Atmospheric Ionization and Clouds as Links Between Solar Activity and Climate

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Observations of changes in cloud properties that correlate with the 11-year cycles in space particle fluxes are reviewed. The correlations can be understood in terms of one or both of two microphysical processes; ion mediated nucleation (IMN) and electroscavenging. IMN relies on the presence of ions to provide the condensation sites for sulfuric acid and water vapors to produce new aerosol particles, which, under certain conditions, might grow into sizes that can be activated as cloud condensation nuclei (CCN). Electroscavenging depends on the buildup of space charge at the tops and bottoms of clouds as the vertical current density (J_z) in the global electric circuit encounters the increased electrical resistivity of the clouds. Space charge is electrostatic charge density due to a difference between the concentrations of positive and negative ions. Calculations indicate that this electrostatic charge on aerosol particles can enhance the rate at which they are scavenged by cloud droplets. The aerosol particles for which scavenging is important are those that act as in-situ ice forming nuclei (IFN) and CCN. Both IMN and electroscavenging depend on the presence of atmospheric ions that are generated, in regions of the atmosphere relevant for effects on clouds, by galactic cosmic rays (GCR). The space charge depends, in addition, on the magnitude of J_z. The magnitude of J_z depends not only on the GCR flux, but also on the fluxes of MeV electrons from the radiation belts, and the ionospheric potentials generated by the solar wind, that can vary independently of the GCR flux. The roles of GCR and Jz in cloud processes are the speculative links in a series connecting solar activity, the solar wind, GCR, clouds and climate. This article reviews the correlated cloud variations and the two mechanisms proposed as possible explanations for these links.

1. INTRODUCTION

Clouds play a key role in the energy budget of Earth's surface and lower atmosphere, and are probably the largest contributor to the uncertainty concerning the global climate change [*IPCC*, 1996]. Small modifications of the amount, distribution, or radiative properties of clouds can have significant impacts on the predicted climate [*Hartmann*, 1993]. The suggestion that space particle fluxes, as modulated by solar activity and the solar wind, affect clouds and climate has a long history. Space particle fluxes include galactic cosmic rays (GCR); MeV electrons from the radiation belts; occasional energetic solar proton events; and the bulk solar wind plasma itself. Following a note by *Ney* [1959], *McDonald and Roberts* [1960] and *Roberts and Olsen* [1973], speculated that their correlations between short-term solar variability and intensification of cyclones in the Gulf of Alaska might involve stratospheric ionization changes, produced by solar particle precipitation, affecting clouds. A more in-depth discussion of theoretical aspects of such a mechanism was made by *Dickinson* [1975]. He pointed out that while direct condensation of water vapor in the ionization produced by GCR would not occur in the atmosphere (it requires supersaturations of several hundred percent) the presence of sulfuric acid vapor in the atmosphere allowed condensation on the ions of H_2SO_4 molecules together with H_2O molecules. Provided that these can grow large enough to act as cloud condensation nuclei (CCN), they could affect the particle size distribution and lifetime of clouds, and thus the radiative properties affecting climate. This process is now termed ion-mediated nucleation [*Yu and Turco*, 2000, 2001] , and in later sections we will discuss it in more detail. *Svensmark and Friis-Christensen* [1997] and *Svensmark* [1998] demonstrated correlations of cloud cover with GCR flux, and speculated that ionization processes could affect nucleation or the phase transitions of water vapor. *Marsh and Svensmark* [2000] first noted the potential importance of IMN for such GCR-cloud links.

A more indirect effect of GCR flux changes was suggested by *Herman and Goldberg* [1978] and *Markson and Muir* [1980], in that GCR and other space particle fluxes modulate the current flow in the global electric circuit [*Israël*, 1973; *Tinsley*, 1996; *Bering et al.*, 1998], and the passage of this current density through clouds affects their initial electrification. There are a number of ways in which weak electrification can affect the microphysics of clouds, with consequences for cloud lifetime, radiative properties, and precipitation efficiency. It was suggested by *Tinsley and Deen* [1991] that electrical processes in clouds containing supercooled water lead to enhanced production of ice, which directly leads to enhanced precipitation by the Wegener-Bergeron-Findeisen process. Recent work [*Tinsley et al. 2000, 2001]* has focused on a process called electroscavenging as an electrical link leading to precipitation. In this case the dominant climatic effect is not due to changes in the atmospheric radiation balance, but rather in dynamical effects on the storm system that is undergoing precipitation, due to the latent heat transfer involved. An increase in precipitation entails a decrease in the amount of cloud water that is available for evaporation into air being entrained into the air mass of a storm. The reduction of this diabatic (non-adiabatic) cooling is equivalent to a diabatic heating, and it increases uplift and redistributes vorticity towards the center of the storm (non-conservation of potential vorticity). For a winter cyclone, that is drawing vorticity from the latitudinal shear in the winter circulation, the vorticity redistribution is likely to enhance feedback processes, and increase the overall strength of the storm.

The effects of concentration and overall increase of vorticity appear to be responsible for the observed correlations of tropospheric dynamics and temperature with particle fluxes on the day-to-day timescale [*Tinsley*, 2000]. There is an increase in meridional transport of heat and momentum associated with increased vorticity in winter storms, that is concentrated in the cyclogenesis longitudes, and the cumulative effects over a winter season have the potential for significant feedback on the circulation itself, e. g., on Rossby wave amplitude and regional changes in temperature and storm track latitude.

Thus *Tinsley and Deen* [1991] suggested that shifts in winter storm track latitude in the North Atlantic, that had been observed to correlate with sunspot number for six solar cycles, were a consequence of cosmic ray induced freezing of supercooled water in storm clouds. The climatic effects of these changes in Rossby wave amplitude are much more regional than global. The regional changes show differences in sign, so that the global mean changes, although not negligible, represent relatively small residuals. The greatest societal impact would appear to be from regional changes on the decadal and longer timescales.

In addition to proposed effects of solar modulation of atmospheric electricity on clouds and climate there are changes in solar spectral and total irradiance that affect ozone and have been proposed to affect climate, as discussed in other sections of this volume and by *Shindell et al.* [2001]. On the decadal and longer timescales one cannot easily distinguish between effects on climate of particle flux changes and effects of irradiance changes. However, the temporal signatures are completely different on the day-to-day timescale. The review by *Tinsley* [2000] describes a long history of studies of day-to-day changes in atmospheric vorticity and temperature that correlate unambiguously with short term space particle flux variations. The latitudinal and seasonal variations are fully consistent with the hypothesis of electroscavenging inducing ice production and precipitation from clouds, affecting atmospheric temperature and dynamics. The same processes must operate on the decadal and longer timescales: - however the relative imporance compared to irradiance effects on those timescales remains to be determined.

In this review we will first discuss observations supporting the two hypotheses - ion mediated nucleation and electroscavenging - as links between solar activity and effects on clouds and climate. The influence of solar variability on atmospheric ionization will then be reviewed, followed by outlines of the ion-mediated nucleation and electroscavenging mechanisms. The consequences for precipitation, atmospheric dynamics and radiation of the cloud changes associated with electroscavenging and ion-mediated nucleation will then be discussed. Thereafter, the extent to which the correlations with atmospheric changes support one or the other or both of the mechanisms will be examined.

2. CORRELATIONS OF CLOUD PROPERTIES WITH DECADAL GCR AND SUNSPOT VARIATIONS

Among the many reported decadal timescale correlations of meteorological parameters with solar activity, one of the least ambiguous as an effect of space particle fluxes on clouds is that shown in Figure 2.1. This is a correlation of precipitation and precipitation efficiency with GCR flux in the Southern Ocean that is greatest at the highest geomagnetic latitudes, where the amplitude of the GCR flux variations and the associated vertical current density (J_z) variations are greatest [*Kniveton and Todd*, 2001]. The location of the geomagnetic pole is marked by an X. The precipitation data were from the Climate Prediction Center Merged Analysis of Precipitation (CMAP) product. The amplitudes of the precipitation and precipitation efficiency variations were 7-9% at 65-75° geomagnetic latitudes and at those latitudes the GCR flux and J_z vary by 15-20% over the solar cycle. The statistical significance of the correlation with GCR flux is better than 95% over a large oceanic region as shown in Fig. 2.1. There is a tendency for reversed correlation at lower latitudes.

Another decadal variation that is very consistent with cloud forcing by space particle fluxes is the correlation of cloud cover with cosmic ray flux that was noted earlier.

Figure 2.1. Correlation coefficients between 12-month moving averaged cosmic ray flux and precipitation. Positive and negative correlation coefficients are shown as solid and broken lines respectively, with a contour interval of 0.2. Shading indicates areas with locally significant correlations at the 95% level, where the amplitude of the response is 7-9% of the mean value. From *Kniveton and Todd* [2001].

Svensmark and Friis-Christensen [1997] and *Svensmark* [1998] analyzed total cloud cover data from the C2 data set of the International Satellite Cloud Climatology Project (ISCCP). Later analyses of the altitude-resolved ISCCP-D2 data set by *Marsh and Svensmark* [2000a,b] and *Palle and Butler* [2000] showed a significant positive correlation with the frequency of low clouds, below about 3.2 km, but not with clouds at higher altitudes. The correlation is strongest at low latitudes (between 45°N and 45°S) with the intertropical convergence zone excluded. The global average amplitude of the cloud cover change is 1.5-2% as illustrated in Figure 2.2.

In addition to the high correlation between low cloud cover and cosmic ray fluxes, *Marsh and Svensmark* [2000a,b] find the cosmic ray fluxe is also strongly correlated with the cloud-top temperatures of low clouds. In this case, a band of significantly high correlation is centered around the tropics where stratocumulus and marine stratus clouds are dominant. Again there is no obvious correlation for middle and high clouds.

