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Mapping Regional Freeze/Thaw Patterns with Satellite Microwave Radiometry

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I Products of this investigation

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II Project objectives

Soil moisture contributes to the energy exchange between the air and ground through latent heats of fusion and vaporization. Whether as boundary conditions for mesoscale climate modeling, or as inputs to an agricultural productivity model, the amount and state of soil moisture are regional parameters that one would like to estimate through satellite remote sensing. There is a large body of literature that addresses the estimation of soil moisture from remotely sensed radiobrightness. The primary objective of this project has been to develop an operational technique, or algorithm, for producing a temporal sequence of maps showing the seasonal distribution of frozen soil, i.e., the state of soil moisture, in the northern Great Plains. These maps would be based upon data from the Defense Meteorological Satellite's Special Sensor Microwave/Imager (SSM/I).

As an unanticipated product of this project, we discovered what may be an effective alternative method for estimating the amount of soil moisture in thawed, prairie terrains. Our models show that day-night differences in radiobrightness are a more sensitive measure of thermal inertia than are day-night differences in thermal infrared brightness, and that thermal inertia, as previously recognized by the thermal infrared community, should be a sensitive measure of the moisture content of soils. Therefore, the testing of our radiobrightness thermal inertia models became an additional objective of the project.

III Review of progress

Continental atmospheric circulation models are driven by radiant energy, sensible heat, latent energy, momentum, and moisture flux boundary forcing at the land-atmosphere interface (e.g., Bhumralkar, 1976; Wilson and Henderson-Sellers, 1985; Abramopoulis et al, 1988; Verstraete, 1989; Dickinson et al, 1989; Avissar and Verstraete, 1990; and Giorgi and Mearns, 1991). There is ample evidence that mesoscale and global atmospheric circulation are particularly sensitive to the hydrologic boundary forcing of moisture and latent energy flux (e.g., Namias, 1958 and 1963; Walker and Rowntree, 1977; Rind, 1982; Shukla and Mintz, 1982; Yeh et al, 1984; Oglesby and Erickson, 1989; and Delworth and Manabe, 1988 and 1989). Hydrologic parameters that are interactive with circulation models have evolved from the bucket model of Manabe (1969) and Budyko (1974) to complex parameterizations like the Biosphere-Atmosphere Transfer Scheme (BATS) (Dickinson, 1984; and Dickinson et al, 1986) and the Simple Biosphere Model (SiB) (Sellers et al, 1986). While BATS involves more than 80 parameters and SiB involves 44, many are predictable for large, relatively homogeneous regions like the northern prairie or the arctic tundra. For example, soil thermal conductivity, density, and specific heat are readily estimated for either environment.

However predictable many of the hydrologic parameters are, the dynamic hydrological variables must be obtained through observation. Of these variables, soil moisture is critical because it represents the integrated effects of rainfall, runoff, and evapotranspiration. Because soil moisture tends to be relatively slowly varying with time at mid- and high-latitudes (e.g., Yeh et al, 1984; and Delworth and Manabe, 1988), its estimation is particularly suited to daily, or even to weekly, satellite observation. Given that some satellite observation might yield an estimate of soil moisture, producing such estimates on something like the 25 km spatial grid of a mesoscale atmospheric circulation model is relatively undemanding of sensor spatial resolution or of satellite data management systems.

Estimates of soil moisture and state would be based upon a combination of satellite observations and a model that places the observations in a physical or temporal context. For example, we expect L-band (1.4 GHz) radiobrightness to correlate well with soil moisture in situations where a physical model of emission from a moist soil halfspace is plausible (e.g., Schmugge, 1980). Complications caused by rough surfaces, by vegetation, or by non-uniform distributions of moisture in the soil are viewed as perturbations to the basic halfspace model. Without such physical models, we cannot readily extrapolate to untested situations. As models become more sophisticated, they might include the dynamical effects of diurnal and annual insolation, of vegetation, and of snow. The Michigan Cold Region Radiobrightness (MCRR) model is a moist soil radiobrightness model that includes the effects of diurnal and annual insolation and the effects of freezing and thawing soils. This model has provided a basis for classifying soils as frozen or thawed and has suggested that soil moisture might be inferred from the diurnal extremes in radiobrightness. Because the MCRR model does not yet include the effects of vegetation or snow, it is not yet suitable links between radiobrightness observations and the land-atmosphere fluxes that drive the continental boundary layer of a General Circulation Model (GCM). That is, the current MCRR model is more a qualitative than a quantitative model.

The Michigan Cold Region Radiobrightness (MCRR) Model

Because so little radiant energy is transferred at microwave frequencies, the problems of thermal modeling and of microwave emission modeling are readily separable. There are several models for diurnal, near-surface soil temperature. These include the Watson model (1975) for diurnal heating of dry soils, the Kahle model (1977) for diurnal heating of moist soils, and the early MCRR model (England, 1990) for diurnal heating of moist soils that are subject to freezing and thawing. The Watson and Kahle models predict thermal infrared signatures, while the MCRR model predicts radiobrightness signatures. We recently expanded the MCRR model to an annual model for dry soils (England and Liou, 1992), and, more significantly, to an annual model for wet, freezing, and thawing soils (Liou and England, 1992b).

The thermal module of the MCRR model solves the one-dimension heat flow equation in terms of soil depth, z, and time, t,

$$\frac{\partial E(T(z,t))}{\partial t} = \frac{\partial}{\partial z}\left(K(T(z,t))\frac{\partial T(z,t)}{\partial z}\right) \tag{1}$$

for a moist soil halfspace. Both E(T), moist soil enthalpy, and K(T), moist soil thermal conductivity, are functions of thermal temperature, T. Permitting the moisture in soil to freeze and thaw means that the heat flow equation cannot be linearized. The problem is particularly difficult because phase boundaries propagate in time, and because soils, particularly clay-rich soils, freeze over a range of temperatures rather than at 0^o C. That is, the phase boundary is diffuse. We employ a modified Chernous'ko method (England, 1990)—a finite element method that tracks isotherms within the soil—to achieve a solution to equation (1). The model typically converges to within 0.01 K after 5 iterations for all 10 minute intervals of an annual cycle. Model predictions are tested by their reasonableness and, for non-freezing soils, by comparison with numerical solutions to a variable time interval form of the Laplace analytical method (Liou and England, 1992a).

Figures 1 and 2 show examples of solutions for prairie soils near Bismarck, North Dakota. The 7% moist soil is atypically dry while the 25% moist soil is atypically wet. We note that diurnal extremes in soil temperature are relatively sensitive to moisture content but annual extremes are not. Moisture dependent anomalies in the annual cycle occur during fall freezing and spring thawing.

