to large, rounded crystals. Snow begins to age and metamorphose immediately after falling due to contact pressure between snow crystals and temperature and vapor pressure gradients within the pack. Wind also increases the density of snowpacks by sublimating and fragmenting snow crystals into extremely small particles, allowing them to repack to high densities, as high as 0.45 g cm^{-3} (Perla and Glenne 1981). Though snow can attain a density similar to that of pure freshwater ice, as is the situation with windblown wet snow, old snowpacks rarely attain a density greater than about 0.54 g cm^{-3} (U.S. Army 1956). During this process, snowpack volume decreases, and crystals become more intimately connected. Other physical processes effective in increasing snowpack density are percolation of snowmelt or rainwater through the pack and freezing and plastic deformation of the pack due to the weight of overlying snow and traffic over the snow.

Bilello (1969) demonstrated the effect of snowpack aging upon snowpack density from measurements at 27 North American stations in a diversity of climates. He found that average seasonal snowpack density increases as average seasonal temperature decreases and wind speed increases. Bilello found densities to be greatest in the cold northern plains and tundra of the continent, which are also locations where roughness is least and wind speeds tend to be higher. In general, snowpack density is also least within forests that are not subject to high winds or thaws, and greatest densities are found on nonvegetated open areas subjected to frequent melt episodes. McKay and Gray (1981) provide a brief overview of expected seasonal snowpack conditions in various specific landscape types.

The ability of snow to act as an insulator decreases as snowpack density and thus snow thermal conductivity increase (Yen 1965). However, the thermal conductivity of snow is nearly always less than that of soil, and as a result, snow is generally an excellent insulator. If soil is covered by a deep, low-density snowpack before temperatures become very cold, the soil will be protected from freezing or at least deep freezing. However, if soil freezes deeply before the snowpack is developed, it will warm little or only slowly until the snowpack has melted, because any thermal energy from the atmosphere will be consumed as latent heat for melting the snowpack. Therefore, soil frost established before the snowpack forms will generally remain for most of the winter, though the frost does degrade slowly from the bottom through the winter due to the release of heat stored deep within the soil. As a result, a soil that is deeply frozen and cold before the establishment of a deep snowpack will warm slowly during the duration of the snowpack.

Keeping snow cleared around fences and near buried IDSs will allow differential soil freezing, which may cause problems with IDSs. If snow removal occurs along only part of a buried IDS's detection zone and the frost depth varies within the zone, then there may be location-dependent differences in detection capability of electromagnetic and ground-motion IDSs. Ground motion is induced by sound as well as by contact forces, although the acoustically coupled component is neglible in hard-frozen soil (Peck 1992c). Sound propagating over or through a snow cover is attenuated and altered in its frequency content by the interaction of the compression waves with the porous snow (Albert 1987).

The albedo of snow, its ability to reflect sunlight, is also affected by the depth and age of the snowpack. Albedo affects the surface energy budget and thus the air temperature. More importantly, however, it affects the brightness of the snow surface when illuminated by the sun. A fresh, deep, dry snowcover has an albedo of about 0.95 to 0.75, whereas an older, wet snow cover's albedo is about 0.45 to 0.35 (Table 2). The difference is primarily due to a reduction in albedo for near-infrared radiation, with little change in snow albedo for visible radiation (Wiscombe and Warren 1980). Since albedo is a function of snow density, snow crystal size, and wetness, it decreases as snow ages. Albedo is also a function of the solar elevation, so the visible and near-infrared contrast of an intruder with a snow background changes with time of day (Peck 1994c), which may introduce a time-of-day dependence to the detection capability of video-motion IDSs. As solar elevation decreases, albedo increases and, if the surface is wet or icy, specular reflection may occur, rendering video-motion IDSs ineffective by saturating the camera scene or causing nuisance alarms.

In general, it takes about three weeks for a snowpack's albedo to decrease from about 0.8 to 0.4 (Riley et al. 1972). In addition, the albedo of snowpacks less than 10 cm deep will be affected, in part, by characteristics of the surface underlying the snow (Geiger 1965). Some sunlight will penetrate completely through a snowpack less than 10 cm thick, heating the substrate and possibly causing melting from below and warming of the soil surface. This is most true for snow that has not been packed by foot or vehicular traffic.

High snow albedos may, at certain times of the day, sufficiently brighten scenes that scene contrast is reduced in imagery from surveillance cameras. High albedo also keeps the snowpack surface cold, thereby increasing the thermal heterogeneity of scenes that include objects that are internally heated (such as buildings) or that absorb more solar radiation and so are warm relative to the snow. The cold snow background may be better suited for intruder detection if the intruder's heat loss and clothing provide a large thermal contrast with the snow cover. Low albedos allow the snow surface to absorb more solar radiation, and so the snow surface warms. If it reaches 0°C and becomes wet, the spatial heterogeneity of the thermal background decreases, and the temporal variation in temperature distribution stabilizes while melting continues. When the near-surface air also is moist and warm, there is little thermal contrast between objects and the snow cover, so the likelihood of nuisance alarms and intruder detection may be less for thermal infrared IDSs (Peck 1993b). A wet snow surface may cause problems for microwave IDSs because it gives a stronger backscatter (return) due to a higher reflection at the air/wet snow interface. Note that wet snow is more absorptive than dry snow, so the microwave return from within the snowpack (volume scattering) will be less when the snow is wet.

Other processes can also cause snowpack surface wetness that will decrease the efficiency of microwave systems. Obviously, rainfall on the snowpack or melting of the snow surface by contact with warm air, such as in the warm sector of a cyclone, can wet the snow surface. High humidity can also wet the snow surface and produce fogs, affecting thermal infrared, near-infrared, and visual systems. If the snowpack surface is cold and a warmer and humid air mass moves over the snowpack, the lower layers of the air mass may be chilled to below the dew point by the snow surface. Condensation will occur onto the snow, wetting or depositing hoarfrost on the snow surface. At the same time, the lower layers of air in contact with the snow will be chilled, producing advection fog, fog that is produced when warm, humid air moves over a cold surface. Advection fogs can be very dense and will persist as long as the air mass and/or the snowpack persist.

Snowpacks ablate in various ways. Areas that develop deep seasonal snow packs have the greatest problems in spring when the snowpack disappears. Ideally, a snowpack will melt slowly, allowing meltwater to infiltrate the soil and percolate to the water table. Instead, often the snowpack ripens, reaching a temperature of 0°C throughout, before melting. This may be due to rainfall that releases large quantities of heat within the pack or to gradual daytime warming with refreezing at night. The result is a snowpack with a large amount of liquid water held within pores between snow grains. Additional rainfall or a series of warm days can cause rapid and destructive melt with high runoff volumes and subsequent flooding and erosion. In general, the longer spring melt is delayed, the greater the danger of destructive runoff and flooding because of stronger solar radiation and increased probabilities of rainfall (Male and Gray 1981). The runoff may jeopardize IDSs in obvious ways, such as disrupting the alignment of line-of-sight sensors or exposing buried IDSs, but a less apparent problem is the destabilization of security fences while the soil around post footings is saturated. Soil will often become saturated during snowmelt because frost within the soil may not allow the surface meltwater to percolate through. Most saturated soils have low bearing capacities and thus are less able to withstand the loading imparted by wind action on security fences or aboveground IDSs. Buried electromagnetic IDSs would experience a reduction in detection capability, with the severity dependent on the thickness of the saturated layer relative to cable depth. Buried IDSs that detect on the basis of ground motion have improved detection capability in wet or damp soil as compared with frozen or hard, dry soil (Peck 1994d).

Even less extreme snowmelt events can affect IDS operations. Melting and refreezing of the pack surface, especially in the presence of rainfall, can cause ponding of water on the snow surface as impermeable ice layers form. The snow surface will often be wet on a daily basis, and even at night if advection or radiation fogs form. The wet snow surface and ponded water produce a thermally uniform surface, near 0°C, which is favorable for reliance on thermal infrared IDSs. Pooled water on the snowpack is potentially disruptive to microwave IDSs because of reflections off the water surface. Ponding also is a frequent winter event during extratropical cyclone passage. As the storm approaches, freezing rain associated with the warm front may seal the snow surface and so prevent infiltration of water. When the precipitation becomes rain as the front approaches, water ponds on the surface. This situation may persist for up to 36 hours until colder air arrives behind the cold front, when the surface refreezes and snow may be deposited on top of the wet or ice-covered snowpack. This sequence can repeat itself frequently in late winter or spring, but it can occur, especially at southerly or coastal locations, at any time during the winter. While the frozen pool remains uncovered by snow, there may be nuisance alarms by thermal infrared IDSs if radiation in the 8–12 μm range reflects off the ice surface into the of the IDS's detection zone. There have been instances at SOROIDS when passing trains that are several hundred yards beyond the well-localized detection zone of a passive infrared IDS have caused nuisance alarms during periods when the surface of the snow cover was ice glazed, but not otherwise. Similar nuisance events occurred with people passing near the IDS's detection zone at locations that never led to alarms in the absence of the glazed surface.

Atmospheric icing

Atmospheric icing is the formation of ice on surfaces from moisture delivered by the atmosphere. Atmospheric icing can result from the direct deposition of ice from water vapor or from the deposition of liquid water droplets either from clouds or precipitation. Accumulations resulting from liquid water droplets in the air are typically most destructive and most likely to disrupt IDSs.

Hoarfrost. Hoarfrost is deposited directly from water vapor and produces a layer of ice crystals on surfaces that have cooled to the frost point of the air (similar to the dew point, but at temperatures below 0°C) (Minsk 1980). Hoarfrost is distinct from frozen dew because to produce frost, water vapor is deposited directly from the gas-

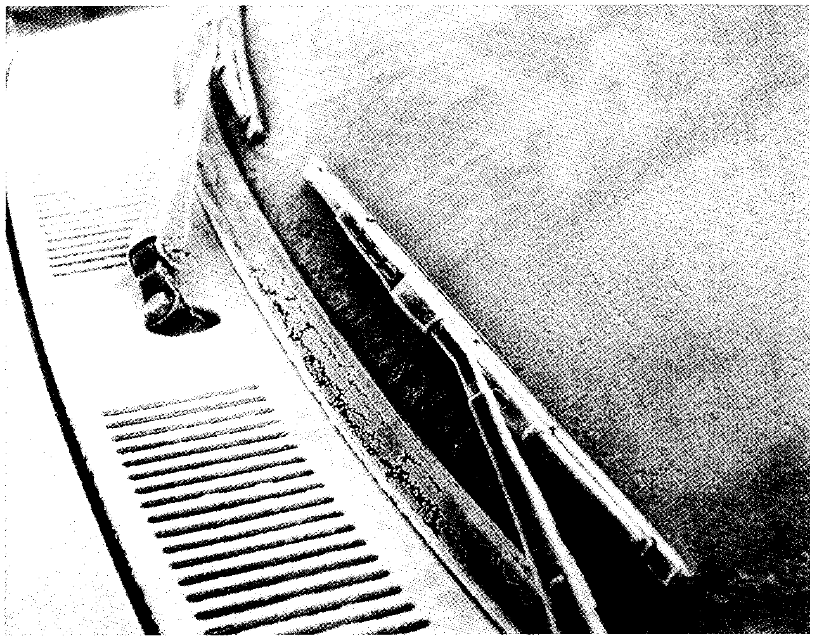


Figure 20. Hoarfrost deposited on an automobile windshield.

eous state to the solid ice state, without passing through the intermediate liquid state. Therefore, frozen dew is and looks like clear, frozen water drops, whereas hoarfrost looks like white ice crystals composed of needles, plates, and other forms, depending upon the frost point temperature.

Hoarfrost forms on calm, clear nights, when objects exposed to the night sky cool radiatively to the sky. The clear night sky is typically radiatively very cold—much colder than the atmosphere surrounding objects on the earth's surface. This is most true when a strong high pressure cell with dry, frigid, arctic air has settled over an area (Ryerson et al. 1994). Atmospheric gases gain and lose energy radiatively less efficiently than solid objects; thus wires, plant stems, and automobile windshields (Fig. 20) all radiatively lose heat more rapidly to the night sky than the surrounding atmosphere does. In addition, the cold sky returns little radiation, thus the net loss of energy, or heat, from these solid objects is large. As a result, these solid objects cool to the frost point before the air around them, causing water vapor to be deposited from the air onto the objects as frost.

The importance of exposure to the cold sky is illustrated best by the windshield of an automobile parked with tree limbs hanging over the right side; the windshield will frost on the left side and remain frost free on the right because the tree limbs are warmer than the clear sky and shield the glass from the sky. For this reason also, frost rarely forms on cloudy nights because clouds are much warmer than the clear sky, so objects at the surface cool more slowly and may never cool to the frost point because they are receiving considerable radiation from the clouds.

Hoarfrost can also form on objects that have become cold-soaked by frigid air residing over the area. If warm, moist air quickly advects into an area cold-soaked by a previous frigid air mass, hoarfrost will form upon all objects that warm slowly because of their relatively high thermal mass. In this case, frost will form in higher wind speeds, and exposure to the sky or orientation has little effect. Meteorologically, this situation may occur as a high pressure cell is retreating to the east and a storm system is approaching from the west, advecting warm, moist air into the area.

Hoarfrost forms a very low-density deposit usually only a centimeter or less thick, though on flat surfaces, such as glass, it is typically less than 0.1 cm thick. The deposit on glass can be soft, or very firm and difficult to remove. On curved surfaces, such as wires and tree branches, and on some flat surfaces, the deposit may consist of large ice needles or crystals that resemble snowflakes in shape. Hoarfrost adds little mass to objects upon which it accretes, because its density is typically less than 0.1 g cm^{-3}. However, it does obscure lenses and windows, and it increases the drag of air passing over wires, though the de-

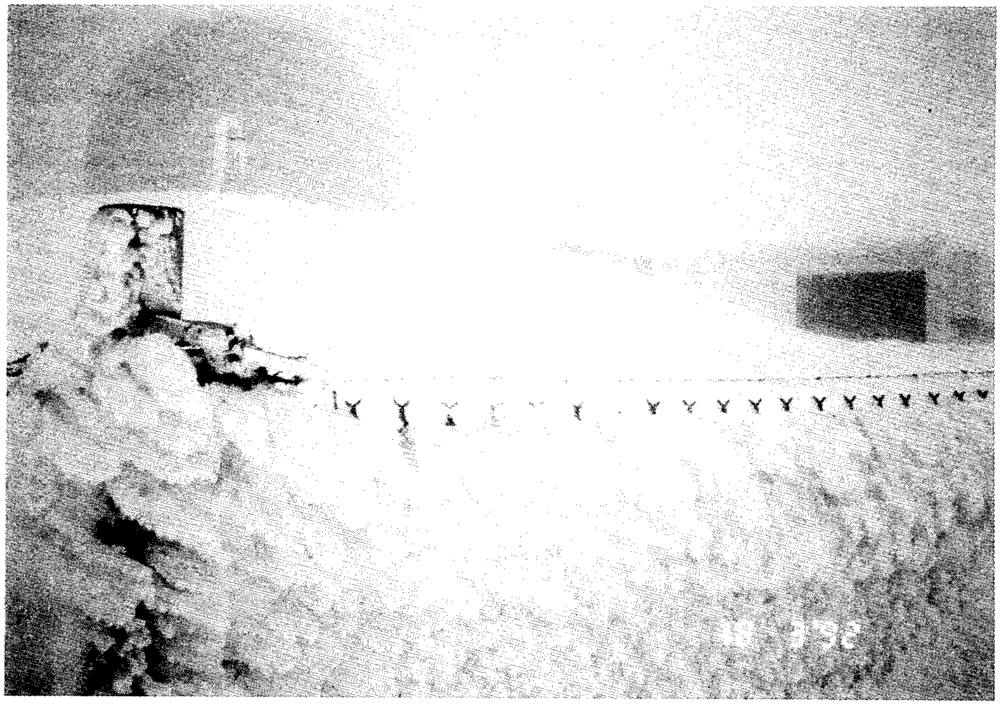


Figure 21. Rime ice deposited on a perimeter fence guarding a radar station. Rime ice closes fence openings and increases wind drag, causing the fence to shake.

posit will rapidly be removed by the wind or by sublimation if wind speeds reach more than a few meters per second. As a result, hoarfrost can be a problem for unheated surveillance cameras.

Hoarfrost typically does not remain on objects for long. Most often, if skies are clear, it disappears completely shortly after sunrise. Under frigid air masses or if clouds move in after the frost has formed, hoarfrost can reside for longer before sublimating away. Typical radiatively formed hoarfrost events lasted from 5.3 to 17.8 hours in New Hampshire (Ryerson et al. 1994). Extreme hoarfrost mass accumulated on a 100.0-cm^2 black metal plate overnight was 3.0×10^{-2} g cm^{-2} (Ryerson et al. 1994).

Rime ice. Rime ice is a result of the movement of supercooled fog droplets across a site (Fig. 21). Fog droplets typically average from 10 to 20 µm in diameter. Droplets this tiny will not typically freeze in the free air until they reach temperatures of –30 to –40°C. They are thus supercooled, but they are not stable. If they strike another object, however, they will rapidly freeze, forming a white (because of air trapped between the rapidly freezing droplets), feathery accretion on the upwind sides of objects.

The amount of accretion on any object is a function of the concentration of liquid water in the air, the size of the droplets, the wind speed, and the shape of the object the droplets strike. Of these factors, the most changeable and most important are fog liquid water content, wind speed, and the initial shape of the object. Rime ice accretes rapidly at rates of several cm hr^{-1} at higher wind speeds (generally greater than 15 m s^{-1}). Rime ice also accretes fastest on objects of small diameter and on sharp edges. For example, wires and pipes less than 2 cm in diameter typically have high accretion rates, as do the edges of beams and corners of structures. Flat surfaces, such as building walls, and round objects 50 cm or larger in diameter will usually have very little accretion. Accretions of several meters can form on objects on high mountains in high winds in as little as 24 to 48 hours.

Rime ice forms when fog and wind are present. This means that at low elevations, such as along a coast or lake, rime ice forms when fog is advected off the water. Rime ice rarely forms to any extent within radiation fogs because air is typically calm in such fogs.

Rime ice is most commonly found where upslope fogs occur on hills or mountains or where clouds either intercept a mountaintop or are created by the presence of the mountain. Wind speeds are commonly higher in these locations, temperatures are colder, and the frequencies of fog or cloud are greater. The occurrence of rime

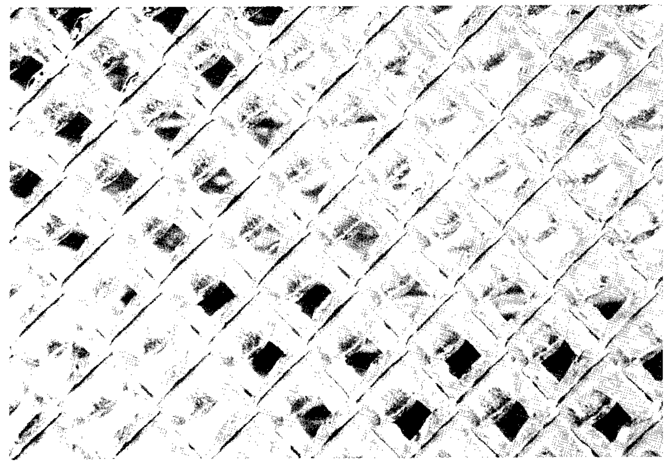


Figure 22. Glaze ice deposited upon a perimeter fence in New York. Glaze ice decreases fence fabric openings, increasing wind drag and shaking the fence.

ice usually increases with elevation, with rime icing being accreted up to 39% of the time during the winter on high mountains in the eastern United States (Ryerson 1988a, 1990). At any elevation, icing is most intense at locations exposed to the wind. Ryerson (1987) illustrated the effects on ice accumulation rates of even small hummocks on a mountainside. In the eastern United States, rime ice forms most frequently on mountains immediately after cold front passage where cloud formation is frequent, wind speeds high, and cold is encroaching (Ryerson 1988b). However, rime is also common within warm fronts that may produce a denser, almost clear accretion.

Rime ice, with a typical density of about 0.6 to 0.7 g cm^{-3} (although 0.2 to 0.9 g cm^{-3} are found), can add considerable mass to IDS hardware. Chain-link fence fabric, with its small-diameter wires, often fills completely with ice, increasing wind drag but perhaps decreasing sensitivity to intruder-caused vibrations. Ice-filled fences also decrease visibility and cause reflections in visual systems. Trip wires and other long-span, small-diameter wires such as are used with taut-wire IDSs will ice rapidly. The accretions may wrap completely around the wire because, as the icing occurs on the side of the wire, the wire, having little torsional rigidity, twists under the asymmetrical weight distribution. Eventually the wire may rotate completely around so that icing occurs on all sides, producing a sleeve of ice.

Rime icing may affect the operation of visual and infrared (both thermal and near-infrared) systems by covering the surfaces of unheated cameras, lights, transmitters, and receivers. Since rime icing forms most rapidly on the sharp edges of objects, if these systems are protected by rainshields with sharp edges, the accreting ice may bridge over windows and radiating surfaces without touching them. In addition, the glare of rime ice is strong in sunlight and may saturate video systems and surveillance cameras. Perhaps the most significant feature of rime ice is that its formation during conditions of fog and high winds will affect IDSs most when security systems are already being compromised by the effects of the fog and wind.

Finally, rime ice often unloads from structures suddenly, causing extreme shaking and vibration if the structures are wires and fences. Such conditions are likely to produce nuisance alarms by fence-mounted IDSs until most of the ice is gone. Unloading begins shortly before temperatures rise above freezing in strong sunlight, and in high winds.

Glaze. Glaze is arguably the most destructive form of atmospheric icing because it affects large geographic areas and can accumulate large masses quickly (Fig. 22). Glaze is a product of rain, perhaps supercooled, falling through air that is colder than freezing onto below-freezing surfaces (Bennett 1959). It is usually a product of warm fronts in extratropical cyclones, but it can occur in other meteorological circumstances. Glaze formation typically requires that an inversion be present in the atmosphere. Glaze is most frequent in the midlatitudes. It can occur at any time of the winter in southern regions of the United States and similar areas in the midlatitudes on other continents. It is also common along coastal areas where warm air is available for overrunning cold air at any time during the winter. However, in higher latitudes, such as along the U.S.–Canadian border, glaze is most frequent in the fall and spring when warm air is more likely to penetrate these regions. Areas frequented by glaze storms, such as the plains of the midwestern United States and south-central Canada, sustain considerable damage as a result of glaze accretion—especially to electrical transmission lines and communications towers.

Glaze produces a clear deposit of ice, with a density of 0.8 to 0.9 g cm^{-3}. It forms in the following way, best illustrated with a warm front (Fig. 16). Warm, moist air overrides cold air at the surface and cools adiabatically to the dew or frost point. If the frost point is reached, snow forms and remains snow as it falls to the earth's surface. As the warm front approaches, the air aloft warms and so melts the falling snow that passes through it, producing rain. When the rain falls from the warmer air into the colder air below, which may be 300 to 1000 m thick, the raindrops cool to the freezing point or colder (supercooled rain). The drops subsequently strike cold objects at the surface and freeze. Freezing occurs sufficiently slowly that air is excluded and clear ice results. Glaze does not require a warm front to form, though that is the most common environment for glaze formation. It can also form as a result of warm air overriding cold air trapped in mountain valleys. Subsequent rainfall may produce glaze in these valleys before the cold air can be eroded away.

Glaze is a very damaging accretion. It adds considerable weight to structures as a result of its high density. Its accretion pattern differs considerably from rime since it is deposited by drizzle or raindrops (Bennett 1959). At these drop sizes, the shape of the accretion surface has less impact on the amount of ice that accretes; the orientation of surfaces is more important. Because glaze is a product of precipitation, it accretes preferentially upon objects oriented skyward. Icing is also greatest on the windward sides of objects. Security fences (excluding outriggers) have little direct exposure to the sky. However, even a slight wind inclines the fallpath of the raindrops toward the fence, allowing the drops to wet the fence fabric or mesh. Glaze accretions are often thicker during drizzle than during periods of higher rain rates because the latent heat of fusion can be removed more quickly from the smaller and less numerous drops found in drizzle. Therefore, ice loads are often greater after drizzle than after rain.

Heavier rain, however, has an additional effect because water that does not freeze runs off, frequently producing icicles. Icicles may form on the edges of rainshields protecting cameras, thermal infrared IDSs, or the transmitters and receivers of near-infrared IDSs, thus blocking their line of sight.

Glaze unloading occurs in approximately the same conditions that promote rime unloading. One difference is that glaze usually persists longer if it has completely encompassed smaller-diameter materials, such as wires. In addition, glaze is less susceptible to wind unloading than rime because of its smaller thickness and less pronounced orientation to the wind.

The potential impact on IDSs is varied: security fences and attached IDSs that become weighted by glaze may collapse or experience other physical damage, fence motion caused by the ice unloading when wind or warming temperatures cause it to break apart and fall through the fence may produce nuisance alarms by fence-mounted IDSs, and the intact ice coating may dampen the fence motion caused by an intruder to the extent that his activities (cutting, climbing) pass undetected. Limited experience with ice-coated fence-mounted IDSs at SOROIDS indicates that a thin, continuous coating of ice on the chain-link fence fabric and attached IDSs sufficiently changes the vibration characteristics of the fence that the IDSs do not alarm at fence taps (a nondestructive substitute for cutting the fence fabric). When the ice coating is manually cracked, lessening its continuity and rigidity, the IDSs again alarm at taps to the fence fabric (Peck 1993a, 1994b). Glaze on the snow surface in a camera scene can direct strong reflected light at a surveillance camera. This may saturate the camera detector and render video-motion IDSs ineffective.

Lightning

Lightning, a product of thunderstorms, is an electrical discharge resulting from charge imbalances between two clouds (the most frequent occurrence) or between a cloud and the ground (Uman 1984). The electrical discharge, which is observed as a lightning stroke and heard after the stroke as thunder, results from the flow of current to satisfy the imbalance.

Theory states that charge imbalances, primarily between the atmosphere and the earth's surface, begin during fair weather when the earth, which is normally negatively charged, attracts positive ions from the upper atmosphere (the ionosphere at an elevation of about 100 km), which is normally positively charged and a conductor. The ionosphere's positive charge is a result of ionization due to the impact of cosmic rays from the sun. These positively charged ions, which are heavier than electrons, sink to the earth's surface because they are attracted by the earth's generally negative charge and because they are heavy. Electrons, at the same time, gradually leak from the earth's surface into the atmosphere, with greatest intensity through pointed, conductive objects such as fence posts. The result is that the earth's surface first gradually becomes neutralized, and then becomes slightly positively charged over time. The lower atmosphere correspondingly becomes slightly negatively charged. Lightning restores the charges to normal by pouring electrons back into the earth's surface (Day and Sternes 1970). The charges do not restore themselves without a lightning discharge because the lower atmosphere is normally a poor conductor of electricity. As a result, a charge buildup occurs, similarly to an electrical capacitor.

As cumulus clouds build prior to becoming thunderstorms, a process of charge separation occurs within the clouds. Lightning cannot occur unless the upper part of a cumulus cloud is below freezing and the bottom above freezing. The top of a developing thunderstorm becomes positively charged and the bottom negatively charged because of charge separation processes, which are theorized but not well understood. One theory states that falling raindrops, which are positively charged because they originate near the top of the cloud (which is itself positively charged because it is nearer to the positively charged ionosphere), attract electrons, which are negatively charged, and drag them to the bottom of the cloud as they fall. Another theory claims that when raindrops that are carried up into the cloud to produce hail freeze, shards of ice splinter off, carrying positive charges higher into the cloud, leaving the rain drop negatively charged (Ahrens 1982).

Eventually the bottom of the cloud is very negatively charged, and the top positively charged. The earth's surface is also positively charged from the transfer of ions during fair weather. As a thunderstorm moves, its negatively charged base attracts positive ions at the earth's surface, which migrate to the area under the storm. This process builds an even stronger charge potential, to about 100,000,000 volts between the earth's surface and the cloud base.

As the charge builds, current tries to flow between the cloud and the ground, but the electrical conductivity of the atmosphere is so poor that this cannot readily occur. However, this begins the process of forming a lightning stroke. The strong cloud–earth electrical potential begins to draw current up from the earth into the atmosphere. This flow is most effective through conductive objects that are high and pointed, such as fence posts and towers (perhaps housing cameras and other electrical equipment). Current flow through these objects into the atmosphere ionizes the atmosphere above, improving its conductivity and occasionally causing a visible glow, called corona, due to the intense collisions of ions with air molecules. They are also locations where a subsequent lightning stroke is highly probable.

At the same time as a corona forms due to the electrical potential at the surface, current flow begins from the cloud base in the form of pilot leaders, which are ionizing thrusts of current down into the atmosphere below the cloud. These thrusts average about 100 m long each, and are about the diameter of a human arm (Day and Sternes 1970). Ionization of the air improves its conductivity. Repeated thrusts of current through the pilot leader allow it to reach closer to the earth's surface. Eventually the pilot leader becomes sufficiently close to an object releasing ions at the ground surface that a full conductive path is made between the cloud and the ground, and a lightning bolt occurs. At that moment, a thrust of negative charge races from the cloud to the ground, replenishing the earth with negative charge. This often creates a charge imbalance between the base of the cloud and another cloud, causing a subsequent cloud-to-cloud discharge. All of this occurs within a few microseconds, and repeated strokes may occur in rapid succession within the same conductive path until all charge potential is dissipated and the conductive path is lost.

