UPSCALING IN GLOBAL CHANGE RESEARCH

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Abstract. This paper reviews the problems of upscaling that arise, in the context of global change research, in a wide variety of disciplines in the physical and social sciences. Upscaling is taken to mean the process of extrapolating from the site-specific scale at which observations are usually made or at which theoretical relationships apply, to the smallest scale that is resolved in global-scale models. Upscaling is pervasive in global change research, although in some cases it is done implicitly. A number of conceptually distinct, fundamental causes of upscaling problems are identified and are used to classify the upscaling problems that have been encountered in different disciplines. A variety of solutions to the upscaling problems have been developed in different disciplines, and these are compared here. Improper upscaling can dramatically alter model simulation results in some cases. A consideration of scaling problems across diverse disciplines reveals a number of interesting conceptual similarities among disciplines whose practitioners might otherwise not communicate with each other. Upscaling raises a number of important questions concerning predictability and reliability in global change research, which are discussed here. There is a clear need for more research into the circumstances in which simple upscaling is not appropriate, and to develop or refine techniques for upscaling.

1. Introduction

This paper and others in this special issue of *Climatic Change* focus on the problem of scaling upward from the point or site-specific scale at which observations are frequently made, to the grid cell size found in global models used to study global environmental change. Following Turner et al. (1990), the term "global environmental change" is meant to mean changes that are global by virtue of the fact that they involve systemic changes in the properties of the atmosphere or ocean, or changes that, although local or regional in scale, are so widespread in their occurrence that they can be regarded as global-scale problems. Climatic change arising from the accumulation of greenhouse gases and the depletion of stratospheric ozone due to the emission of halocarbons are the prime examples of the former, while tropospheric ozone buildup, acid rain, land use changes that result in degradation or destruction of the original vegetation and soils, and biodiversity and habitat loss, are the prime examples of the latter.

The models used to study global change and the corresponding "grid" cells differ widely. In the case of regional climatic change due to a given increase in greenhouse gas concentrations, the models are coupled, three-dimensional atmosphere-ocean general circulation models (AOGCMs) or atmospheric general circulation models (AGCMs) with sea ice and land surface modules. The grid

cell is a latitude-longitude element ranging from 2.5° x 2.5° to 5° x 5°. Some of the key processes that require upscaling are (i) the partitioning of rainfall into runoff, infiltration into the soil, and evaporation; (ii) the transfers of heat, momentum, and water vapour between the surface and atmosphere; and (iii) the formation of cumulus clouds and cumulus cloud systems, and their effects on grid-scale atmospheric properties. In the case of regionally-distributed terrestrial biosphere models, the standard grid cell size is 0.5° x 0.5°, which is 25 to 100 times finer a resolution than that found in AOGCMs. Processes often observed at the scale of a leaf - the regulation of stomatal conductance, photosynthesis, and the fluxes of water and carbon - have to be simulated at the scale of an entire grid cell. In the case of global economic models used to project the costs of greenhouse gas emission abatement, the analog to the grid cell - the smallest resolvable element in the model -- is an individual large nation or geopolitical block. These models have been developed from the top down using relationships already developed at the scale of the model resolution, so in a sense the upscaling has already been performed but not necessarily correctly. The opposite approach, a bottom-up approach based on the analysis of greenhouse gas abatement options at the scale of the individual or firm, requires explicit upscaling to the scale of economic units that are resolved by global economic models.

The problem of interest here - upscaling - is distinct from the problem of downscaling, which also arises in global change research. Upscaling is concerned with the development of relationships that are applicable at the grid-cell scale of models, so that they can be implemented in such models as part of the process of developing projections for the future and/or assessing vulnerabilities to a range of environmental stresses. Downscaling is concerned with taking the output of global change models and deducing the changes that would occur at finer scales than resolved by the model. The two problems are not entirely independent, however, in that common processes underlie both scaling problems. In this paper, the term "small scale" is meant to mean having a small spatial extent. This is opposite to the meaning in cartography, where "scale" refers to the ratio of map length to real-world length, such that a small-scale map pertains to a larger spatial extent than a large-scale map.

This paper is organized as follows: Section 2 identifies the major scaling problems in a wide range of disciplines, explains why they arise, and outlines solutions that have been applied so far. Section 3 presents tables which summarize the scaling problems by causal factor, and summarize the solutions that have been adopted. Section 4 closes with a discussion of research needs and the implications of upscaling for predictability and other issues of concern in global change research.

2. Upscaling by Discipline

In this section, I give examples of upscaling problems related global change that arise in the disciplines involved in global change research. In many cases, the upscaling problem consists of correctly applying point process models to a large grid cell, when there is significant spatial heterogeneity in the controlling parameters or variables, and the relationships describing the processes are highly nonlinear. King (1991) has discussed this particular upscaling problem in the context of population ecology, but his discussion is applicable to many of the specific disciplinary examples to be presented below. It is therefore useful to summarize the main points from King (1991) before proceeding with a disciplinary survey of scaling problems.