The radiobrightness module of the MCRR model is a first-order approximation to emission from a semi-transparent halfspace whose temperature varies with depth (e.g., England 1990). Radiobrightness with time, t, becomes

$$T_b(t) = e(t)\left\{T(0,t) + \mu(t)\eta(t)\left(\frac{\partial T}{\partial z}\right)_0\right\} \tag{2}$$

where radiobrightness, T_b, and emissivity, e(t), are implicit functions of polarization (V or H) and incidence angle. T(0,t) is thermal temperature at the soil surface, $\mu(t)$ is the direction cosine with respect to vertical of the ray in soil, $\eta(t)$ is optical depth, and $\left(\frac{\partial T}{\partial z}\right)_0$ is the thermal gradient at the surface. Emissivity is based upon Fresnel coefficients for a plane interface. Permittivity of the soil varies with moisture content and temperature so that emissivity, ray direction, and optical depth become time dependent. The model predicts V- and H-polarized radiobrightness for the incident angle and frequencies of the Nimbus 7 Scanning Multichannel Microwave Radiometer (SMMR), (50^o and 10.7, 18.0, and 37.0 GHz [Gloerson and Hardis, 1978]), or the Defense Meteorological Satellite's Special Sensor Microwave/Imager (SSM/I), (53.1^o and 19.35, 37.0, and 85.5 GHz [Hollinger et al, 1987]).

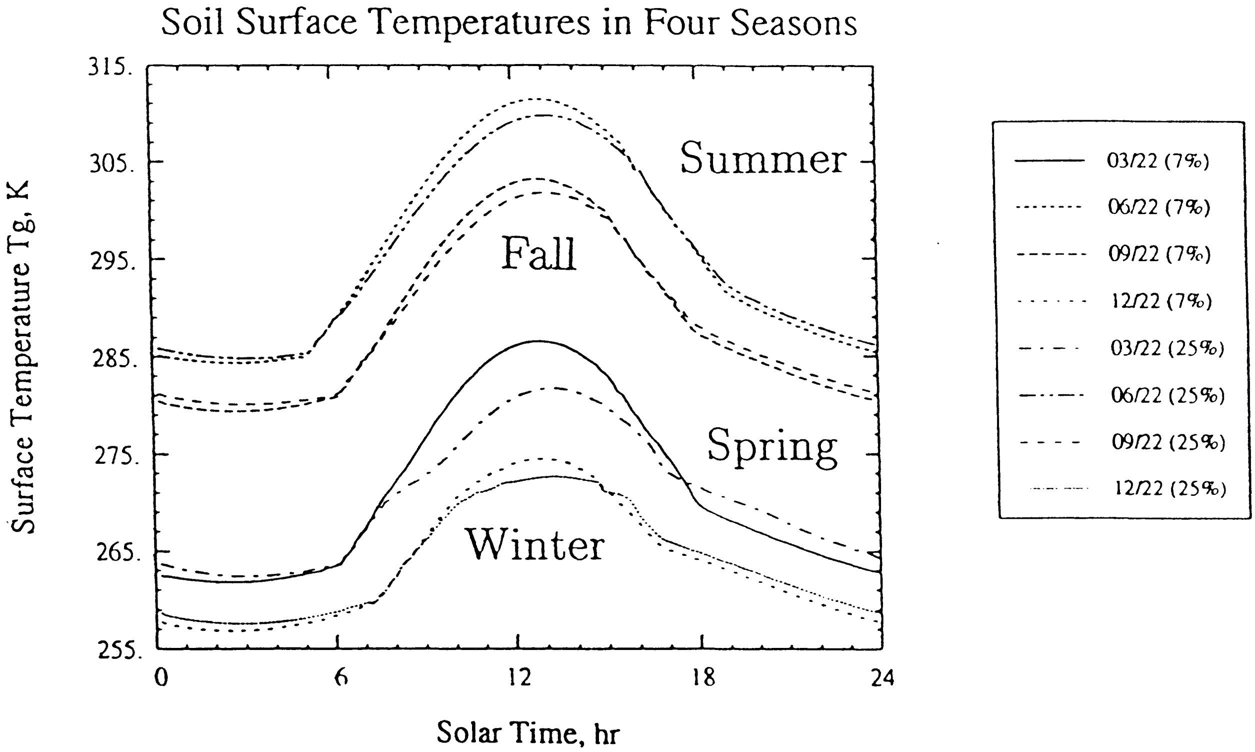

Fig 1. Predicted diurnal surface temperature for prairie near Bismarck, North Dakota.

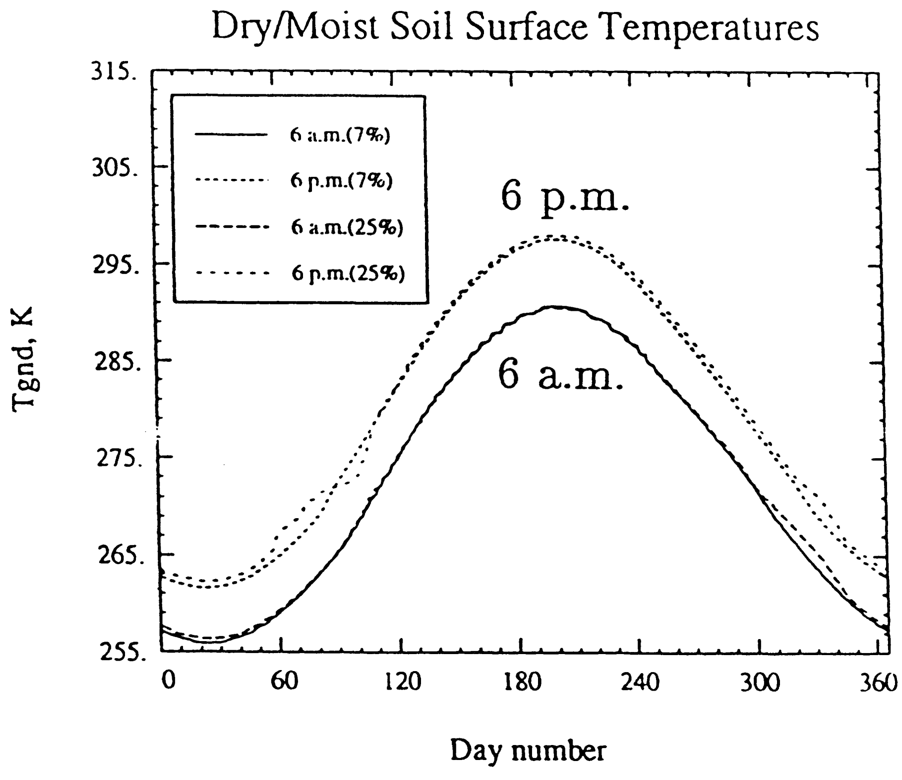

Fig 2. Predicted annual surface temperature for prairie near Bismarck, North Dakota.

Figure 3 shows predicted diurnal 19 GHz, H-polarized radiobrightness. Note that moisture causes an extreme variation for winter and spring days when freezing and surficial thawing occur. Figure 4 shows predicted annual radiobrightness at 6:00 a.m. and 6:00 p.m.—times of SSM/I overflight. The precipitous decrease during spring thaw or increase during fall freeze should be evident in the SSM/I data.