The current discharged to the ground continues to flow as ground currents, because when a lightning bolt discharges, its surge of electrons flows into the soil. The resulting current flows most efficiently through moist soil, which during a thunderstorm is usually surface soil because it has been wetted by rainfall. As a result, ground currents flow along the soil surface. Ground currents have the potential for damaging IDS electronics and injuring personnel and often kill cattle standing under trees for shelter (Viemeister 1972).

Lightning is most common in thunderstorms, which are a warm-weather phenomenon. Thunderstorms are produced by mechanical lifting of warm, humid air by cold and warm fronts and topography, and by convection within warm, humid, unstable air masses. Thunderstorms are not found where temperatures are subfreezing throughout the atmosphere.

Thunderstorms may be found anywhere on earth. In the United States, they are most common in Florida, with about 90 days per year reporting thunder. Frequencies decrease radially from Florida to the north and west, with up to 60 storm days per year in the Colorado Front Range area. Frequency drops to about 25 thunderstorm days per year in New England and to only 5 days per year on the West Coast (Ahrens 1982). In general, the frequency of lightning strikes also increases with elevation (Viemeister 1972). Thunderstorms are most frequent in the late summer and in the late afternoon at the times of maximum heating. Seventy percent of all lightning deaths in the United States occur from June through August and between noon and 6 p.m.

SEASONAL WEATHER EFFECTS

Seasonal weather effects cycle once annually and are caused by changes in solar radiation received at the earth's surface due to the tilt of the earth on its axis. Daily and quasi-periodic weather are superimposed upon seasonal weather and, as a result, the weather on any day is an expression of some combination of changes on all three time scales.

The impact of seasonal changes in weather is dependent on location. The results of seasonal weather change are clearly recognizable at mid- and high-latitude sites that may experience winter conditions of frozen soil, persistent snow cover, and below-freezing temperatures and summer conditions of high heat, humidity, and rainfall. Other locations may experience less obviously pronounced seasonal weather changes that have equally significant effects on IDSs.

For many IDSs it is the transitional periods of seasonal change, rather than the more stable conditions of roughly midwinter or midsummer, that are the most troublesome. This is because IDSs generally can be adapted, through changes in operator-selected parameters, to work well under certain categories of conditions. Security personnel with long tenure at a site may know from experience that a buried ground-motion IDS needs to be set at a higher sensitivity during winter when the soil is likely to be frozen (Peck 1994d) or that the sensitivity of a fence-mounted IDS has to be reduced during summer because the thermally expanded chain-link fence is looser than in winter and wind-induced alarms would be too numerous otherwise (Peck 1993a, 1994b).

The reliability of IDSs is less certain during periods of frequent change in site conditions. Depending on a site's location, topography, and weather, its IDSs may experience a fairly rapid succession of different operating conditions in the spring and autumn. The overall trend of warming in spring and cooling in autumn may be interrupted by exceptionally warm or cold days. It is impractical to respond to these relatively short yet pronounced changes in site conditions by adjusting IDS sensitivities. Instead, for the duration of the episodes of substantial weather change, security personnel must revise their reliance on the IDSs in use.

This section explains weather changes that are primarily a result of the seasons. It includes consideration of the typical major changes in site conditions that may occur.

Solar radiation

The earth's orbit lies on the plane of the ecliptic, which extends approximately through the sun's equator. The earth's axis is tilted 23°27′ from a perpendicular to the plane of the ecliptic. The axis is always tilted in the same direction relative to the cosmos, and as a result at all times of the year the North Pole axis of rotation is pointed nearly at the star Polaris. This phenomenon, parallelism, provides the seasons because, as the earth revolves around the sun, at times the North Pole is oriented towards the sun and at times away from the sun.

The seasons result as the orientation of the earth's axis changes in its position relative to the sun (Strahler and Strahler 1983). During the Northern Hemisphere (NH) summer, the North Pole is tilted toward the sun, causing longer daylight and higher elevation of the sun above the horizon relative to winter. Maximum tilt of the NH to the sun occurs during the NH summer solstice, which is on or about 22 June. At this same time, the Southern Hemisphere (SH) is experiencing the winter season, after the winter solstice on or about 22 June, because the South Pole is tilted away from the sun. As a result, SH day lengths are shortest and solar elevations are lowest.

The reverse situation occurs six months later. At this time, during the NH winter and the SH summer, the earth's axis is oriented the same direction in the cosmos as six months earlier, due to parallelism, but the earth is on the opposite side of the sun, making the tilt reversed relative to the sun. The date of maximum tilt, and thus the summer and winter solstices in the respective hemispheres, is on or about 22 December.

Halfway between the solstices, at the equinoxes, the earth's axis is nearly parallel to the sun's axis. This characterizes the start of the transition seasons of spring and fall, on or about 21 March and 23 September, respectively, in the NH. On the equinoxes, solar radiation is entering the top of the NH and SH atmospheres equally, and day length is everywhere 12 hours in duration.

Some effects of day length and sun angle have been demonstrated in the diurnal section of this document. Methods of computing day length and solar angle are available either mathematically (Walraven 1978) or from sun charts (List 1958, Mateer and Godson 1959, Brown 1973). Sun charts allow day length to be estimated for any day of the year, and solar elevation and solar azimuth to be estimated for any hour of any day of the year. However, sun charts are only usable for the latitude for which they are constructed. The analemma* may be used to calculate solar declination for any day of the year, and from that the elevation of the sun above the horizon for any location, but only at solar noon (Strahler 1960). (Declination is the angular distance in degrees from the equator to the latitude where the sun's rays strike the earth's surface orthogonally at solar noon. The analemma provides the declination for any day of the year.) The analemma also allows the equation of time to be computed, which, after adjusting for location within a time zone, allows time corrections to be made for seasonal changes of the earth's speed in its orbit (List 1958).

The astronomical seasons, defined by earth–sun relations, produce the meteorological seasons, which are the atmospheric response to the astronomical changes (Tremberth 1983). The meteorological seasons are delayed, because of thermal lags, about 27.5 days on the average for the conterminous United States (Tremberth 1983). Lag is larger in more oceanic locations, such as most of the Southern Hemisphere. The impact of the astronomical seasons on weather are dramatic because of their effect upon insolation (incoming solar radiation).

Insolation varies seasonally because day length is seasonally dependent. Day length is always 12 hours at the equator, where there is no notable seasonal change. On the equinoxes, day length is 12 hours at all latitudes. On the solstices, however, day length changes are greatest with latitude. For example, during the NH summer solstice, the day is 24 hours long poleward of 66.5°N, whereas there are no hours of daylight at this time poleward of 66.5°S.

Those areas that experience no daylight receive essentially no insolation, gaining energy primarily from atmospheric advection of sensible and latent heat and from ocean current transport of sensible heat. The radiation budget is negative, and the heat budget is negative as well. That is, more heat is lost than is gained. Areas of negative energy budgets often extend nearly to the equator in the winter hemisphere during the solstice. The summer hemisphere experiences an excess of radiative energy, with more radiative energy received than lost.

Sun angle also varies with the astronomical seasons, with lower average sun angles in the winter hemisphere than in the summer hemisphere. Solar intensity, discounting atmospheric effects, is a function of the trigonometric sine of the angle of the solar beam above the horizon. At noon on the solstice, the mean sun angle in the winter hemisphere is 26° (assuming negative sun angles are zero), and in the summer hemisphere the mean sun angle is 59°. The shallower the sun angle, the lower the solar intensity on a horizontal surface. It follows, therefore, that on the solstices the mean solar intensity on a horizontal surface in the summer hemisphere, from sun angle alone, is approximately twice the solar intensity on a horizontal surface in the winter hemisphere.

Seasonal differences in day length and solar intensity create dramatic changes in local energy budgets. In the winter hemisphere, both sun angle and day length decrease as latitude increases, producing a rapid decrease in insolation with latitude, resulting in a steep insolation gradient with latitude. Since radiation drives the energy budget, and thus sensible heating and cooling, a thermal deficit is created that increases with latitude. General atmospheric circulation responds to energy gradients, and the gradients are largest in the winter hemisphere, so mean wind speeds and general storm activity are greater in the winter hemisphere.

In contrast to the winter hemisphere, the summer hemisphere is relatively quiet because there are no strong energy gradients with latitude. Again, this is because of the coincident relationship between day length and sun angle with latitude. In the summer hemisphere, day length increases with latitude, which offsets the decrease in sun angle with latitude. Sun angles are never very low on the summer solstice at solar noon in the summer hemisphere. Consequently, in the lower latitudes where day length is only about 12 hours, sun angles are very high, creating high insolation. As latitude increases, the sun angle decreases, but day length increases to 24 hours,

* A graph shaped like the number 8 and often printed on the equator in the Pacific Ocean on globes.

which compensates for the lower sun angle. Because of this, the summer hemisphere does not experience the strong energy gradient with latitude experienced on the solstice in the winter hemisphere. In fact, there is almost no change in daily insolation with latitude in the summer hemisphere near the equinox (Barry and Chorley 1970). Therefore, there is a weak energy–thermal gradient with latitude, and thus no need for the atmosphere to transport large amounts of energy meridionally. As a result, the atmosphere of the summer hemisphere is not as active, on the synoptic scale, as that of the winter hemisphere. The atmosphere of the summer hemisphere generally has fewer extratropical cyclones, the cyclones are not as strong, and general atmospheric wind speeds are slower. Summer weather, as a result, is more local in origin, whereas winter weather is more synoptic in origin. For these reasons, the winter hemisphere is considered the more storm-ridden, with more changeable weather. The seasons thus control the overall level of weather activity within each hemisphere because of the changing seasonal needs to transport energy poleward and equatorward.

Other seasonal effects also change the intensity of weather with season. The winter hemisphere, because of the lower overall sun angles, experiences greater problems with topographically created shadows. Poleward-facing slopes are shadowed more in the winter than in the summer and therefore many mountain valleys receive even less of the already low solar radiation provided by the meager day lengths and low sun angles. Some deep mountain valleys may actually receive no direct-beam solar radiation on some days during the winter (Garnett 1935). Those days of deepest shadow may occur on or near the solstice, depending upon the arrangement of topography in the area.

In contrast, steep equatorward-facing slopes in the winter hemisphere may actually experience higher solar intensities in the winter than in summer because they may more directly face the solar beam at that time, which may be near the horizon. Vertical cliffs may receive several times more intense solar radiation than nearby horizontal surfaces (see Shortwave Solar Radiation in the diurnal weather discussion above). For example, at solar noon in January at 45° north latitude, a 60° south-facing slope receives direct-beam solar radiation with twice the intensity of a horizontal surface. This suggests that fences, depending upon their orientation, may receive far more radiation on their vertical fabric in midwinter than does the snow surface around them. For IDSs, higher instantaneous solar loads—and thus differential heating—may be greater on vertically exposed surfaces than on horizontally exposed surfaces.

Vegetation

Vegetation structure of most areas of the earth changes dramatically with season. During the spring, deciduous tree species leaf and annuals begin to grow. By mid- to late summer, deciduous trees reach their maximum leaf area, and annuals have reached their maximum height and growth density. In autumn, deciduous species lose their leaves, the canopy thins, and the land surface becomes covered with dead and dried leaves. Annuals die and dry, with some collapsing to the ground after first frost. During the winter, most deciduous woody plants have lost all leaves, and annuals and the ground surface are commonly covered with snow that reduces surface roughness.

This scenario repeats itself annually in most mid- and high-latitude locations, the major differences by location being the length of each season and the plant species composition. However, variants of the midlatitude patterns also occur in regions without true "winter" and "summer" climates. These areas include the wet-winter, dry-summer Mediterranean climates of the north coast of the Mediterranean Sea and Southern California; the dry-winter, moist-summer regime of the tropical wet and dry areas on the poleward sides of the equatorial tropics in Africa, South America, and Indonesia; and the monsoon climates of South and Southeast Asia. In these latter climates, deciduous trees often lose their leaves during the dry season and annuals die.

The changing seasonal structure of vegetation has large impacts on local energy and heat budgets and on wind regimes at the earth's surface (Schwartz 1992). Most of these seasonal energy budget impacts were discussed under Diurnal Weather Effects.

Seasonal changes in the vegetation and the land surface produce large changes in a site's energy budget (Schwartz 1992). When deciduous trees lose their leaves in the fall, the canopy opens (Frodigh 1967). Few or no leaves remain on the trees to intercept solar radiation, and abundant sunlight can penetrate to the forest floor. The forest floor is also more open during the autumn, allowing the soil to be more effectively warmed. In addition, little solar radiation is being used in evapotranspiration, which consumes large amounts of energy to transpire water from leaf stoma. As a result, nearly all solar radiation can be used for sensible heating of soil, air, and tree trunks instead of being consumed to change liquid water to water vapor.

The amount of solar radiation reaching through the canopy is a function of the leaf area index (LAI). This is a dimensionless measure of the foliage density or, equivalently, of the plant's ability to expose photosynthesizing leaf surfaces within a given area. During the summer, a deciduous forest canopy may have an LAI of 4.9 (Arya 1988). Most crops have LAIs from 2 to 6, whereas pasture, sugar cane, and rice may have LAIs as high as 12 (Chang 1968). In general, as LAI increases so does the utilization of solar radiation falling upon a plant.

During the winter, the LAI of most plants is 0. As a result, energy not used in the canopy for photosynthesis and evapotranspiration is now used for sensible heating and some evaporation. The progression of plant stands or canopies from leafed to leafless changes the use of solar radiation from photosynthesis to other uses, and it changes the distribution of energy from the canopy to the forest or crop floor. In addition, changes in canopy quantity and quality also change the albedo, thus affecting the amount of solar radiation actually absorbed within the vegetation stand.

Geiger (1965) cites measurements in old stands of four deciduous species (red beech, oak, ash, and birch) indicating that with foliage 25% of insolation incident at the top of the canopy reaches the forest floor, whereas without foliage 54% reaches the forest floor. Another study cited by Geiger (1965) showed that the percentage of sunlight in the visible spectrum reaching the forest floor in a mid-European deciduous forest decreased from 51% in March, before buds opened, to 6% in June with leaves opened. Approximately 80 to 90% of the reduction in light intensity is due to the crown area and its branches; the rest of the depletion is caused by trunks (Geiger 1965). Independently of the season, only an average of 21% of light incident at the top of evergreen trees (silver fir, spruce, and pine) reached the forest floor.

The amount of solar radiation reaching through annual field crops varies, primarily with crop height. Chang (1968) summarizes work by Stanhill that shows that radiation reaching the base of a crop is linearly related to its height, with 80 to 90% of sunlight reaching the base of 20-cm-high crops, but only 40 to 50% of sunlight incident at the crop crown reaching the base of 60-cm-high crops. Oke (1978) reports that only 5 to 10% of radiation incident above mature plant stands reaches the ground beneath. Young plants without fully developed crowns allow more radiation to reach the surface, of course. For example, at noon the amount of light reaching the base of a corn crop forty-two days after planting declined from 40% to only 5% seventy-three days after planting (Chang 1968). After crops are cut, such as mowing hay or harvesting corn plants into shocks, or after plants collapse due to frost or drying, radiation reaching the ground surface is most often near 100% of that available, depending upon the amount of litter at the surface.

The ratio of direct beam to diffuse radiation, or shadow illumination, also changes significantly with season. For example, Geiger (1965) reports ratios of light intensity (ratio of direct sunlight intensity to intensity within shadow) within a European oak forest of 17:1 during leaf-out, with ratios increasing to 180:1 without leaves during winter. This indicates less diffuse light within the winter forest, with more "hard" shadows and greater light–shadow contrast.

Even though more radiation can reach the ground surface beneath deciduous trees or in areas of primarily herbaceous vegetation in winter than in summer, solar intensity is lower because of day length and solar elevation. The ground may also remain cool relative to vertical objects such as tree trunks, which are oriented broadside to the direct radiance. Tree trunks are particularly likely to be warmer than the surface of snow-covered ground because they can absorb large amounts of solar energy and heat to higher temperatures than the snow surface surrounding them (Fig. 23; Munis 1992). As a result, an infrared IDS can be presented a thermally complex scene even over snow if defoliated deciduous trees are part of the view.

For microwave IDSs, vegetation consists of plant matter, bound water (water molecules tightly held to the organic compounds by physical forces), and free water (water molecules freer to move within the vegetative material and rotate in response to an applied electromagnetic field). The internal water content of vegetation

Figure 23. Visible and thermal (3–5 μm) images of trees on a sunny day. The tree trunks present a complex thermal background as a result of absorbing insolation (Munis 1992).

varies with the available soil moisture, which influences whether the vegetation is undergoing moisture stress and wilting. Both attenuation of microwaves propagating over a vegetation cover and backscattering of microwaves from vegetation that extends into the IDS detection zone depend on the surface wetness and the internal moisture content of the vegetation. Seasonal differences arise from progressive changes in the amount of foliage, plant height, amount of exposed ground and its moisture content, and the wetness of the vegetation. Grant and Yaplee (1957) document seasonal differences in backscattering for a number of vegetation types.

Wind

Seasonal changes in hemispherical thermal and pressure gradients, as discussed above, dramatically affect the strength, persistence, and directions of regional winds. However, several other seasonal influences on winds can affect the reliability of IDSs.

Seasonal changes in vegetation affect the strength of winds in the lower atmosphere through variation in the frictional resistance to wind motion caused by the plant canopy and through changes in the energy budget that allow more or less convection at some times of the year. These changes potentially affect IDS performance because wind-loading of security fences and the local microclimate of nearby IDSs will be different. Such seemingly innocuous activities as removing or planting trees in the vicinity of IDSs or harvesting nearby crops may have an impact on IDSs prone to wind-related alarms.

Atmospheric motion over a surface is dependent on the relative smoothness of the surface. As the surface becomes rougher, the drag between the air and the surface increases. Accordingly, surfaces are characterized by their roughness length, a measure of the irregularity of a surface, obtained by measuring wind speed at various heights above a surface in neutral stability conditions and finding the intercept where wind speed approaches zero when height is plotted on a logarithmic scale (Nicholas and Lewis 1980). Roughness length is the height at which a neutrally buoyant wind profile extrapolates to a zero wind speed (Oke 1978). The greater the irregularity of a surface, the larger the roughness length and the higher above the surface is the zone of zero wind speed.

Very little seasonal change in roughness length is observed over treeless urban areas, but changes are dramatic in areas dominated by deciduous trees such as parks, open areas, and residential areas (Nicholas and Lewis 1980). Typical roughness values are 0–0.1 cm for sand, 0.02–0.6 cm for open water and snow surfaces, 0.6–4.0 cm for short grass, and 4.0–10.0 cm for long grass (Chang 1968). Roughness length generally increases with vegetation height. Chang (1968) provides relationships between plant height and roughness that show fir forests of 20 m height having a roughness length of 256 cm, orange orchards of 5 m height with roughness lengths of 64 cm, and brush of about 150 cm height with a roughness length of 20 cm.

Wind conditions at an IDS site should be monitored near the ground at the level of IDSs or security fences. The actual wind force on a fence or above-ground IDS is the product of the velocity pressure, a gust response factor (which accounts for the additional loading effects due to wind turbulence), a force coefficient (that depends on the ratio of solid area to gross area of the fence or IDS), and the area of all exposed fence or IDS members projected on a plane normal to the wind direction (ASCE 1990). The velocity pressure, which is proportional to the square of the wind speed, varies directly with a parameter called the exposure coefficient, which depends upon the terrain and obstructions in the vicinity of the fence or above-ground IDS.

An increase in roughness length increases turbulence at the surface, which in turn lowers maximum daytime air temperature, raises minimum nighttime air temperature, increases evaporation, and decreases general wind speed at the surface (Chang 1968). The result is that, as annuals (such as a corn crop) grow, an IDS downwind of the corn a short distance will experience decreasing wind speeds over the growing season. As a crop grows, the ground surface where wind speeds are effectively zero is raised because of the roughness of the plant surface. This zero-plane displacement, roughly 67% of the crop height for most crops, is smaller for crops that sway with the wind and higher for crops that stay erect in the wind. Therefore, the zero plane will increase from 0 m when the corn emerges in the spring to about 1.5 m before the dry stalks are cut in the fall. On the day harvesting occurs, the upwind zero-plane displacement decreases again to near 0 m (because of the remaining stubble it may not be exactly 0 m), and wind speed will increase considerably at the surface. Wind loading of nearby security fences and free-standing IDSs, such as bistatic microwave or near-infrared beambreak units, for which alignment is important, will thus change as dramatically.

Even apparently minor changes in plant height can have significant effects on winds and IDS performance, perhaps in ways not expected. It is apparent that increases in wind speed, or increases in turbulence, may affect fence shaking. Simply allowing grass length to increase from 3 to 7.5 cm increases turbulence sufficiently at the grass crown height to increase evaporation rates by as much as 50% (Sellers 1965). As a result, soil moisture is depleted more rapidly and the soil around buried IDSs will dry more rapidly. If the soil dries too quickly, grass may wilt and dry, thus changing the complexity of the thermal environment monitored by thermal infrared systems.

Seasonal changes in wind activity over deciduous forests are not as large as over cropped land, because tree trunks and branches are always present to cause roughness. Roughness decreases significantly, however, when leaves are shed. Geiger (1965) demonstrated that wind speed decreases rapidly within the upper portions of a 24-m high, 115-year-old oak forest canopy when leaves are out. Wind speeds decreased from over 4 m s^{-1} to about 1 m s^{-1} within the upper 4 m of the canopy and were nearly constant from that height to the surface. Before buds opened in the spring, wind speeds of about 4 m s^{-1} above the canopy did not decrease to 1 m s^{-1} until reaching the forest floor. At a height of 4 m above the forest floor, the number of calm hours rose from 67% before leaf-out to 98% after leaf-out (Geiger 1965).

In summary, there is a sharp drop in wind speed within the upper canopy of a leafed forest, but little change of wind speed from the level of maximum leaf area to the surface. An exception to this is an occasional peak near the forest floor if the forest floor is clear. After leaves are shed, there is less forest-induced drag, and wind speeds increase throughout the depth of the forest (Oke 1978). As additional illustrations, in Vienna, Austria (Landsberg 1981), and in Nashville, Tennessee (Frederick 1961), wind speeds are reported to decrease 20 to 30% when the leaves are out on deciduous trees.

Whereas the growth of vegetation during the summer generally results in a larger roughness length, the development of a snow cover in winter reduces the roughness length. For example, a summer surface covered with short grass with a roughness length of 0.6 to 4.0 cm, or long grass with a roughness length of 4.0 to 10.0 cm, may be covered during the winter with a deep snow cover having a roughness length of 0.02 to 0.6 cm. The result of this seasonal change in roughness length is that turbulence may be lower in the winter, thereby reducing buffeting of fences, but wind speeds may be higher because of the lower surface friction.

IDS sites may create roughness in a snow cover where none naturally exists. Plowing snow and creating snowbanks increases the roughness of a site, and thus the wind turbulence and gustiness. Depending upon the location, orientation, and height of snowbanks, the turbulence regime created may be sufficient to cause nuisance alarms in fence-mounted IDS systems. Leaving snow unplowed, or disposing of it carefully away from security zones, may reduce the incidence of nuisance alarms in windy conditions.

Several other wind-related seasonal effects can affect IDS performance. As discussed earlier, wind speeds are often higher during winter because of stronger meridional temperature, and thus pressure, gradients during the winter. Diurnal wind cycles are also less likely to be evident during the winter because the cold ground or snow surface does not heat the overlying air sufficiently to cause convective air movement and the resulting advective air flow that produces some local winds. Moreover, the nature of winds changes.

The highest summer winds occur within thunderstorms, producing often destructive force. Destructive thunderstorm winds are, however, geographically localized and usually last only a few minutes to a few hours. High winter winds are often of much longer duration and cover larger geographical areas, but they are less likely to be of destructive force than high summer winds. In the summer, therefore, a fence-mounted IDS may be driven closer to alarming because of high gusty winds, but the winds are usually of short duration. During the winter, the winds may be high for longer periods of time because they are cyclonically driven, but they may not push the IDS as close to alarming as during the summer (Peck 1993a). A contributing factor to whether the IDS experiences wind-related alarms is the seasonal difference in fence stiffness due to thermal contraction or expansion of the fence fabric. There are also seasonal differences in soil rigidity that influence the likelihood of wind-related nuisance alarms, with dry or frozen soil providing better support for fence posts than does deeply saturated soil or thawing, frost-heaved soil.

Dominant wind directions may also change seasonally. The magnitude of wind loading of a fence depends in part on the fence's orientation relative to the wind direction (ASCE 1990). Consequently, the portions of a fence-mounted IDS perimeter that are most severely affected by high winds may change seasonally with the change in dominant wind direction. For example, in the southeastern United States, southerly winds are more frequent during the summer months, but northerly winds are more common in winter, with the highest winds coming behind cold fronts. Depending on the topography and structures at an IDS site, the ~180° change in wind direction may significantly alter the relative sheltering from wind loading that a security fence or aboveground IDS experiences. This problem is most dramatic in areas with monsoon winds, winds that dramatically change direction with season, such as occur in South Asia.

Fog

Fog is classified by its mechanism of formation into six types: advection, steam (convection), valley, radiation, frontal, and upslope fog. All have some seasonal extremes, and several were discussed earlier in this report. The most seasonal are advection, steam, and radiation fogs.

The obvious impact of fog on security operations is decreased visibility with surveillance cameras, which may render video-motion IDSs ineffective. Fog also reduces transmission of thermal infrared and near-infrared radiation. The high propagation loss in fog leads to nuisance alarms by near-infrared beambreak IDSs (Peck 1994a); conversely, the thermal uniformity of a fog background reduces the probability of nuisance alarms by passive infrared IDSs. Although an intruder may have a high inherent thermal contrast in fog, transmission loss may be sufficiently high that the intruder's thermal contrast apparent at a passive infrared IDS is below detection and/or alarm level.* Fog has a negligible impact on microwave transmission over distances of less than several kilometers, so microwave IDSs are unaffected by fog.

Advection fog occurs when warm, moist air moves over a cold surface, chilling to the dew point the lower air layers in contact with the surface. Over land surfaces, especially if covered by snow, and over water bodies, advection fog is most common in the late winter and spring; the season extends through at least early summer over larger, deeper water bodies. During this time, land and water surfaces are cold, and warm, humid air advecting poleward, such as in the warm sector of a cyclone, produces dense and deep fog. The probability of advection fog can be predicted if it is known that the ground surface is colder than the advecting air's dew point temperature. For example, if the dew point temperature of air moving over a ripe, melting snow cover is 4°C, then fog is likely to form. This is especially true at night with the added benefit of radiational cooling.

The extinction of thermal infrared radiation (8–12 μm) in a heavy advection fog is slightly wavelength dependent (Fenn et al. 1985). The overall reduction in thermal contrast due to scattering and absorption would be ~95% over a 100-m path. Extinction of near-infrared radiation, such as that of beambreak IDSs, and visible radiation would be similar in magnitude, but almost totally due to scattering (negligible absorption).

Convection fog (steam fog—Appendix D) forms when cool air moves over warm, moist surfaces. This is most likely to occur in the fall when cold air masses from continental interiors move equatorward over warm bodies of water. Since a warm surface is beneath cold air, the lowest layer of air in contact with the warm surface heats, creating instability. As the warm air rises from the surface, water vapor is carried aloft, which promptly cools to the dew point, producing wisps of fog having the appearance of filaments, or gossamer, rising from the surface. It rapidly dissipates with height and evaporates after rising only a few meters.

On the Great Lakes, steam fog forms within the cold, dry air that follows cold fronts moving from the west or northwest. The cold, dry air may be 5 to 40°C colder than the water and may be moving at gale force (Eichenlaub 1979). By the time the fog reaches the southern and eastern shores of the lakes there is sufficient moisture in the air and sufficient instability to cause rain or snow squalls. In the early winter these are the infamous lake-effect storms well-known east of Cleveland, Ohio, and Buffalo, New York, and in the Watertown, New York, area. The fogs producing these storms may be up to 1500 m thick (Eichenlaub 1979).

Less dramatically, steam fog occurring at night over smaller lakes, such as Lake Champlain in upper New York and Vermont, produces a thick deck of cumulus clouds about 300 m above the water surface. In calm conditions, and if temperatures are not too cold, the clouds burn off by late morning. If there is wind, however, the clouds move onshore, producing snow squalls and rime ice deposits in the Green Mountains about 30 km to the east. If the water–air temperature difference is very large, steam devils (rotating masses of air meteorologically similar to dust devils) will rise off the surface. If they reach land, they are deprived of the necessary surface heating and so do not survive long.