Inferred cloud cover changes over Russia that correlate with the 11-year cycle of cosmic ray flux changes have been reported by *Veretenenko and Pudovkin* [1999, 2000]. The cloud cover changes were inferred from 1961-1986 observations by a network of actinometric stations, and were negatively correlated with GCR flux below about 57º geographic latitude (about 50º geomagnetic latitude) and positively correlated at higher latitudes.

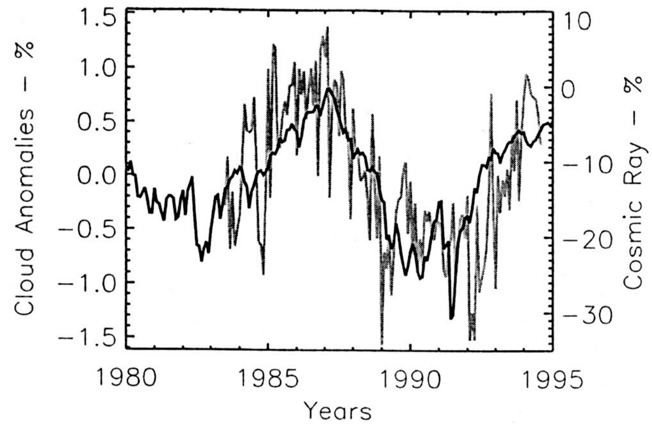

Figure 2.2. Monthly mean values for global anomalies of low (>680 hPa) cloud cover (grey), and GCR fluxes from Climax (black, normalized to May 1965). The global average temporal mean low cloud cover was 28%. From *Marsh and Svensmark* [2000a].

Cloud cover changes over the USA that correlate with the 11-year solar cycle have been reported by *Udelhofen and Cess* [2001] from cloud and actinometric observations in 1900-1987. Relative to GCR variations the correlations were negative below about 42º geographic latitude (about 50º geomagnetic), with a tendency to positive correlations at higher latitudes. Thus, the continental-low-latitude cloud cover variations over the USA and Russia have the same phase relative to GCR variations, and about the same geomagnetic latitude at which the correlation reverses. These continental-low-latitude variations have opposite phase to the mainly oceanic low-latitude cloud cover observations of *Svensmark and Friis-Christensen* [1997].

Changes in cloud cover on the day-to-day timescale that correlate with Forbush decreases of GCR have been reported by *Todd and Kniveton* [2001]. We do not discuss them here as the present focus is on climate change on the decadal and longer timescales.

3. THE INFLUENCE OF SOLAR VARIABILITY ON ATMOSPHERIC IONIZATION

3a. Space Particle Fluxes

The ionization at cloud level that determines the rate of electroscavenging and IMN is influenced by space particle fluxes as illustrated in Figure 3.1. These inputs are in the form of GCR; MeV electrons precipitating from the radiation belts with associated X-ray bremsstrahlung; the bulk solar wind plasma with its embedded magnetic fields that determines the horizontal distribution of potential across the polar cap ionospheres; and occasional energetic

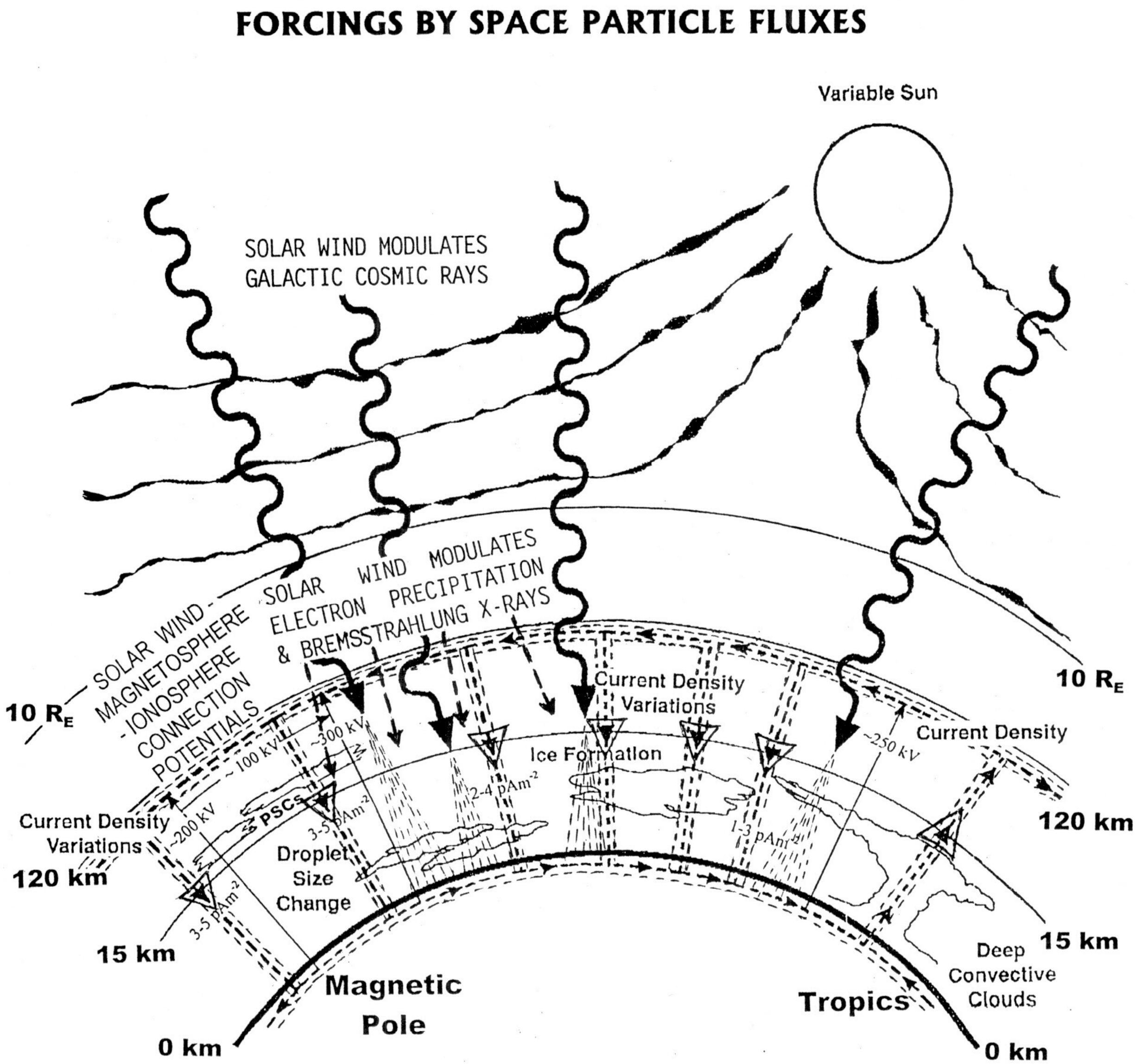

Figure 3.1. Solar wind variations modulate the fluxes of GeV galactic cosmic rays and the MeV electrons coming into the atmosphere, and the ionospheric potential in the polar caps. The fluxes of energetic particles change the vertical column resistance between the ionosphere and the surface, particularly at mid-high latitudes, and this together with the variations in ionospheric potential, change the electrical currents flowing down from the ionosphere into clouds.

solar proton events. The latter occur too infrequently to have a significant effect on climate, and are not illustrated.

The GCR flux is responsible for almost all of the production of ionization below 15 km altitude, that determines the conductivity in that region. The MeV electrons and their associated X-rays produce ionization in the stratosphere, and affect the conductivity there. The current flow in the global electric circuit is generated mainly by charge separation in deep convective clouds in the tropics, and maintains the global ionosphere at a potential of about 250 kV. Variations above and below this value occur in the high latitude regions due to solar wind - magnetosphere - ionosphere coupling processes. The current density J_z varies horizontally due to variations in the local vertical column resistance (this is affected by the GCR and MeV electron fluxes) and by variations in the local ionospheric potential (especially to those in the high latitude regions). Because J_z flowing through clouds in the troposphere responds to conductivity and potential changes occurring all the way up to 120 km altitude, it is a very effective coupling agent for linking inputs in the stratosphere and ionosphere with cloud levels.

3b. Ion concentrations

The GCR flux in the troposphere produces positive air ions of concentration n^+, and negative air ions of concentration n^-, and controls the variations of the total ion concentration $n_t = n^+ + n^-$. Both electroscavenging and IMN depend on these ion concentrations, with IMN production rates, in current theory, dependent on n_t. For electroscavenging it is the difference between positive and negative ion concentrations, that gives the space charge density $\rho = (n^+ - n^-)e$, that is more important than the total ion concentration. The ions are taken as being singly charged, with charges of $\pm e$, where e is the elementary charge.

The space charge density is dependent J_z, that creates vertical electric fields $E_z = J_z / \sigma(z)$, where $\sigma(z)$ is the local conductivity at altitude z that depends on both n_t and the ion mobility. Where there are local vertical gradients in conductivity there will be gradients in E_z and space charge will be generated according to Poisson's equation:

$$\rho = \varepsilon_o \nabla \cdot E = \varepsilon_o J_z \, d/dz(1/\sigma) \qquad (1)$$

Strong conductivity gradients covering a significant horizontal extent are found for example, at the upper and lower boundaries of stratus, stratocumulus, and cirrus clouds, stratified dust and aerosol layers, and at the top of the mixing layer [*Sagalyn and Burke*, 1985]. The space charge density variations, along with the ion production variations, are modulated by solar activity influencing the space particle fluxes on the day-to-day and decadal timescales with relative amplitude 3-20%, as discussed below. The amplitude variations increase to a factor of two or more on the century timescale. They are ubiquitous throughout the troposphere, with greater amplitude at high latitudes. They constitute the solar forcing agent with the largest relative amplitude of any that might affect climate.

3c. GCR and ion production variations.