Figure 5a shows SSM/I-derived, 19 GHz, H-polarized, radiobrightness averages during August through December, 1988, for a 100 km square region centered on Fargo, North Dakota. Radiobrightness was estimated by correcting 6:00 a.m. SSM/I scene temperatures for a standard atmosphere using differences between 19.35 and 22.235 GHz data, and assuming a water vapor scale height of 3 km. A 7-day running boxcar filter was used to reduce daily scatter. Precipitation, snowpack thickness, and minimum and maximum air temperatures also appear in Figure 5. Qualitatively, rainfall in mid-September correlates with reduced radiobrightness during the second half of September; the snowpack during the second half of November correlates with the anomalously low radiobrightness during the same period; low air temperatures beginning in mid-November and extending through December correlate with the rise in radiobrightness during December; and the new snowpack in late December correlates with the precipitous drop in radiobrightness in late December. We note that the period of cold temperatures in December that would freeze snow-free soil does correspond to an increase in radiobrightness as predicted by the model. We also note the scatter darkening in dry snow during November and during late December.

Snow and of vegetation must be incorporated in the MCRR model. With vegetation comes evapotranspiration, scattering, and emission, and with snow comes scattering. We will borrow from the GCM land-atmosphere parameterizations like BATS to incorporate evapotranspiration. Many volume scattering models might be applied to microwave emission from snow (e.g., Gurvich et al, 1973; England, 1975; Tsang and Kong, 1976; Fung and Chen, 1981; and Tsang and Ishimaru, 1987), and from grass and sedge vegetation (e.g., Lang and Sidhu, 1983; Ulaby et al, 1990a; and Sarabandi et al, 1990). The simplest approach for vegetation would be to treat grass blades as independent scatters. The most complex would be to develop scattering matrices for leaves and stems based upon their dielectric properties, size, orientation, and distribution, and incorporate these scattering matrices in a multi-layer, radiative transfer model, e.g., the Michigan Microwave Canopy Scattering Model (MIMICS) (McDonald et al, 1989, and Ulaby et al, 1990b). The independent scatterer model would be unrealistically simple, but the MIMICS model may be unnecessarily complex. MIMICS was designed to predict radar backscatter where scattering is the dominant process. Emission, absorption, and scattering are equally important to canopy radiobrightness so that a true radiative transport formulation may be unnecessary. We are currently developing a strong fluctuation model for emission from prairie grass (e.g., Tsang and Kong, 1981). Dry snow is not absorbing so that one of the many radiative transfer models must be used. There is, as yet, no adequate model for wet snow.

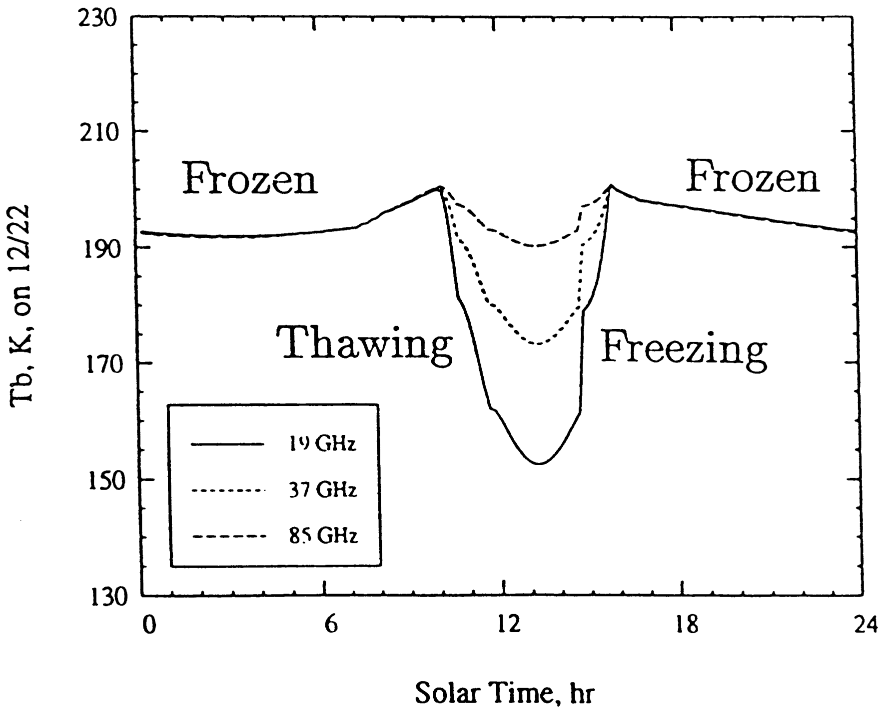

Fig 3. December 19.35, 37.0, and 85.5 GHz, H-pol diurnal radiobrightness of a 25% moist soil.

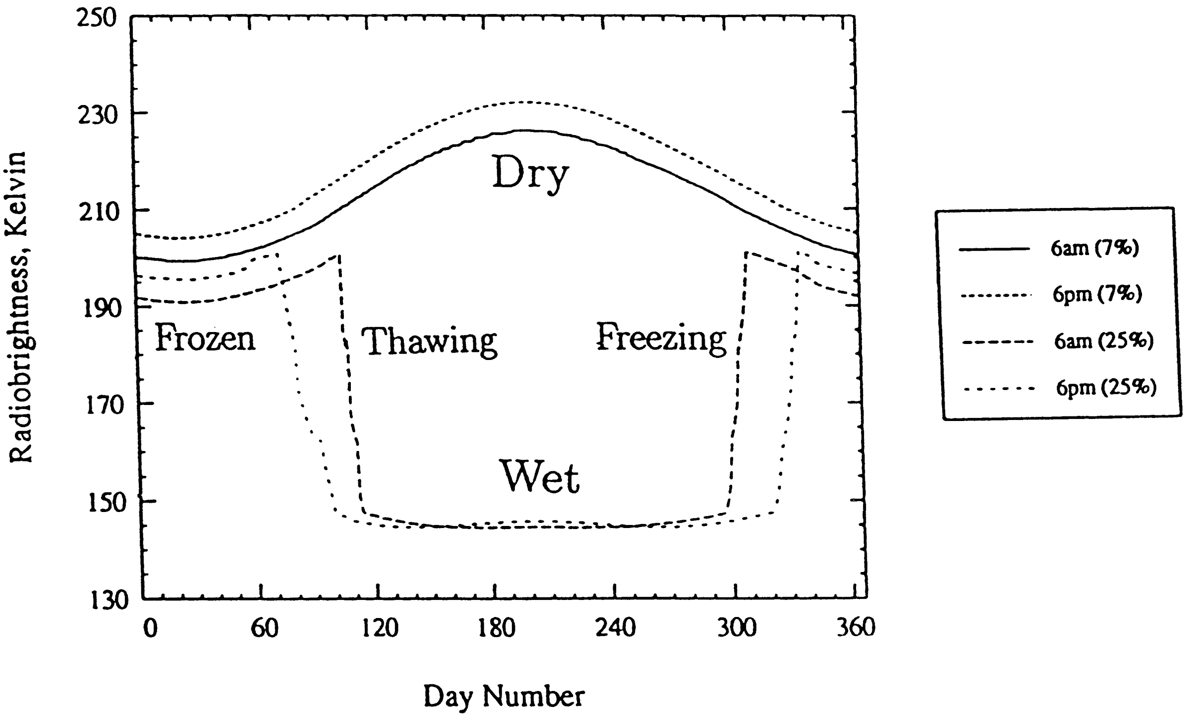

Fig 4. 19.35 GHz, H-polarized annual radiobrightness.