Radiation fog (Appendix D) is the least dramatic seasonal fog. It can occur at any time of the year, predominantly at night. However, winter is the preferred time for the formation of radiation fog because cold, dry air masses invading from continental interiors during the fall and winter allow longwave radiation to readily escape the earth's surface. The result is rapid cooling at the surface and the formation of shallow surface inversions, usually less than a few meters to 10 m thick, but occasionally hundreds of meters thick. Radiation fog often starts shortly after sunset in calm winds and reaches its maximum development by sunrise. It most often forms over grass surfaces where moisture is available and rapid cooling can occur. It rarely forms over pavements or in urban areas where energy is added to the

* J. Lacombe, 1994, CRREL, Hanover, N.H., personal communication.

atmosphere by the surface as quickly as it is lost radiatively. Occasionally the fog forms within a valley bottom where abundant moisture is available from streams, hence it is called a valley fog. Radiation fog typically burns off within a few hours of sunrise.

The wavelength-dependent extinction of thermal infrared (8–12 μm) radiation in light to moderate radiation fog is an order of magnitude less than that in a heavy advection fog (Fenn et al. 1985). Consequently, the reduction in thermal contrast over a 100-m path is only ~25% in a light to moderate radiation fog, compared with ~95% in a heavy advection fog. The extinction coefficients of near-infrared and visible radiation are 2 to 3 times that of thermal radiation, so this type of fog is potentially more detrimental to near-infrared beambreak IDSs (~60% transmission loss) and video-motion detection IDSs (~60% reduction in visual contrast) than to passive infrared IDSs.

Soil moisture

The moisture content of soil at any time is the result of a dynamic interaction between sources of water supply to the soil and mechanisms removing water from the soil. The water remaining within soil after supply and demand have acted is soil moisture, that moisture residing between soil particles within soil capillaries.

Methods have been developed to account for soil moisture supply and demand. In addition, time-series patterns of soil moisture throughout the seasons of the year have been meticulously constructed (Mather 1978). Understanding the effects of soil moisture upon IDS reliability requires an understanding of soil moisture supply and demand processes.

Soil moisture is water held between soil particles by capillary attraction. It is water found within the phreatic zone above the saturated water table or, more explicitly, within the root zone of plants growing at the soil surface. Soil above the water table is rarely saturated—the state of there being only water and soil particles and no air between the soil particles—and soil is rarely ever completely dry with no water and only air between soil mineral particles. There is nearly always some water within the soil, even if it is bound in molecular thin layers to soil mineral and humic particles. Most often, soil moisture exists as a mixture of air and water held within soil pores between grain contacts by capillary attraction. It is this water that undergoes the largest seasonal variation and so has the greatest impact upon IDS operations. The amount of water held in soils that are neither "oven dry" nor saturated is typically soil-type-dependent, but varies between about 5 and 30% of the soil volume (Carson 1969).

Soil moisture is an important control of soil thermal properties. Soil heat capacity is a summation of the heat capacities of individual soil components (mineral, organic, gaseous, and liquid) in the proportion of their composition within the soil (Table 3). The heat capacities of most soil components other than water are small. Therefore, changes in soil water content generally cause large changes in overall soil heat capacity. For example, with an unfrozen sandy loam soil, as moisture content increases 7% (from 10 to 17% moisture content, by weight) or 8% (from 17 to 25%), its heat capacity increases 22% in each case. Overall, soil heat capacity increases almost linearly as soil moisture increases (Buckman and Brady 1969).

Thermal diffusivity is the ratio of the thermal conductivity (see Soil Temperatures in the diurnal weather discussion above) of a material to its heat capacity; it represents the rate of temperature change to be expected with a given temperature gradient. The highest value of thermal diffusivity, and thus the greatest ability for soil temperature to change under a given temperature gradient, occurs at intermediate soil moisture contents (Higashi 1953). This means that extremely dry or extremely wet soils change temperature slowly in response to a temperature change imposed by the atmosphere.

Soil moisture also affects soil thermal characteristics by controlling the type and health of vegetation that grows at the soil surface. Most plant species flourish within only a narrow range of soil moistures. The plants in turn control soil moisture through root depth, evapotranspiration rates, and their ability to shade the soil surface. Plants also enable the soil surface to absorb greater quantities of precipitating water. Plants, therefore, influence soil moisture by affecting moisture infiltration and by their own consumptive processes.

Moisture content also strongly influences the electrical properties of soils. The reflection of microwaves at the air/soil interface depends on the dielectric constant of the soil (also referred to as the real part of the soil's complex permittivity), and the penetration of microwaves within the soil is determined by the imaginary part of the complex permittivity or its related parameter, the soil's electrical conductivity. (Note that reflection coefficients cited in the literature are often implicitly for near-normal incidence of microwaves at the ground (or snow) surface. With IDSs, the microwaves typically are incident at near grazing angles (~2°), and so both the amount of reflected radiation and its relative change with the moisture content of the ground cover can be significantly different.) The dielectric constant and electrical conductivity of a soil depend primarily upon its moisture content and less significantly on soil textural composition (Ulaby et al. 1982). They vary with the frequency of the electromagnetic radiation, with temperature, and with salinity because the dielectric constants of bound and free water depend on these factors, and they vary with the bulk soil density, shape of the soil particles, shape of the water inclusions, total volumetric water content, and relative fractions of bound and free water (which is itself related to the soil's particle size distribution) because these factors relate to the amount of water held within the soil (Ulaby et al. 1986). The strength of the aboveground electromagnetic field set up by a buried electromagnetic IDS depends on the soil's electrical conductivity: the smaller the electrical conductivity (the smaller the unfrozen moisture content) of the soil, the less attenuated the field is within the soil with distance from the transmitting cable, the stronger the aboveground electromagnetic field is that establishes the detection zone, and the larger the IDS response to an intruder's movement within the zone.

In a similar manner, moisture content influences ground motion, only now it is differences in the elastic properties of air and water that lead to the dependence of soil mechanical properties, such as compressional wave speed and deflection under loading, on moisture content.

Soil moisture sources

Precipitation is the primary source of soil moisture, with snowmelt, dew, fog drip, and direct condensation the other major contributors. Primary factors that determine the effectiveness of rainfall are the soil's antecedent moisture condition, the rate or intensity of rainfall, the amount of rainfall, and the soil's infiltration capacity. Soils with high moisture contents prior to rainfall cannot readily accept additional moisture. If pore spaces are nearly full of capillary water, additional water will drain through the soil to the water table, only temporarily increasing the soil moisture content within the phreatic or root zone. The rate of water drainage from the root zone to the water table is dependent upon the soil's ability to percolate moisture through it. Dense clay, silt soils, and soils that have been compacted generally have low percolation rates. These soils also typically have a low infiltration capacity, which is the ability of soil to absorb water that is on the surface. Clay and silt soils typically allow less than 5–15 mm of water to infiltrate per hour, whereas healthy loam and loamy sand soils allow up to 25–50 mm of water to infiltrate per hour (Morgan 1969). In fact, soils inadequately protected by vegetation cover that are subjected to heavy rains are often sealed at the surface by the impact of the raindrops, thereby limiting infiltration capacity when it is most needed (Morgan 1969).

If infiltration capacity and percolation rates are low, water will pond at the surface and eventually run off, perhaps causing erosion and carrying plant nutrients away, as well as possibly exposing buried IDS sensor cables. While they last, such localized pools of water cause the surface reflection of electromagnetic radiation to be different, and they change temperature at rates different from that of the surrounding soil. Particularly if the water surface ripples in the wind, there is a high likelihood that microwave, buried electromagnetic, and thermal infrared IDSs will be subject to nuisance alarms. Visual assessment using surveillance cameras, as well as video-motion IDS reliability, may be degraded if there is specular reflection off the water surface. Should the pooled water freeze, or at least skim over with an ice coating, the surface is likely to be highly reflective and cause extraneous light, microwave, and thermal infrared radiation to be reflected into IDS detection zones.

Other factors affecting soil infiltration capacity are the slope of the land and vegetation type and condition. Rugged, steep slopes with low and incomplete vegetation cover are conducive to rapid water runoff, low infiltration, and rapidly changing soil moisture conditions. Negligible slopes with dense, healthy vegetation cover of forest or grass are conducive to good infiltration and stable soil moisture conditions.

Intense rainfall typically occurs within thunderstorms and during the passage of cold fronts. Heavy rainfall is least likely to change the moisture content of the soil: intense rainfall is likely to run off rather than infiltrate the soil. Light rainfall, which is conducive to recharging soil moisture, generally occurs from warm-frontal situations and nimbostratus clouds. The timing and intensity of rainfall that occurs at each time of the year are critical in determining the amount of moisture in the soil at any time.

Snow delays the recharge of soil with moisture until melting occurs. Deep snowpacks that either supply moisture slowly throughout the winter during intermittent thaws, or that melt slowly in the spring, are ideal for recharging soil moisture. Rapid melt in the spring may not recharge soils well, and it promotes erosion. When snowmelt is an important source of soil moisture, the timing and intensity of snowstorms is not as important as with rainfall. The primary importance of snowfall timing is its effect upon soil freezing. If the snowpack forms after the soil has frozen deeply, recharge may not be as effective in the spring or during intermittent winter thaws. Formation of a deep snowpack prior to prolonged cold periods may prevent or reduce soil freezing and therefore enhance the recharge of soil moisture in the spring.

Soil moisture consumption

Soil moisture can be considered, in concept, as the balance that results from a budgeting process, with precipitation as the source of soil moisture and evapotranspiration as the consumer of soil moisture (Thornthwaite and Mather 1955). Evapotranspiration, the evaporation of water from the soil surface and transpiration of water from plant stoma, is controlled by different processes than is precipitation; it is the change of phase of liquid water to water vapor and the transport of the vapor into the atmosphere either from plant stoma or from the soil surface. As a result, evapotranspiration is an important mechanism for transporting water and energy from the soil to the atmosphere.

Evapotranspiration cannot occur unless the vapor pressure of liquid water in the soil or within plant stoma is greater than the vapor pressure in the atmosphere. Vapor pressure is controlled by the state of water (solid, liquid, or gas) and by its temperature. At a given temperature, the vapor pressure of water vapor is greater over liquid water than over ice. Vapor pressure over water or ice also decreases with temperature. If the temperature of water within a leaf stoma is sufficient for the vapor pressure of that water to be higher than the atmosphere's vapor pressure, evaporation will occur from the stoma. However, as evaporation occurs, latent heat is removed from the water that remains to change the liquid to a gas, thus cooling the remaining water to the point where its vapor pressure eventually equals that of the atmosphere. As long as there is a vapor pressure gradient from the stoma water to the atmosphere, evaporation occurs. Once the vapor pressures of the atmosphere and the stoma water are equal, a vapor pressure gradient no longer exists and evapotranspiration stops until more energy is added to the leaf water to raise its vapor pressure. In a leaf, that energy comes primarily from solar radiation, with some contribution from longwave radiation and advection of sensible heat.

Intuitively it might seem that evapotranspiration would be more likely to occur when the air is dry, but evaporation of water from the soil and transpiration from plants are not primarily functions of the relative humidity of the air. Energy is the mechanism that drives evapotranspiration. If energy and moisture are available in sufficient quantities to raise the vapor pressure of a water surface above the vapor pressure of the air and to keep it raised even during the consumption of energy from latent heat, evapotranspiration will occur. If sufficient energy (and moisture) is not available, then evapotranspiration will not continue, even if atmospheric humidity is low.

Overall, the following conditions control evaporation rates (Thornthwaite and Mather 1955): 1) the availability of energy, primarily from solar radiation, at the evaporating surface, 2) the atmosphere's capacity to remove water vapor and maintain a vapor pressure gradient through wind or turbulent diffusion removing moist air and introducing dry air, 3) the vegetation's ability to absorb solar radiation and to obtain moisture from the soil through its roots, and 4) the amount of moisture available in the root zone of the soil. This list indicates that evapotranspiration rates are greatest in those locations and at those times marked by availability of energy (primarily from insolation), wind and air movement, plants with roots that penetrate the soil to some depth, large leaf surfaces that grow thickly, and soils with the capacity to hold and provide to plants large quantities of water. All of these conditions may not occur together, lowering evapotranspiration rates.

During the winter there is often little energy for evapotranspiration, yet there may be abundant plant cover and moisture available within the soil. This is the situation in Southern California during the winter, and in many of the southern states of the United States. Evapotranspiration rates are also low in Southern California during the summer, but this is because there is insufficient moisture. At the Bonneville Salt Flats in Utah there is abundant moisture available just below the soil surface, even during portions of the summer, but because there is no plant cover to bring the moisture to the surface, evapotranspiration is negligible. In the plains of eastern Colorado there is abundant energy during the summer and the native plant cover of grasses covers the area thickly, but there is little soil moisture because of the low rainfall rate (less than 50 cm per year), and so evapotranspiration rates are low.

In the eastern and midwestern United States during the summer, there are abundant soil moisture, plant cover with deep roots and large transpiring surfaces, energy, and excellent moisture-holding soils. Evapotranspiration rates are high in these areas during the summer, but unless frequent and abundant rainfall occurs, the ability to evaporate water will quickly exceed the ability to supply water. With insufficient soil moisture, the ability to evaporate water exceeds the water available for evapotranspiration, and soil moisture deficits occur. This dries the soil and wilts and dries plants, thus changing the operating environment of buried IDSs and changing surface temperature patterns in the field of view of infrared IDSs. These changes occur rapidly during the summer because abundant energy is available to evaporate water.

Another example of the seasonal differences in evapotranspiration is the situation in Berkeley, California, in January when on average there is sufficient energy available to evaporate and transpire about 28 mm of water from the soil. However, during that month about 177 mm of precipitation falls. After the energy available removes 28 mm of moisture, 149 mm remain to run off and recharge soil moisture. In contrast, during July there is sufficient energy available to evaporate and transpire about 89 mm of water, but less than 1 mm typically falls as rainfall, and less than 50 mm is available in typical soils under healthy plant cover. This indicates that the soil is drying because there is a demand for 38 mm more water than is available. In reality, even the full 50 mm is not readily available because, as the soil dries, additional volumes of water are more difficult to remove from the soil. The excess energy not used for evapotranspiration is used, therefore, for sensible heating.

Another geographic influence on seasonal differences in evapotranspiration is illustrated by Seabrook, New Jersey, where there is sufficient energy in January to evaporate and transpire only about 1 mm of water, yet about 88 mm of precipitation falls. The excess precipitation may recharge the soil (if needed), run off in streams, or remain stored on the soil surface as snow. In contrast, there is sufficient energy in July to evaporate and transpire about 155 mm of water, but July precipitation is only about 112 mm, leaving a deficit of 43 mm. However, there is some moisture within the soil, typically about 36 mm, some of which can be drawn out. The result is a monthly deficit of about 7 mm of water, enough to cause some drying of plants and soil, but barely noticeably.

The message is that the availability of neither water nor energy alone can increase evapotranspiration rates; both must occur together. When moisture availability (from precipitation) and energy availability (the potential for evapotranspiration) do not coincide and match from month to month, then soil moisture deficits and surpluses occur, generally at about the same time year after year. Therefore, rainfall alone does not determine soil moisture conditions. Seasonal rainfall amounts must be sufficient to supply the changing demands of moisture made by seasonal changes in energy, or deficits and surpluses of soil moisture will occur.

The water budget

The budgetary process of computing monthly soil moisture by accounting for precipitation and evapotranspiration is called the water budget. This is an arithmetic tabulation scheme that allows the monthly demand for water from potential evapotranspiration (the energy available for evapotranspiration), the amount of precipitation falling that month, and the amount of soil moisture available for evapotranspiration to be accounted for. The procedure then determines how much moisture plants must draw from the soil during each month if potential evapotranspiration exceeds rainfall, or how much water will be recharged into a partially dry soil if rainfall exceeds potential evapotranspiration. It also calculates actual evapotranspiration and runoff for the month.

The most comprehensive reviews of the water budget for determining soil moisture at a location without the use of special instrumentation are presented by Thornthwaite and Mather (1955, 1957) and Mather (1978). Soil moisture can also be measured directly using a variety of instrumentation (McKim et al. 1980, Ungar et al. 1992).

If IDSs are sensitive to soil moisture, then the water budget or soil moisture measurement instruments can be used to determine the soil moisture condition. With that information, action can be taken, if desired, to control soil moisture. The type and cost of control depends upon the climate, the soil type, the size of the area requiring control, and whether the problem is moisture surplus or deficit.

Moisture deficits could logically be controlled by reducing evapotranspiration and supplying additional water. Unfortunately, evapotranspiration is difficult to control. Mulches are occasionally suggested as moisture controls, but they are generally considered to be ineffective (Buckman and Brady 1969). Watering, or irrigation, is probably the only effective way to reduce soil moisture deficits. The timing of water applications is important, as is the method of water application. Generally, watering should be done at night when less energy is available to evaporate the applied water before it infiltrates the soil. Water is also best applied by systems that do not spray water into the air, since spraying water promotes evaporation.

Excess soil moisture can be drained. However, some soils, such as clays and silts, drain slowly and may never dry sufficiently. Plants that use large quantities of water and that have deep roots, such as phreatophytes, may be useful for drying some persistently wet soil.

Seasonal soil frost

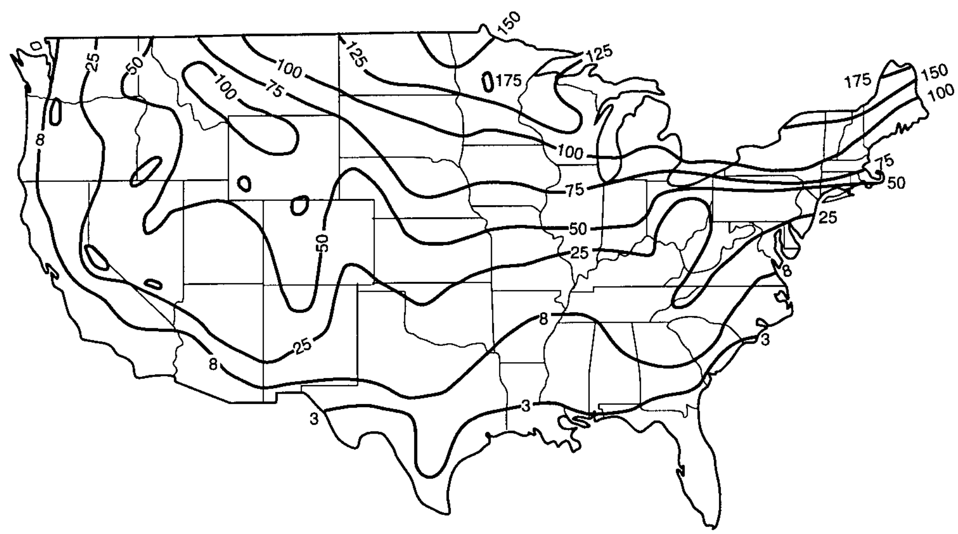
Seasonal soil frost completely thaws every spring and returns again in the fall or winter. It occurs annually over extensive areas of the mid- and high latitudes and at most high-altitude locations. In the Northern Hemisphere it occurs from roughly latitude 30° northward to 60° (Woods et al. 1960). Within this zone, soil frost exhibits continuity, frequency of occurrence, and depth gradients. At higher latitudes, seasonal soil frost is a deeply freezing, continuous winter feature, found on all exposures and slopes and under most soil and vegetation conditions. At lower latitudes, seasonal soil frost is usually shallow and discontinuous, and at its lowest latitudinal limits it is extremely shallow and found only where thermal and moisture conditions are optimal.

Permafrost is a high-latitude (poleward of the seasonal frost zone) and high alpine variant of seasonal frost that is found where soil remains frozen without a complete thaw for two or more consecutive years (Price 1972, Washburn 1973). Technically, the active layer (the zone of soil that thaws seasonally above permafrost and freezes again the following winter) is similar to seasonal soil frost (Washburn 1973). However, the active layer freezes only as deeply as the depth of the previous summer thaw, with further subfreezing temperatures only lowering the temperature of the permafrost beneath. Therefore, seasonal soil frost is seasonal freezing of soil in regions not underlain by permafrost (Strock and Hotchkiss 1939, Woods et al. 1960, Washburn 1973).

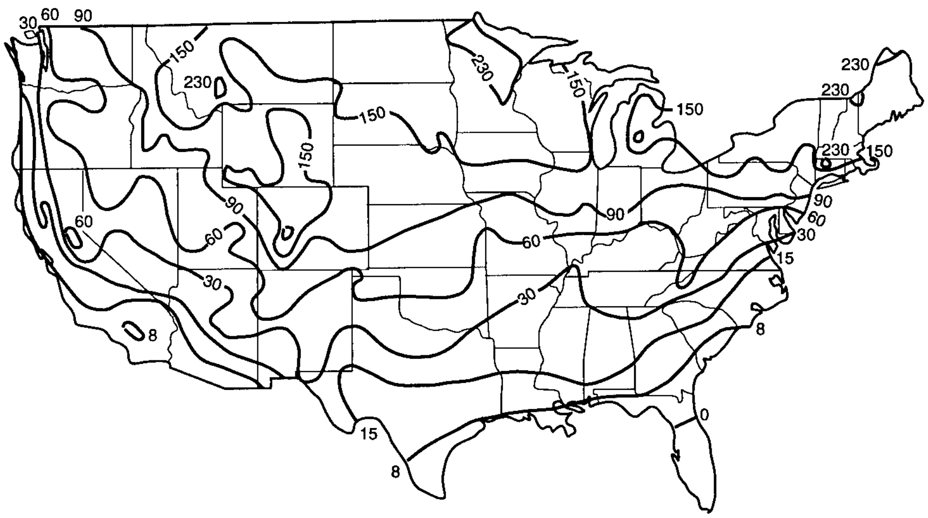
Two measures of seasonal frost variability are frost depth and the number of freeze–thaw events. Figure 24 illustrates the great range of possible frost depths in the United States. For example, St. Louis, Missouri, may expect on the average 35 cm of frost, but in the extreme it may receive 65 cm. Freeze–thaw frequencies are greatest near the equatorward limit of frost. In the U.S. they are generally greatest in the central states and decrease poleward and equatorward (Visher 1945, Williams 1964) (Fig. 8). Thus, the freeze–thaw threshold is crossed most frequently and frost depths are of greatest variability in this region. This imparts a wintertime geographic dependence to the reliability of IDSs that are affected by the unfrozen moisture content of the soil. The number of episodes of change in IDS detection capability would be expected to correlate highly with the frequency of freeze–thaw events. An exception to this would be if installation procedures, such as surrounding buried IDSs with sand, or site-specific conditions, such as described under Diurnal Weather Effects, inhibit variation in the frozen–unfrozen status of the soil.

Soil frost occurs primarily in two forms: that which freezes solid and forms a nearly impermeable layer, known as concrete frost, and that which heaves and produces a loose, porous structure, known as honeycomb, stalactite, or granular frost (Storey 1955, Trimble et al. 1958). Concrete frost is usually found in the open and where soil moisture is high and forms a hard, impervious layer with nearly all soil pores filled with solid crystals of ice. Honeycomb or granular frost, usually found in woodlands, creates numerous pores in the soil because it heaves surface soil layers and occurs under lower soil moisture conditions than concrete frost.

Seasonal soil frost can severely affect the soil condition. Soil infiltration capacity and soil erosion may all change when the soil freezes. Soil bearing capacity and tractionability change with both seasonal and diurnal soil frost occurrences. In addition, soil freezing often draws moisture to the soil surface, which increases the liquid water



a. Average depth of frost penetration.



b. Extreme depth of frost penetration.

Figure 24. Seasonal soil frost penetration, in cm (Visher 1954).

content of the soil as it thaws. This affects the detection capability of IDSs sensitive to soil moisture content. For example, bistatic microwave systems with a detection zone that extends to the ground surface experience a propagation loss that depends on the near-surface soil wetness, with the signal loss becoming a larger portion of tolerable variation in signal strength in cases of large propagation distance (greater weakening of the signal due to beam spreading) and/or poor antenna alignment (Peck 1992a). The detection capability of buried electromagnetic IDSs, which depends on the unfrozen moisture content of the soil, improves when the soil becomes frozen and degrades when the soil thaws (Peck 1992b, 1994d). The poor end-of-winter performance in thawing and thawed soil persists until the water released by snowmelt and pore ice melt has evaporated or percolated away.

The detection capability of buried ground-motion IDSs, however, improves as the soil transitions from being hard frozen to damp after thawing and snowmelt (Peck 1994d), but this higher level of reliability is again lost if the soil becomes hard packed upon drying.

Soil freezing, coupled with heaving, creates a host of problems. Heaving is most severe in locations with moist, silty soils. Capillary action carries moisture from below to the freezing front. If the freezing front encounters a discontinuity in

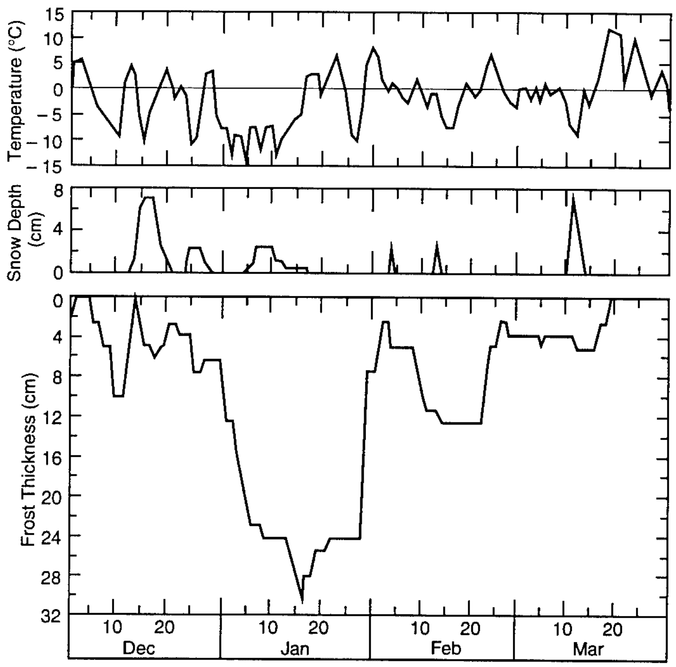


Figure 25. Daily fluctuations of air temperature, snow depth, and frost thickness at Coshocton, Ohio, during the winter of 1968–69. Air temperature was measured at 2 m above the soil surface, and frost thickness, not frost depth, was measured with a frost tube (after Ryerson 1979).

soil structure, an ice lens is initiated. The freezing front halts its advance as the ice lens forms and heaves the soil. The ice lens grows as long as a supply of capillary water can be drawn to it from deeper in the soil. To accommodate the ice volume, the soil displaces in the direction of the lowest confining pressure. Pavements that contain IDS cables may be heaved out of vertical alignment, causing the pavement surface to break and decreasing the bearing strength of the surface. Fence posts may be heaved, causing loss of alignment and large changes in fabric tension. As soil thaws and settles, support for pavements and fence posts is reduced; the soil is often saturated with moisture drawn to the ice lenses, which further decreases its ability to support loads. Fence posts are more firmly anchored in frozen soil in winter or in hard, dry soil in summer than they are in saturated or thawing, frost-heaved soils. There is thus a seasonality to the ability of fence posts to resist fence motion due to wind loading.

Two precautions during fence installation can reduce the likelihood that fence posts in heavable soil will be displaced as the soil heaves. The first is to pour a post's concrete footing into a smooth-sided, cylindrical form set in the ground, such as a Sonotube. This makes it easier for the soil to slip past the footing rather than dragging it upward as the soil heaves. The second precaution is to ensure that no excess concrete is allowed to spill beyond the diameter of the footing. Any lip of concrete is a bearing surface against which the heaving soil pushes.* The concrete footings should extend below the active frost zone if feasible.

The daily variance of weather variables may modify freeze–thaw rates rapidly. Figure 25 illustrates the daily variations of frost thickness, snow depth, and mean daily air temperature from 1 December 1968 through 31 March 1969, at Coshocton, Ohio (Ryerson 1979). Comparison of frost thickness and air temperature indicates that there is a positive relationship between air temperature fluctuations and frost thickness. Air temperature fluctuations alone do not explain all of the variations in frost thickness, however. For example, on 11 and 18 December, maximum frost thicknesses of 10.2 and 6.4 cm were attained. Minimum temperatures prior to attaining those maxima were identical, yet frost thicknesses were different. The second event may have been shallower because the duration of freezing temperatures was shorter than for the first event, and a 7.6-cm snow cover insulated the soil versus a 0.25-cm cover during the first event.

Bouyoucos (1916) broadly classified factors affecting soil temperature as external and intrinsic. External factors are those variables above the soil surface that modify freeze–thaw rates. These include slope, aspect, the presence of insulators at the soil surface such as snow and vegetation, and all meteorological variables. Intrinsic factors are all those within the soil that affect freeze–thaw rates such as soil moisture, soil density, and soil mineral composition.

Frost depths are generally greater on north slopes than south slopes, and steeper slopes have a more dramatic effect on frost depths than gentler slopes (Goodell 1939, Post and Dreibelbis 1942). Slope and aspect also advance or retard the times of frost formation and melt (Kienholz 1940). Regions with rugged topography may include a wide spectrum of freeze–thaw rates as a result of slope and aspect effects.