The GCR flux is modulated in interplanetary space by changes in magnetic fields embedded in the solar wind, with larger modulation amplitude for the lower energy component (less than 5 GeV/nucleon). The Earth's magnetic field confines the low energy particles to higher geomagnetic latitudes, so that only GCR of order 10 GeV/nucleon penetrate into the atmosphere at the magnetic equator. The filtering effect of the geomagnetic field is also variable in time, with the magnetospheric currents that grow during periods of magnetic activity allowing particles of a given energy to penetrate to lower latitudes [*Flückiger et al.*, 1987].

The latitude and time variations of GCR flux into the atmosphere, together with altitude profiles of the resulting ion production, have been reviewed by *Herman and Goldberg* [1978]; *Bazilevskaya* [2000]; and *Bazilevskaya et al.* [2000]. The altitude profiles of ion concentration peak at about 12 km, where the bulk of the primary cosmic ray flux, mainly protons of energy 1-5 GeV, is stopped. The ion pair production rates at 12 km vary from about 20 to 50 ion pairs $cm^{-3}s^{-1}$, with the higher production at higher latitudes and weaker solar activity. Below the 15 km level the production is by cascades of secondary particles, particularly μ mesons. Atmospheric attenuation of secondaries reduces the production rate to about 10 $cm^{-3}s^{-1}$ at 5 km and 2 $cm^{-3}s^{-1}$ near sea level. Over continents the ion production in the lowest few km is several times greater, due to surface radioactivity [*Hoppel et al.*, 1986].

At geomagnetic latitudes of about 40° (e.g, Climax Colorado) the amplitude of the solar cycle modulation of the GCR flux above about 1GeV/nucleon, as measured by neutron monitors, is about 20%. At equatorial latitudes the modulation is only 3-5% because of the weaker modulation of the 10 GeV/nucleon particles. A similar decrease of modulation amplitude with decreasing latitude applies to Forbush decreases. These are short-term reductions in the GCR flux, with onset time usually less than a day and recovery time about a week, that occur when coronal mass ejections pass over the Earth. So both for the solar cycle variations and the Forbush decreases, the ion pair production rates in the stratosphere and troposphere and their effects on n_t and J_z and ρ follow the GCR variations, and are consequently also modulated most strongly at the higher geomagnetic latitudes.

In the polar stratospheres the ion production is mainly by lower energy (0.1-0.5 MeV) GCR that are modulated more strongly in interplanetary space than the 1-5GeV component. the balloon measurements of *Bazilevskaya et al.* [2000] show a factor of two modulation of the 500 MeV component at 25-30 km altitude on the 11-yr solar cycle. On the century time scale the GCR flux changes in the 1-5 GeV component are considerably larger than the 11-yr cycle variations. *McCracken and McDonald* [2001] review ^{10}Be data from polar ice cores and find increases, compared to present solar cycle averages, of a factor of two for epochs of low solar activity near 1900 and 1800 AD, and a factor of three near 1700 AD. They suggest that for more extended solar minima the flux could rise a further 60%.

4. ION-MEDIATED NUCLEATION

4a. General

The effects of aerosols on chemistry and climate are sensitive to particle size and concentration, which are influenced significantly by nucleation processes that are not well understood. It has been demonstrated recently that air ions may play an important role in the production of new particles under typical tropospheric conditions [*Yu and Turco*, 2000, 2001]. Essentially, the production of H_2SO_4 in the gas phase by photochemical reactions in the atmosphere results in supersaturated H_2SO_4 vapor. There is competition between deposition of H_2SO_4 on pre-existing aerosols and on the ions. The competition results in a complicated dependence of the nucleation rates on the concentrations of pre-existing aerosols, ions, and H_2SO_4 vapor. The charged molecular clusters, condensing around ions, are much more stable and can grow significantly faster than corresponding neutral clusters, and thus can preferentially achieve stable, observable sizes. The proposed ion-mediated nucleation (IMN) theory can physically explain the enhanced growth rate (a factor of ~ 10) of sub-nanometer clusters as observed by *Weber et al.* [1997], and seems to account consistently for ultrafine aerosol formation in jet plumes [*Yu and Turco*, 1997, *Yu et al.*, 1998, 1999; *Kärcher et al.*, 1998], in motor vehicle wake [*Yu*, 2001a], in marine boundary layer [*Yu and Turco*, 2001], in clean continental air as well as for the diurnal variation in the atmospheric mobility spectrum [*Yu and Turco*, 2000]. The IMN theory adds another important parameter–ion concentration (n_t) dependent on the ionization rate (Q)–to the list of known parameters controlling the nucleation rate of atmospheric particles [*Yu*, 2001b, 2002]. Galactic cosmic rays (GCRs) are the dominant sources of ionization in the oceanic troposphere, and in the continental troposphere except in the boundary layer where ionization by radioactive nuclides from the soil usually dominates in the lowest 1 km [*Reiter*, 1992]. Recently, the hypothesis that IMN affects cloud properties has received increasing attention, as a result of the high correlation noted earlier between total cloud cover over midlatitude ocean and GCR intensity [*Svensmark and Friis-Christensen*, 1997], and the development of the IMN theory [*Marsh and Svensmark*, 2000a; *Yu and Turco*, 2000, 2001]. More recent results, indicating the presence of a significant positive correlation between GCR intensity and the frequency of low clouds, below about 3.2 km but none with clouds at higher altitudes [*Marsh and Svensmark*, 2000a, b; *Palle and Butler*, 2000] is at first sight surprising, since the solar modulation of the cosmic ray intensity is a maximum around the tropopause and decreases with decreasing altitudes. However, IMN theory does predict strong altitude dependencies of ultrafine aerosol production, as we now discuss.

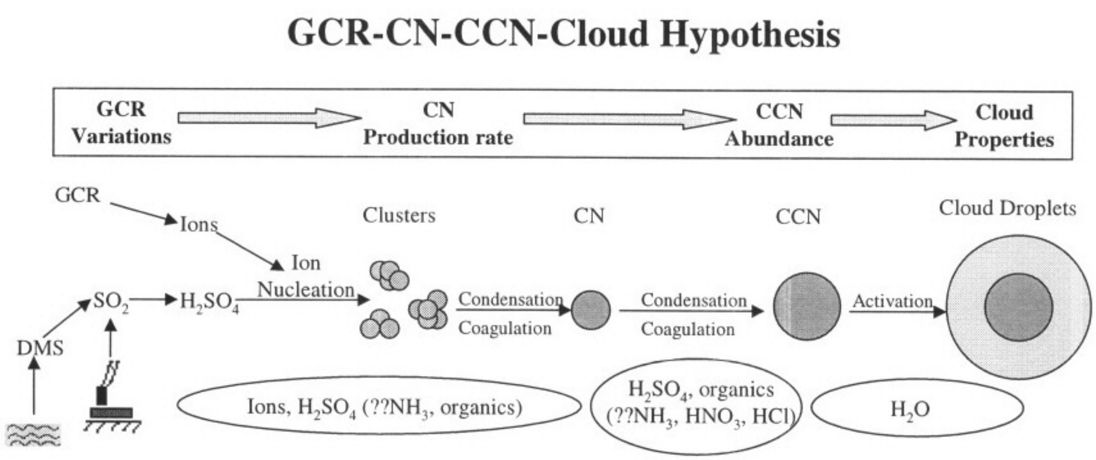

Figure 4.1. Schematic illustrating of GCR-CN-CCN-Cloud hypothesis that, if confirmed, might explain the correlation between variations of GCR flux and low cloud cover. The possible dominating species involved in the different phases of CN formation and growth processes are also indicated. The organics species may play an important role in growing the CN into the size of CCN. (From *Yu*, 2002).

Figure 4.1 shows three key steps involved in the IMN-Cloud hypothesis, which can also be called the CGR-CN-CCN-Cloud hypothesis. If confirmed, they might offer a physically-based mechanism connecting GCR fluxes and global cloud properties. First, the modulation of galactic cosmic radiation by the solar cycle will cause an observable variation in aerosol production and condensation nuclei (CN) population in the lower atmosphere. The term CN is an alternative name for ultrafine particles, or stable clusters of diameter a few nanometers that can act as condensation nuclei in cloud chambers where the supersaturation is several hundred percent. Second, to act as cloud condensation nuclei (CCN) in the atmosphere, where the supersaturation is nearly always less than 2%, they must grow in diameter by a factor of about 10 to act as condensation nuclei. If the number of CCN thus produced is comparable to the total CCN abundance, a systematic change in the ultrafine particle production rate will affect this abundance. Third, a change in CCN abundance will affect the cloud properties.

Yu and Turco [2001] demonstrated that systematic variations in ionization levels due to the modulation of galactic cosmic radiation by the solar cycle are sufficient to cause an observable variation in condensation nuclei (CN) production in the marine boundary layer. Since the dominating number of cloud condensation nuclei (CCN) over ocean evolves from newly formed ultrafine particles [e.g., *Fitzgerald*, 1991], it is physically plausible that a systematic increase in ultrafine production rate will increase the CCN abundance, though the magnitude of such effect is currently unknown. It is well known that the CCN abundance affects cloud properties [*Twomey*, 1977, 1991; *Albrecht*, 1989; *Hobbs*, 1993]. Clouds that form in air containing high CCN concentrations tend to have high droplet concentrations, which lead to an increase in both albedo and absorption. Increases in the CCN concentration also inhibit rainfall and therefore increase cloud lifetimes (cloud coverage). These effects – which are due to more, smaller droplets at fixed liquid water content – are particularly significant in marine air, where the CCN concentrations are generally quite low.

In order to assess the magnitude of the effect of GCR variations on global cloudiness properties, all three steps have to be quantitatively understood. This is not an easy job and requires much further study. However, as far as the sign of the effect (e.g., the correlation between GCR variations and cloudiness) is concerned, the first step is critical. An increase in CN production is expected to increase the CCN abundance and cloud cover, but an increase in GCR fluxes does not always lead to an increase in CN production [*Yu*, 2001b, 2002]. The dependence of CN production on Q is a complex function of Q and [H_2SO_4], as well as ambient conditions (T, RH, surface area of pre-existing particles, etc). Since Q, [H_2SO_4], and ambient conditions vary significantly with altitudes, it is of interest to understand how systematic change of ionization rates as a result of solar activity will affect the particle production at different altitudes.