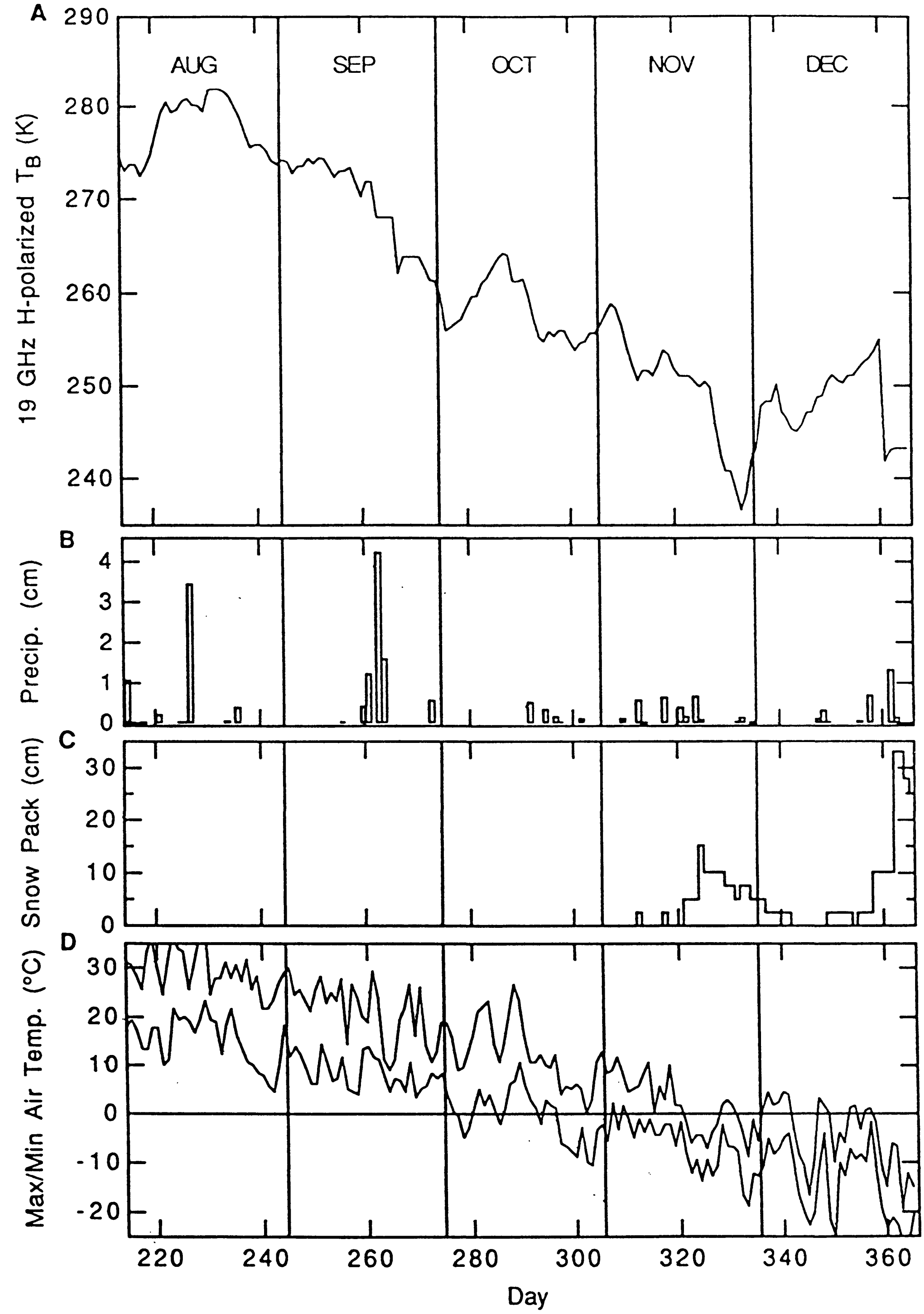

Fig 5. (a) 19 GHz, H-polarized, 6:00 am, SSM/I radiobrightness near Fargo, North Dakota (b) daily precipitation, (c) snow pack depth, and (d) high and low air temperatures, August through December, 1988.

The Remote Measure of Soil Moisture

Moisture influences soil reflectivity and emissivity at all wavelengths so that, potentially, there are several remote sensing methods that might be used to infer moisture content or state. For example, soil wetness governs the visible reflectance of bare soil (Idso et al, 1975a). However, such estimates are limited by a dependence on soil type, by surficial drying, by vegetation cover, and by clouds.

Radiobrightness is sensitive to soil moisture in prairie and arctic tundra—regions of relatively short vegetation—through the dominant influence of liquid water upon microwave emissivity in soil and in the grass. The Debye relaxation of liquid water (12 GHz at 280 K [Hasted, 1972]) causes the microwave emissivity of moist soil to increase with frequency, f, so that the radiobrightness spectral gradient, $\frac{\partial T_b}{\partial f}$, becomes increasingly positive with moisture content. Single frequency estimates of moisture content are possible at microwave frequencies near to or below the Debye relaxation frequency if the thermal temperature of the soil is known. L-band, single frequency estimates are least prone to errors caused by vegetation. Volume scatter darkening in senescent grass causes a slightly negative bias in the 1-30 GHz spectral gradient (England et al, 1991).

Vegetation is a primary physical link between soil moisture and the atmosphere in northern prairie. Any correlation between soil moisture and atmospheric moisture would involve the integrated moisture content from the soil's surface through the root zone of its vegetation at a spatial scale which is smaller than a drainage basin (significantly less than 100 km). High correlation between radiobrightness and root zone moisture occurs at frequencies below 5 GHz (e.g., Burke et al, 1979; and Wang et al, 1982), but moderate spatial resolution at frequencies below 5 GHz from satellite altitudes requires large antennas. This combination of frequency and spatial resolution from satellites has not been achieved with passive imaging systems. For example, the lowest frequency and spatial resolution of the SMMR was 6.6 GHz and 150 km, and that for the SSM/I is 19.35 GHz and 43 km. At these frequencies, scattering and emission by plant canopies certainly competes with soil moisture in governing the spectral gradient.