* F. Crory, 1992, CRREL, Hanover, N.H., personal communication.

In general, frost is deeper and more frequent in open land, deeper in plowed land than pasture, and less deep and less frequent in forests. Within forests, conifers allow deeper frosts than hardwoods. Primary responsibility for these differences with cover type is the vegetative influence upon snow depth, with a secondary influence by litter and soil humus layers.

The single intrinsic variable that has the greatest impact upon frost depth is soil moisture (Chang 1958). Water affects soil temperatures through latent heat exchanges during phase change, through modification of soil heat capacity and thermal conductivity, and through infiltration of rainfall into soil in sufficient quantity to modify and homogenize soil temperature (see Diurnal Weather Effects above).

A particular physical situation may be encountered with buried IDSs located in a subsurface layer of sand that was laid down when the soil was excavated for installation of the IDSs. The sand layer used with buried electromagnetic IDSs typically is located between 15 and 30 cm beneath the surface, with the IDS cables in the middle of the sand layer. Ground-motion IDSs are customarily buried at shallower depth with a thinner sand layer. Sand is dryer than all but extremely dry soil; at SOROIDS, sand in cable trenches typically has a moisture content of ~3% in autumn although the sandy loam surrounding the sand layer has a moisture content of ~17%. Sand grains are larger than most soil particles, so they do not pack as closely and there are relatively large spaces between sand grains. Because of this and the dryness of the sand, the bulk thermal conductivity of sand is low compared to many soils.

Freezing of the soil above the sand layer proceeds as it would if the sand layer were not present. Once the freezing front reaches the top of the sand layer, freezing proceeds more rapidly through the sand layer than it would through silty soil at the same depth. Unfrozen sand of 3% moisture content by weight has a latent heat and a sensible heat capacity that are 34% and 82%, respectively, of that of a silty soil with 10% moisture content. Frost penetration through the sand is a tradeoff between the smaller amount of heat (relative to freezing silty soil) that must be released to cool and then freeze the sand and the lower thermal conductivity of the frozen sand (77% of that of frozen silty soil), which determines the transport of the released heat away from the freezing front. For the sand and silty soil considered here, the net result is that the freezing front reaches the base of the sand layer sooner than it reaches an equivalent depth in homogeneous silty soil (no sand layer) (Peck and O'Neill 1995).

If the weather conditions are such that frost penetration extends deeper than 30 cm, then the presence of the sand layer has two effects on the subsequent freezing history of the silty soil beneath it. First, maximum frost depth is less than it would be if the sand layer were not present (Peck and O'Neill 1995) because the low thermal conductivity of the sand retards upward flow of heat from warm soil at several meters' depth. Heat is retained at the base of the sand layer, and the soil immediately below the sand warms. Because more heat must be released to cool and freeze this warmer soil, and because what heat is released propagates upward more slowly through the frozen sand (compared with what its flow would be in frozen silty soil), the freezing front does not penetrate as deeply below the sand layer. Second, the drier the soil, the greater the frost depth in silty soil beneath the sand layer at a given time (Peck and O'Neill 1995). This is the net result of the smaller amount of heat that must be released to cool and then freeze drier silty soil and of the relatively slow removal of heat through the frozen sand layer.

Just as frost penetration will vary along a buried IDS's detection zone if the moisture content of the soil varies laterally, so will it be nonuniform if the sand layer is not continuous, all other factors (weather, exposure, soil type, etc.) being the same. Soil freezing affects the reliability of buried IDSs, enhancing the detection capability of electromagnetic IDSs and reducing the detection capability of ground-motion IDSs (Peck 1994b, d).

The rate of freeze–thaw ultimately is a function of the soil–atmosphere temperature gradient. All other factors only modify the heat flow rate. Of those other factors, soil moisture has the greatest impact through its influence on heat flow through the soil and because it determines the amount of latent heat that must be removed or added for a given depth of freeze or thaw.

Snow and duff damp the rate of heat flow into and out of soil. Snow is spatially and temporally very variable, and it is important in determining the depth and timing of frost events at any given location. Vegetation, slope, and aspect are also great modifiers of frost depth indirectly through their effect upon snow depth and directly through their effect on solar radiation incident at a location.

SUMMARY

The environment within which IDSs of different phenomenologies at the same site operate is often spatially diverse because of the geographic size of the guarded facility. This fact is readily appreciated by security personnel because they can see the variety of physical conditions represented by the different zones of IDSs. It is less obvious that the IDSs' environment is temporally diverse from zone to zone, with the site weather creating different, location-dependent operating conditions for the IDSs. Although many temporal variations in an IDS's operating conditions may be obvious to a casual observer, less noticeable weather-related changes in its environment may be overlooked when security personnel try to establish the causes of differences in performance of IDSs at the same site or try to interpret conflicting information on any given IDS's detection capability from many sites.

The environmental parameters affecting IDS operations depend on an IDS's detection phenomenology. They may not necessarily be the weather and environmental parameters observed or noted by security personnel, and they may not be extreme conditions or among the conditions typically monitored by meteorologists. Indeed, IDSs may not respond negatively to a change in any one environmental parameter. Concurrent changes in several parameters together may be required to compromise IDS reliability.

The key is that subtle changes, over time, of multiple environmental conditions not readily apparent to humans may cause an IDS to vary in detection capability. The subtlety of change may be due less to the magnitude of change than to the rate of change being slower than humans can readily detect. A good analogy is the way the hour hand of a clock appears not to move when viewed continuously, but it actually advances significantly over the course of an hour.

The goal of this document has been to demonstrate the variety of conditions that can affect IDS operations, explain the temporal (and where possible, geographic) patterns of these conditions, and explain why or how these conditions occur. To actually detect when these conditions are occurring requires careful observation or monitoring of changes in the environment within which an IDS operates or direct monitoring of the IDS's proximity-to-alarm.

One way in which security designers try to attain a high degree of weather independence with their IDSs is to use a suite of IDSs representing several detection phenomenologies. The intent is to guarantee that at least one system is operating reliably when others are erratic due to the weather. The difficulty, however, is determining which systems are compromised due to weather and which systems should be relied upon.

Guidelines presented within this report should help security personnel decide when a system may be suspect due to environmental conditions. Determining when new systems, or systems in unusual environments, are unreliable requires that security personnel have an understanding of how that environment varies. It also requires that they be willing to become sleuths and to carefully note the time, location on the perimeter, and type of degraded performance to try to identify patterns of unreliability that may be correlated with environmental changes, however subtle. Although not all IDS types and environments are discussed in this report, a sufficient understanding of environmental systems has been developed to provide a basis for understanding other environments. Essentially the same processes occur in all environments, from energy budget to water budget. What changes from one environment to another is the magnitude and timing of the processes.

This report has stressed temporal diversity and has provided only modest discussion of the potential geographical weather diversity along a perimeter. This is not because the geographical problem is trivial or lacks complexity; in fact, geographical change along a perimeter may be large. Even if a site has little topographic variation, the path of the sun through the sky is certain to introduce differences in energy budgets along the perimeter. As a result, perimeters should be carefully mapped, with characteristics of the environment noted as it changes with location (see Appendix F). Among the conditions to note are directional orientation, elevation, soil type, vegetation cover and type, proximity to structures, and exposure to seasonal winds. This will provide an awareness of the variation in environment among IDS zones and will assist security designers in selecting suitable IDSs for a site; it will also assist site personnel in solving problems with IDS reliability as they arise.

This report is intended to provide an understanding of the complexity and subtlety of the environments within which IDSs operate and to identify how common classes of IDSs respond to changes in their environment. It is a resource to be consulted whenever attempting to solve an IDS problem resulting from the environment. It does not, and is not intended to, answer all environmental problems that may be encountered by security designers and personnel working with IDSs. Obvious catastrophic weather conditions, such as tornadoes and hurricanes, for example, have not been addressed because their effects are so dramatic and obvious that they do not warrant discussion.

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APPENDIX A: ENVIRONMENTAL INFLUENCES ON EXTERIOR INTRUSION DETECTION SYSTEMS

The environmental parameters that affect the detection capability of exterior intrusion detection systems (IDSs) are reviewed here for general IDS classes; they are summarized in Table A1. The intent of this summary is to assist security personnel in identifying those site conditions that significantly alter the probability of detection or the nuisance alarm rate of exterior IDSs using experience gained from research conducted at SOROIDS, the CRREL IDS research site in South Royalton, Vermont. With that awareness they can then determine from this report when those conditions are likely to change.

Weather-related IDS damage or electronics failure or unreliability is not considered here.

Seagraves (1989) assesses environmental influences on electro-optical devices.

A summary of the principles of extinction of electromagnetic radiation by obscurants (fog, rain, airborne snow) concludes this appendix.

Microwave radar IDSs

There are two types of microwave radar IDSs: bistatic systems, which have separate transmitter and receiver units, and monostatic systems that combine the transmit and receive functions in one unit. The radars incorporated in these IDSs typically have antenna polarization that is E(electric field)-plane vertical, a configuration of the radiation field for which there is a relatively small reflected contribution from the ground surface to destructively interfere with direct and forward-scattered radiation. These IDSs generate an alarm based on the characteristics of a change in the received microwave field, such as would be caused by microwaves scattering off an intruder. An intruder's disturbance to the microwave field is time-varying as he crosses the detection zone, scattering microwaves back toward the transmitting antenna and, for bistatic IDSs, also toward the receiving antenna at the opposite end of the detection zone. Changing site conditions that also cause variations in the microwave field are potential causes of nuisance alarms.

The received microwave field for bistatic IDSs is dependent on microwave transmission over the length of the detection zone. One potential cause of nuisance alarms is rapid fluctuations in direct transmission, which can occur during rain and falling snow. Extinction of the microwaves depends on the number of water or ice particles in the air and on the distribution of particle sizes, so as those factors change, the amount of radiation incident at the receiving antenna changes. For bistatic microwave IDSs, the length of the detection zone remains defined by the separation between transmitter and receiver even when there is severe transmission loss. A walking intruder crossing the detection zone close to the receiving unit should still be detected, although the intrusion alarm may occur among nuisance alarms. The received microwave field also changes because of variable microwave reflection at the ground cover/air interface and at any other reflecting surfaces. The type of ground cover and its moisture content primarily determine the magnitude of the ground-reflected microwave field, while reflections from other objects, such as a chain-link fence or standing water, alter when the position of the object changes, as when the fence fabric moves under wind-loading or the water ripples. Reflections from metal objects or water can vary rapidly and so cause nuisance alarms. A change in the reflective properties of the ground surface typically occurs slowly, so its effect is more likely to be a change in the magnitude of propagation loss over the length of the detection zone, rather than nuisance alarms. If the propagation loss, together with a reduction in signal strength due to beam spreading (dependent on the transmitter–receiver separation) as well as loss due to antenna misalignment, become larger than the automatic gain control of the receiver's processor can compensate for, then the validity of subsequent processing of the signal for evidence of an intruder is suspect. Vegetation and snow on the ground that intrude into the microwave field also cause transmission losses, but their more detrimental effect is to shield a crawling intruder from detection; the relative loss of detection capability depends on the wetness of the snow or vegetation and on the proportion of the intruder's body that is concealed by the snow or vegetation.

The received microwave field for a monostatic microwave IDS is composed of radiation scattered and reflected back to the unit. An intruder crossing the IDS's detection zone alters the mi-

Table A1. Classes of exterior IDSs and the primary environmental factors influencing their reliability.

Type of IDS	*Detection criterion*	*Environmental components*	*Affected by*
Passive (thermal) infrared	Thermal contrast	Ground cover thermal radiance (magnitude, rate of changes, spatial uniformity) Transmission of thermal contrast	Ground cover surface temperature (insolation, exposure, ground cover characteristics [solar albedo, thermal properties, uniformity, wetness]) Wind-induced movement of objects Fog, rain, airborne snow*
Video-motion detection	Visual contrast	Ground cover reflection of visible and near-infrared radiation (magnitude, uniformity) Visibility	Illumination (direct, diffuse, shadowing) Ground cover optical properties Fog, rain, airborne snow* Wind-induced movement of objects
Near-infrared beambreak	Beam interruption	Transmission loss	Fog, rain, airborne snow* Wind-related alignment loss Blockage by snow, vegetation
Microwave radar (bistatic, monostatic)	Characteristics of received micro–wave field	Scattering by airborne particles Reflection, attenuation at ground cover	Rain, airborne snow, fog* Wind-related alignment loss Wind-induced motion of reflecting surfaces Moisture content of ground cover Blockage by snow, vegetation
Buried electromagnetic	Disturbances to aboveground electromagnetic field	Unfrozen moisture content of soil (electrical conductivity)	Soil moisture (rainfall, thawing, influx of snow meltwater) Soil temperature (frozen, unfrozen) Wind-induced motion of standing water
Buried ground-motion (optical fiber)	Changes in light pattern	Unfrozen moisture content of soil (elastic properties, rigidity) Snow cover properties	Soil moisture content (rainfall, thawing, influx of snow melt water) Soil temperature (frozen, unfrozen) Characteristics of snow cover (density, stratigraphy, depth) Wind-induced motion of surface objects
Fence vibration	Fence motion (mechanical, optical, electric sensors)	Wind loading Precipitation impact Temperature-dependent stiffness of fence fabtric Temperature-dependent coupling between sensor cables and fence fabric Ice coating Improved anchoring of fence posts in dry or frozen soil	Wind Rain Temperature Icing, snow adhesion (affect vibration characteristics; nuisance alarms when shed) Soil temperature (frozen, unfrozen) and dryness
Taut wire	Wire displacement	Temperature-dependent changes in wire tension Snow and ice loading of wires	Temperature Icing, snow adhesion (collection efficiency of wire)

* Severity of transmission loss within fog, rain, or airborne snow depends upon rate of precipitation and particle sizes, on wavelength of radiation, and on length of IDS's detection zone.

crowave field by causing an early return of some microwave radiation and diverting other radiation. For this type of microwave IDS, severe transmission loss in rain or falling wet snow can reduce the effective length of the IDS's detection zone. The stronger the backscatter (return of microwave radiation toward the unit), the less the penetration of microwaves through the rain or snow to maintain the extent of the detection zone that exists under clear-sky conditions. Nuisance alarms occur when there is variation in the backscattering from rain or airborne snow, such as when the precipitation rate (which relates to particle size and quantity) changes during a storm, or when there is movement of reflecting objects, such as metal fences or standing water. Snow on the ground and tall vegetation can shield a crawling intruder from detection.

At SOROIDS, a bistatic microwave IDS operating over a 120-m path and a monostatic microwave IDS operating at a range of ~45 m, both 10.525 GHz radars, infrequently experienced alarms during rain or snowfall if winds were calm. Strong winds, whether or not accompanied by precipitation, caused the bistatic microwave IDS to have a high proximity to alarm; this was attributed to disturbances of the microwave field caused by motion of the nearby chain-link fence, which was oriented obliquely to the line-of-sight of that IDS. The monostatic microwave IDS was oriented perpendicular to the fence, with its detection zone (defined by walkthroughs) ostensibly terminating just in front of the fence. It experienced nuisance alarms during high winds, with fewer alarms occurring at low air temperature when the stiffer fence was more resistant to wind-induced motion. It was concluded that both the bistatic and the monostatic microwave IDSs would have had acceptable nuisance alarm rates during precipitation if they had been operating in an open area and not near chain-link fences (Peck 1992b).

Thermal (passive) infrared IDSs

The alarm criteria for thermal infrared IDSs generally are the nonuniformity, magnitude, and rate of change of thermal radiance within the detection zone. A single IDS typically has two thermal detectors. Simultaneous, similar changes in thermal radiance within both detectors' fields of view produce no net sensor response. A differential change of input to the two detectors, such as may be caused by an intruder crossing first one's, then the other's field of view, causes a response that is determined by the magnitude and rate of change of thermal radiance received by each detector. An intruder's thermal contrast with his background depends both on their relative temperatures (temperature difference between intruder and the background he moves across) and their absolute temperatures, since the warmer an object is, the greater the difference in thermal radiance resulting from a 1°C difference in temperature. A thermally uniform background is the ideal situation for thermal infrared IDSs, which is why snow-covered terrain is so favorable. In addition, because of its low temperature (not exceeding 0°C), the thermal radiance of snow is less than that of other surfaces. Diverse backgrounds—those with a distribution of materials with different thermal properties, such as grass, thatch, and exposed soil—are likely to cause nuisance alarms during periods of alternating sunlight and overcast because the materials will heat and cool differentially as they absorb solar radiation and emit longwave radiation, respectively.

Even surfaces that visually are uniform, such as a grass cover or soil, may be thermally complex. Under strong solar warming, the top portion of grass blades is warmer than the lower portion, which is sheltered from direct solar radiation. If the grass is tall enough to blow in the wind, a surface with a different thermal radiance is momentarily exposed to the IDS. Heat transfer in soil depends significantly on its moisture content. Damp portions of any soil type heat and cool more slowly than does dry soil under the same weather conditions, affecting the temperature distribution at the soil surface.

The thermal contrast of warm objects with a cold snow surface is high. Under conditions of overcast sky and melting snow, however, the thermal contrast is diminished as the air, snow surface, and near-surface objects approach the same temperature. Examples of high threat intrusions using countermeasures against passive infrared detection, and an analysis of environmental conditions required for successful intrusions, are given in Lacombe and Peck (1994) and in a CRREL video, *Thermal Infrared Imaging for Perimeter Security in Winter* (P90019). When melting is not occurring, snow temperatures in the interior of the snow pack can be much warmer than the surface snow. An intruder burrowing through snow can have a relatively low thermal contrast with the warm interior of the snow pack.

Fog presents a thermally uniform background for passive infrared IDSs, so the likelihood of nuisance alarms during fog is low. Although the thermal contrast of an intruder (using no countermeasures against detection) in fog may be relatively high, that contrast is diminished with distance from the intruder due to path attenuation. Consequently, the thermal contrast apparent at a passive infrared IDS may be insufficient for detection. How much detection capability is reduced depends on the type of fog and the intruder–IDS separation. Similar considerations of the loss of thermal contrast over distance apply to rain and falling snow, for which the relevant parameters are the quantity and sizes of the particles (raindrops, snowflakes) in a volume of air.

Near-infrared beambreak IDSs

Near-infrared beambreak IDSs are active systems that alarm when a near-infrared continuous or pulsed beam (between transmitter and receiver units) is interrupted by a sufficient amount for a certain duration. Beambreak sensors located near to the ground to detect a crawling intruder are vulnerable to nuisance alarms when accumulating snow or growing vegetation intrude into the beam.

Transmission of near-infrared and visible radiation through falling snow is similar. The likelihood of nuisance alarms due to falling snow depends on the beam diameter and the amount of beam interruption tolerated. A near-infrared IDS at SOROIDS with a small beam (transmitting and receiving sensors with ~2-cm-diameter lenses) experienced sufficient transmission loss during most snowstorms that nuisance alarms were common. A second near-infrared IDS, having a larger-diameter beam (~9-cm-diameter sensors) and tolerating a 98.5% loss in signal strength between transmitter and receiver, did not alarm during snowfall (unless snow accumulated in front of a sensor lens), but did have nuisance alarms during blowing snow events. Blowing snow is distinct for its generally smaller particle size and for being concentrated near the ground, where pre-existing partial obstructions of the sensor beam would increase the likelihood of nuisance alarms during blowing snow events. Similarly, in summer birds on the ground or blowing objects such as leaves are more likely to cause nuisance alarms if a sensor beam is already partially blocked by rough terrain or vegetation.

Weather-related transmission loss or beam blockage may cause nuisance alarms, depending on the amount of signal reduction tolerated by the IDS receiver before generating an alarm. Transmission loss through fog can be a severe problem and limits the maximum detection zone length for avoidance of fog-related nuisance alarms. Although falling snow in midwinter did not cause nuisance alarms for a 50-m-long detection zone, late-winter storms during which snow and fog occurred concurrently did. Before a nuisance alarm occurs, a potential benefit of environmentally caused transmission loss is that it may reduce the margin of error of an intruder trying to avoid interrupting a beam sufficiently to cause an alarm, and so may increase detection capability. Alternatively, if near-infrared radiation that normally is not incident at the receiver is scattered to the receiver by water drops or snowflakes, this may partially compensate for the beam interruption caused by the intruder.

Buried electromagnetic IDSs

The buried electromagnetic IDS responds to disturbances to an electromagnetic field set up between two active cables, one that transmits (leaky cable) electromagnetic energy and one that receives electromagnetic energy. The electromagnetic field extends above the ground surface, establishing a volumetric detection zone. Nonintruder disturbances to the field are caused by the motion of water or metallic objects. The strength of the electromagnetic field depends on the electrical properties of the soil (dielectric constant and electrical conductivity), which change significantly with the soil's moisture content. Electromagnetic IDSs in frozen or dry soils, which have a small electrical conductivity because of their small unfrozen water content, have improved detection capability over those in wetter soil. This IDS has a larger margin of reliability (stronger response to an intruder) when operating in frozen or dry soil. The combination of thawing of the soil, which releases water when interstitial ice melts, and the influx of water during snow melt causes thaw periods to be troublesome to this class of IDSs. This is particularly true if the thaw is interrupted by freezing episodes so that the appropriate sensitivity setting of the IDS corresponds alternately to high (unfrozen) or low (frozen) moisture content.

Fence-mounted IDSs

This is a broad category of IDSs that are designed to alarm when the security fence they are attached to is being cut or climbed. They detect a fence disturbance mechanically by means of lost contact when a mass is bounced off its support; electrically by means of a friction-generated charge transfer between the inner and outer portions of a cable attached to the fence (triboelectric), by means of a charge transfer generated in a dielectric within a sensor cable subjected to mechanical stress (piezoelectric), or by means of relative motion between a conductor and a charge-storing dielectric (electret); and optically by changes in the pattern of standing waves of light in optical-fiber cables attached to the fence.

The characteristics of the fence motion depend on how well the fence posts are anchored in the soil and on the stiffness of the fence panels. Posts in frozen or dry soil move less under loading than do posts in deeply saturated or thawing, frost-heaved soil. The stiffness of the fence panels is temperature-dependent because of thermal contraction and expansion and time-dependent because of relaxation, particularly under the effects of thermal cycling.

The detection capability of cable systems depends on the time of installation, which is attributed to temperature-related changes in the degree of coupling between IDS cables and the fence. With a taut sensor cable in close contact with the fence, there is greater consistency between the motion of the fence and that of the sensor cable. A loose cable is more isolated from the fence motion (degraded detection capability) and freer to move on its own under wind loading or precipitation impact (higher nuisance alarm rate). Cable systems that have been installed on the chain-link fences at the SOROIDS site in winter generally have poor detection capability the next summer, whereas IDSs installed during the summer to fit snugly to the fence remain reliable through the following winter.

More than with any other class of IDSs, the quality of the fence strongly determines the likelihood of nuisance alarms. The vibratory response of a fence to dynamic loading, such as climbing or cutting panels, depends on the stiffness of the fence. All fences experience some variation in stiffness caused by thermal contraction and expansion, perhaps accompanied by abrupt relative displacement of structural components of the fence that are in direct contact. The dominant environmental cause of fence motion is wind loading. Other causes are the impact of precipitation (falling snow, hail, or rain) or ice or adhered snow that disturbs the fence as it is shed. An intact ice coating on the fence, however, may change the vibration characteristics of the fence panels such that intruder disturbances go undetected.

Normal stiffness and transverse stiffness refer to the displacement of a fence panel (that portion of the fence bounded by adjacent fence posts) under loading applied perpendicular to and parallel to, respectively, the plane of the fence at the center of the panel (Walsh and Peck 1990). With a chain-link fence, for example, the stiffness of a panel depends on the tension of the fabric (initial tension at installation and time-dependent relaxation of the fabric) and the number and location of stiffening elements such as cables or pipes, as well as the temperature-dependent contraction and expansion of the metal. The chain-link fence at SOROIDS (2.4-m-high, 3-m-wide panels) was installed in 1987, originally with 5 horizontal stiffening cables spaced at ~0.6-m intervals from the base to the top of the fence fabric. In September 1990, the normal stiffness of 4 panels, under a load of 132-N (30-lbf) push or pull at the center of the panel, ranged from 50 to 70 mm (14°C air temperature). In November, after the fabric had been removed from the fence posts, retensioned, and reattached to the posts, with a metal pipe replacing the lowest stiffening cable, the normal stiffness of the same panels ranged from 10 to 50 mm (12°C air temperature), a reduction in displacement of 18, 20, 40, or 48 mm (Peck 1993a). By June 1991, the normal stiffness of a heavily braced corner panel had changed from 10 to 29 mm, and that of an interior panel from 50 to 65 mm. The larger displacements in June compared with November are attributed to the fence being warmer (air temperature 20°C in June) and so thermally expanded and to loosening caused by long-term relaxation and thermal cycling over the 7 months since the fence was retensioned (Peck 1993a).

During the winter of 1990–91, it was observed that the occurrence of wind-related nuisance alarms by fence-mounted IDSs on the newly retensioned chain-link fence at SOROIDS depended on air temperature, which was taken as an index of relative fence stiffness. Under similar high wind conditions, the colder the air, the greater the likelihood of nuisance alarms was (Peck 1992b). The magnitude of a temperature-related change in fence stiffness may lessen with time, however, as the chain-link fence loosens under the effects of thermal cycling and overall relaxation (Peck 1993a). This means that IDSs mounted to a newly installed or retensioned chain-link fence may be more variable in detection capability and nuisance alarm rate until the fence fabric has loosened sufficiently that temperature-related changes in overall fence stiffness are less pronounced. The possibility of increased consistency in IDS performance with time contradicts the intuitive expectation that a stiff fence, especially a new, high-quality fence, will present fewer difficulties to the reliable operation of fence-mounted IDSs. As the fence stiffness varies, the differences in fence motion due to wind loading or precipitation impact as well as cutting or climbing activity take the form of different vibration spectra and different displacement maxima. If the fence fabric or stiffeners loosen to the extent that they move freely and strike posts or other fence components during windy periods, however, then the greater likelihood of wind-induced nuisance alarms offsets any benefits of reduced temperature-dependence in fence stiffness.

The actual wind force on a fence is the product of the velocity pressure, a gust response factor (which accounts for the additional loading effects due to wind turbulence), a force coefficient (which depends on the ratio of solid area to gross area of the fence), and the area of all exposed fence members projected on a plane normal to the wind direction (ASCE 1990). The velocity pressure, which is proportional to the square of the wind speed, varies directly with a parameter called the exposure coefficient, which depends upon the terrain and obstructions in the vicinity of the fence. If the terrain obstructions vary seasonally, as for example due to changes in vegetation height or fullness, then this imparts an environmental difference to the wind loading of the fence.

Taut-wire IDSs

Taut-wire IDSs monitor the displacement of a strand of wire under tension. This IDS is installed as a physical barrier with only a few cm of vertical clearance between a second wire or the top of a wall or fence. An intruder cannot pass his body through the gap without deflecting the wire.

Variation in wire tension due to seasonal temperature changes is the main environmental effect on these IDSs. Icing and snow accumulation on the wires are also potential problems: the severity depends on the collection efficiency of the wire (barbed wire has a higher collection area than smooth wire) and on the weather conditions.

Two taut-wire systems at SOROIDS are used in conjunction with a vertical array of six strands of barbed wire on top of a chain-link fence. Between 1 November 1990 and 30 April 1991, the tension of the barbed wires first decreased relative to what it was upon installation (25 October), which is attributed to the initial relaxation of the wires and their hardware; then increased to maximum values in January, the coldest time that tension measurements were made; and finally progressively decreased through March under the combined effects of thermal expansion and relaxation due to thermal cycling. The tension of the wires in June 1991, on a day that was 10°C warmer than 25 October, was 20–25% lower than the tension immediately following installation. That autumn, the onset of low temperatures led to an increase in wire tension to levels comparable or slightly lower than those of the previous November.

Neither taut-wire IDS had nuisance alarms over the winter (November to May). The IDSs showed either no overall seasonal variability in detection capability or an increase in their sensitivity, such that a smaller displacement of a taut wire was necessary for an alarm to occur, despite the changes in the tension of the wires. What varied was the ease with which intruders could deflect a wire and then keep it displaced while slipping through the enlarged gap between the two adjacent wires.

Buried optical-fiber IDSs (ground-motion IDSs)

Buried optical-fiber IDSs detect ground motion optically by changes in the pattern of standing waves of light in optical fiber cables buried at shallow (~5 to 9 cm) depth. A cable is generally laid in a serpentine pattern to give dense coverage. It may be attached to plastic construction webbing to give greater coupling to the burial medium, which is typically soil or gravel. If gravel is used, sand may be placed directly on the cable/

webbing to protect the cable from abrasion by the gravel.