4b. Altitude and solar cycle dependencies

Ambient ions are continuously generated by galactic cosmic rays as noted earlier, with the magnitude of the ionization rate variations being a function of latitude and altitude. During a solar cycle, the values of Q vary by ~20-25% in the upper troposphere and ~5-10% in the lower troposphere for high latitudes, and by ~4-7% in the upper troposphere and ~3-5% in the lower troposphere for low latitudes [*Ney*, 1959]. The effect of such systematic change in ionization rate on the altitude profile for the production of ultrafine particles has been studied by *Yu* [2002].

The advanced particle microphysics (APM) model which Yu [2002] employed for the study simulates a size-resolved multi-component aerosol system via a unified collisional mechanism involving both neutral and charged particles down to molecular sizes. The size-resolved ion-ion recombination coefficients, ion-neutral collision kernels, and neutral-neutral interaction coefficients calculated in the model are physically consistent and naturally altitude (temperature T, pressure P, and relative humidity RH) dependent [*Yu and Turco*, 2001]. The baseline values of Q at different altitudes are from observations [*Millikan et al.*, 1944; *Neher*, 1971], and the temperature and pressure are according to the US standard atmosphere. In Yu's study, the [H_2SO_4] and RH were parameterized in a way so that they are constant in the lowest 2 km of atmosphere ($2\text{x}10^7/\text{cm}^3$ and 90%, respectively) and gradually decrease with altitude above 2 km. These parameterizations are reasonable and are within the range of the observed values in various field campaigns where significant nucleation events have been identified [*Weber et al.*, 1999; *Clarke et al.*, 1999]. The ion concentration is initialized as $\sqrt{Q/\alpha}$ where α is ion-ion recombination coefficient. The pre-existing particles are initialized as two log-normal modes with total number densities of $19.5/\text{cm}^3$ and $0.6/\text{cm}^3$, median dry diameters of 0.09 μm and 0.3 μm, and standard deviations of 1.6 and 1.5, respectively. This gives an initial wet surface area of ~ 4.2 $\mu\text{m}^2/\text{cm}^3$ at 90% relative humidity, corresponding to a cloud-processed clean air mass where typical significant aerosol nucleation has been observed.

Figure 4.2 shows the total condensation nuclei bigger than 3nm ($N_{d>3\ nm}$) after three hours of simulations at different altitudes. The line with open circles is for the baseline Q values while the line with filled circles is for Q values 20% over the corresponding baseline values. The shaded areas in Figure 4.2 are low, middle, and high cloud regions as defined in ISCCP cloud data. [H_2SO_4], Q, T, and RH at each altitude are fixed during the three-hour simulations. The production rates of ultrafine particles are sensitive to both [H_2SO_4] and n_t (or Q). The growth rate of the ion clusters is controlled by [H_2SO_4], while n_t determines the lifetime of charged clusters as well as the availability of ions. The neutralization by ion-ion recombination will make the growing charged clusters lose their growth advantage and the resulting neutral clusters may dissociate if smaller than the critical size. At typical [H_2SO_4] where significant nucleation has been observed, for very low Q most of the ion clusters have sufficient time to reach the larger stable sizes prior to recombination and the nucleation rate is limited by Q. As Q (or altitude) increases, ion concentration increases but the lifetime of ions decreases and hence the fraction of ions having sufficient time to grow to the stable sizes decreases. As a result, the total number of particles nucleated first increases rapidly but later on decreases as Q (or altitude) increases. The altitude of the turning point is around 4 km under the vertical profiles assumed in this study.

It is clear from Figure 4.2 that an increase in GCR ionization rate associated with solar activity leads to an increase in the ultrafine production rate (i.e., $dN/dQ > 0$) in the lower troposphere (as indicated by the arrows) but a decrease in the ultrafine production rate (i.e., dN/dQ < 0) in the upper troposphere (as indicated by the arrows). In the middle troposphere, dN/dQ changes sign and the average value of dN/dQ is small compared to that of lower and upper troposphere.

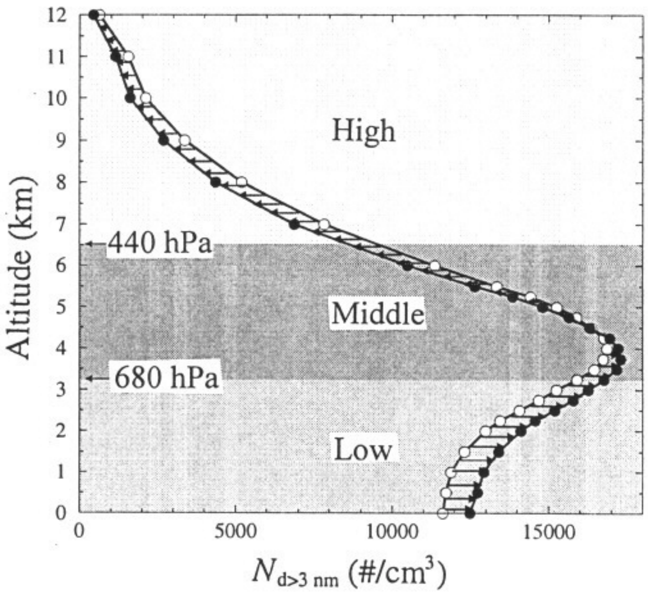

Figure 4.2. Simulated concentrations of total condensation nuclei larger than 3nm after three hours of simulations at different altitudes. The open circles are for the baseline Q values while the closed circles are for Q values 20% over corresponding baseline values. Thus the arrows indicate the changes in $N_{d>3\ nm}$ as ionization rates increase by 20%. The shaded areas are the ranges of Q corresponding to low (>680 hPa), middle (440-680 hPa), and high (<440 hPa) cloud regions as defined in ISCCP cloud data. (From *Yu*, 2002)

As we pointed out earlier, the magnitudes of ionization rate variations during one solar cycle are smaller than 20% for low latitudes. If a smaller value of ionization rate variation was used, the difference between two curves shown in Figure 4.2 would be reduced but the altitude-dependent behavior of changes in CN production would not change. Based on the GCR-CN-CCN-Cloud hypothesis and the influence of GCR ionization change on particle formation rate at different altitudes as shown in Figure 4.2, we can expect that if GCR variations have any impact on cloudiness, they should correlate positively with low cloud amount and negatively with high cloud amount. For middle clouds, such a correlation (if any) is likely to be weak.

We conclude that solar-modulated GCR fluxes can affect the CN abundance and such effect is likely altitude-dependent. Thus, the first key process (i.e, influence of GCR variations on nucleation and CN abundance) in the proposed GCR-CN-CCN-Cloud hypothesis seems to be valid. While, the second and third processes in the proposed hypothesis are logical and physically plausible, we currently do not know how much the GCR produced CN will affect the CCN abundance and cloud properties. Laboratory and field measurements, as well as theoretical studies are needed to validate the predicted dependent-behaviors of nucleation on ionization rates at different altitudes, to investigate the effect of GCR variations on CCN abundance, and to clarify the complex microphysics of aerosol/cloud interactions.

As discussed earlier, the ion-mediated nucleation is sensitive to the ion concentration. Since the space charge density variation as a result of the ionosphere-earth current density change is due to a difference in the concentrations of positive and negative ions, it may also have an impact on ion-mediated nucleation. Charge of mainly one sign in the interaction regions at the boundaries of clouds, as opposed to equal numbers of charges of both signs away from the clouds in clear air, may actually favor the ion-mediated nucleation for clusters of that one sign, because the lifetime of those ion clusters will increase due to the reduced recombination. Recently, significant nucleation has been observed in the top boundaries of clouds [e.g., *Keil and Wendisch,* 2001; *Weber et al*., 2001]. Thus, ion-mediated nucleation may also respond to the ionosphere-earth current density changes. Further research on this issue is obviously required.

5. THE GLOBAL ELECTRIC CIRCUIT AND ELECTROSCAVENGING

5a. Modulation of J_z in the global circuit.

The global electric circuit was illustrated pictorially in Figure 3.1, and a schematic circuit diagram is given in Figure 5.1. General properties of the circuit have been reviewed by *Bering et al.* [1998[. Earlier comprehensive reviews have been given by *NAS* [1986] and *Israël* [1973]. The polar potential pattern is superimposed on the thunderstorm-generated potentials. In a given high latitude region the overhead ionospheric potential, V_i is the sum of the thunderstorm-generated potential and the superimposed magnetosphere-ionosphere generated potential for that geomagnetic latitude and geomagnetic local time. During magnetic storms the changes in V_i from the mean can be as high as 30% within regions extending up to 30° of latitude out from the geomagnetic poles [*Tinsley et al.*1998].

The vertical current density J_z is determined both by the local ionospheric potential V_i and by the local ionosphere-earth column resistance R, with $J_z = V_i /R$, and R given by $R = \int_o^{120km}(1/\sigma(z))dz$. In Figure 5.1 the tropospheric contributions to the vertical column resistances at given latitudes are designated T_E, T_L, T_H, and T_P, for equatorial, low, high and polar geomagnetic latitudes, with corresponding designations S_E, S_L, S_H, and S_P for the stratospheric contributions. The conductivities in the mesosphere and lower thermosphere are considered to be so high that their contribution to the column resistance can be neglected. At any latitude the current density is given to a good approximation by $J_z = V_i/(T+S)$.