Radar measures of soil moisture have been examined by, for example, Ulaby and Batlivala (1976), Blanchard and Chang (1983), and McDonald et al (1988). When compared to imaging radiometers, operational satellite radars, like ESA's ERS-1, the Japanese JERS-1, and, eventually, the Canadian Radarsat, offer the advantages of much better spatial resolution (typically a few tens of meters), and measurements that are, to first order, independent of thermal temperature. Their disadvantage includes significantly greater cost for both the spaceborne system and for subsequent data processing. Beyond economics, radar differs from radiometry in that radar backscatter is more strongly influenced by scattering in the plant canopy and by rough soil surfaces. If canopy over soil is pictured as a scattering layer over a roughsurfaced halfspace, then a radar signal has to pass through the scattering layer at least twice in the process of being reflected by the soil. Radiation emitted by the soil passes through the scattering layer only once. The difference is analogous to viewing an object through frosted glass when the light is on the viewer's side (radar), in contrast to viewing the object when the light is on the object side (radiometry).

Differences caused by surface roughness can be even more striking. For example, if a vegetation-free, moist soil were effectively a homogeneous, quasi-specular halfspace at the microwave wavelength, then there would be no radar backscatter at off-nadir incidence for any moisture content. The equivalent radiobrightness would decrease monotonically with increasing wetness. Radar and microwave radiometry are complementary. Radar should provide a better estimate of biomass and radiobrightness is more usefully sensitive to soil or plant moisture.

There are alternative methods for obtaining soil moisture from satellite imaging radiometers. Moisture increases the apparent thermal inertia of soil by increasing its thermal conductivity, density, and specific heat, and by daytime cooling through evapotranspiration and nighttime warming through condensation. That is, as the moisture content of soil increases, its day-night difference in thermal temperature tends to decrease, and, consequently, its day-night difference in radiometric brightness also decreases. These effects have been examined in the thermal infrared spectrum (e.g., Idso et al, 1975b; Reginato et al, 1976; Price, 1980; Heilman and Moore, 1980; and Vleck and King, 1983), and were the basis of the Heat Capacity Mapping Mission (HCMM) (e.g., Heilman and Moore, 1981 and 1982), a thermal infrared experiment that, in part, used differences in the near-surface storage of moisture to discriminate among various rock and soil types.

The thermal microwave day-night signature will exceed the equivalent thermal infrared signature because soil moisture reduces microwave emissivity, but increases (slightly) thermal infrared emissivity. While a thermal infrared technique more easily achieves high spatial resolution, the twice-daily coverage required by either thermal inertia method favors a radiobrightness technique for its tolerance to cloud cover. In essence, the Radiobrightness Thermal Inertia (RTI) method uses the diurnal thermal pulse to probe the soil. Its effective penetration depth comes from the thermal pulse while the radiometric sensitivity comes from thermal and dielectric changes in the surficial soils. RTI is a product of the MCRR model (England et al, 1992). Predicted sensitivities to soil moisture are shown in Figure 6.

The RTI method is particularly well suited to satellite remote sensing by Sun-synchronous satellites because they provide successive day and night observations for properly phased orbits. Overflight times of SMMR were local midnight and noon which are nearly optimum for observing maximum day-night temperature differences. Overflight times of the current SSM/I are 6:00 a.m. and 6:00 p.m. which are particularly poor for RTI because they correspond to thermal crossover.

Vegetation will tend to mask both thermal infrared and microwave soil signatures. Diurnal variations of moisture within extensive plant canopies have complex signatures of their own (e.g., Burke and Schmugge, 1982; and Wang, 1985). Additionally, there are diurnal changes in the vertical distribution of moisture in soil. Njoku and O'Neill (1982) have investigated the diurnal biases of single frequency measures of soil moisture caused by diurnal variations in the effective emission depth for frequencies of 0.6-0.9, 1.4, and 10.7 GHz. At the SSM/I frequencies of 19.35, 37.0, and 85.5 GHz, penetration is slight, but such variations may be important. The effects of vegetation—volume scattering, emission, and absorption—are expected to be less severe in prairie or in arctic regions where vegetation is short or sparse, or when there is little moisture in the canopy. Because RTI uses the difference between day and night radiobrightness, an underlying dependence upon moisture should emerge through a bias caused by the canopy if the bias is relatively constant over time. In fact, day-to-night effects are not constant because canopy moisture varies diurnally.

The RTI measure of soil moisture was tested in our Prairie Experiment near Sioux Falls, South Dakota, during the fall and winter of 1992-93. The data have not yet been analyzed.

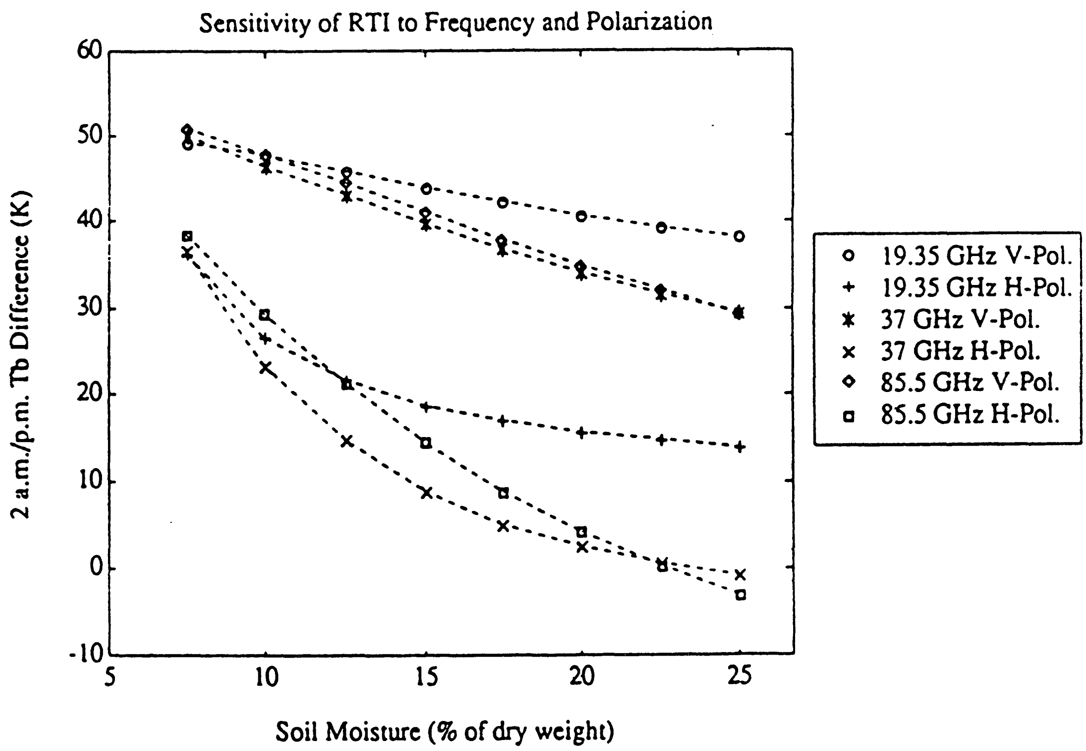

Figure 6. Sensitivity of RTI to soil moisture (England et al, 1992). These are summertime day-to-night differences derived from the MCRR model.