The essential characteristic of the burial medium is that it be displaced sufficiently under the intruder's weight that the induced ground motion meets the alarm criteria of the IDS. The criteria generally involve the magnitude and frequency spectrum, and perhaps the repetition rate, of changes in the light pattern as the cable is displaced within the burial medium. In the winter, gravel is superior to soil as a covering for the cables. The soil freezes from the surface down, becoming stiffer and displacing differently under an intruder. The gravel remains loose unless, because of poor drainage, it eventually becomes locked in ice. For this reason, an optical-fiber IDS in gravel typically maintains a more acceptable detection capability longer into the winter than one buried in soil. If a snow cover is present, its depth, the number of rigid freeze–thaw layers, and density determine whether the motion induced by an intruder walking in or on the snow in turn causes ground motion sufficient to meet the alarm criteria of the IDS. An IDS buried in soil has a high detection capability when the ground is wet following the spring thaw and snow melt. Its detection capability will diminish as the soil drains and dries and will be even lower if the soil becomes hardpacked.

Wind-induced motion of surface objects, whose motion couples into the ground, is the primary cause of weather-related nuisance alarms.

Video-motion detection systems

Video-motion IDSs analyze a camera scene for changes indicative of an intruder's presence. The system typically digitizes a camera scene and converts its elements to a gray-scale representation. A subsequent scene is similarly processed, and the gray-scale representations are compared. An alarm is generated if the magnitude, extent, and rate of change of brightness meet criteria based on characteristics of intruder motion.

Nuisance alarms are caused by natural changes in scene brightness, such as moving cloud shadows, the appearance and disappearance of shadows due to intermittent cloud cover, and wind-induced motion of sunlit objects.

Detection capability is diminished by precipitation and fog (decreased visibility, loss of visual contrast) and by low visual contrast (diffuse illumination under cloud cover). High levels of reflected solar radiation may saturate the camera detector, rendering the video-motion IDS ineffective.

Radiation attenuation through obscurants (fog, rain, falling or blowing snow)

Within the main text and this appendix, certain weather effects are expressed as the extinction, transmission loss, or attenuation of radiation in the presence of fog, rain, or airborne snow. These terms are used interchangeably to represent the diminishing of radiation intensity with distance from its source because of interaction between the radiation and the medium it is propagating through. (This distinguishes extinction from a reduction in intensity caused by geometrical spreading of the radiation.) The source may be part of an IDS, such as a microwave radar or a near-infrared transmitter; the radiation may originate from an intruder, as with thermal radiance; or it may be the visible light (natural or artificial illumination) that constitutes the camera scene used with video-motion IDSs. The reduction in intensity of radiation is proportional to the distance over which it is observed, so that the larger the separation between a (microwave, near-infrared) IDS's transmitter and receiver units, the greater the decrease in the proportion of radiation that transmits through the obscurant to be incident at the receiver. For passive infrared IDSs, a transmission loss between intruder and IDS reduces the intruder's thermal contrast apparent at the IDS. Surveillance cameras and video-motion IDSs are less effective when obscurants are present because the extinction of visible radiation reduces the intruder's visual contrast in the camera scene.

Extinction of radiation occurs through two processes, absorption and scattering. Absorption of radiation by aerosols or airborne particles, with kinetic energy being converted to heat, permanently removes the radiation from an IDS's detection zone. Scattering refers to radiation being incident on a particle from one direction but leaving that particle in another direction. Radiation that is initially scattered out of the line-of-sight of a microwave or near-infrared IDS may be returned to that path line as the net result of subsequent scattering events. Similarly, visible and thermal radiation that is outside the original path line may be scattered into an IDS's line of sight; this is one of the causes of reduction in visual and thermal contrast encountered with videomotion and passive infrared IDSs.

Under clear sky conditions, absorption and scattering are unimportant to IDS operations by design. For a particular class of IDS (microwave, near-infrared, thermal infrared), the wavelength(s) at which the IDS operates are selected in part to take advantage of atmospheric windows characterized by low loss of radiation at those wavelengths. When airborne water drops (fog, rain) or ice particles (falling or blowing snow) are present to potentially scatter and/or absorb the radiation, the radiation loss attributable to these obscurants is expressed through such parameters as a coefficient of extinction or attenuation, or as a propagation loss or a transmission loss. By Beer's Law,

$$dE/E = -k_{ext}dx, \quad (1)$$

where E is the incident irradiance (radiant flux per unit cross section) of a parallel beam of radiation, dE is the irradiance removed from the parallel beam in the distance dx the beam travels in the medium, and k_{ext} is the extinction coefficient of the medium, which may be a function of wavelength of the radiation. After propagating a distance x through fog, rain, or snow, the irradiance of the beam is

$$E_x = E_o\, e^{-(k_{ext}x)}, \quad (2)$$

where k_{ext} is the extinction coefficient that characterizes the obscurant and E_o is the irradiance at the source. The extinction coefficient is generally expressed in km^{-1}; when it is given in dB km^{-1}, it must first be converted to km^{-1} [k_{ext} (dB km^{-1}) ≈ 4.343 k_{ext} (km^{-1})] before substituting in eq 2. The proportion of radiation remaining in the original beam is E_x/E_o. This ratio is the transmittance of the radiation through a particular obscurant. The transmittance loss or propagation loss is $1 - E_x/E_o$. All these terms quantify the change in radiation intensity under the net effects of absorption and scattering; for a given wavelength of radiation, and type and size of obscurant particle, one process may dominate the other such that effectively the extinction coefficient is a scattering coefficient (k_s) or an absorption coefficient (k_a) (the processes are linear, so $k_{ext} = k_s + k_a$).

Calculations of extinction of electromagnetic waves by a dielectric sphere are based on Mie theory. A dimensionless size parameter, $X = 2\pi (r/\lambda)$ conveys the size of the obscurant particle (radius r) relative to the wavelength (λ) of the incident radiation. For $X << 1$, the Mie equations reduce to simpler Rayleigh approximations. For $X > \sim 30$, extinction is well represented by calculations of reflection, refraction, and diffraction of the radiation on or within the particle. Otherwise, the complete Mie formulation must be used.

The extinction of radiation through scattering and absorption by a particle depends on the particle's size relative to the wavelength of the incident radiation. The particle has a scattering cross section, Q_s (distinct from its physical cross section), that determines the total energy scattered in all directions by the particle. Conceptually, the amount of scattered radiation is equal to that portion of the incident beam that would be interrupted by a particle having an actual cross-sectional area equal to Q_s. There is an angular dependence to the scattered radiation such that its intensity in any direction can be calculated. In addition to the wavelength of the incident radiation and the radius of the particle, Q_s depends on the complex index of refraction of the material of the particle. Accordingly, when calculating Q_s of fog, rain, and snow, the dielectric properties of water and ice must first be known for wavelengths corresponding to the radiation of interest. A similar quantity, Q_a, can be defined to express the absorption cross section of the particle. The sum of a particle's scattering cross section and its absorption cross section is the particle's extinction cross section, Q_{ext}.

The ratio of extinction cross section to physical cross section is a particle's extinction efficiency, ξ_{ext}. Scattering and absorption efficiency factors are similarly defined. Because the physical cross section of a sphere of radius r is easily calculated (πr^2), water drops and snowflakes are often represented as spheres of equivalent mass. [Several papers present corrections to the scattering calculations to take into account that, while Mie theory was developed for spherical scatterers, raindrops and snowflakes are not spheres (e.g., Bohren and Koh 1985, Seagraves 1986).] The extinction coefficient for radiation of a given wave length through a particular obscurant is directly proportional to the corresponding extinction efficiency of the fog, rain, or snow particles. When particles of a range of sizes are present concurrently, the net scattering efficiency of the medium is determined jointly by the particle size distribution (number of particles of each size) and the extinction efficiency associated with each particle size. Extinction efficiency is often plotted as a function of X. As X becomes very large, ξ_{ext} approaches 2.

Eq 3 expresses the extinction coefficient, k_{ext}, of an ensemble of particles (water droplets or snowflakes) in terms of the extinction cross sections of the particles:

$$
\begin{aligned}
k_{ext} &= \int_0^\infty Q_{ext}(r) N(r)\, dr \\
&= \int_0^\infty \left[Q_s(r) + Q_a(r)\right] N(r)\, dr
\end{aligned}
\tag{3}
$$

where $N(r)$ is the number of particles per unit volume with radii between r and $r + dr$.

The larger the number of particles, the greater the likelihood of a scattering or absorption event. This is why, in the text, the effect of fog, rain, or snow on IDS reliability is cited in terms of both the particle size and the density of fog or rate of rainfall or snowfall. From the rate information, the number of rain or snow particles in a given volume of air in a given time interval may be calculated. Raindrop size and precipitation rate are related through empirical determinations such as the Laws–Parsons and Marshall–Palmer distributions, which present, as a function of rain rate, the percentage distribution of drops of various sizes in a volume of air.

Field measurements of transmission of radiation over a known distance through obscurants sometimes indicate that there is a spectral (wavelength) dependence to the extinction of the radiation when, by extinction theory, the transmission loss should be independent of wavelength because the particles are large relative to wavelength (recall that the extinction efficiency is 2 for large values of X). An example is the measured extinction of near-infrared and thermal infrared radiation in falling snow, which often is greater than that of visible radiation (e.g., Curcio et al. 1981). This is explained by the occurrence of forward scattering, which refers to radiation leaving the snow particle at an angle that differs only slightly from its original path line. Since the instruments used in transmission experiments have receivers with finite apertures, some of the forward-scattered radiation is collected by the receiver. Because visible light interacting with snowflakes scatters more strongly in the forward direction than does near-infrared or thermal infrared radiation, more of the scattered visible radiation is received, and so the apparent extinction coefficient (or transmission loss) of visible radiation is less than that at the other wavelengths. With IDSs (near-infrared beambreak, passive infrared) whose detection phenomenology is based on the magnitude of changes in the radiation received, the apparent extinction coefficients may be a more realistic indication of the interaction of the radiation with obscurants than are coefficients that have been corrected to remove the effect of forward scattering of radiation. The forward-scattered radiation will differ in phase from unscattered radiation, however, which may be significant to IDSs (microwave) whose alarm criteria also depend on certain phase relationships between received radiation.

In summary, the relative impact of a given type of obscurant (fog, rain, or snow), on different IDSs (video-motion detection, near-infrared beambreak, thermal infrared, microwave) depends on the number of possible extinction events, which in turn depends on the size of the obscurant particles relative to the wavelength of the radiation and on the number of particles of a given size. Because both the size and the quantity of particles in a volume of air can vary during a fog event or a rain- or snowstorm, transmission loss need not be constant. In some instances, it may be the variability in extinction, rather than the absolute extinction, that results in nuisance alarms. This can explain nuisance alarms that occur sporadically during a fog event or a rain- or snowstorm.

APPENDIX B: INDEXING FROM AN IDS PERSPECTIVE

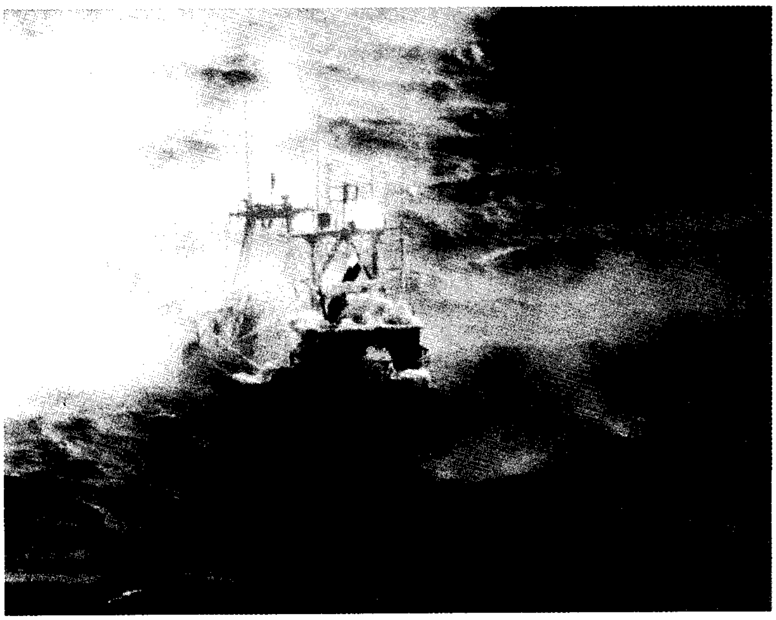
The purpose of this index is to allow readers to locate special IDS topics of interest easily. The index is organized to locate only IDS-related information within the main text. This allows all discussions of passive infrared IDSs, for example, to be located to determine the impact of weather upon the reliability of passive infrared IDSs.

APPENDIX C: INDEXING FROM A WEATHER PERSPECTIVE

This index allows readers to locate special weather topics of interest. It is organized to locate only weather phenomena within the main text. This allows all discussions of blowing snow, for example, to be located to determine the impact of blowing snow on IDS reliability.

APPENDIX D: COMMON FOGS AND CLOUDS

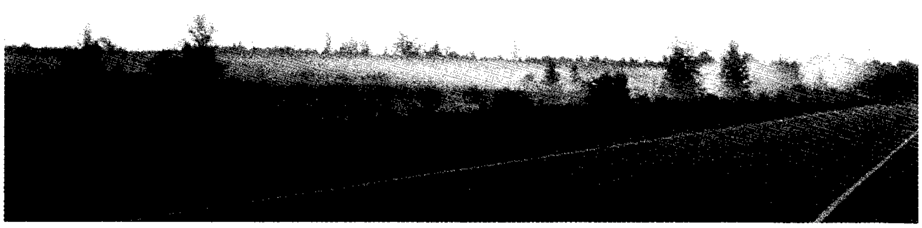
The following photographs illustrate common varieties of fogs (Fig. D1) and clouds (Fig. D2).



D1a. Steam fog rising from the Bering Sea.

D1b. Upslope fog in the Great Smoky Mountains.

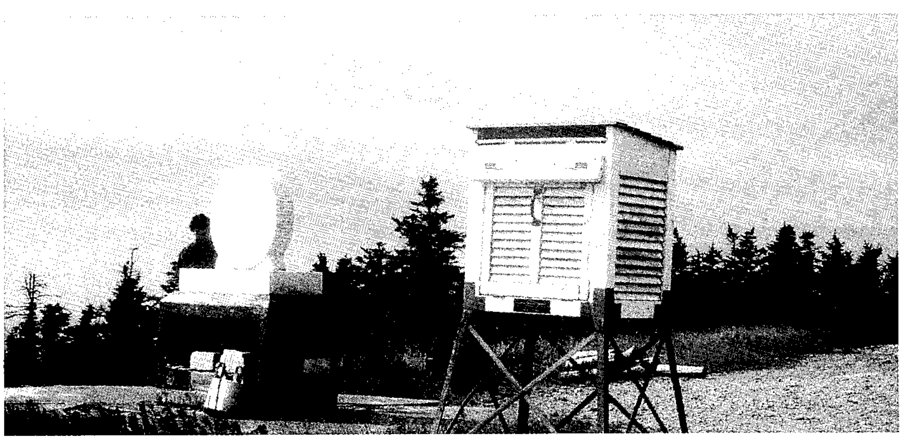
D1c. Valley fog near Waterbury, Vermont.



D1d. Radiation fog, Upper Peninsula, Michigan. The clouds are altostratus and cirrostratus.

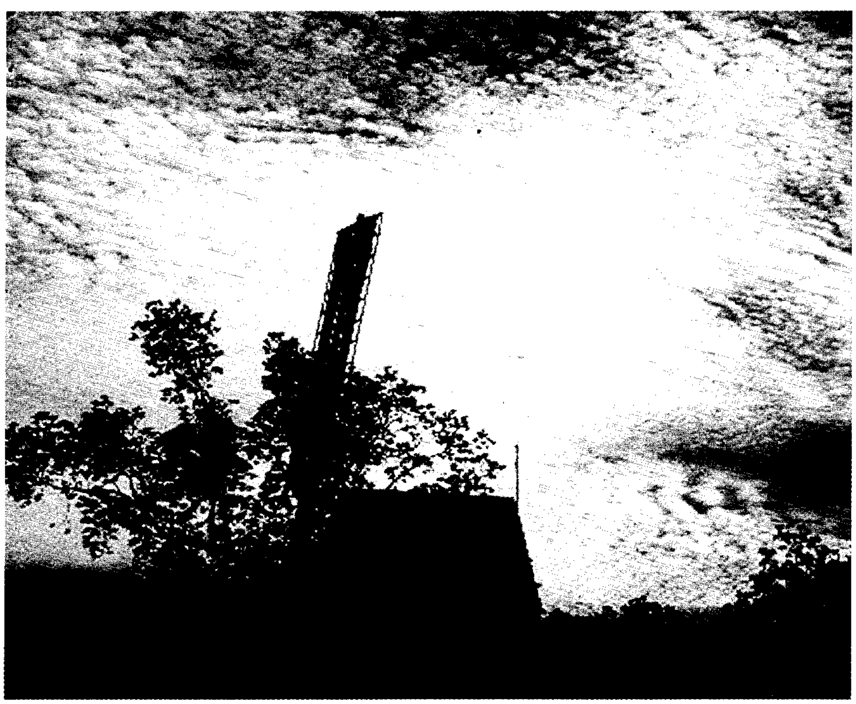
D2a. Cumulus clouds (below) and cirrostratus clouds (above), Melbourne, Florida.

D2b. Cirrus clouds, Jungfraujoch, Switzerland.



D2c. Stratus clouds, Mt. Mansfield, Vermont.

D2d. Stratus clouds and valley fog below, and altostratus clouds above, Mt. Mansfield, Vermont.



D2e. Altocumulus (below) and cirrocumulus (above) clouds, Williamsburg, Virginia.

APPENDIX E: RELATIONSHIP OF WEATHER FORECASTS TO WEATHER PATTERNS

This report addresses weather patterns, especially as observed on weather maps, and their impacts upon local weather conditions that may affect IDS reliability. However, security designers and security personnel must overcome several difficulties in attempting to relate discussions in this report to weather forecasts.

1) Weather patterns discussed in this report are often ideal, simplified, "textbook" weather situations. Weather is a very dynamic phenomenon that rarely is simple and is often chaotic. Simplification helps by providing an understanding of the patterns that often occur. However, the weather on any one day or in any season can, and does, often deviate significantly from these idealized patterns.

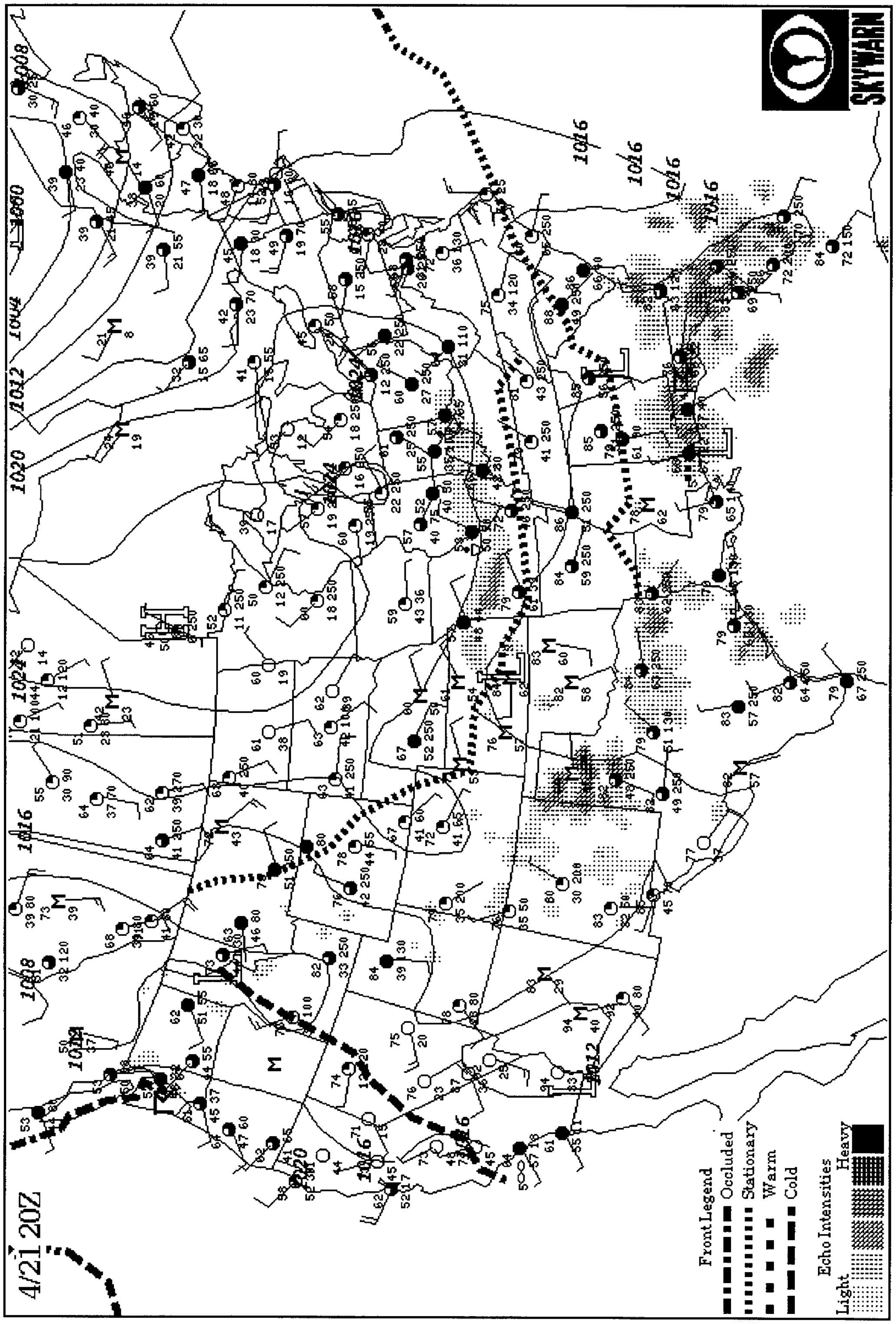
2) Weather patterns evolve not only temporally, but also spatially. Storms increase or decrease in vigor depending upon their location—depending upon whether they are in a region that supports cyclone strengthening, such as off Cape Hatteras, or in an area that often reduces the strength of cyclones, such as the Rocky Mountains. Cold fronts, for example, are often more vigorous in the Great Plains where warm, moist air from the Gulf of Mexico can clash directly with cold, dry air originating in the Canadian prairies. This is where severe thunderstorms and tornadoes are common summer phenomena. Mountains and high relief slow the progress of air masses and often lessen the consequences when they do produce fronts. The geographic influence, therefore, which also varies with season and time of day, also determines the impacts of these weather patterns at different locations.

3) Security personnel responsible for IDSs often do not have access to daily weather maps and are usually not trained to read them. They generally have little more than printed weather forecasts to assist them in determining how conditions will change. However, the conditions indicated in the forecasts as changing are not usually the conditions that will directly affect IDS reliability. It is weather associated with the forecasted changes, but perhaps not noted by the forecasts, that will affect IDS reliability.

The purpose of this appendix is to relate weather forecasts at selected cities to mapped weather patterns at the time the forecasts were issued. Forecasts only describe the weather at one location, but they do provide a temporal perspective by providing predictions of the expected changes in weather over the next 24 to 72 hours. Since forecasts are created by weather services from weather maps, they do provide insight to general weather patterns that can be related to this report.

The purpose of this appendix is not to provide comprehensive insight into every possible weather situation. It is to demonstrate how forecasts relate to national weather patterns and to demonstrate that forecasts for a location can provide some guidance to the map patterns without even seeing a map.

Five weather maps are shown and eleven weather situations are discussed below with forecasts for selected cities. The forecasts are interpreted and related to the maps. See the key at the end of this appendix to read the weather symbols on the maps.


4/21 20Z
SKYWARN
Front Legend
Occluded
Stationary
Warm
Cold
Echo Intensities
Light
Heavy

Weather situation: High pressure **21 April 1994**

Example: Green Bay, Wisconsin

High pressure is located northwest of Green Bay, and light winds are rotating clockwise around the high. The resulting winds at Green Bay are from the northwest. Skies are clearing as the high approaches, as indicated by the sky coverage circles at each station. Regional wind speeds are about 10 knots or less. The forecast is for clear and cold tonight as a result of cloudless skies, which allows strong loss of longwave radiation to space. These calm, clear conditions may produce light hoarfrost during the evening. Friday warms to the 60s (°F) also because the sky is clear, allowing strong solar radiation. If regional winds remain calm, winds should increase after sunrise and calm again after sunset, because of convective coupling of surface air with winds aloft.

Regional winds eventually become southerly as the high passes to the east. Clockwise flow around the high causes southerly flows to its west, advecting warmer and more humid air. The southerly winds developing on Friday, only partly sunny skies on Saturday, and chance of showers on Sunday are all evidence of this. Note that Friday night temperatures are warmer than Thursday night temperatures because of the southerly flow and the associated partial cloudiness, which inhibits loss of longwave radiation to space.

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*************************************************************************
ADAMS-BROWN-CALUMET-FOND DU LAC-GREEN LAKE-JUNEAU-MARATHON-MARQUETTE-
OUTAGAMIE-PORTAGE-SHAWANO-WAUPACA-WAUSHARA-WINNEBAGO-WOOD-INCLUDING THE
CITIES OF...WAUSAU...GREEN BAY
345 PM CDT THU APR 21 1994
TONIGHT CLEAR AND COLD. LOWS IN THE UPPER 20S. LIGHT WINDS.
FRIDAY MOSTLY SUNNY. HIGHS IN THE LOWER 60S. LIGHT WINDS BECOMING
SOUTHERLY AROUND 10 MPH.
FRIDAY NIGHT FAIR. LOWS AROUND 40.
SATURDAY PARTLY SUNNY AND WARM. HIGHS IN THE UPPER 60S OR LOWER 70S.
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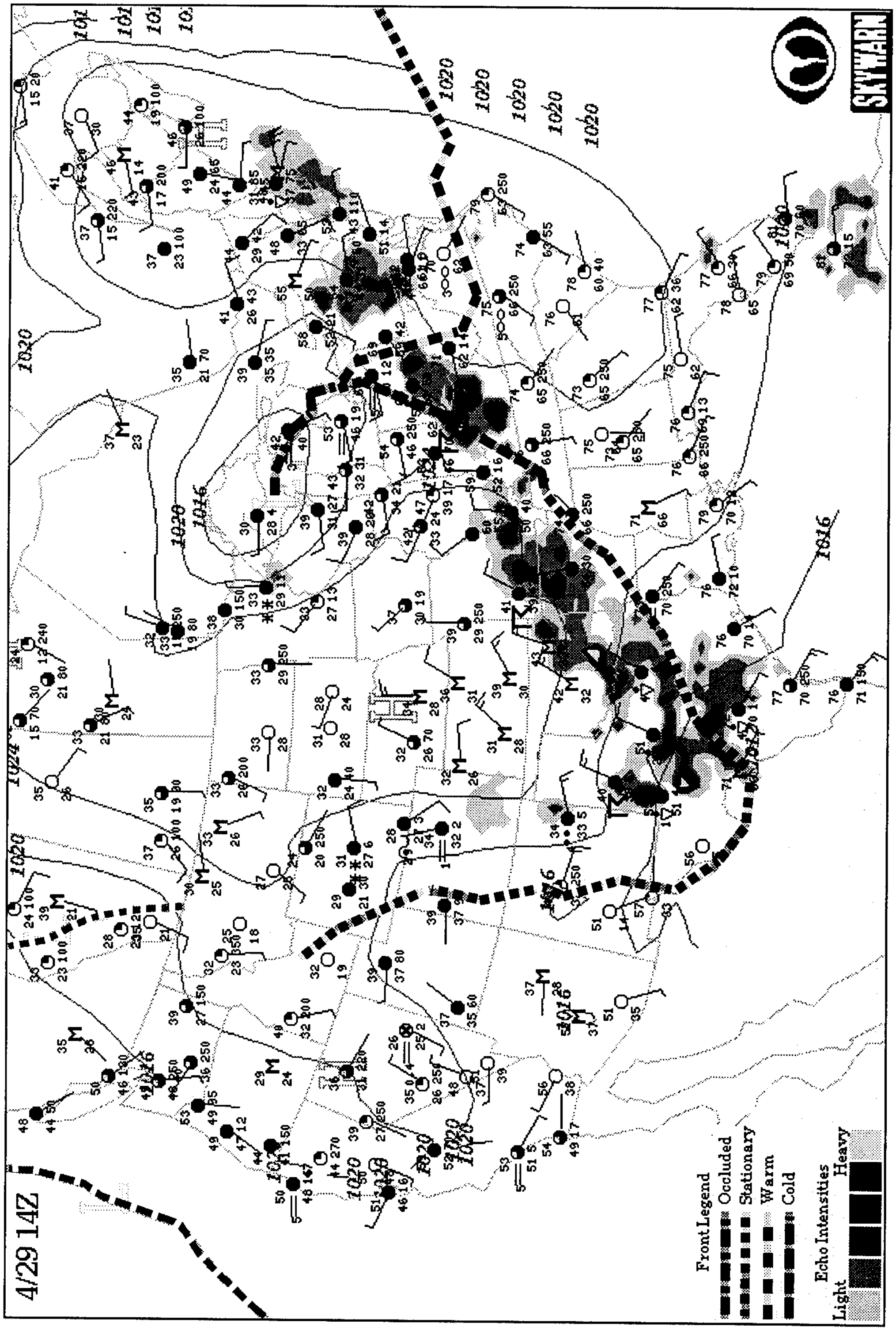
4/29 14Z
SKYWARN
Front Legend
Occluded
Stationary
Warm
Cold
Echo Intensities
Light
Heavy

Example: Indianapolis, Indiana; Cincinnati, Ohio; Nashville, Tennessee

The weather map represents 1000 local daylight time in these cities. The forecasts below were made about 6 hours after the map was constructed. The vigorous cold front on the map has crossed Cincinnati, Indianapolis, and Nashville. Hail, lightning, and damage to trees were reported in Cincinnati and Nashville. Occasional rain and thunderstorm activity will continue for several days because most of the frontal system is stationary south of the cold front. The stationary front in the south indicates that the cold air–warm air boundary is not moving, and storms will continue to be generated along the front, affecting the area for several more days. Note that wind directions are mostly southerly in the warm sector of the storm, easterly ahead of the warm front, and westerly and northwesterly after the cold front passes and as the high pressure cell in Nebraska approaches. Temperature contrasts in the storm are strong, a cause of its intensity. Temperatures are in the 70s (°F) in the warm sector, but only in the 30s (°F) behind the cold front.