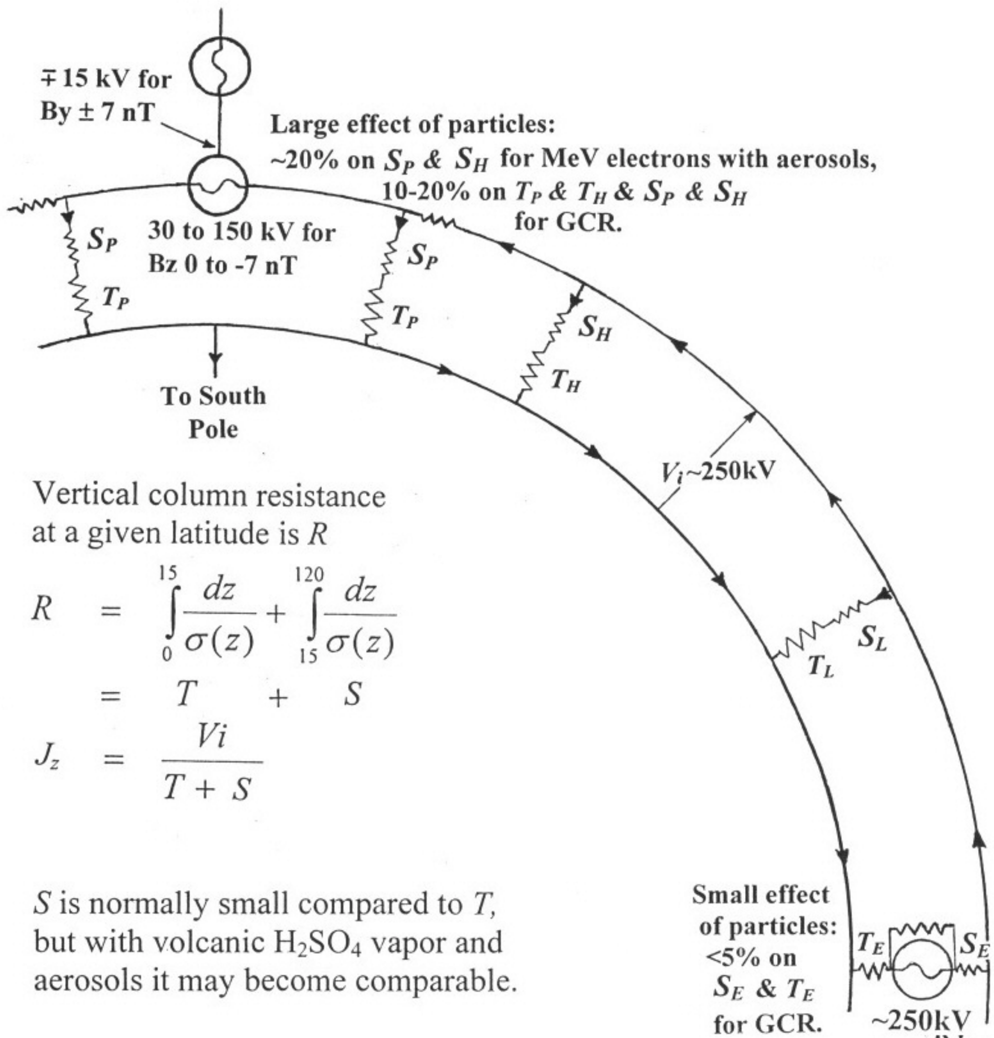

Figure 5.1. Schematic diagram of the global electric circuit. The geometry is essentially plane-parallel and the equatorial and low latitudes have much smaller column conductivity changes, due to energetic particle influx, than high or polar latitudes. The main generator is tropical thunderstorms, producing an ionospheric potential of order 250 kV over most of the globe, except in the polar regions, where superimposed dawn-to-dusk potentials and a pole-to-pole potential are produced by solar wind-magnetosphere-ionosphere interactions.

The GCR flux modulates T and S continuously. The more intermittent fluxes of MeV electrons precipitating into the stratosphere with their associated X-ray Bremsstrahlung modulate S_P and S_H, and the occasional solar proton events modulate S_P. The flux of MeV electrons in the magnetosphere is strongly correlated with solar wind velocity, as discussed by *Li et al.* [20001a,b] who also describe the time and the latitude variation of the precipitating flux. The S_P and S_H contributions to R are normally small, on account of the relatively high conductivity in the stratosphere under normal conditions. But their contribution can be significant when the general stratospheric conductivity has been greatly reduced by H_2SO_4 liquid aerosol particles and vapor following volcanic eruptions [*Tinsley*, 2000]. Polar stratospheric clouds may have the same effect. Measured potential gradients (proportional to J_z) near the surface at the South Pole have been analyzed by *Tinsley et al.* [1998] and show the expected correlation with the solar wind-driven changes in the overhead ionospheric potential. Measured potential gradients and associated J_z variations responding to MeV electron flux changes are discussed by *Tinsley* [2000].

The schematic shown in Figure 5.1 is for one quarter of a meridional section through the three dimensional global circuit. Comprehensive numerical models of the circuit have been constructed by *Hays and Roble* [1979] and *Roble and Hays* [1979]. A model by *Sapkota and Vareshneya* [1990] includes in addition a first attempt to take into account the effect of volcanic H_2SO_4 aerosols and vapor on the stratospheric column resistance, as well as a more detailed version of the latitude variation of GCR modulation.

As indicated in Figure 5.1, horizontal potential differences of order 100 kV are generated, high on the dawn side and low on the dusk side, producing corresponding changes in V_i and J_z. The dawn-dusk potential difference has a strong dependency on the product of the solar wind velocity, v_{sw}, and the B_z(GSM) north-south solar wind magnetic field component [*Boyle et al.*, 1997]. Fig. 5.1 includes not only a horizontal potential generator in the northern polar cap, but a solar wind-driven generator of a potential difference between the northern polar cap ionosphere and the southern polar cap ionosphere. This potential depends on the product of v_{sw} and By(GSM), the east-west solar wind magnetic field component. These products are measures of the east-west and north-south components of the electric field in the solar wind. The electric field is due to the motion of the solar wind with its embedded magnetic field relative to the Earth. The resulting potential differences across the magnetosphere are partially coupled onto magnetospheric field lines and conducted by them down into the polar ionospheres.

5b. Variation with latitude of J_z responses to GCR changes.

The latitude variation of J_z responses to GCR changes is important because models show that it changes sign between high and low latitudes, and therefore any cloud response to GCR, that is linked via J_z changes, would also be expected to change sign.

Roble and Hays [1979] showed that during a Forbush decrease the resulting conductivity decrease at high latitudes (increase of R) results in a diversion of the (assumed constant) current output of the tropical thunderstorm generators from high latitudes to low. This is the case even with conductivity decreases at middle and low latitudes as well, since these decreases are smaller than at high latitudes.

The treatment of this effect by *Sapkota and Varshneya* [1990] is illustrated in Figure 5.2, for a relatively large decrease of GCR flux, of 35% at high latitudes and 12% at equatorial latitudes. The values of ΔJ_z and ΔE_z, the changes in J_z and E_z from values before the GCR flux change, are plotted against latitude for longitude 72.5°E. The effect of changes in surface altitude (orography) and thus in near-surface resistivity with latitude and longitude cause the departures from symmetry about the equator. The effects are especially important for E_z that is proportional to the altitude-dependent near-surface resistivity.

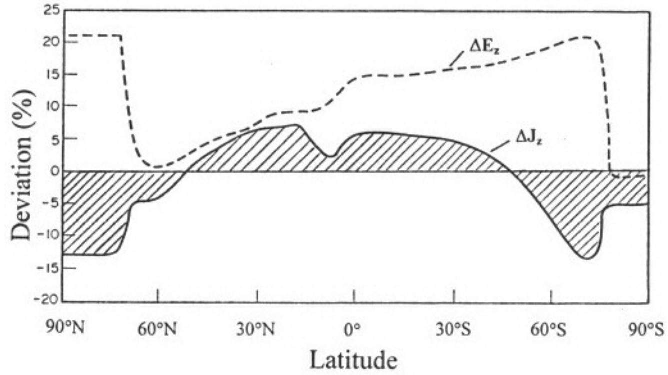

Figure 5.2. Percentage changes in J_z and near-surface E_z for a Forbush decrease of 35% (polar) to 12% (equatorial) for longitude 72.5°E. The departures from symmetry about the equator are due to the variations in surface altitude, that have greatest effect on the near-surface E_z. Adapted from *Sapkota and Varshneya* [1990].

5c. Charges on droplets and aerosol particles.

As noted earlier, a region of space charge develops at the tops and bottoms of clouds because of the flow of current through the gradients of conductivity there. In the limiting case of the cloud conductivity being zero and the cloud of very large horizontal extent the full ionosphere-earth potential difference of ~250 kV would appear between the top and bottom of the cloud. The gradients of conductivity at cloud boundaries arise because of the decrease in the concentration of the high-mobility light ions, due to the presence droplets and condensation on nuclei (forming haze particles) that increase the surface area for attachment and recombination of the ions. Also the humidity approaching 100% ensures that water molecules cluster on ions, reducing their mobility. The conductivity changes have been discussed by *Anderson and Trent* [1996], *Reiter* [1992], and *Dolezalek* [1963]. Observations and theory have been compared by *Pruppacher and Klett* [1997], herinafter PK97, who give a range of between a factor of 3 and 40 for the reduction of conductivity within clouds. *Rust and Moore* [1974] found a factor of 10 reduction in clouds compared to clear air at the same altitude.

To estimate the approximate space charge density ρ for cloud tops we assume that the conductivity σ decreases by a factor of 10 from outside the cloud at 5 km, where $\sigma \sim 10^{-13}\ \Omega^{-1}m^{-1}$ [*Gringel et al.*, 1986], to inside where $\sigma \sim 10^{-14}\ \Omega^{-1}m^{-1}$. If the width of the interface is 1 m, then application of equation (1) gives $\rho \sim 10^4 e\ cm^{-3}$ where e is the elementary charge. If the width is 10 m thick then $\rho \sim 10^3 e\ cm^{-3}$. The layer of charge density in either case amounts to a surface charge of 1.6 x 10^{-9} C m^{-2} that is supplied by J_z in a time of order 15 minutes.