Remote Classification of Frozen Soils

The radiobrightness signatures of frozen and thawed soils were first studied theoretically and through satellite observations by Zuerndorfer et al (1989 and 1990), and experimentally by Wegmüller et al (1989). The satellite observations refer to data from SMMR for a test area that included North and South Dakota, and parts of surrounding states and southern Canada.

Classification of frozen and thawed soils is possible because moisture state in soil is expressed through unique radiometric characteristics. Diurnal radiobrightness at the SMMR frequencies of 10.7, 18, and 37 GHz were computed for one day of each month. The MCRR/Diurnal model predicted that the 37 GHz radiobrightness would best track soil surface temperature, the 10.7-37 GHz spectral gradient of thawed soil would be strongly positive, the spectral gradient of frozen soil would be slightly negative, and the daytime spectral gradient of frozen soil would be more positive than the nighttime gradient. SMMR data for relatively snow-free days confirmed these predictions except that the observed frozen soil spectral gradients were often strongly negative (Fig 7). England et al (1991) explained the observed negative gradients as being caused by increased scatter darkening in frozen soils and vegetation at higher microwave frequencies.

The decision criteria for classifying frozen soil were empirically derived from migrating means clustering of the noon and midnight SMMR data shown in the scatter diagrams of Figures 7a and 7b. These data are from sixteen SMMR passes over the Dakotas during August through December, 1984. The 18 and 37 GHz radiobrightness data were resolution compensated to the (coarse) resolution of the 10.7 GHz channel before computing spectral gradients. Spectral gradients were obtained from linear regressions of radiobrightness versus frequency. Data at all frequencies were averages of vertical and horizontal polarization. Surfaces were classed as frozen, wet, hot, or mixed. There were few wet surfaces at midnight during the test period so that wet and mixed surface types became inseparable. Frozen surfaces were characterized by relatively low spectral gradients and low 37 GHz radiobrightnesses. Wet surfaces were characterized by high positive spectral gradients and low 37 GHz radiobrightnesses. Hot surfaces exhibited high 37 GHz radiobrightnesses but "dry", nearly neutral spectral gradients.

Through Mahalanobis minimum distance classification (maximum likelihood classification) based upon assumed bivariant normal distributions, we obtained the cluster centroids shown in Table 1. Constant-deviation, single-class ellipses were drawn in decision space for frozen, wet, and hot surfaces (at noon) and for frozen and hot surfaces (at midnight). The freeze/thaw criteria were obtained by allowing the deviation of these ellipses to expand equally until all ellipses intersected as shown in Figures 7a and 7b. The corresponding freeze/thaw criteria appear in Table 2. Note that the gradient criterion was more positive at noon than at midnight as predicted by the MCRR/Diurnal model.

Table 1. Cluster centroid in decision space				
	NOON DATA		MIDNIGHT DATA	
Surface Type	37 GHz (K)	Gradient (K/GHz)	37 GHz (K)	Gradient (K/GHz)
Frozen	227	-.43	234	-.35
Hot	277	0.11	258	-.01
Wet	238	0.37		
Mixed	250	0.14	243	0.015

If the freeze/thaw criteria from clustering are viewed as initial estimates, then they can be refined for greater self-consistency by requiring minimum scatter of the 37 GHz radiobrightness, Tb(37), along freeze/thaw boundaries. The process involves adjusting the 37 GHz freeze/thaw criterion, T37, to minimize the sum square error, SSE,

$$SSE = \sum_{i} [Tb_i(37) - T37]^2 \tag{3}$$

where subscript i refers to the i^{th} pixel along a freeze/thaw boundary. Refined T37 are also shown in Table 2. The process is first-order since we do not iterate SSE minimization with refined criteria. An interpreted history of ground-freezing for the Dakota region during the fall of 1984 appears in a paper by Zuerndorfer and England, 1992.

Table 2. Freeze/thaw criteria in decision space; σ are standard deviations of the data within the ellipses.				
		Refined		Deviation σ
at				
	37 GHz(K)	37 GHz(K)	Gradient(K/GHz)	Intersection
Noon	252	249	0.0625	3.1
Midnight	247	244	-0.044	2.55

The MCRR model, the satellite observations, and the experimental data support use of SMMR's 37 GHz radiobrightness and its 10.7-37 GHz spectral gradient as discriminants in frozen soil classification of high latitude prairie. A similar argument applies to the SSM/I's 37 GHz radiobrightness and its 19.35 - 85.5 GHz spectral gradient (Dahl et al, 1992). The thawed soil gradient is less positive because the lowest SSM/I frequency lies well above the Debye relaxation frequency of liquid water. In contrast, the frozen soil spectral gradient is more negative because of increasingly severe scatter darkening at the higher SSM/I frequencies. An example of the sensitivity of the SSM/I data to frozen soils is shown in Figures 8a and 8b. These are atmospherically corrected observations near Fargo, North Dakota, for the fall and early winter of 1988. Radiobrightnesses at 37.0 and 85.5 GHz were spatially filtered to achieve an effective beam pattern equivalent to that of the 19.35 GHz channel. Small dots in Figures 8a and 8b represent all data points. The squares in

Figure 8a correspond to periods that are known from meteorological reports to be snow-free and cold enough for frozen soil. The crosses in Figure 8b correspond to periods known to have greater than 2 cm of snow and air temperatures cold enough to expect the snow to be dry. We include both figures to emphasize that frozen soil and dry snow appear radiometrically similar and, consequently, may be misclassified.

The boundaries in the SSM/I decision space between frozen and thawed soil shown in Figure 8 are based upon meteorological data. A better estimate of these boundaries will be a product of our Prairie Experiment. Figures 9a, 9b, and 9c are examples of frozen soil classification maps for the Dakotas based upon the boundaries shown in Figure 8.