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BARTHOLOMEW-BROWN-CLAY-GREENE-HANCOCK-HENDRICKS-JOHNSON-
MARION-MONROE-MORGAN-OWEN-PUTNAM-SHELBY-SULLIVAN-VIGO-
INCLUDING THE CITIES OF...INDIANAPOLIS...TERRE HAUTE... SHELBYVILLE...
BLOOMINGTON...COLUMBUS
330 PM EST FRI APR 29 1994
THIS EVENING... RAIN ENDING. NORTHWEST WIND AROUND 10 MPH.
TONIGHT... CLOUDY WITH A 40 PERCENT CHANCE OF RAIN LATE TONIGHT. LOW IN THE
UPPER 40S. NORTHWEST WIND AROUND 10 MPH.
SATURDAY... PERIODS OF RAIN AND SCATTERED THUNDERSTORMS. HIGH NEAR 60. EAST
WIND AROUND 10 MPH. CHANCE OF RAIN 100 PERCENT.
SATURDAY NIGHT... RAIN LIKELY AND TURNING COOLER. LOW AROUND 40.
LIKELIHOOD OF RAIN 70 PERCENT.
SUNDAY... A 30 PERCENT CHANCE OF LIGHT RAIN IN THE MORNING THEN
DECREASING CLOUDINESS. HIGH IN THE MIDDLE 50S.
*************************************************************
CINCINNATI AND VICINITY FORECAST
NATIONAL WEATHER SERVICE CINCINNATI OH
350 PM EDT FRI APR 29 1994
TONIGHT... SHOWERS LIKELY. THUNDERSTORMS ALSO POSSIBLE. LOW 50 TO 55. WEST
WINDS ABOUT 10 MPH BECOMING NORTHEAST. CHANCE OF RAIN 70 PERCENT.
SATURDAY... OCCASIONAL RAIN AND A CHANCE OF THUNDERSTORMS. HIGH IN THE MID
60S. EAST WINDS 10 TO 15 MPH. CHANCE OF RAIN 100 PERCENT.
SATURDAY NIGHT... SHOWERS AND THUNDERSTORMS. LOW 50 TO 55. CHANCE OF RAIN
100 PERCENT.
SUNDAY... CLOUDY. CHANCE OF SHOWERS MAINLY IN THE MORNING. HIGH AROUND 60.
CHANCE OF RAIN 50 PERCENT.
**********************************************************
Special weather statement in effect for CVG
**********************************************************
SPECIAL WEATHER STATEMENT... CORRECTED
NATIONAL WEATHER SERVICE CINCINNATI OH
406 PM EDT (306 PM EST) FRI APR 29 1994
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STRONG THUNDERSTORM HAS DEVELOPED ACROSS SOUTHERN OWEN COUNTY KENTUCKY. AT 405 RADAR INDICATED A STRONG THUNDERSTORM 5 MILES SOUTHEAST OF OWENTON KENTUCKY. THE STORM WAS MOVING EAST AT 40 MILES AN HOUR AND WILL MOVE INTO SOUTHERN GRANT COUNTY NEAR CORINTH BY 420 PM EDT. BRIEF GUSTY WINDS TO 40 MILES AN HOUR... SMALL HAIL AND VERY HEAVY RAIN CAN BE EXPECTED WITH THIS THUNDERSTORM.

****

Storm damage report for CVG

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LOCAL STORM REPORT
NATIONAL WEATHER SERVICE CINCINNATI OH
537 PM EDT FRI APR 29 1994
PRELIMINARY SEVERE WEATHER REPORTS FROM 4/29/94

ALL TIMES	TYPE OF DAMAGE	LOCATION	SOURCES ARE LOCAL
04/29 0505 EST	GOLF-BALL-SIZED HAIL	DEARBORN COUNTY IN	REPORT FROM TV STATION
04/29 0635 EDT	NICKLE SIZED HAIL	BOONE COUNTY KY	COUNTY POLICE
04/29 0708 EDT	NICKLE SIZED HAIL IN DRY RIDGE	GRANT COUNTY KY	COUNTY POLICE
04/29 0645 EDT	GOLF-BALL-SIZED HAIL NEAR WARSAW	GALLATIN COUNTY KY	COUNTY POLICE
04/29 0645 EDT	DIME-SIZED HAIL IN ANDERSON TOWNSHIP...	HAMILTON CO OH SOUTHEAST PART OF THE COUNTY ALONG BEECHMONT AVENUE.	SKYWARN

****

SPECIAL WEATHER STATEMENT AND RADAR SUMMARY
NATIONAL WEATHER SERVICE NASHVILLE TN
437 PM CDT FRI APR 29 1994
SEVERE THUNDERSTORM WATCH UNTIL 9 PM...
AT 430 PM CDT... THUNDERSTORMS WERE SCATTERED OVER MIDDLE TENNESSEE... THE CUMBERLAND PLATEAU... AND CENTRAL KENTUCKY.
MOVEMENT WAS NORTHEAST AT 35 MILES AN HOUR. THE HEAVIEST RAIN IN MIDDLE TENNESSEE WAS JUST SOUTH OF LEWISBURG. SOME OF THESE STORMS MAY CONTAIN SMALL HAIL... GUSTY WINDS... DEADLY LIGHTNING AND VERY HEAVY DOWNPOURS OF RAIN. STAY TUNED TO NOAA WEATHER RADIO FOR UPDATES OR WARNINGS.

****

Storm damage report for BNA

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STORM REPORT
NATIONAL WEATHER SERVICE NASHVILLE TN
500 PM CDT FRI APR 29 1994

STORM REPORT FOR FRIDAY APRIL 29 1994

ALL TIMES IN CDT...
140 PM... HICKMAN CO... TREES BLOWN DOWN IN SOUTHERN PART OF COUNTY FROM CENTERVILLE TO INTERSTATE 40. REPORTED BY CENTERVILLE POLICE DEPT.
150 PM... ROBERTSON CO... METEOROLOGIST IN CHARGE REPORTED TREES BLOWN DOWN IN GREENBRIER AREA.

240 PM... DAVIDSON CO... GOLF BALL SIZE HAIL AT SHONEYS RESTAURANT IN GOODLETTSVILLE.
245 PM... DAVIDSON CO... EMERGENCY MANAGEMENT REPORTED NUMEROUS TREES BLOWN DOWN ACROSS THE COUNTY.
****

Weather situation: Weak stationary front **29 April 1994**

Example: Denver, Colorado

The stationary front in the southern Rocky Mountains is giving weak warm-frontal-type weather to the Denver area. A stationary front is a cold front or warm front that has ceased moving across the landscape. This stationary front was probably a warm front before it halted. The front may be tied to a weak low pressure cell located in Alberta, Canada. However, moisture is being supplied from the Gulf of Mexico where warm, moist air is overrunning the stationary front in Texas.

Skies are overcast in the Denver area. Though winds are weak and confused, most winds (except Denver) are generally easterly (also associated with warm fronts). Rain is falling in eastern New Mexico, with fog in Denver, freezing rain in Cheyenne, Wyoming, and snow falling in central Wyoming. The forecast indicates that some snow has fallen in the Denver area, but the weakening system has caused the forecast to be revised, and snow advisories have been cancelled. The forecast indicates that thunderstorms may occur as the front slowly moves into the high plains, and that drier air will follow as the high in Nevada moves in.
****

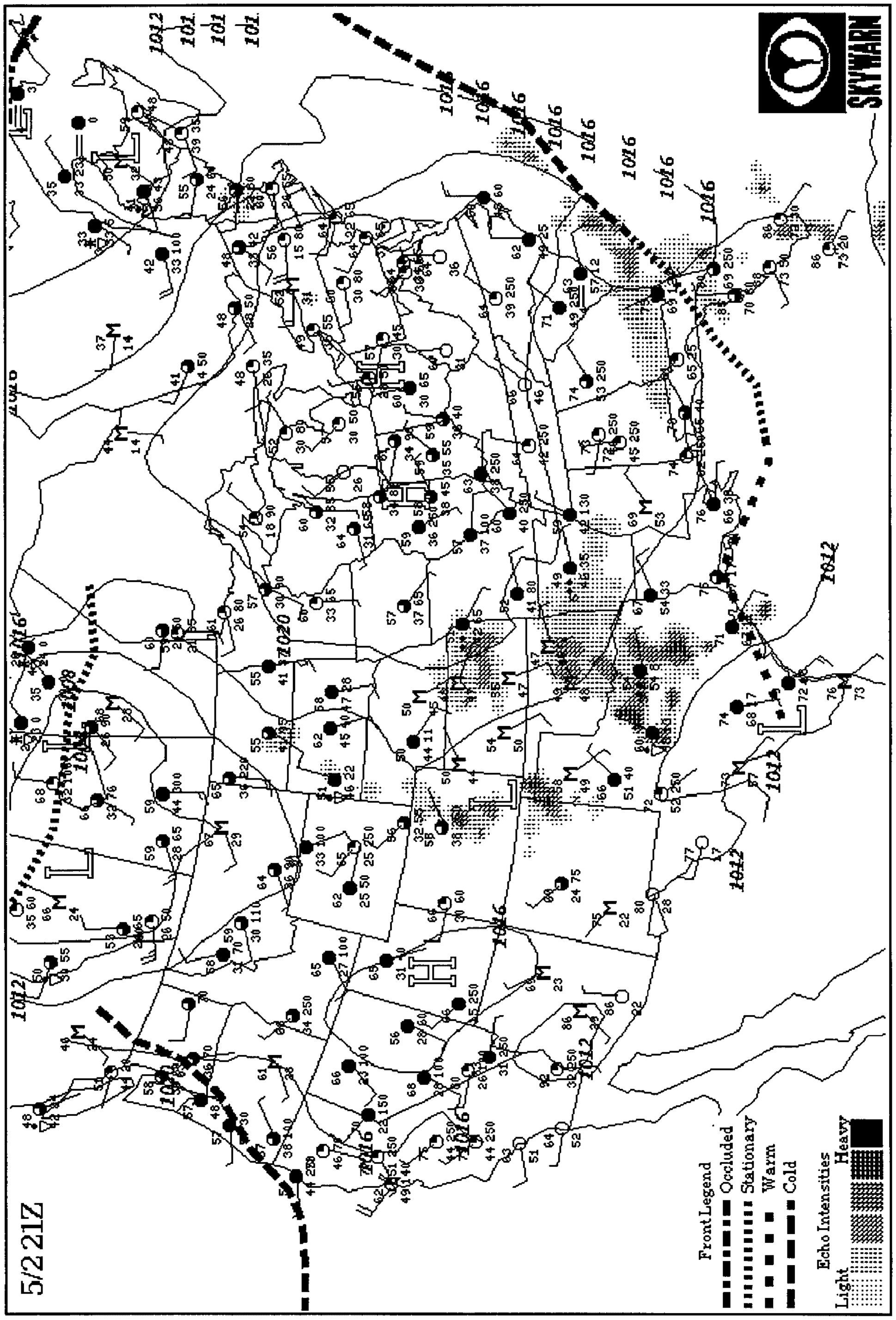
DENVER AND VICINITY-
330 PM MDT FRI APR 29 1994

TONIGHT... LIGHT SNOW LIKELY WITH AREAS OF FOG. SNOW ENDING SHORTLY AFTER MIDNIGHT... CLOUDS DECREASING LATER. TOTAL ACCUMULATION OF 1 TO 2 INCHES... MAINLY ON GRASSY AREAS. LOWS AROUND 30. CHANCE OF SNOW 70 PERCENT.
SATURDAY... PARTLY CLOUDY WITH SOME MORNING FOG. HIGHS AROUND 50.
SATURDAY NIGHT... MOSTLY CLEAR. LOWS 35 TO 40.
SUNDAY... PARTLY CLOUDY. HIGHS 55 TO 60.
****

Special weather statement in effect for DEN
****

SPECIAL WEATHER STATEMENT
NATIONAL WEATHER SERVICE DENVER CO
300 PM MDT FRI APR 29 1994

..ALL ADVISORIES CANCELLED FOR SOUTHERN COLORADO...
THE SNOW ADVISORIES THAT WERE ISSUED EARLIER FOR THE SANGRE DE CHRISTO MOUNTAINS... WET MOUNTAINS AND THE PIKES/TELLER COUNTY AREA HAVE BEEN CANCELLED. ALTHOUGH PERIODS OF HEAVY SNOW HAVE OCCURRED TODAY... MUCH OF THE SNOW THAT HAS FALLEN HAS MELTED AS IT HAS REACHED THE SURFACE. AN ADDITONAL 2 TO 4 INCHES OF SNOW ARE POSSIBLE IN THESE AREAS TONIGHT... SO ROADWAYS ARE LIKELY TO BE SLICK. MOTORISTS TRAVELLING INTO THESE MOUNTAINOUS AREAS TONIGHT SHOULD STILL USE EXTRA CAUTION.
****


5/2 21Z
Front Legend
Occluded
Stationary
Warm
Cold
Echo Intensities
Light
Heavy
SKYWARN

Example: Boston, Massachusetts; Burlington, Vermont; Cleveland, Ohio

The vigorous storm system that dominated the 29 April map has passed out to sea, though the center of the low is still lingering over New Brunswick and Newfoundland, Canada. The tail of the cold front, which is still stationary in the deep south, is causing cloud cover over northern Florida and southern Georgia. Boston is clearing as the cold high moves in with brisk, westerly winds. Temperatures are expected to range from lows in the 30s and 40s (°F) Monday night to highs in the 60s (°F) on Tuesday because of longwave radiation losses to the clear night air and strong insolation gains during the following sunny day. Similar patterns are expected in Burlington, Cleveland, and Roanoke, with frost expected in all three cities. Daytime wind speed peaks could also be expected in Burlington, with its expected clear skies and calm regional winds on Tuesday, because of solar-driven convective coupling with upper winds.

Boston and Roanoke are expected to begin clouding by Wednesday, with a chance of rain. Cleveland and Burlington are not expected to have rain until Thursday. The source of rainfall is probably the developing lows in Texas and Colorado, areas of frequent cyclogenesis. The high over the East will cause the delay in rainfall in Cleveland and Burlington by blocking and steering the storm through the Ohio River valley and the Appalachians toward the coast.

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GREATER BOSTON METROPOLITAN AREA FORECAST
NATIONAL WEATHER SERVICE BOSTON MA

258 PM EDT MON MAY 2 1994
TONIGHT... BECOMING CLEAR AND CHILLY. LOW NEAR 40 IN THE URBAN
CENTERS AND MID 30S SUBURBS.  WEST WIND 10 TO 20 MPH.
TUESDAY... SUNNY. HIGH 60 TO 65... BUT TEMPERATURES FALLING THROUGH THE 50S
IN THE AFTERNOON ALONG SOUTH FACING COASTAL POINTS. WEST WIND 10 TO 20 MPH
BECOMING SOUTH IN THE AFTERNOON.
TUESDAY NIGHT... INCREASING CLOUDS. LOW 40 TO 45.
WEDNESDAY... SUNSHINE EARLY OTHERWISE THICKENING CLOUDS WITH A CHANCE OF
RAIN LATE IN THE DAY. HIGH 60 TO 65.  CHANCE OF RAIN 30 PERCENT.
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THE OFFICIAL FORECAST FOR THE BURLINGTON AREA
NATIONAL WEATHER SERVICE BURLINGTON VT

245 PM EDT MON MAY 2 1994

REST OF THIS AFTERNOON AND EARLY THIS EVENING... CLOUDY PERIODS WITH SCATTERED
RAIN... SLEET OR SNOW SHOWERS. WESTERLY WIND 10 TO 20 MPH AND GUSTY.
TONIGHT... CLEARING AND COLD WITH WIDESPREAD FROST AND FREEZING
TEMPERATURES. LOW 25 TO 30. NORTHWEST WIND DIMINISHING.
TUESDAY... SUNNY AND WARMER. HIGH AROUND 60.
TUESDAY NIGHT... CLEAR. LOW AROUND 40.
WEDNESDAY... MOSTLY SUNNY AND PLEASANT. HIGH 65 TO 70.
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CLEVELAND AND VICINITY FORECAST
NATIONAL WEATHER SERVICE CLEVELAND OH

330 PM EDT MON MAY 2 1994

..FROST WARNING TONIGHT...

TONIGHT... PARTLY CLOUDY WITH FROST LIKELY. LOW 30 TO 35. VARIABLE WINDS 5 TO 10 MPH BECOMING SOUTHEAST.

TUESDAY... MOSTLY CLOUDY. HIGH IN THE LOWER 60S. SOUTHEAST WINDS 10 TO 15 MPH.

TUESDAY NIGHT... PARTLY CLOUDY. LOW IN THE MIDDLE 40S.

WEDNESDAY... PARTLY SUNNY. HIGH IN THE LOWER 60S.

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Special weather statement in effect for CLE

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SPECIAL WEATHER STATEMENT AND AFTERNOON OHIO WEATHER STORY
NATIONAL WEATHER SERVICE CLEVELAND OH

337 PM EDT MON MAY 2 1994

A FROST WARNING REMAINS IN EFFECT FOR NORTH CENTRAL AND NORTHEAST OHIO TONIGHT.

SOME CITIES IN THE WARNING AREA INCLUDE CLEVELAND... MANSFIELD... AKRON-CANTON... AND YOUNGSTOWN.

AFTER A COLD AND FROSTY START TO THE DAY THIS MORNING... SOME SUNSHINE PUSHED TEMPERATURES UP THIS AFTERNOON. HOWEVER... READINGS REMAINED BELOW NORMAL WITH TEMPERATURES AT MID AFTERNOON IN THE MID TO UPPER 50S ACROSS THE STATE. NORMAL AFTERNOON HIGHS FOR THIS TIME OF THE YEAR RANGE FROM THE MID 60S IN THE NORTH TO THE LOWER 70S IN THE SOUTH.

MOSTLY CLEAR SKIES AND LIGHT WINDS SHOULD LEAD TO A SECOND CONSECUTIVE FROSTY NIGHT ACROSS NORTHEAST OHIO WITH LOWS FROM 30 TO 35 DEGREES. AGRICULTURAL INTERESTS AND PEOPLE WITH OUTDOOR PLANTS IN THIS AREA SHOULD TAKE MEASURES TO PROTECT TENDER VEGETATION FROM FROST OVERNIGHT.

ACROSS THE REST OF OHIO... INCREASING CLOUDS AND WINDS SHOULD KEEP TEMPERATURES A LITTLE WARMER. OVER NORTHWEST OHIO TEMPERATURES SHOULD FALL INTO THE MID 30S WHICH COULD ALLOW FOR THE FORMATION OF SOME SCATTERED AREAS OF FROST... WHILE THE REMAINDER OF OHIO WILL SEE LOWS FROM THE UPPER 30S TO MIDDLE 40S.

SOME SUNSHINE WILL BE EVIDENT ACROSS NORTHERN OHIO TUESDAY WHILE CLOUDS WILL BE WIDESPREAD ACROSS SOUTHERN OHIO ALONG WITH A RISK OF SHOWERS. HIGH TEMPERATURES WILL BE IN THE LOWER TO MIDDLE 60S STATEWIDE.

ON THE LATEST WEATHER MAP... HIGH PRESSURE WAS CENTERED OVER OHIO. THE HIGH WILL MOVE EAST TO PENNSYLVANIA BY TUESDAY WHILE A WEAK REGION OF LOW PRESSURE DEVELOPS OVER EXTREME SOUTHERN OHIO. BY WEDNESDAY MORNING THE HIGH WILL HAVE MOVED OFF THE EAST COAST WHILE THE WEAK LOW PRESSURE AREA REMAINS IN PLACE ACROSS EXTREME SOUTHERN OHIO.

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ROANOKE AND VICINITY FORECAST
NATIONAL WEATHER SERVICE ROANOKE VA

431 PM EDT MON MAY 2 1994

TONIGHT... MOSTLY CLEAR. COLD WITH PATCHY FROST POSSIBLE. LOW NEAR 40. LIGHT WIND.
TUESDAY... INCREASING CLOUDS. HIGH NEAR 60. SOUTHEAST WIND 10 MPH.
TUESDAY NIGHT... PERIODS OF RAIN. LOW AROUND 50. CHANCE OF RAIN 80 PERCENT.
WEDNESDAY... CLOUDY AND COOL WITH SHOWERS LIKELY. HIGH AROUND 60.
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Weather situation: Cyclogenesis and warm frontal conditions **2 May 1994**

Example: Little Rock, Arkansas

The Gulf Coast, especially in the Texas Panhandle, Oklahoma, and Arkansas area, is a region of frequent cyclogenesis because it is a location where warm, moist air from the Gulf of Mexico and cool, dry air from the northern plains readily and frequently clash. Movement of the upper winds poleward to the Northeast from this area also encourages cyclogenesis.

The weather map shows two lows in the vicinity. The low in southwest Colorado is not organized yet, but it is a source of some precipitation. It may organize into a well-developed storm with fronts if it moves closer to the Gulf of Mexico. The low located on the Texas–Mexico border is better organized and is associated with the front trailing the storm currently over Newfoundland and New Brunswick. A warm front has formed ahead of the low that is producing some of the shower activity in Arkansas and northeastern Texas. As in most warm fronts, surface winds at Little Rock are easterly, temperatures are cool, and conditions are cloudy. The warm front is fairly vigorous because it contains embedded thunderstorms. Conditions will not improve for several days in Little Rock. As the storm and warm front move over, temperatures rise and cloud cover diminishes, perhaps because the warm sector moves over the city after the warm front passes. Note that winds are northeasterly ahead of the front and southeasterly south of it, both to be expected in warm frontal weather.

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LITTLE ROCK AND VICINITY FORECAST
NATIONAL WEATHER SERVICE LITTLE ROCK AR

445 PM CDT MON MAY 2 1994

TONIGHT... CLOUDY AND COLD. A 60 PERCENT CHANCE OF RAIN OR THUNDERSHOWERS. LOW 45 TO 50. NORTHEAST WIND 5 TO 10 MPH.
TUESDAY... MOSTLY CLOUDY AND COOL WITH A 50 PERCENT CHANCE OF RAIN OR THUNDERSHOWERS. HIGH 60 TO 65. NORTHEAST WIND 5 TO 10 MPH.
TUESDAY NIGHT... PARTLY CLOUDY AND COOL. LOW 50 TO 55.
WEDNESDAY... PARTLY CLOUDY AND WARMER. HIGH IN THE MID 70S.
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Example: Seattle and Spokane, Washington

A cold front has passed over Seattle within the last few hours and is due to cross Spokane on Wednesday. Winds have become southwesterly to westerly in the Seattle area, whereas Spokane is experiencing southerly winds (in the forecast) since it will be within the warm sector ahead of the cold front. The warm front of the storm is in the Canadian prairies. Temperatures cool only slightly behind the cold front because on the West Coast maritime air follows the cold front. The colder air following cold fronts in the Northwest is milder as a result of its oceanic origins. West Coast cold fronts are also usually not as vigorous as those found east of the Rocky Mountains because the temperature contrasts are much smaller due to the maritime influence. As a result, thunderstorms are rare on the West Coast, occuring fewer than 5 days per year on the average.

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SEATTLE TACOMA EVERETT AND VICINITY FORECAST
NATIONAL WEATHER SERVICE SEATTLE WA

1030 AM PDT MON MAY 2 1994

THIS AFTERNOON... BECOMING PARTLY SUNNY. A FEW SHOWERS, MAINLY NORTH OF
SEATTLE. A LITTLE COOLER. HIGHS NEAR 60. WIND SOUTHWEST 5 TO 15 MPH,
EXCEPT NORTHWEST 10 MPH IN THE NORTH PART.
TONIGHT... PARTLY CLOUDY. LOWS IN THE LOWER TO MID 40S. LIGHT SOUTHWEST
WIND.
TUESDAY... AREAS OF LOW CLOUDS AND PATCHY FOG OTHERWISE PARTLY SUNNY.
INCREASING CLOUDS AFTERNOON WITH A SLIGHT CHANCE OF RAIN. HIGHS IN THE
MID 60S.
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SPOKANE AND VICINITY FORECAST
NATIONAL WEATHER SERVICE SPOKANE WA

400 PM PDT MON MAY 2 1994

TONIGHT... PARTLY CLOUDY. LOW ABOUT 38. SOUTH WIND 5 TO 15 MPH.
TUESDAY... PARTLY SUNNY AND WARMER, THE HIGH NEAR 66. VARIABLE WIND 5 TO
15 MPH.
TUESDAY NIGHT... INCREASING CLOUDS AND A 20 PERCENT CHANCE OF SHOWERS
LATE. LOW ABOUT 42.
WEDNESDAY... MOSTLY CLOUDY WITH A 30 PERCENT CHANCE OF SHOWERS OR AN
AFTERNOON THUNDERSTORM. HIGH AROUND 64.
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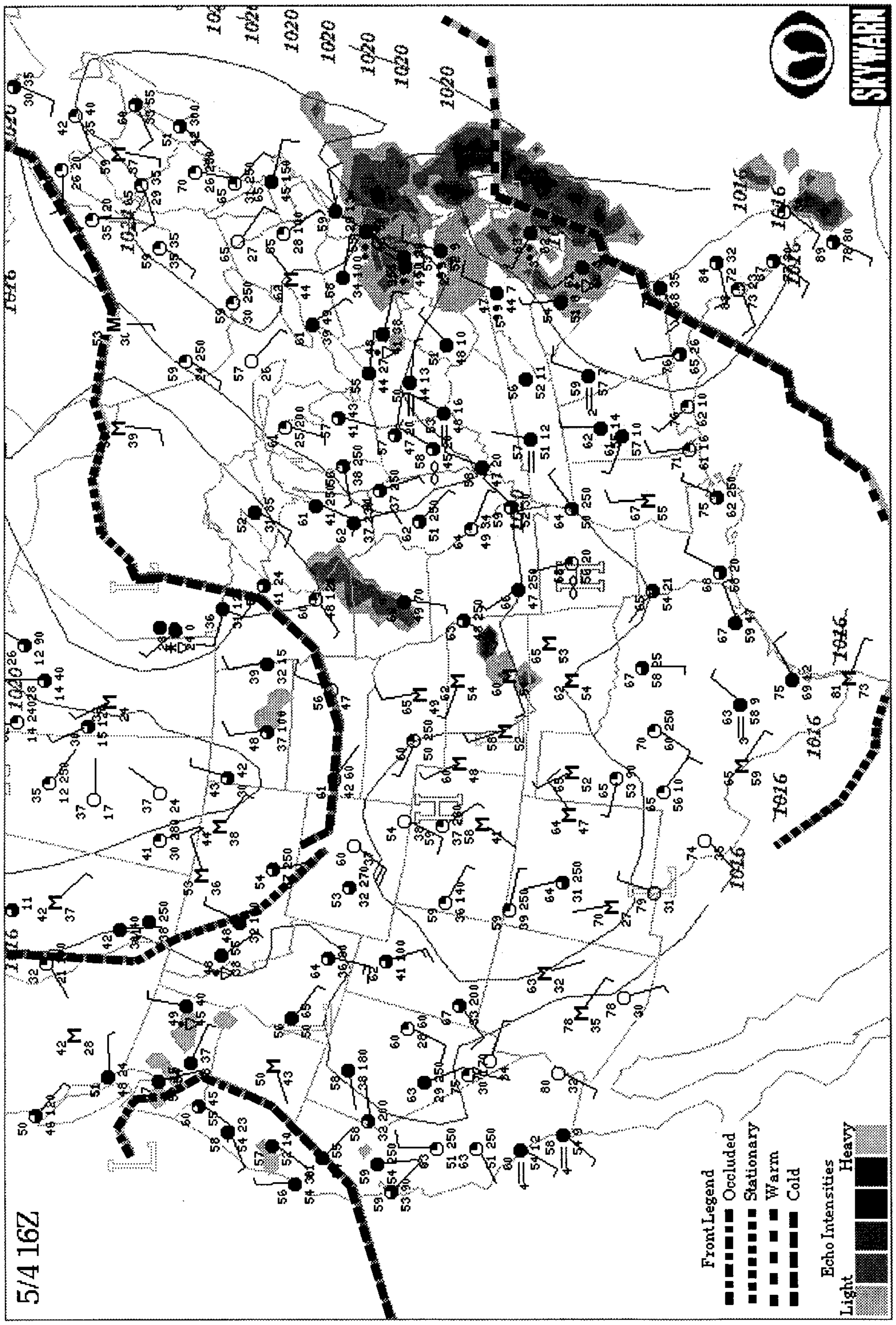
5/4 16Z
SKYWARN
Front Legend
Occluded
Stationary
Warm
Cold
Echo Intensities
Light
Heavy

Example: Boston, Massachusetts; Norfolk, Virginia

Coastal storms, storms that travel along the coast from south to north, are common along the East Coast. The southeast coast, off the Carolinas, is a frequent area of cyclogenesis, especially in the winter. Cold air moving offshore is heated at its base and humidified by the warm Gulf Stream. In addition, supporting upper winds are often moving poleward in this area. These storms develop and strengthen quickly, often producing high winds and heavy precipitation. The storm off Cape Hatteras on the map formed along the cold front off the southeast coast visible on the 2 May map.