The turbulence at the interface between clear and cloudy air mixes parcels containing charge density, whether on air ions or aerosol particles or droplets, both inwards and outwards. On a longer timescale convection and mixing within the cloud will move parcels containing negative charge from cloud base to cloud top, and parcels containing positive charge from cloud top to cloud base. In general the amounts of positive and negative charge distributed in the cloud will scale as J_z. This applies also to space charge convected into the cloud from other regions in the troposphere where there are conductivity gradients, e.g., on dust layers or from near the surface. However, if ice is eventually produced by processes other than electroscavenging, thunderstorm processes involving ice particle collisions may provide a great deal of additional charge independently of J_z variations.

Many measurements have been made of droplet charges, and a summary is given in Figure 18.1 of *PK97*. For clouds in which no ice has formed, and for droplets of radius about 10 μm, charges of about 100e per droplet are typical. The charge would initially be deposited as space charge in interface layers at the tops and bottoms of clouds, by the current J_z flowing through the conductivity gradients in those locations. The charges migrate from air ions to the droplets, which are the largest objects present. The space charge deposited is positive at cloud top and negative at cloud base, and becomes mixed through the cloud depending on the amount of convection and turbulence present.

Charges on aerosol particles are also highly variable. Charge is transferred from air ions to particles, but the greatest charge is present on the residues of charged droplets, immediately after the droplets have evaporated, when they retain essentially all of the charge of the original droplets. As discussed by *Tinsley et al.* [2000] they lose this charge with an initial decay time constant estimated to be about 15 minutes for mid level clouds. The charge does not decay to zero, but asymptotically approaches an equilibrium value that varies with the amount of space charge present, or more precisely with the ratio n^+/n^-. For example, in the interface layer at cloud top with a moderate amount of mixing and evaporation a droplet with a charge of say 100e creates a charged aerosol particle with 100e that can be scavenged by other droplets. For a typical aerosol particle of radius 0.3μm and with n^+/n^- = 10 the particle charge decays to an equilibrium value of 10e.

5d. Electroscavenging

Small electric charges on aerosol particles induce the ‘scavenging’, or collection of such particles by falling droplets. Figure 5.3 illustrates the movement across streamlines caused by the electrical forces). Aerosol particles, including CCN, will be retained by the droplet after making contact, and these will then be unavailable for further condensation cycles. If the scavenged particle is an ice-forming nucleus, (IFN), and the droplet is supercooled (at a temperature below freezing but still liquid) the droplet is likely to freeze when contact is made.

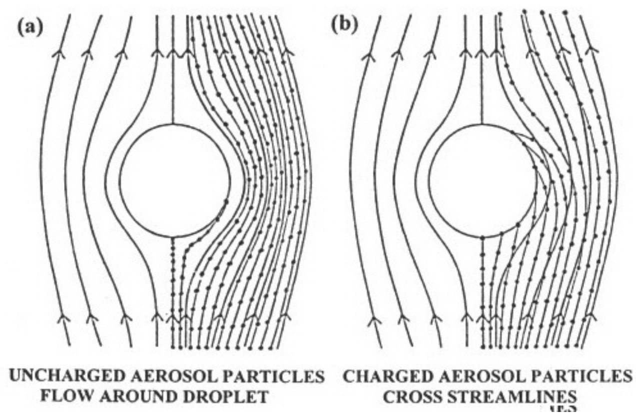

Figure 5.3. (a) Schematic of aerosol flow around a falling droplet in the absence of electrical forces. (b) Schematic of effect of electrical forces in moving aerosol particles across streamlines.

Electroscavenging rates are determined by calculating trajectories of charged aerosol particles relative to falling droplets, under the influence of electrical and other forces. Such trajectories have been calculated by *Tinsley et al.* [2000, 2001]. The simulations take into account the previously neglected force between the particle charge and its image charge on the droplet, and the results show that this is very important. Collision efficiencies are calculated, being defined as the fraction of those particles in a cylindrical volume swept out by a falling droplet, that actually make contact with the droplet. The tendency for particles to be carried around the droplet by the flow results in collision efficiencies for uncharged particles in the range 10^{-2} to less than10^{-3} for particles of radius *a* in the range 0.1 μm to 1 μm in the path of typical cloud droplets.

For particles in this size range, which is the most important one for contact ice nucleation, the electrical effects were found in most cases to dominate the scavenging, even with charges of the same sign on the droplet. Although there is a long-range repulsion between charges of the same sign, the flow carries these larger particles against the repulsion close to the droplet, so that the short-range attractive force, due to the attraction between the particle charge and its image charge, ensures particle collection. Figure 5.4 shows the variation of collection rate with droplet radius for droplet charges of 100e and particle charges of 20e of the same sign. The rapid increase of collection rate with droplet radius is due to the volume swept out in a given time varying approximately as the fourth power of the droplet radius (both fall velocity and cross sectional area increase as radius squared), although the collection efficiencies actually decrease with increasing droplet radius for the larger particles. This variation of collection rate means that electroscavenging processes are likely to be more important for clouds where the droplet size distributions include an appreciable fraction with droplet radii greater than about 10 μm. The low concentrations of aerosol particles and condensation nuclei over the oceans generally results in larger average droplet sizes there, as observed by *Bréon and Colzy* [2000]. Thus, electroscavenging will be more important for oceanic than for continental type clouds.

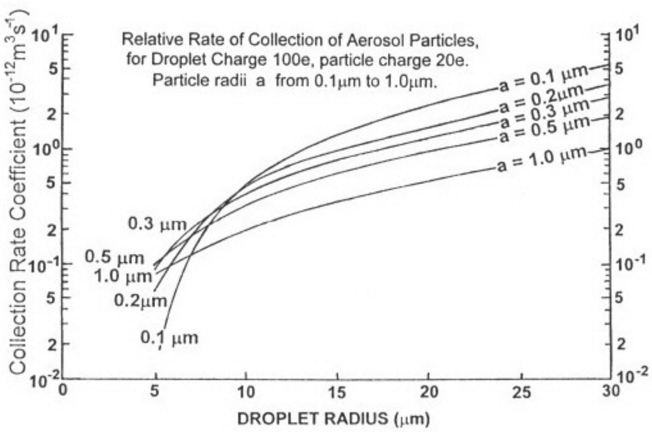

Figure 5.4. Collection rate coefficients as a function of droplet radius for particle radii a as labeled. The particle charges are all 20e and the droplet charges are 100e.

5e. Electroscavenging and ice formation

When clouds cool below 0°C there is usually little production of ice until temperatures of -10° to -20°C are reached [*Rogers and Yau*, 1989] and for very pure air and water it is necessary to cool below - 40°C before droplets freeze (homogeneous nucleation). In the middle and lower troposphere with temperatures above -40°C the freezing of these supercooled droplets is due to a variety of heterogeneous processes. Contact ice nucleation occurs when an aerosol particle with a suitable surface, making it an IFN, makes contact with a supercooled droplet. The formation of ice is an important first step in the production of precipitation in supercooled clouds. It is only necessary that the supercooling be in the region near the cloud top, as the frozen droplets grow by absorbing vapor there, and when they are large enough to fall below that region they will continue to accrete liquid water while in the cloud, even if they melt and continue to do so as droplets.

The clouds where contact ice nucleation is likely to be the most important are those where slow uplift has recently produced cloud top temperatures below, but not greatly below 0°C, and where a moderate amount of mixing is occurring (as in marine stratoculuslus) that results in the evaporation of charged droplets. Mixing at the interface between the cloudy and clear air at cloud top provides an optimum location for electroscavenging, leading to contact ice nucleation and enhanced precipitation. This interface region is where droplets become charged from the space charge deposited in the conductivity gradient, and where evaporation of these charged droplets is occurring, and where there is further mixing, so that other supercooled droplets can fall through the air parcels containing charged evaporation residues and electroscavenge them.

5f. Electroscavenging and CCN removal

A consequence of electroscavenging is that the scavenged aerosol particles (both IFN and CCN) become incorporated in the droplets, and the remaining particle size distributions become depleted in the component removed. If the droplets precipitate the CCN or IFN will be removed from the air mass, and if the droplets evaporate a number of individual scavenged particles are likely to be consolidated into larger sized particles, although the volatile components will tend to evaporate and be available for further cycles of IMN. There is a tendency for preferential removal of the smaller condensation nuclei and generation of giant nuclei.

In maritime clouds the observed concentrations of CCN are considerably lower than over continents as noted, and so the recycling of evaporation residues as CCN is likely to be greater than for continental clouds. Repeated cycles of cloud formation and evaporation occur, so that removal by electroscavenging of CCN during evaporation is followed by effects on the size distribution of the droplets that are formed during the next condensation cycle. The size range of interest for these hygroscopic particles extends down to roughly 0.02 μm (20 nm) radius, below which they would not grow large enough by absorbing water to be activated for droplet formation. Because of uncertainty in CCN size distributions and surface coatings we can only estimate roughly the time required for a significant reduction in CCN population by electroscavenging.

Using the data of Figure 5.4 and the six droplet size distributions typical of oceanic clouds discussed by *Tinsley* [2001] the estimated lifetime of condensation nuclei against scavenging ranges from 0.5 hr to 5.5 hr. The estimate quoted earlier for the initial time constant for decay of the charges on newly formed evaporation residues is about 15 minutes, and this leaves insufficient time for significant effects on the CCN population, on account of the decay of the particle charge. However, more time is provided by the repeated cycles of condensation and evaporation. *Pruppacher and Jaenicke* [1995] reviewed the subject of aerosol processing by clouds, and pointed out that most clouds evaporate and do not precipitate, and that aerosol particles and water in the atmosphere typically undergo at least three cycles of condensation and evaporation before being removed. Also, the probability of significant electroscavenging of CCN is increased by particles having a smaller charge than 20e, but a finite value maintained for the lifetime of the cloud in regions of finite space charge. Clearly, there is a need for time-dependent modeling that includes charging, mixing, and evaporation, to arrive at more quantitative evaluations of electroscavenging effects on CCN.