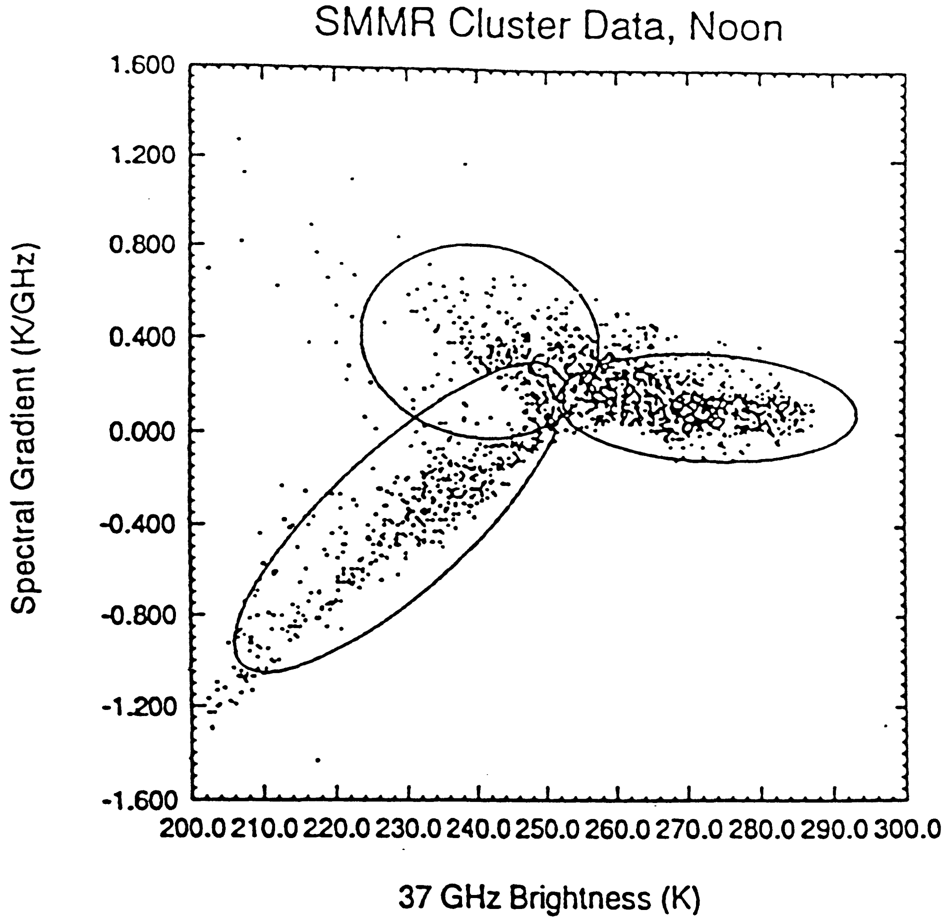

Fig 7. Scatter diagram of $\partial Tb/\partial f$ versus Tb(37) throughout North Dakota and the surrounding region. Noon data were collected from 8/1/84 to 12/31/84 (Zuerndorfer and England, 1992).

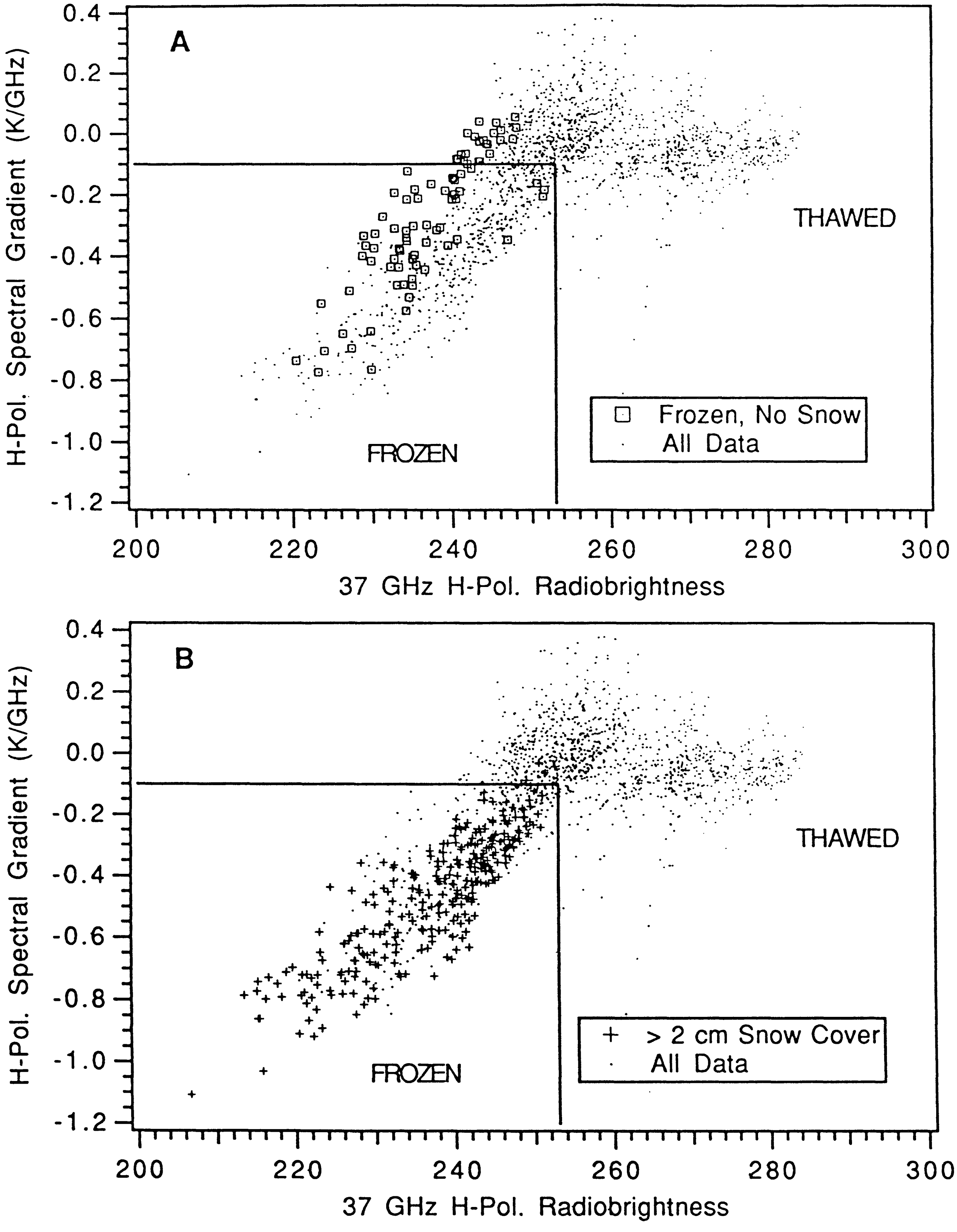

Fig 8. Scatter diagrams of SSM/I, H-polarized data for a 100 km square region centered on Fargo, North Dakota. Data were collected from 8/1/88 to 12/31/88 at 6:00 am local solar time. Data marked are for (a) frozen ground, no snow, and (b) more than 2 cm snow, as reported at Fargo.

Remote Detection of Permafrost

Periglacial processes associated with the freezing and thawing of boreal soils are fundamental to understanding high latitude, land environments (e.g., Washburn, 1973, and Tedrow, 1977). Frozen soils appear dry to land-atmosphere processes. Furthermore, boundaries between continuous and discontinuous permafrost represent divisions between climate regions which are sufficiently cold that they support perennial ground ice, and regions which are less cold. Because permafrost is the integrated effect of many cold seasons, migrations of these boundaries are potential indicators of climate change (Washburn, 1980).