The storm is rapidly approaching Boston. Conditions are currently partly sunny, but clouds are moving in quickly, and winds will become southeasterly. As the storm approaches with its counterclockwise rotation, initial winds will be easterly. Clearing occurs within 48 hours of the storm's initial appearance.

Winds are northeasterly in Norfolk as the storm approaches. The storm also lasts less than 48 hours there. Clearing after the storm in both cities is caused in part by the high pressure following from Arkansas.

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GREATER BOSTON METROPOLITAN AREA FORECAST

NATIONAL WEATHER SERVICE BOSTON MA

1117 AM EDT WED MAY 4 1994

THIS AFTERNOON... PARTLY SUNNY. HIGH IN THE UPPER 60S BUT TEMPERATURES
FALLING INTO THE UPPER 50S ALONG THE COAST AS THE WIND BECOMES SOUTHEAST
10 TO 20 MPH.
TONIGHT... RAIN DEVELOPING TOWARD MIDNIGHT AND BECOMING WINDY. LOW NEAR
45. WIND BECOMING NORTHEAST 20 TO 30 MPH. CHANCE OF RAIN NEAR 100
PERCENT.
THURSDAY... RAIN LIKELY. WINDY WITH A HIGH NEAR 50. CHANCE OF RAIN 70
PERCENT.
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HAMPTON ROADS AND VICINITY FORECAST

NATIONAL WEATHER SERVICE NORFOLK VA

417 AM EDT WED MAY 4 1994

 ..SMALL CRAFT ADVISORY ON THE BAY...
 ..HEAVY SURF ADVISORY AND GALE WARNING ALONG THE COAST...
TODAY... OCCASIONAL RAIN POSSIBLY HEAVY AT TIMES. WINDY AND COOLER WITH
HIGHS AROUND 60. WINDS NORTHEAST 15 TO 25 MPH. CHANCE OF RAIN NEAR 100
PERCENT.
TONIGHT... MOSTLY CLOUDY WITH A GOOD CHANCE OF LIGHT RAIN OR DRIZZLE
ENDING AROUND MIDNIGHT. LOWS IN THE MID TO UPPER 40S. WINDS NORTH 15 TO
25 MPH. CHANCE OF RAIN 90 PERCENT.
THURSDAY... DECREASING CLOUDINESS. HIGHS IN THE MID TO UPPER 60S. WINDS
BECOMING WEST 15 MPH.
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Weather situation: Approaching cold front with following high pressure **4 May 1994**

Example: Milwaukee, Wisconsin; Bismark, Williston, and Fargo, North Dakota

A well-developed cold front is approaching Milwaukee from the northwest. A strong high pressure cell follows that is currently (off the map) located in northern Saskatchewan. The front passed the three North Dakota cities hours ago.

Winds are from the south in Milwaukee as the storm approaches because the city is located within the warm sector. Showers and thunderstorms are expected overnight, probably from a squall line ahead of the front. Temperatures fall slightly as the following high pressure arrives, but they rise again near the end of the extended forecast period as southerly winds bring warmer temperatures on the backside of the high as it moves eastward.

Temperatures are colder on the leading side of the high because the high's clockwise rotation brings in northwesterly winds. Warmer, southerly winds trail the high because of this same clockwise rotation. Note the large (30°F) temperature range expected on Saturday as the high moves cool, dry, and clear air into the area after the front passes, with its expected rapid radiative cooling at night and radiative heating during the day. In North Dakota, the southerly air that follows after the center of the high has passed is also humid and is expected to produce airmass thunderstorms from Friday through Sunday as insolation heats the soil surface, producing strong, localized convection.

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KENOSHA-MILWAUKEE-OZAUKEE-RACINE-
1030 AM CDT WED MAY 4 1994

THIS AFTERNOON... MOSTLY CLOUDY WITH A 30 PERCENT CHANCE OF SHOWERS AND
THUNDERSTORMS. HIGHS IN THE MID TO UPPER 60S INLAND... BUT ONLY AROUND 60
LAKESIDE. SOUTH TO SOUTHWEST WINDS 10 TO 15 MPH.
TONIGHT... A 40 PERCENT CHANCE OF SHOWERS AND THUNDERSTORMS... MAINLY EARLY...
THEN TURNING PARTLY CLOUDY. LOWS IN THE MID TO UPPER 40S. WINDS BECOMING
WEST TO NORTHWEST 5 TO 10 MPH.
THURSDAY... PARTLY TO MOSTLY CLOUDY. HIGHS IN THE MID 60S.
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BURLEIGH-EMMONS-GRANT-KIDDER-MERCER-MORTON-OLIVER-SIOUX-
INCLUDING THE CITIES OF... BEULAH... MANDAN... BISMARCK
1045 AM CDT WED MAY 4 1994

THIS AFTERNOON... PARTLY SUNNY. HIGH NEAR 60. WIND NORTH 10 TO 20 MPH.
TONIGHT... CLEAR TO PARTLY CLOUDY. LOW AROUND 30. WIND LIGHT.
THURSDAY... PARTLY SUNNY. HIGH AROUND 60. LIGHT EAST WIND
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DIVIDE-DUNN-MCKENZIE-WILLIAMS-
INCLUDING THE CITY OF... WILLISTON
1045 AM CDT WED MAY 4 1994

THIS AFTERNOON... PARTLY SUNNY. HIGH IN THE UPPER 50S. WIND NORTHEAST 5 TO
15 MPH.
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TONIGHT... CLEAR TO PARTLY CLOUDY. LOW 30 TO 35. WIND LIGHT.
THURSDAY... PARTLY SUNNY. HIGH AROUND 60. WIND SOUTHEAST 10 TO 15 MPH.
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CASS-CLAY MN-NORMAN MN-RANSOM-RICHLAND-SARGENT-STEELE-TRAILL-
INCLUDING THE CITIES OF... FARGO... LISBON... WAHPETON... MOORHEAD MN
1045 AM CDT WED MAY 4 1994

THIS AFTERNOON... MOSTLY CLOUDY WITH A FEW SPRINKLES. HIGH NEAR 50. WIND
NORTHWEST 15 TO 25 MPH.
TONIGHT... PARTLY CLOUDY. LOW IN THE LOWER 30S. WIND LIGHT.
THURSDAY... PARTLY SUNNY. HIGH IN THE UPPER 50S. WIND NORTH 5 TO 15 MPH.
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Weather situation: Approaching low and occluded front **4 May 1994**

Example: Olympia, Walla Walla, Spokane, Washington

An occluded front has entered western Washington and, on the map, is located approximately over Puget Sound. Occluded fronts, not discussed in the main report, occur when a cold front overtakes a warm front, producing a combination of weather associated with each type of front. They are typically no more vigorous than either front individually, but they do provide weather sequences that are modified versions of the warm front–warm sector–cold front scenario.

This occluded front is particularly vigorous, however, especially for the West Coast, because thunderstorms are expected. The vigorous front, coupled with the mountainous terrain of the region, is producing particularly strong storms accompanied by high winds and small hail. The thunderstorm conditions at Olympia are expected a few hours before similar conditions in Walla Walla and Spokane. This is a rapidly moving system, indicated by the similarity of the forecasts in the western and eastern portions of the state, and because clearing begins very rapidly, with partly sunny or partly cloudy skies forecast for Thursday at all locations.

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OLWA-OLYMPIA AND VICINITY FORECAST
NATIONAL WEATHER SERVICE OLYMPIA WA

430 AM PDT WED MAY 4 1994

TODAY... OCCASIONAL SHOWERS AND PARTIAL CLEARING. POSSIBLE
AFTERNOON THUNDERSHOWER. HIGH IN THE LOWER 60S. WIND SOUTH
5 TO 15 MPH.
TONIGHT... OCCASIONAL SHOWERS AND PARTIAL CLEARING. POSSIBLE EVENING
THUNDERSHOWER. LOW 40 TO 45. WIND SOUTHWEST 5 TO 15 MPH.
THURSDAY... PARTLY SUNNY WITH A CHANCE OF SHOWERS. HIGH NEAR 65.
CHANCE OF MEASURABLE RAINFALL IS 90 PERCENT TODAY, 60 PERCENT TONIGHT,
AND 30 PERCENT THURSDAY.
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Special weather statement in effect for OLM
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SPECIAL WEATHER STATEMENT

NATIONAL WEATHER SERVICE OLYMPIA, WA

830 AM PST WED MAY 04 1994

..MOIST AND UNSTABLE AIR MOVING OVER SOUTHWEST WASHINGTON...
THE AIR FLOWING OVER WESTERN WASHINGTON TODAY IS VERY MOIST AND UNSTABLE. MORNING TEMPERATURES ARE ALREADY INTO THE UPPER 50S, AND THAT SUGGESTS THAT THERE WILL PROBABLY BE A FEW THUNDERSTORMS OVER THE SOUTHWEST CORNER OF THE STATE THIS AFTERNOON. SOME OF THESE STORMS MAY PRODUCE SHORT PERIODS OF HEAVY RAIN, SOME SMALL HAIL AND GUSTY WINDS.
IF YOU ARE PLANNING ANY OUTDOOR ACTIVITIES THIS AFTERNOON, YOU SHOULD STAY ALERT FOR QUICKLY DEVELOPING THUNDERSTORMS. THESE STORMS ARE ALWAYS ACCOMPANIED BY DANGEROUS LIGHTNING. IF THUNDERSTORMS APPROACH SEEK APPROPRIATE SHELTER.

STAY TUNED TO NOAA WEATHER RADIO OR YOUR FAVORITE NEWS SOURCE FOR FURTHER STATEMENTS AS THIS WEATHER DEVELOPS.
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WALLA WALLA AND THE WALLA WALLA VALLEY FORECAST
NATIONAL WEATHER SERVICE PENDLETON OR

445 AM PDT WED MAY 4 1994

TODAY... RAIN AT TIMES DECREASING TO SCATTERED SHOWERS WITH POSSIBLE THUNDERSTORMS THIS AFTERNOON. HIGHS MID TO UPPER 60S. VARIABLE WIND TO 15 MPH.
TONIGHT... SHOWERS AND A CHANCE OF AN EVENING THUNDERSTORM. LOWS UPPER 40S. VARIABLE WIND TO 15 MPH.
THURSDAY... PARTLY SUNNY BUT CHANCE OF SHOWERS AND THUNDERSTORMS MAINLY IN THE AFTERNOON. HIGHS IN THE 60S.
CHANCE OF MEASURABLE RAINFALL... 60 PCT TODAY, 50 PCT TONIGHT AND THURSDAY.
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SPOKANE AND VICINITY FORECAST

NATIONAL WEATHER SERVICE SPOKANE WA

400 AM PDT WED MAY 4 1994

TODAY... A 70 PERCENT CHANCE OF SHOWERS AND A CHANCE OF AFTERNOON THUNDERSTORMS. MOSTLY CLOUDY AND COOLER WITH A HIGH AROUND 60. LIGHT WINDS WITH HIGHER GUSTS NEAR THUNDERSTORMS.
TONIGHT... STILL A 60 PERCENT CHANCE OF SHOWERS WITH POSSIBLE EVENING THUNDERSTORMS. MOSTLY CLOUDY WITH A LOW NEAR 47. LIGHT WINDS, EXCEPT GUSTY VICINITY OF THUNDERSTORMS.
THURSDAY... PARTLY CLOUDY AND A LITTLE WARMER, THE HIGH ABOUT 63. A 40 PERCENT CHANCE OF SHOWERS.
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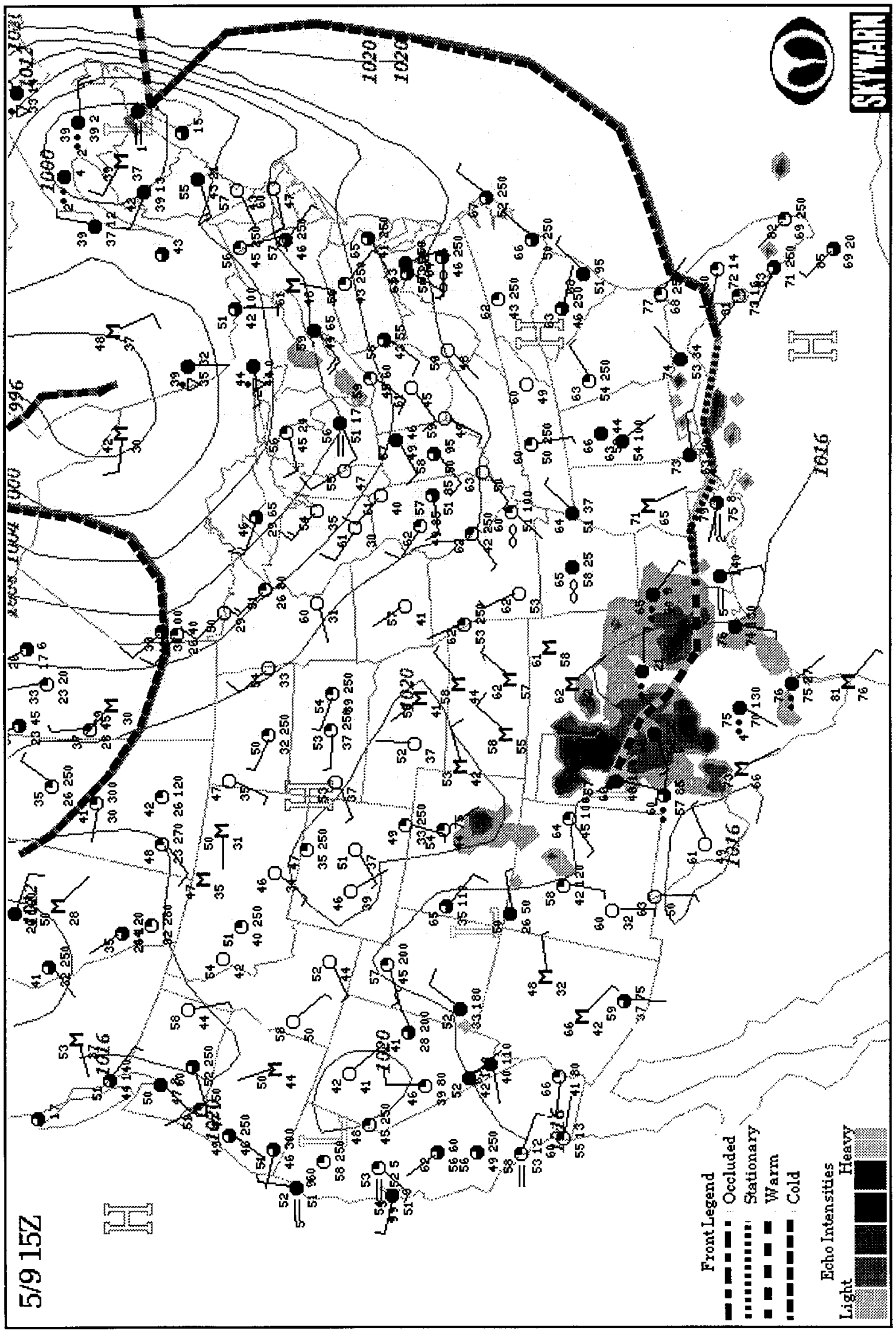

5/9 15Z
Front Legend
Occluded
Stationary
Warm
Cold
Echo Intensities
Light
Heavy
SKYWARN

Example: Dallas–Ft. Worth and Lubbock, Texas; Shreveport, Louisiana; Oklahoma City, Oklahoma; Little Rock, Arkansas

Warm fronts are a result of warm, moist air overrunning cooler air at the surface. As a result, they are associated with widespread stratus clouds, steady rains, and occasional thunderstorms. The rainshield within the warm front bisecting Texas extends about 250 miles to the northeast. Rain is occurring ahead of the front to the northeast, and frontal fog is found along the frontal boundary at Lake Charles and New Orleans, Louisiana. There is evidence of heavy rains on the map, especially near Lubbock, Texas. The Dallas–Ft. Worth forecast indicates that there is a high probability of severe thunderstorms, though little chance for tornadoes. The Dallas–Ft. Worth forecast also indicates that a cold front will be advancing from Kansas to produce some of the rain, though that front is not in evidence on the map, unless the forecast is expecting the cold front in the southern Canadian prairies to advance quickly. Another possibility is that the cold front is forming between the northerly winds on the eastern side of the high between South Dakota and Wyoming, and the southerly winds in Texas and Oklahoma. Warm fronts often have embedded thunderstorms if the overrunning air is very unstable and if there is a jet stream overhead, as there is in this case (as stated in the Dallas–Ft. Worth forecast).

Shreveport, Oklahoma City, and Lubbock are each forecasted for heavy rains and thunderstorms, with conditions to persist for 48–72 hours. Storm development is most vigorous near the storm center, with severe thunderstorms expected in Oklahoma. All winds are easterly to southerly, typical for locations ahead of warm fronts. Little Rock will escape most of the severe weather because of its distance to the northeast. The map shows haze in Little Rock. Little Rock weather for the next few days is more typical of warm, humid summer airmasses than of warm fronts, probably because of its distance from the front.

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DALLAS-FORT WORTH METROPLEX FORECAST

NATIONAL WEATHER SERVICE FORT WORTH TX

515 AM CDT MON MAY 9 1994

TODAY... A 70 PERCENT CHANCE OF THUNDERSTORMS... A FEW POSSIBLY SEVERE WITH
HEAVY RAIN. A HIGH IN THE LOWER 80S. SOUTHEAST WIND 10 TO 15 MPH.
TONIGHT AND TUESDAY... RAIN AND THUNDERSTORMS LIKELY WITH A FEW STORMS
POSSIBLY SEVERE WITH HEAVY RAIN. A LOW IN THE MIDDLE 60S AND A HIGH IN
THE UPPER 70S. WIND BECOMING EAST TO NORTHEAST 5 TO 15 MPH. CHANCE OF
RAIN 70 PERCENT TONIGHT... 50 PERCENT TUESDAY.
THE EXTENDED FORECAST... TUESDAY NIGHT THROUGH FRIDAY... MOSTLY CLOUDY
TUESDAY NIGHT WITH THUNDERSTORMS SLOWLY DIMINISHING AND A LOW NEAR 60.
PARTLY CLOUDY WITH A CHANCE OF THUNDERSTORMS WEDNESDAY THROUGH FRIDAY.
LOWS IN THE 60S... HIGHS IN THE 80S.
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Special weather statement in effect for FTW

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SEVERE WEATHER OUTLOOK FOR NORTH TEXAS

NATIONAL WEATHER SERVICE FT WORTH TX

600 AM CDT MON MAY 09 1994

THERE IS A SLIGHT RISK OF SEVERE THUNDERSTORMS TODAY AND TONIGHT ACROSS ALL OF NORTH TEXAS.

THE MAIN RISKS THROUGH TONIGHT WILL BE FROM HEAVY RAINFALL... LARGE HAIL... AND DAMAGING WINDS. AT THIS TIME THE RISK OF TORNADOES APPEARS TO BE LOW.

AN UPPER LEVEL LOW PRESSURE CENTER WILL BECOME STATIONARY OVER ARIZONA TODAY. EAST OF THE LOW... A DIFLUENT UPPER LEVEL WIND PATTERN ..FAVORABLE FOR THUNDERSTORM FORMATION... WILL BECOME ESTABLISHED ACROSS NORTH TEXAS.

AT THE SURFACE A SLOW-MOVING WARM FRONT EXTENDED FROM SOUTH OF LUFKIN TO NEAR HILLSBORO AND THROCKMORTON AROUND SUNRISE. A WEAK COOL FRONT WAS ADVANCING SOUTH FROM KANSAS. THESE FRONTAL SYSTEMS WILL BECOME FOCUSING BOUNDARIES FOR THUNDERSTORM REDEVELOPMENT THIS AFTERNOON AND EVENING. THUNDERSTORM OUTFLOW BOUNDARIES FROM THIS MORNING'S ACTIVITY ALSO MAY SERVE TO TRIGGER LATE DAY THUNDERSTORMS.

UNLIKE THE EVENT WHICH PRODUCED THE DEVASTATING TORNADOES TWO WEEKS AGO... THIS SLOW-MOVING PATTERN SHOULD BE MORE OF A HEAVY RAIN THREAT... WITH ONLY A FEW STORMS PRODUCING LARGE HAIL AND DAMAGING WINDS. ALTHOUGH HIGH INSTABILITIES AND STRONG UPPER LEVEL JET STREAM WINDS ARE EXPECTED... THE LOW-LEVEL WINDS AND WIND SHEARS SHOULD BE TOO WEAK TO SUPPORT SIGNIFICANT TORNADOES. THE GREATEST VERTICAL WIND SHEAR VALUES... POSSIBLY SUFFICIENT TO SUPPORT ONE OR TWO BRIEF SUPERCELL STORMS... WILL BE IN WESTERN SECTIONS OF NORTH TEXAS THIS AFTERNOON AND EARLY EVENING.

EMERGENCY MANAGEMENT OFFICIALS AND STORM SPOTTERS SHOULD BE READY FOR POSSIBLE ACTIVATION THROUGHOUT TODAY AND THIS EVENING.

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LUBBOCK AND VICINITY FORECAST

NATIONAL WEATHER SERVICE LUBBOCK TX

1115 AM CDT MON MAY 9 1994

THIS AFTERNOON... SCATTERED SHOWERS AND THUNDERSTORMS. LOCALLY HEAVY RAINFALL POSSIBLE. HIGH 80-85. WIND BECOMING EAST 10-20 MPH. RAIN CHANCE 50 PERCENT.

TONIGHT... SHOWERS AND THUNDERSTORMS LIKELY. LOCALLY HEAVY RAINFALL POSSIBLE. LOW AROUND 55. EAST WIND 10-20 MPH AND GUSTY. RAIN CHANCE 60 PERCENT.

TUESDAY... A 50 PERCENT CHANCE OF SHOWERS AND THUNDERSTORMS. LOCALLY HEAVY RAINFALL POSSIBLE. HIGH NEAR 70. EAST WIND 10-20 MPH AND GUSTY.

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SHLA-SHREVEPORT/BOSSIER CITY METROPOLITAN FORECAST

NATIONAL WEATHER SERVICE SHREVEPORT LA

1100 AM CDT MON MAY 9 1994

THIS AFTERNOON... CLOUDY WITH PERIODS OF RAIN... SHOWERS AND A FEW THUNDERSTORMS. HIGH NEAR 70. EAST WIND 5 TO 10 MPH. CHANCE OF RAIN 80 PERCENT.

TONIGHT... CLOUDY WITH SHOWERS AND THUNDERSTORMS LIKELY. LOW IN THE LOWER 60S. NORTHEAST WIND 5 TO 10 MPH. CHANCE OF RAIN 50 PERCENT.

TUESDAY... CLOUDY WITH A CHANCE OF SHOWERS AND THUNDERSTORMS. HIGH IN THE UPPER 70S. NORTHEAST WIND NEAR 10 MPH. CHANCE OF RAIN 40 PERCENT.

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ARKLATEX WEATHER SUMMARY AND FORECAST

NATIONAL WEATHER SERVICE SHREVEPORT LA

1100 AM CDT MON MAY 9 1994

CLOUDS PERSISTED ACROSS THE ARKLATEX MONDAY... WITH RAIN AND SOME THUNDERSHOWERS SPREADING INTO THE REGION THROUGHOUT MONDAY MORNING. HIGHS SUNDAY WERE FROM NEAR 70 TO THE MID 70S WHILE LOWS MONDAY MORNING WERE IN THE 60S.

FORECAST...

THIS AFTERNOON... CLOUDY WITH NUMEROUS AREAS OF RAIN... SHOWERS AND THUNDERSTORMS SPREADING ACROSS THE REGION. HIGHS IN THE MID 60S WEST TO THE MID 70S EAST.

TONIGHT... CLOUDY WITH SCATTERED TO NUMEROUS SHOWERS AND THUNDERSTORMS. LOWS FROM NEAR 60 TO THE MID 60S.

TUESDAY... CONTINUED CLOUDY WITH SCATTERED SHOWERS AND THUNDERSTORMS... MORE
NUMEROUS SOUTH. HIGHS IN THE UPPER 70S TO LOWER 80S.

****

OKLAHOMA CITY METROPOLITAN FORECAST... UPDATED

NATIONAL WEATHER SERVICE OKLAHOMA CITY OK

1145 AM CDT MON MAY 9 1994

THIS AFTERNOON... MOSTLY CLOUDY WITH SCATTERED SHOWERS AND THUNDER-STORMS. HIGH IN THE LOWER 70S. EAST TO SOUTHEAST WIND AROUND 10 MPH. CHANCE OF RAIN 40 PERCENT.

TONIGHT... CLOUDY WITH A 40 PERCENT CHANCE OF SHOWERS AND THUNDERSTORMS. LOW IN THE MID 50S. EAST WIND AROUND 10 MPH.

TUESDAY... MOSTLY CLOUDY WITH A 30 PERCENT CHANCE OF THUNDERSTORMS. HIGH IN THE MID 70S. EAST WIND 10 TO 15 MPH.

TUESDAY NIGHT... MOSTLY CLOUDY WITH A CHANCE OF SHOWERS AND THUNDER-STORMS.
LOW IN THE MID 50S. CHANCE OF RAIN 30 PERCENT.

****

Special weather statement in effect for OKC

****

THUNDERSTORM OUTLOOK

NATIONAL WEATHER SERVICE OKLAHOMA CITY OK

630 AM CDT MON MAY 9 1994

THUNDERSTORMS MAY APPROACH SEVERE LIMITS TODAY AND TONIGHT IN SOUTHERN OKLAHOMA AND THE WESTERN PART OF NORTH TEXAS. THE AREA IS GENERALLY SOUTH OF A LINE FROM ERICK TO PAULS VALLEY TO CLAYTON. THIS INCLUDES ALTUS... ARDMORE... AND HUGO IN OKLAHOMA... AND KNOX CITY... ARCHER CITY... AND WICHITA FALLS IN NORTH TEXAS.

EARLY THIS MORNING... A WEAK COLD FRONT WAS MOVING SOUTHWARD THROUGH NORTHERN OKLAHOMA WHILE ENERGY FROM AN UPPER LEVEL JET HELPED PRODUCE THUNDERSHOWERS IN SOUTHERN OKLAHOMA AND NORTH TEXAS. AS THE DAY PROGRESSES... THE COLD FRONT WILL CONTINUE TO MOVE SLOWLY SOUTHWARD INTO SOUTHERN OKLAHOMA AND NORTH TEXAS. THIS WILL PROVIDE A FOCUS FOR ADDITIONAL THUNDERSTORMS THIS AFTERNOON AS MORE ENERGY MOVES EAST FROM AN UPPER LEVEL STORM SYSTEM IN ARIZONA.

INSTABILITY AND MARGINAL LOW-LEVEL VERTICAL WIND SHEAR WILL BE LIMITING FACTORS FOR SEVERE THUNDERSTORM DEVELOPMENT. THERE IS A WARM AND MOIST AIR MASS SOUTH OF THE FRONT BUT LOW CLOUDS AND ONGOING RAINFALL IN SOUTHERN OKLAHOMA AND NORTH TEXAS WILL LIKELY INHIBIT AFTERNOON HEATING SIGNIFICANTLY.

EMERGENCY MANAGEMENT AND STORM SPOTTER GROUPS IN NORTH TEXAS AND SOUTHERN OKLAHOMA MAY WANT TO MONITOR THESE WEATHER DEVELOPMENTS.

****

LITTLE ROCK AND VICINITY FORECAST

NATIONAL WEATHER SERVICE LITTLE ROCK AR

1045 AM CDT MON MAY 9 1994

THIS AFTERNOON... PARTLY SUNNY AND MILD. A 20 PERCENT CHANCE OF SHOWERS OR THUNDERSHOWERS. HIGH IN THE MID 70S. EAST WIND 5 TO 10 MPH.

TONIGHT... MOSTLY CLOUDY AND COOL WITH A 30 PERCENT CHANCE OF SHOWERS OR THUNDERSHOWERS. LOW IN THE UPPER 50S. NORTHEAST WIND 5 MPH.