6. EFFECTS ON PRECIPITATION AND LIFETIMES OF CLOUDS

The macroscopic properties of clouds such as albedos, lifetimes, fractional cloud cover and amount of precipitation are affected by the microphysical processes, such as the addition or removal of cloud condensation nuclei by ion-mediated nucleation or electroscavenging respectively, or from the production of ice by contact ice nucleation due to electroscavenging. The importance of the effects depends on cloud temperature, thickness, and type.

For cold clouds (those with tops colder than 0°C) there are two effects by which electroscavenging can increase precipitation. The first is the production of ice as described previously. The frozen droplets grow by the deposition of vapor until they reach a radius of ~25 μm, at which stage they will fall fast enough to grow more rapidly by collision and coalescence with smaller droplets and ice crystals in their path (*PK97*). If the cloud depth is great enough they attain radii of ~100 μm and have large enough fall speeds to leave the cloud within a typical cloud lifetime. The timescale for electroscavenging and contact ice nucleation to increase the precipitation in cold clouds, following a change in ionosphere-earth current density J_z, is hours or less. This is because the time constant for the space charge ρ to develop on clouds and on droplets that evaporate to generate charged evaporation residues is less than one hour, and the mixing processes and contact ice nucleation processes and timescales for growth of ice particles have similar short timescales.

The second way that electroscavenging can increase precipitation in cold clouds is by the removal of CCN as described earlier, that leads to an increase in the average droplet size. With larger droplets there is a higher rate of electroscavenging, as can be seen from the increase in collection rate coefficient with droplet size in Figure 5.4. This increases the rate of contact ice nucleation that initiates precipitation as described above. The time scale for removal of CCN by electroscavenging is likely to be that of days rather than hours, as repeated cycles of cloud formation and evaporation would appear to be necessary to significantly change the CCN concentrations.

In cold clouds as well as in warm clouds ion-mediated nucleation can affect the CCN concentration. Increases in the CCN concentration producing decreases in the average droplet size were shown by *Twomey* [1977] to produce an increase in cloud albedo. This effect is well known as a result of anthropogenic production of CCN. From the results reported by *Schwartz* [1996] the maximum increase in cloud top reflectance is for clouds of about 300 meters thickness, where there is an increase in cloud top reflectance by about 5% (from 50% to 55%) for a doubling of the droplet concentration from 100 cm^{-3} to 200 cm^{-3}. This would suggest a quite small effect for ~5% change in droplet concentration.

In warm clouds another effect due to changes in CCN concentration, that would appear to be more important than albedo effects, is to change the probability of precipitation. The smaller average droplet size with increased CCN concentration reduces the droplet fall speeds and increases the time necessary for collision and coalescence; processes that are required for the production of precipitation in thick clouds and drizzle in thinner clouds [*Beard and Ochs*, 1993]. With less drizzle production the cloud lifetime and cloud cover are likely to increase.

In thick storm clouds changes in the heavy precipitation that contributes significantly to latent (diabatic) heat transfer can occur as a result of either electroscavenging or IMN changes. We discuss the consequences of this in the next section. Again, there is a need for time dependent cloud modeling to arrive at more quantitative evaluations of the effects of IMN and electroscavenging on CCN concentrations, ice production, precipitation rates, cloud lifetimes and cloud cover.

7. DYNAMICAL AND RADIATIVE EFFECTS OF CLOUD CHANGES

7a. Effects of precipitation on diabatic heating and dynamical changes

From the foregoing theoretical considerations it is apparent that an increase in electroscavenging can lead to increased precipitation, and an increase in ion-mediated nucleation can lead to decreased precipitation in certain types of clouds and air masses. When precipitation removes water from a cloud there is less cloud water available to cool the air mass by evaporation when the cloud eventually dissipates. In winter storms where there is continuing uplift of air from lower levels the continuing precipitation entails a reduction of the evaporation of water into unsaturated air being entrained into the storm in updrafts and in the mixing at cloud tops. Again, there is less cooling of the air mass than would otherwise have been the case without precipitation. The effect of precipitation is thus one of diabatic (non-adiabatic) heating of the air mass, and this can have important effects on the dynamics of the air mass, (non-conservation of potential vorticity) because of the large latent heat content of moist air in comparison with the specific heat of the same mass of dry air.

In a winter cyclone the primary driver of the dynamics is the baroclinic instability in the winter circulation, with the storm extracting vorticity from the latitudinal shear in the circulation, and converting it to the vorticity of the cyclone. The effective diabatic heating associated with precipitation and reduced cooling of entrained air amounts to an increase in potential vorticity and uplift in the air mass, and is likely to concentrate the vorticity near the cyclone center. In addition, by enhancing the feedback processes inherent in the baroclinic instability, it can increase the overall vorticity of the cyclone. It has been demonstrated analytically by *van Delden* [1989] and from numerical storm simulations by *Zimmerman et al.* [1989] and *Mallet et al*. (1999) that a positive feedback exists between the storm dynamical configuration and the diabatic processes. Thus precipitation changes explain the many reported examples of correlations of the vorticity area index (VAI) with GCR flux change and J_z reviewed by *Tinsley* [2000].

Winter cyclones grow in the presence of a well-developed baroclinic winter circulation, and their growth tends to reduce the baroclinicity, by dissipating momentum and by transporting heat poleward. These effects are concentrated in cyclogenesis longitudes and downstream from them. The changes associated with external forcing and reflected in the VAI variations, when sustained over a season, have the potential for significant effects on the circulation itself. Such changes should appear as changes in Rossby wave amplitude, and in regional variations in average storm track latitude.

Figure 7.1 compares observed changes in (a) GCR flux above about 1 GeV measured by the Climax, Colorado neutron monitor [*NGDC,* 1989] and above about 500 MeV measured by balloons at 18 km altitude in the Arctic [see *Bazilevskaya,* 2000]; (b) the number of winter storms in the western North Atlantic crossing 60°W between 35°N and 65°N in the west phase of the QBO [from *Labitzke and van Loon,* 1989]; (c) the latitude shift in storm tracks in the eastern North Atlantic crossing 5°E above 50°N for both QBO phases (smoothed) and for the west QBO phase [see *Tinsley and Deen,* 1991]; (d) the 700 hPa wind speed and direction in the North Atlantic and European regions, for west QBO phase, and cosmic ray minimum winters compared with cosmic ray maximum winters 1950-1988 [from *Venne and Dartt*, 1990]; (e) the mean Jan-Feb. temperatures for three eastern USA stations for west QBO phase winters [*van Loon and Labitzke,* 1988]. These correlated decadal variations were postulated by *Tinsley and Deen* [1991] to be due to increased cyclogenesis in the western north Atlantic responding to increased cosmic ray flux and cloud responses at solar minimum relative to those at solar maximum, in accordance with the general scenario that we have now described in more detail above. This general scenario does not imply that the effects of cosmic ray flux on clouds are fully responsible for the observed decadal atmospheric variations, as such variations at these longitudes have been attributed to an internal coupled atmosphere-ocean oscillation known as the North Atlantic Oscillation [*Walker and Bliss*, 1932; *van Loon and Rogers*, 1978]. The influence of the GCR effect on clouds need only be strong enough to phase lock the NAO to the solar cycle, during periods of stronger solar activity, to account for the six consecutive cycles of storm track latitude that have remained in phase with the solar cycle in the mid-20th century.

An effect of increased cyclogenesis is to increase the meridional transport of heat and momentum and weaken the prevailing westerly zonal winds, as we see in Figure 7(d) for central western Europe. Thus increased cyclogenesis in the North Atlantic due to greater GCR flux during the Maunder Minimum may have contributed to the reduced zonal winds and colder winters in Europe at that time [*Luterbacher et al*., 2001]. The main effect has been attributed by *Shindell et al*. [2001] to reduced solar UV during the Maunder Minimum.

The quantitative study of the dynamical consequences of externally forced cloud and diabatic heating changes has received little attention so far, but the physical theory needed for numerically modeling such effects seems to be reasonably well established. Such models would provide a good test of the general scenario describe qualitatively above.

7b. Effects of cloud cover changes on radiative balance

Clouds play a key role in the energy budget of Earth's surface and lower atmosphere. The net cloud radiative forcing depends on the altitude and optical thickness of clouds. Optically thin clouds at high and middle altitudes tend to warm due to their relatively high transparency at short wavelengths but significant opacity in long wavelengths, while optically thick clouds tend to cool as a result of the dominance of the increased albedo of shortwave solar radiation. Overall, clouds reflect more solar radiation than they trap, leading to a net cooling of ~27.7 W/m^2 from the mean global cloud cover of ~63.3% [*Hartmann*, 1993].