Regional shifts in permafrost boundaries reflect regional variations in dominant weather patterns. Relating such regional shifts to global change will require monitoring continuous and discontinuous permafrost boundaries on a polar scale over many years. Polar scale phenomena are best monitored by operational satellite sensors like the SSM/I. It is not enough to monitor changes of periglacial morphology or of boreal forest tree lines because such features will not reflect recent gains or losses of permafrost. There are no satellite remote sensing techniques that would directly probe soil depths below an active layer. However, there may be a temporal surface temperature signature associated with underlying permafrost.

Thermal characteristics of permafrost terrains have been modeled by Lachenbruch (1959), and Kazemi and Perkins (1971) among others, but their primary objective was to understand the gross thermal structure at depth. Their approximations to the land-atmosphere energy budget were too coarse to provide reliable ground surface temperatures. Predictions from the MCRR/Annual model suggest that permafrost will influence the temporal characteristics of the surface temperature and, consequently, of the temporal radiobrightness signature. Qualitatively, permafrost increases the apparent thermal inertia of soil. Spring thaws in permafrost terrains require more thermal energy than they would in seasonally frozen terrains, and the thaw process that creates the active layer actually extends throughout the summer. Similarly, a fall freeze occurs more quickly because the permafrost acts as a thermal sink. The effect is that permafrost terrains should exhibit later thaws, earlier freezes, and persistently lower surface temperatures throughout the growing season. An evolutionary version of the MCRR model will be tested in a field experiment on tussock tundra near the Toolik Field Station, Alaska, during the summer and fall of 1994.

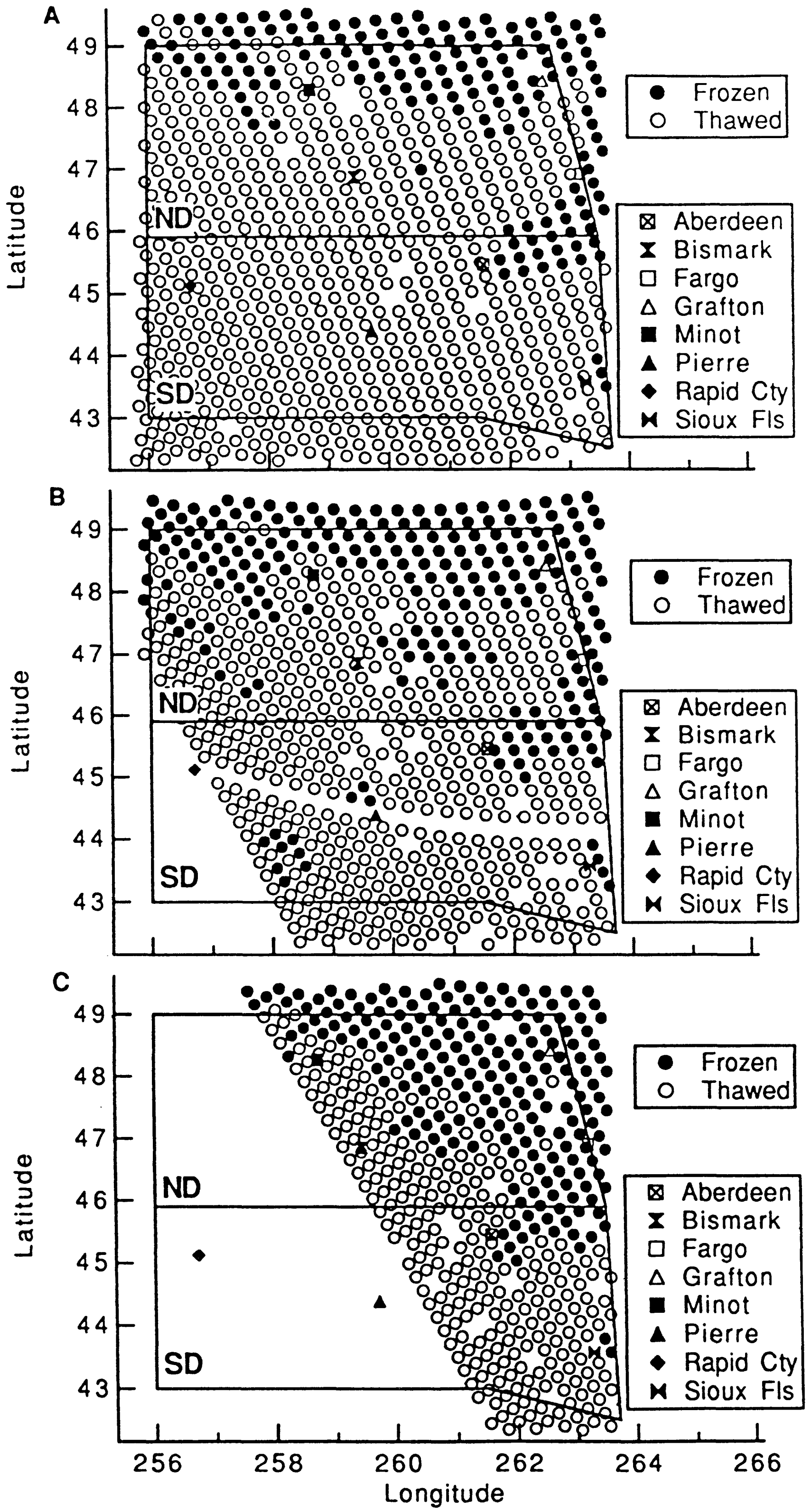

Fig 9. SSM/I-derived freeze-thaw patterns for the Dakotas. Filled circles indicate frozen soil. These patterns are consistent with the meteorological data. Data taken at 0600 local solar time on (a) 12/7/88, (b) 12/8/88, and (c) 12/9/88.

IV List of accomplishments

Under the essential impetus of this project, we have:

1. Developed the diurnal MCRR model (England, 1990).

2. Developed a variable time Laplace method for analytical modeling (Liou and England, 1992a).

3. Developed an analytical annual model for dry soil (England and Liou, 1992).

4. Developed a numerical annual model for wet soil (Liou and England, 1992b).

5. Developed a volume scattering model for emission from frozen soil (England et al, 1991).

6. Refined the classification of frozen soils based upon SMMR data (Zuerndorfer and England, 1992).

7. Extended the classification of frozen soils to SSM/I data (Dahl et al, 1992).

8. Developed the Tower Mounted Radiometer System (Galantowicz and England, 1992).

9. Completed the first Radiobrightness Energy Balance eXperiment (REBX-1) near Sioux Falls, South Dakota during August, 1992, through April, 1993 (Galantowicz and England, 1993).

10. Developed the Radiobrightness Thermal Inertia (RTI) method for estimating soil moisture (England et al, 1992).

11. Developed a Backus-Gilbert interpolation scheme now used by the National Snow and Ice Data Center for resampling SSM/I data (Galantowicz and England, 1991).

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