TUESDAY... PARTLY CLOUDY AND MILD. HIGH IN THE UPPER 70S. NORTHEAST WIND 5 TO 10 MPH.

****

Weather situation: Trailing side of high pressure, trough, and warm sector of storm **9 May 1994**

Example: Burlington, Vermont; Boston, Massachusetts

High pressure in the Carolinas is following the large storm system on the northern Maine coast. The counterclockwise flow of air around the Maine low, and the clockwise flow around the Carolina high, are complementing one another and pumping warm, humid air into the Northeast. In addition, this southerly flow is reinforcing the warm sector of a storm near James Bay from which a trough of low pressure emanates, extending through the lower Great Lakes (dip in isobars in Michigan, Ohio, and Kentucky). The warm, humid airmass, coupled with the easterly moving trough, will initiate general shower activity in the Burlington and Boston areas. Winds in both cities are forecast to be southerly and strong because of the steep pressure gradient, signified by the closeness of the isobars on the map.

****

THE OFFICIAL FORECAST FOR THE BURLINGTON AREA

NATIONAL WEATHER SERVICE BURLINGTON VT

1025 AM EDT MON MAY 9 1994

THIS AFTERNOON... SUNNY... BREEZY AND WARM. HIGH AROUND 70. SOUTH WIND 15 TO 25 MPH. INCREASING CLOUDS LATE THIS AFTERNOON WITH A 40 PERCENT CHANCE OF SHOWERS TOWARDS EVENING.

TONIGHT... MOSTLY CLOUDY WITH A CHANCE OF SHOWERS. LOW AROUND 45. CHANCE OF RAIN 40 PERCENT. SOUTH WIND 10 TO 15 MPH.

TUESDAY... PARTLY CLOUDY. HIGH NEAR 60.

****

GREATER BOSTON METROPOLITAN AREA FORECAST
NATIONAL WEATHER SERVICE BOSTON MA
1053 AM EDT MON MAY 9 1994

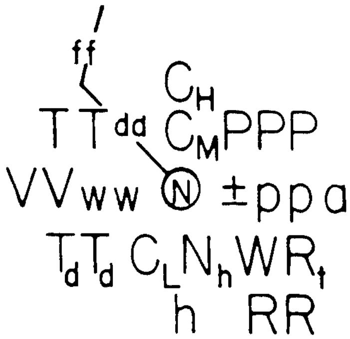
THIS AFTERNOON... MOSTLY SUNNY AND BREEZY. HIGH IN THE UPPER 70S. SOUTHWEST WIND 15 TO 25 MPH.

TONIGHT... INCREASING CLOUDINESS WITH A 30 PERCENT CHANCE OF SHOWERS AFTER MIDNIGHT. LOW 50 TO 55. SOUTHWEST WIND 10 TO 20 MPH.

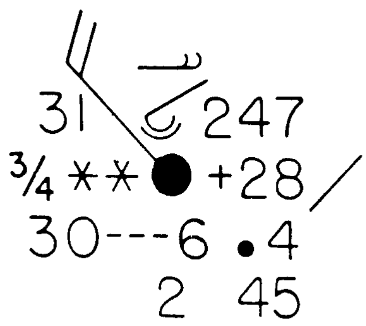
TUESDAY... PARTLY CLOUDY WITH A 30 PERCENT CHANCE OF A SHOWER. HIGH IN THE UPPER 60S.

****

SYMBOLIC STATION MODEL



SAMPLE PLOTTED REPORT



WW PRESENT WEATHER 8

Fog, sky NOT discernible, no appreciable change during past hour	Slight or moderate drifting snow, generally low
	Intermittent drizzle
Haze	Continuous drizzle

U. S. DEPT. OF COMMERCE
WEATHER BUREAU

SURFACE WEATHER MAP AND STATION WEATHER

EXPLANATION OF SYMBOLS

N	Total amount of cloud 8 = completely covered block 6
dd	True direction from which wind is blowing 32 = 320° = NW
ff	Wind speed in knots 20 = 20 knots block 9
VV	Visibility in miles and fractions 12 = 12/16 or 3/4 miles
ww	Present weather 71 = continuous slight snow (block 8)
W	Past weather 6 = rain (block 11)
PPP	Barometric Pressure (in millibars) reduced to sea-level 247 = 1024.7 mb.
TT	Current air temperature 31 = 31° F.
N_h	Fraction of sky covered by low or middle cloud 6 = 7 or 8 tenths block 7
C_L	Cloud type (block 3) 7 = Fractostratus and/or Fractocumulus of bad weather (scud)
C_M	Cloud type 9 = Altocumulus of chaotic sky
C_H	Cloud type 2 = Dense cirrus in patches
h	Height of base of cloud 2 = 300 to 599 feet
T_dT_d	Temperature of dewpoint 30 = 30° F.
a	Characteristic of barograph trace 2 = rising steadily or unsteadily block 10
pp	Pressure change in 3 hours preceding observation 28 = 2.8 millibars
RR	Amount of precipitation 45 = 0.45 inches
R_t	Time precipitation began or ended 4 = 3 to 4 hours ago

N SKY COVERAGE 6 (Total Amount)

No clouds	Six-tenths
Less than one-tenth or one-tenth	Seven-tenths or eight-tenths
Two-tenths or three-tenths	Nine-tenths or overcast with openings
Four-tenths	Completely overcast
Five-tenths	Sky obscured

9 ff KNOTS

ff	KNOTS
◎	Calm
	1 - 2
	3 - 7
	8 - 12
	13 - 17
	18 - 22
	23 - 27
	28 - 32
	33 - 37
	38 - 42
	43 - 47
	48 - 52
	53 - 57
	58 - 62
	63 - 67
	68 - 72
	73 - 77

h HEIGHT IN METERS 5 (Approximate)

h	HEIGHT IN METERS (Approximate)
0	0 - 49
1	50 - 99
2	100 - 199
3	200 - 299
4	300 - 599
5	600 - 999
6	1,000 - 1,499
7	1,500 - 1,999
8	2,000 - 2,499
9	At or above 2,500, or no clouds

W PAST WEATHER 11

W	PAST WEATHER
	Clear or few clouds (Not Plotted)
	Partly cloudy (scattered) or variable sky (Not Plotted)
	Cloudy (broken) or overcast (Not Plotted)
	Sandstorm or duststorm, or drifting or blowing snow
≡	Fog, or smoke, or thick dust haze
,	Drizzle
•	Rain
*	Snow, or rain and snow mixed, or ice pellets (sleet)
▽	Shower(s)
	Thunderstorm, with or without precipitation

Intermittent rain (NOT freezing), slight at time of observation	Slight rain shower(s)	Moderate or thick freezing drizzle	Intermittent fall of snow flakes, slight at time of observation	Slight or moderate thunderstorm without hail, but with rain and/or snow at time of obs.
Continuous rain (NOT freezing), slight at time of observation	Moderate or heavy rain shower(s)	Slight freezing rain	Continuous fall of snow-flakes, slight at time of observation	Heavy thunderstorm with hail at time of observation
Continuous rain (NOT freezing), moderate at time of observation	Slight snow shower(s)	Ice pellets (sleet, U. S. definition)	Continuous fall of snow-flakes, moderate at time of observation	

C_L	DESCRIPTION (Abridged From W.M.O. Code)
	Cu of fair weather, little vertical development and seemingly flattened
	Cu of considerable development, generally towering, with or without other Cu or Sc bases all at same level
	Cb with tops lacking clear-cut outlines, but distinctly not cirriform or anvil-shaped; with or without Cu, Sc, or St
	Sc formed by spreading out of Cu; Cu often present also
	Sc not formed by spreading out of Cu
	St or Fs or both, but no Fs of bad weather
	Fs and/or Fc of bad weather (scud)
	Cu and Sc (not formed by spreading out of Cu) with bases at different levels
	Cb having a clearly fibrous (cirriform) top, often anvil-shaped, with or without Cu, Sc, St, or scud

3 CLOUD ABBREVIATION

- St or Fs-Stratus or Fractostratus
- Ci-Cirrus
- Cs-Cirrostratus
- Cc-Cirrocumulus
- Ac-Altocumulus
- As-Altostratus
- Sc-Stratocumulus
- Ns-Nimbostratus
- Cu or Fc-Cumulus or Fractocumulus
- Cb-Cumulonimbus

C_M	DESCRIPTION (Abridged From W.M.O. Code)
	Thin As (most of cloud layer semi-transparent)
	Thick As, greater part sufficiently dense to hide sun (or moon), or Ns
	Thin Ac, mostly semi-transparent; cloud elements not changing much and at a single level
	Thin Ac in patches; cloud elements continually changing and/or occurring at more than one level
	Thin Ac in bands or in a layer gradually spreading over sky and usually thickening as a whole
	Ac formed by the spreading out of Cu
	Double-layered Ac, or a thick layer of Ac, not increasing; or Ac with As and/or Ns
	Ac in the form of Cu-shaped tufts or Ac with turrets
	Ac of a chaotic sky, usually at different levels; patches of dense Ci are usually present also

N_h	SKY COVERAGE 7 (Low And/Or Middle Clouds)
0	No clouds
1	Less than one-tenth or one-tenth
2	Two-tenths or three-tenths
3	Four-tenths
4	Five-tenths
5	Six-tenths
6	Seven-tenths or eight-tenths
7	Nine-tenths or overcast with openings
8	Completely overcast
9	Sky obscured

a	BAROMETRIC TENDENCY 10	
	Rising, then falling	
	Rising, then steady; or rising, then rising more slowly	Barometer now higher than 3 hours ago
	Rising steadily, or unsteadily	
	Falling or steady, then rising; or rising, then rising more quickly	
	Steady, same as 3 hours ago	
	Falling, then rising, same or lower than 3 hours ago	
	Falling, then steady; or falling, then falling more slowly	Barometer now lower than 3 hours ago
	Falling steadily, or unsteadily	
	Steady or rising, then falling; or falling, then falling more quickly	

C_H	DESCRIPTION (Abridged From W.M.O. Code)
	Filaments of Ci, or "mares tails," scattered and not increasing
	Dense Ci in patches or twisted sheaves, usually not increasing, sometimes like remains of Cb; or towers or tufts
	Dense Ci, often anvil-shaped, derived from or associated with Cb
	Ci, often hook-shaped, gradually spreading over the sky and usually thickening as a whole
	Ci and Cs, often in converging bands, or Cs alone; generally overspreading and growing denser; the continuous layer not reaching 45° altitude
	Ci and Cs, often in converging bands, or Cs alone; generally overspreading and growing denser; the continuous layer exceeding 45° altitude
	Veil of Cs covering the entire sky
	Cs not increasing and not covering entire sky
	Cc alone or Cc with some Ci or Cs, but the Cc being the main cirriform cloud

APPENDIX F: ENVIRONMENTAL SITE SURVEYS OF IDS ZONES

Site surveys are critical to designing and maintaining IDS-based security. An environmental site survey is a map of an IDS-protected perimeter that focuses on spatially changing topographic/climatic factors. It is a survey that requires knowledge about how IDSs operate, an understanding of how the environment operates, and keen senses of observation and intuition. Such a map can be a general map of the environment around the perimeter, simply identifying wetlands and woodlands, for example. Or the end product can be a map displaying the spatial perimeter environment as it relates to specific intrusion detection phenomenologies. The latter requires an understanding of what environmental factors affect operation of specific classes of IDSs, problems that have specifically been addressed at SOROIDS, the CRREL IDS research site, and by this report. This appendix is a guide to identifying locations within IDS zones where site conditions affecting IDS detection capability or nuisance alarm rate are likely to be different.

Site conditions along a perimeter protected by IDSs are likely to exhibit a spatial variability resulting from nonuniform exposure to the weather and differences in surface characteristics. For example, it is generally obvious where rainwater accumulates because of topography; low spots along a perimeter are probably well-identified as wet areas. Security personnel may know from experience that the detection capability or nuisance alarm rate of certain IDSs covering those wet areas is unacceptable until the surface has dried, or conversely, is better while the soil remains damp. It is less apparent that soil in the shadows of buildings or other obstructions of sunlight remains damp longer than exposed soil where evaporation proceeds at a higher rate. Usage of the soil surface also introduces spatial variability into soil properties. For example, depending on the rainfall rate, water pools or runs off as surface flow with little change in soil moisture content where the exposed soil is hard packed due to vehicle traffic, but it infiltrates adjacent soil that is both less compacted and unsaturated.

Such spatial heterogeneity in soil moisture content in turn imparts a temporal variability to changes in temperature at the surface. Soil with a higher moisture content will not undergo sensible heating from solar loading until the available moisture has been evaporated, whereas drier soil will fluctuate in temperature more rapidly in response to the diurnal cycle of solar radiation. Similarly, if there is a vegetation cover, vegetation obtaining moisture from the damp soil will transpire while vegetation on dry soil may experience stress due to insufficient moisture supply and thus undergo sensible heating. IDSs that are affected by the moisture content or the temperature of the ground surface will experience inconsistencies in performance within and between zones due to such relatively subtle spatial variability in the relevant site conditions.

Diurnal and seasonal changes in spatial patterns suggest that a site survey cannot be completed in one afternoon. Ideally, many traverses of the site throughout a year will produce the most complete site survey. In fact, site surveys could be made for individual seasons, and even for different times of the day within seasons. This could be the responsibility of security personnel on their normal patrols.

Performing a site survey

Security designers experienced with buried electromagnetic IDSs routinely require that a site survey be conducted that is directed specifically at that type of IDS. This should be done for all IDS classes. Miller et al. (1992) provide comprehensive guidance in conducting surveys directed at buried electromagnetic IDSs. They stipulate the observation and documentation of overall site characteristics such as topography, drainage patterns, and changes in soil type. This information is also a necessary part of an environmental survey identifying the spatial variability in site conditions. Following are examples of the types of information that could be noted on a base map delineating IDS zones.

Locations where exposure to solar radiation is affected by obstructions

Along a length of fence, one portion may be shadowed by a building at the same time that an adjacent portion is exposed to the sun. The sunlit fence portions, and any attached IDSs, experience changes in thermal expansion and contraction due to temperature fluctuations caused by solar warming and radiational cooling. Exposed IDSs are likely to experience greater variability in detection capability, due to differences in temperature-related coupling to the fence, and in nuisance alarm rates, because wind-induced fence motion is dependent on the temperature-related stiffness of the fence, than will sheltered IDSs. Ground or pavement in exposed locations will experience a greater range of surface temperatures, which in turn affects an intruder's thermal contrast with his background and so the probability of his detection by a passive infrared IDS.

Locations where structures create corridors that affect wind speeds

Any IDS sensitive to wind motion because of alignment requirements (beambreak IDSs, bistatic microwave IDSs), extraneous fence motion (fence-mounted IDSs), motion of nearby objects or vegetation (microwave IDSs, passive infrared IDSs, video-motion IDSs, ground-motion IDSs), or motion of surface water (microwave IDSs, buried electromagnetic IDSs) would experience location-specific nuisance alarms related to localized high winds. Windswept areas that are clear of snow experience deeper frost penetration, which would improve the detection capability of buried electromagnetic IDSs but lessen that of ground-motion IDSs. Blowing snow in high-wind areas reduces the effectiveness of surveillance cameras and video-motion detection IDSs and may cause nuisance alarms with near-infrared beambreak IDSs.

Locations characterized by differences in soil moisture

Soil moisture may vary considerably within short distances because of delayed evaporation (limited exposure to solar radiation), relatively low infiltration (soil compacted by vehicle or foot traffic), or higher snow accumulation with subsequent high water influx during snow melt. Variation in soil moisture along a perimeter may cause location-dependent differences in the detection capability and/or nuisance alarm rate of passive infrared IDSs, which are well-suited by a thermally uniform background; buried ground-motion IDSs, which are well-suited to damp, less-rigid soil; and buried electromagnetic IDSs, which are well-suited to dry, low electrical conductivity soil.

Locations where snow accumulates

Unless snow is removed, soil or pavement beneath the snow will warm because heat from the subsurface flows more slowly through the snow. Soil freezing, which impairs the detection capability of buried ground-motion IDSs, may be prevented or delayed by the presence of a snow cover. The detection capability of buried electromagnetic IDSs is better in situations of deep frost penetration with no surface thawing. Characteristics of the snow cover—depth, density, and presence of rigid freeze–thaw layers—determine whether buried ground-motion IDSs detect walking intruders. Near-infrared beambreak IDSs will generate nuisance alarms as the snow surface extends into the lowest beam. A snow cover screens crawling intruders from detection by microwave IDSs. Nuisance alarms by passive infrared IDSs are less frequent when the ground cover in the detection zone is snow.

Locations of irregular frost penetration

Among the factors influencing frost penetration are soil type and moisture content, exposure to solar radiation, presence of a snow cover, and proximity to heat sources such as building foundations. Soil freezing reduces the detection capability of buried ground-motion IDSs and improves that of buried electromagnetic IDSs.

Locations of fog formation

Dramatic spatial heterogeneity is found in fog that is produced as a result of radiational cooling on clear nights and that forms as a result of cold-air drainage. Radiation fog is typically greatest over surfaces that are well-insulated from the soil and are moist, such as fields of high grass. Radiation fog is generally less frequent over pavements where soil heat can readily heat the lower air layers. Cold-air drainage fog tends to form in low, moist areas such as over streams and bogs. Fogs not generally affected by microtopographic factors include frontal fog and advection fog. Extinction of electromagnetic radiation by water droplets affects the usefulness of surveillance cameras and video-motion IDSs (decreased visibility), passive infrared IDSs (perhaps higher inherent thermal contrast, but lower detection capability caused by reduced apparent thermal contrast at the IDS), and near-infrared beambreak IDSs (increase in nuisance alarms caused by greater transmission loss between transmitter and receiver).

Locations of seasonally changing vegetation structure and agricultural activity

The height and structure of vegetation often changes with season and with the growth and harvest of crops. Deciduous forests change leaf cover through the year, which affects radiation, thermal, moisture, and wind patterns. Annuals grow and die, changing height and density over time, which in turn affects wind speed and gustiness near fences. Agricultural activity has the potential for dramatic changes. Corn crops near an IDS-protected fence may grow to over 2 m in height. Cutting the corn stalks reduces the roughness of the field and increases wind speeds, a change that can occur in less than 24 hours. This suggests that a site survey should include the conditions and activities of land adjacent to the IDS facility in addition to the land it is located upon.

Locations of changing wind patterns

Wind speed and direction may change diurnally and seasonally at many locations. This can cause winds to approach IDS zones at different speeds and from different directions at different times of the day or year. For example, an IDS zone located at the mouth of a valley may experience diurnally reversing wind directions. Some regions have monsoon winds, seasonally changing in direction and strength. In addition, diurnal wind speed increases may occur at midday as a result of convective heating patterns. Wind loading affects security fences and fence-mounted IDSs as well as any aboveground IDSs for which loss of alignment is a problem (near-infrared beambreak and bistatic microwave IDSs). Wind-induced motion of objects in their detection zones causes nuisance alarms by video-motion IDSs (variation in brightness within the camera scene), by passive infrared IDSs (variation in thermal radiance if the moving object is not uniformly the same temperature as its background), microwave IDSs (changes in the amount and direction of radiation reflected from the moving surfaces), buried ground-motion IDSs (motion of surface objects is coupled into the ground), and buried electromagnetic IDSs (variation in aboveground radiation field by moving metallic objects or water).

Bibliography

These sources may assist in conducting site surveys, in addition to the microclimatic references in the main reference list.

Geiger, R. (1951) Microclimatology. In *Compendium of Meteorology* (Thomas F. Malone, Ed.), p. 993–1003.

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Landsberg, H. and W. Jacobs (1951) Applied climatology. In *Compendium of Meteorology* (Thomas F. Malone, Ed.), p. 976–992.

Miller, C., C. Malone and C. Blount (1992) Guide for the conduct of predeployment site characterization surveys for the AN/GSS–34(Y) Ported Coaxial Cable Sensor (PCCS). USAEWES Instruction Report EL–92–1.

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APPENDIX G: WEATHER AND CLIMATE INFORMATION SOURCES

Weather and climate information are available from many sources. This appendix lists sources of daily weather forecast and observation information current at time of publication, from which to derive expected weather effects on IDS reliability as well as sources of archived climatological data for IDS-based security planning. We make suggestions about what kinds of information to ask forecasters and climatologists about IDS sites when designing and operating them. Finally, several references are cited that will help you install a weather/environment monitoring system to assess changes in conditions as they occur.

National and international weather agencies

The National Climate Data Center in Asheville, North Carolina, is the nation's archive for past weather and climate information. Data are available for all U.S. territory in paper or digital form. Civilians should call (704) 259-0682 for a copy of the *Selective Guide to Climatic Data Sources,* NCDC's comprehensive catalog of products. Assistance about product availability is also available by telephone. Military users should contact the Environmental Technical Applications Center (ETAC) at (704) 271-4218 or fax (704) 271-4334.

One of the goals of the World Meteorological Organization (WMO), headquartered in Geneva, Switzerland, is to further the application of meteorology to aviation, shipping, water problems, agriculture, and other human activities. Information about the organization exists in more detail in its quarterly publication, the *WMO Bulletin.* Copies may be obtained from:

The Secretary-General
World Meteorological Organization
Case postale 2300
CH–1211 Geneva 2
Switzerland

Journals

Weatherwise, a journal affiliated with the American Meteorological Society, explains weather and climate phenomena for professional meteorologists and the lay audience. For information, contact:

Weatherwise
Heldref Publications
1319 Eighteenth Street, NW
Washington, DC 20036-1802
(800) 365-09753

Bulletin of the American Meteorological Society, a monthly journal, contains research articles about weather forecasting and many ads for weather forecasting services and instrumentation. For information see below.

Journal of Climate is a journal of the American Meteorological Society that contains research papers about general climate and microclimatic problems. For information see below.

Journal of Applied Meteorology is another journal of the American Meteorological Society containing research papers about general weather, climate, and microclimatic problems, with an emphasis upon applied problems.

Executive Director, AMS
45 Beacon Street
Boston, MA 02108-3693
(617) 227-2425

Laboratories

The U.S. Army Corps of Engineers Cold Regions Research and Engineering Laboratory is the largest facility in the Western Hemisphere devoted to the study of cold region and cold season phenomena. Though some climatic and weather information are available at the lab, its primary aid is as a source of information for weather and climate problems in cold regions and for information on how to monitor weather in cold regions. For additional information contact:

USA CRREL
72 Lyme Road
Hanover, NH 03755-1290
(603) 646-4100

Consulting meteorologists and climatologists

A list of professional consulting meteorologists certified by the American Meteorological Society is listed in the back of each issue of the *Bulletin of* *the American Meteorological Society.* Each listing includes a description of the type of services provided.

Universities are a source of climatologists and meteorologists for consulting. Meteorology, geography, physics, forestry, and botany are all departments often employing such professionals.

Every state has a State Climatologist, who may be located through the Association of State Climatologists or through the local National Weather Service Office. Contact the American Association of State Climatologists at:

American Association of State Climatologists
Pam Naber Knox, State Climatology Office
University of Wisconsin
1225 Dayton St.
Madison, WI 53706
(608) 263-2374

Federal and state agencies such as forestry programs, agriculture, aviation agencies, and specialty labs like CRREL often employ meteorologists.

Books

See the bibliography of this report for comprehensive reference works.

In addition, see MIL-STD-210C, 9 January 1987, *Climatic Information to Determine Design and Test Requirements for Military Systems and Equipment.*

Sources of daily weather observations and forecasts for IDS operations

The U.S. Navy Fleet Numerical Oceanography Center distributes weather maps through its Navy Oceanographic Data Distribution System (NODDS). Maps, satellite imagery, and radiosonde flights may be copied from the system via modem using software available from NODDS. The service is available only to U.S. government agencies and their contractors. For more information, contact:

Commanding Officer
Fleet Numerical Oceanography Center
ATTN: Code 342
Monterey, CA 93943-5005
(408) 656-4431

The Weather Channel is a commercial cable television network devoted to presenting current and forecast national weather. See local cable listings and *Weatherwise*, vol. 45, April/May 1992, p. 9–15, and vol. 35, August 1982, p. 156–163.

Internet weather bulletin boards

Several universities provide current weather analyses on computer bulletin boards and on the World Wide Web. Most files can be copied directly to a personal computer.

Miscellaneous

Refer to the advertisements in the *Bulletin of the American Meteorological Society* and *Weatherwise* for addresses and telephone numbers of distributors of weather instrumentation and information.

A new organization devoted to observing weather, International Weather Watchers, may be able to provide assistance with data and observing problems in your area. They publish *Weather Watchers Review*. Contact the organization at:

International Weather Watchers
P.O. Box 77442
Washington, DC 20013

What to ask for in forecasts and services

This report emphasizes change in micrometeorological conditions as the cause of variability in IDS reliability. Systems respond to intruders because intruders cause changes in what the IDS is monitoring. Since weather can cause similar changes, IDSs respond to weather. An IDS that is insensitive to changes in its operating conditions may also be insensitive to intruders. Security managers should be prepared to look for weather forecasts that indicate whether changes relevant to their IDSs are going to occur. These may be indications of frontal passage, seasonal change, and/or the intrusion of high pressure with clear skies that encourages some weather conditions, such as air temperature, to change over a wide range diurnally.

Unfortunately, weather forecasters do not observe or forecast all site conditions that affect IDS reliability. Therefore, either security personnel or weather consultants must interpret standard forecasts to provide information about related micrometeorological changes that might affect IDSs, such as soil moisture, leaf drop, freeze–thaw, snowmelt, needle ice, and hoarfrost. In addition, security personnel must be aware of the changing position of the sun with time of day through the seasons and the impact that these changes in the energy budget have upon the IDS environment.

How to install a weather station

Security managers who wish to install weather instruments must seek advice on how to select and site instruments and make subsequent observations. Though it may be desirable to install a tradional weather station configured to be useful for weather forecasting, traditional weather measurements may not adequately represent site conditions that affect IDS reliability. Therefore, measurements should be made perhaps at several locations near IDSs. Shielded thermistors can be placed on fences at various locations and heights and at various depths within the soil around the perimeter. Anemometers can be placed on fence posts, and soil moisture can be measured around the fence perimeter. Positioning of instruments will be unique for each site and IDS installation, so operators must be both informed about how to properly select and site weather instruments and creative with respect to placing them where they will be of greatest use. Weather measurements require considerable effort and cost, because instruments must be purchased, installed, and maintained, and data must be recorded, quality checked, and analyzed. Therefore, weather monitoring systems must be carefully designed and justified before any installation occurs.

The most useful sources for installing a traditional weather station are National Weather Service Science and Operations Officers, military meteorologists and technicians, and university meteorologists. Universities, and perhaps U.S. Department of Agriculture Experiment Station scientists, may be the most useful sources for help with micrometeorological measurements. Other sources of more traditional observing information are the World Meteorological Organization and International Weather Watchers. See above for addresses.

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These publications explain how to select and site instruments. See references in these sources for additional information.

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13. ABSTRACT (*Maximum 200 words*)

Fundamentally, an electronic exterior intrusion detection system (IDS) cannot directly detect intruders; it detects a variation in the condition being monitored, extracts characteristics of that variation, and assesses whether such a variation probably is caused by an intruder. However, exterior IDSs do not operate in a benign natural environment. Their environment is constantly changing as a result of solar-driven energy and moisture fluxes that create the weather. These weather changes often cause variations in the conditions being monitored by IDSs. The challenge, therefore, is to recognize how and when IDSs are responding to some change in their natural environment, rather than to intruders.

This report is a technical analysis of causes of weather-driven temporal changes in the environment that impact the operational efficiency of IDSs. The report is intended to assist security designers in selecting suitable IDSs for a site and to assist security managers in operating IDSs at the required level of reliability. This is accomplished by identifying temporal variations in weather that are sufficiently general to be identified as patterns, and by identifying how different IDS phenomenologies respond to these patterns. The result is an understanding of how weather conditions influence the ability of types of IDSs to detect reliably activities representative of an intruder while successfully discriminating against weather-created conditions within a detection zone.

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The main body of the report is organized by temporal scale: diurnal, quasi-periodic, and seasonal. Within each temporal scale, weather processes common at that scale are explained Topics covered include air and soil temperature, soil moisture, precipitation, snow cover, winds, fog, storms, urban and topographic effects, vegetation effects, and solar radiation. Appendixes include a summary of weather impacts on IDSs, extensive indexes by IDS and weather topics, cloud and fog illustrations, explanations of the relationship of weather patterns to forecasts, suggestions about how to perform environmental site surveys, and sources of additional weather information.

14. SUBJECT TERMS

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Beambreak
Climate
Countermeasure
Detection capability
Diurnal
Electromagnetic
Energy budget
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Fence-mounted
Fog
Fronts
Geography
Icing
Intrusion detection
Meteorology
Microwave radar
Nuisance alarms
Optical-fiber
Passive infrared
Precipitation
Seasonal
Security fences
Snow cover
Soil freezing
Soil moisture
Storms
Taut-wire
Temperature
Thermal infrared
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