As discussed in the earlier sections, particle flux induced changes in CCN abundance (via electroscavenging and/or ion-mediated nucleation) may affect the cloud albedo, opacity, and lifetime (hence cloud cover fraction). An increase in CCN tends to increase the cloud albedo and lifetime. Based on the average low cloud radiative forcing of –16.7W/m^2, *Marsh and Svensmark* [2000a] estimated a change in net low cloud forcing of ~ 1.2W/m^2 associated with a ~2% absolute change (~7% relative change) in low cloud cover over a solar cycle. These cloud cover changes are sufficient to change the atmospheric heating profile, as discussed by *Yu*, [2002]. Associated with the radiative effects of such large-scale cloud cover changes are likely to be changes in atmospheric circulation. *Dickinson* [1975]

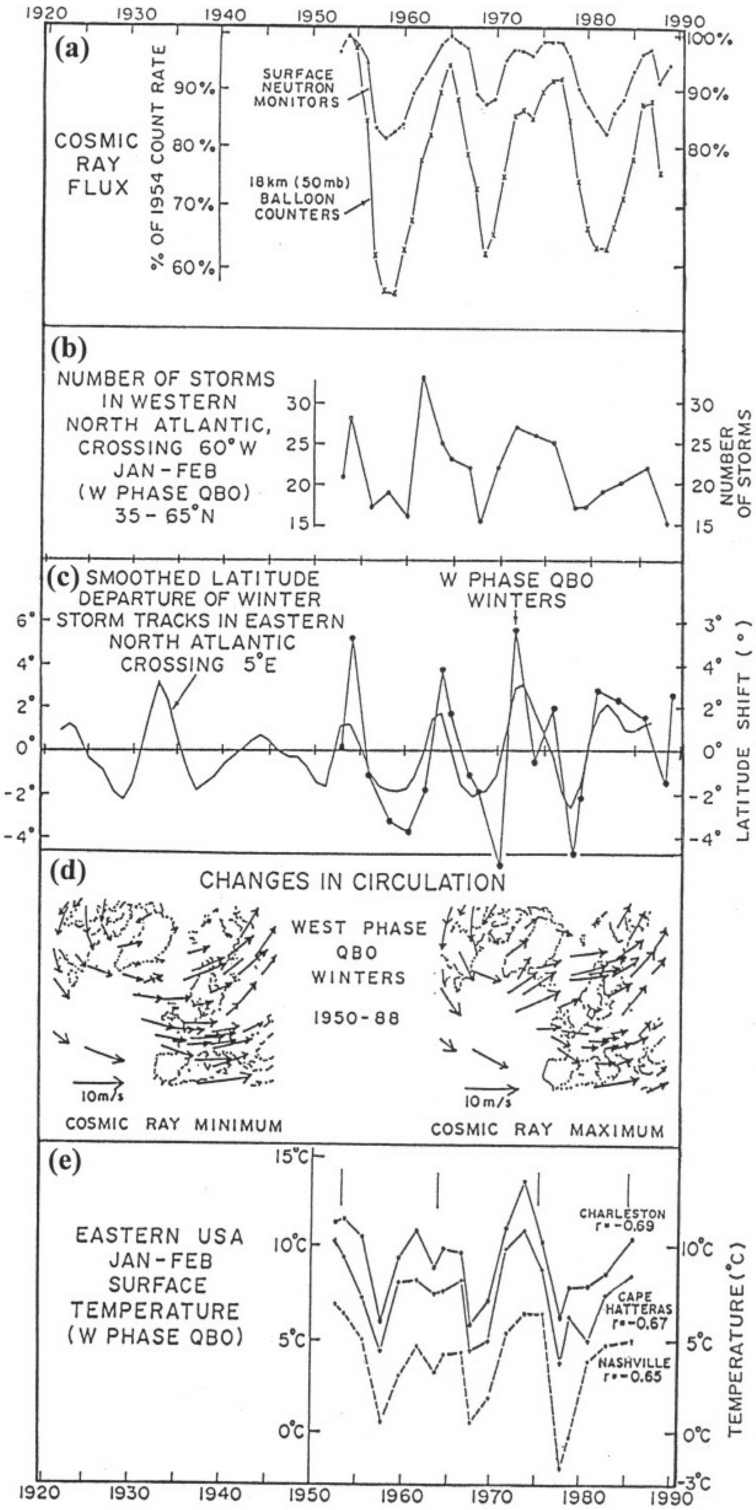

Figure 7.1. Comparison of GCR variability with tropospheric dynamics and temperature in the North Atlantic-Europe region. (a) GCR flux above about 1 GeV/nucleon at Climax, Colorado and above about 500MeV/nucleon in the Arctic Stratosphere; (b) storm frequencies in the western North Atlantic; (c) storm track latitudes in the eastern North Atlantic; (d) 700 hPa wind speeds and directions in the North Atlantic and European regions for GCR maximum and GCR minimum winters; (e) winter temperatures for three eastern USA locations. For data sources and details see *Tinsley and Deen* [1991].

pointed out that a change in upper cloud opacity by 20% would produce heating rates in the column below of order 0.1°C per day, which as a differential across a zone 15° in latitude would lead to changes in zonal winds at the tropopause of order 2 m s^{-1}.

8. CONSISTENCY OF CORRELATIONS WITH MECHANISMS

While the energy inputs into the troposphere from the space particle fluxes are orders of magnitude less than those in the correlated atmospheric variations, the free energy available in the supercooled liquid water and supersaturated water vapor are adequate to provide the needed energy amplification, and this would be supplemented by modulation of solar and terrestrial radiation. There is adequate amplitude, relative to their means, of variations in the particle fluxes to drive linear amplitude changes.

At high latitudes both ion production Q and space charge ρ are positively correlated with both GCR changes and J_z changes. (Although parameter $\sigma(z)$ is in the denominator of equation (1), it does not change this result. It applies to the tropospheric cloud level, and varies by a smaller percentage with GCR changes than does J_z. This is because J_z is determined by the reciprocal of R, and at higher altitudes the contributions to R vary by considerably larger percentages.) At low latitudes Q is positively correlated with GCR changes whereas J_z and ρ are negatively correlated, as shown in figure 5.2.

Thus the decadal low latitude and low altitude oceanic cloud cover changes [*Marsh and Svensmark*, 2000a] are consistent with ion-mediated nucleation responding to changes in Q. In addition, the anticorrelation of J_z with GCR at low latitudes means that at solar minimum, while there is a maximum of production of CCN by ion-mediated nucleation due to Q changes, there would also be a minimum of removal of CCN by electroscavenging due to opposite ρ changes. So both ion-mediated nucleation and electroscavenging mechanisms could be acting together to increase cloud albedo (and possibly absorption), cloud lifetime, and cloud cover at low latitudes.

The decadal precipitation changes at high latitudes observed by *Kniveton and Todd* [2001] are consistent in sign with J_z and ρ increasing electroscavenging and precipitation at solar minimum, but not with Q and IMN changes decreasing it. The tendency for negative correlations below about 50° geomagnetic latitude is consistent with electroscavenging responding to the low latitude anticorrelation of J_z with GCR flux, as well as with the effects of IMN, as in the low latitude low altitude cloud cover variations. Effects of IMN could contribute for high latitudes if there were sufficient reduction of H_2SO_4 vapor concentration at high latitudes to give an anticorrelation of IMN with GCR flux there.

Recent papers by *Ram and Stolz* [1999] and *Donarummo et al.* [2002] have found 11 and 22-yr and longer solar periods in dust concentrations in a high latitude (Greenland) ice core extending over more than 90,000 annual layers. They attribute the changes in dust concentration to changes in precipitation and soil moisture. The presence of the 22-yr cycle, that is a feature of neutron monitor records, suggested that GCR changes, rather than total or spectral irradiance changes were responsible. Using only data for the last 400 years, *Donarummo et al.* [2002] compared the phase of the 11-yr variation to that of the sunspot cycle. The phase showed a tendency to reverse during periods following explosive volcanic eruptions, when the upper tropospheric H_2SO_4 content was greatly increased. They attributed the 11-yr dust variations in years without volcanic H_2SO_4 to increases (at solar minimum) of electroscavenging, ice production, precipitation and soil moisture. They attributed the 11-yr variations with opposite phase, in years with high volcanic H_2SO_4 to a greater effect of enhanced IMN that (at solar minimum) inhibited precipitation, with resulting low soil moisture.

The continental cloud cover variations over Russia and the USA have opposite variations with GCR flux to those over the oceans, that we attribute to ion-mediated nucleation and electroscavenging. It is of interest that both *Udelhofen and Cess* [2001] and *Veretenenko and Pudovkin* [1999] suggested that circulation changes rather than direct microphysical effects could be responsible. As discussed above, circulation changes are to be expected as a result of oceanic cloud changes, which are present over a considerably larger fraction of the globe.

9. CONCLUSIONS

Effects of space particle flux variations on clouds and climate have been reviewed in the context of two proposed mechanisms - electroscavenging and ion-mediated nucleation. The particle flux changes directly affect atmospheric concentrations of ions, and indirectly affect the density of space charge in the troposphere via modulation of current flow in the global electric circuit. The two proposed mechanisms respond to changes in space charge and/or ion concentration, and are likely to affect properties of clouds such as droplet size distributions, cloud albedo, infrared opacity, precipitation rates, cloud lifetimes, and hence fractional cloud cover. These cloud and precipitation changes in turn affect atmospheric temperature and dynamics. The effectiveness of the two mechanisms depend on latitude as well as altitude, cloud type and atmospheric sulfate content.

Considered together, the two proposed mechanisms have the following strengths and weaknesses. Strengths:- the observations of the decadal variations of clouds and atmospheric dynamics are consistent in a number of details (such as variation with latitude, and effects related to the presence of atmospheric sulfate) with one or other or both mechanisms. There is adequate available energy and relative amplitude to drive the observed atmospheric changes. Weaknesses: - there is no decisive result at present to determine how much of the observed decadal variations are due to particle flux inputs as compared to total or spectral irradiance changes. (However, there is no such ambiguity concerning the correlations of atmospheric dynamics with particle fluxes on the day-to-day timescale.) Quantitative evaluations of various aspects of the cloud microphysics have not yet been made. For example, for IMN the growth of the ultrafine particles to CCN size has not yet been modeled, nor has the effect of changes in droplet size on eventual cloud cover. For electroscavenging, the cloud processes linking ice production rates with precipitation rates for various types of clouds has not yet been modeled. Field measurements on the global circuit as well as on cloud microphysical parameters are needed, together with laboratory measurements of IMN and electroscavenging.

Clearly there is a great deal of modeling that is needed in order to provide quantitative relationships between atmospheric ionization and macroscopic clouds properties. However, models of the radiative and dynamical consequences for climate of estimated precipitation and cloud cover changes could be made with present capabilities. Improved cloud cover and precipitation data covering more solar cycles would be useful for validating the present observational results, and as more accurate inputs into global climate models.

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