When spatial heterogeneity is combined with process nonlinearity, there are two steps required for valid upscaling: (i) correctly defining the spatial heterogeneity, and (ii) correctly integrating or aggregating this heterogeneity. King (1991) identifies four ways of performing the latter, and gives a number of specific examples from population growth models. These four methods, with slightly altered terminology, are: *(i) Use of a lumped model.* A point process model is applied to the entire domain, with no change in structure, using spatially averaged input data. This approach is equivalent to assuming that the large-scale system behaves like the small-scale system. *(ii) Use of a deterministically distributed model.* The process model is applied at (or distributed over) a large number of sites or patches within the domain, and the results are summed. The patch sizes and/or distribution are chosen such that the variation within each patch is small. This approach requires that one know the specific spatial distribution of the input parameters. *(iii) Use of a statistically distributed model.* Rather than applying the point process model to real patches in physical space, the model is applied to a set of patches whose associated parameter values span the range of values found in the entire domain. In summing the results, the area of each patch is set to be proportional to the probability of finding parameter values within the range of values represented by the patch. Thus, the point process model is integrated over a joint probability distribution function for the input parameters. This approach can greatly reduce the computational effort compared to (ii), but is valid only if there is no interaction between patches. The specific spatial distributions of the input variables do not need to be known, only their probability distributions. *(iv) By explicit spatial integration.* This solution is possible only if the equations governing the process model can be analytically integrated, which is rarely the case. Consideration of those special (simplified) cases where analytical integration is possible highlights the fact that correct upscaling of nonlinear process models leads to a change in the model structure.

The foregoing discussion is applicable only to those scaling problems that arise due to the combination of spatial heterogeneity and nonlinear processes; as discussed below, the need for upscaling can arise for other reasons as well. Furthermore, if scale-dependent phenomena arise, then the above methods are

not applicable. Rather, the model has to be reformulated so as to be applicable at the larger scales.

In addition to these four explicit methods of upscaling, upscaling is sometimes done implicitly. *Implicit upscaling* can be defined as follows: a parameterization is altered to use parameter values (usually thresholds) that are not the correct, physically justified values when applied to a single point, in order to account for subgrid scale variability. An example of implicit upscaling is the practice in AGCMs of allowing clouds to form in a grid box at relative humidities of less than 100%, or of allowing some snowfall to occur prior to the temperature dropping to below the freezing point.*

2.1   PHYSICAL LAND SURFACE - ATMOSPHERE MODELLING

The treatment of atmosphere-surface interactions and the associated scaling problems can be subdivided in three separate but interconnected realms: surface hydrology; surface-air fluxes of heat, water vapour, and momentum; and boundary layer processes, cloud formation, and precipitation.

2.1.1   Surface hydrology

Scaling issues arise in surface hydrology for three fundamentally different reasons: due to the existence of strong spatial heterogeneity in surface processes and rainfall intensity combined with strongly nonlinear processes; because different processes require a different minimum scale in order to occur in the first place; and because different processes can dominate the overall system at different scales.

Spatial heterogeneity combined with nonlinearity is particularly important for interception of rainfall by vegetation, and for infiltration and runoff. Subgrid-scale variations can occur in rainfall intensity, antecedent soil moisture, soil hydraulic properties, and in vegetative properties. The usual practice in AGCMs is to distribute the precipitation during a given time step uniformly within the grid box, resulting in a drizzle that rarely exceeds the soil infiltration capacity. If a rainfall distribution were to be allowed, some regions within a given grid cell could experience saturated soil conditions and generate runoff in response to a rainfall event, while other regions, and the grid cell as a whole, might remain unsaturated. Similar arguments are applicable to soil properties and antecedent moisture. Allowing the precipitation to be concentrated on 10% of a grid cell rather than uniformly distributed can change the grid cell runoff by a factor of two (Pitman et al., 1993), and can dramatically change the simulated surface climatology (Pitman et al., 1990).

Both deterministically and statistically distributed models have been widely used for upscaling in hydrology. In climate modelling applications where

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* The term *implicit upscaling* is also used by Bugmann (this issue), but he defines it slightly differently than here.

distributed models have been applied, the statistical rather than deterministic approach has been used. Entekhabi and Eagleson (1989) combined spatial distributions of both soil moisture content and the precipitation distribution, and derived an analytical expression for the grid-averaged runoff. Johnson et al. (1993) applied this parameterization in the Goddard Institute for Space Studies AGCM. Successful upscaling required adjusting the parameters describing the spatial distribution of rainfall on each time step based on whether synoptic and convective rainfall was occurring, and required accounting for the grid size of the AGCM. The parameter values cannot readily be determined a priori, so the process of upscaling requires some degree of empirical "tuning" or arbitrary adjustment of the parameter values. However, once this was done, significant improvements were made in the simulated global and zonal mean precipitation, evaporation, and runoff.

Probability distribution functions involving rainfall amount or intensity and vegetation water holding capacity have also been used in computing grid-mean interception of rainfall by the vegetation canopy (Eltahir and Bras, 1993; Ramírez and Senarath, 1998). In current AGCMs, a fixed fraction of the rainfall is intercepted based on the vegetative properties, and this fraction is assumed to be the same everywhere in the grid cell. Current AGCMs tend to predict too much interception. Allowing for the interception to depend on rainfall amount or intensity, and allowing this to vary within the grid cell, significantly reduces the interception of rainfall. Correct computation of the subsequent evaporation of intercepted water requires distinguishing between wet and dry canopy fractions in computing the canopy temperature, rather than using a single canopy temperature to drive evaporation (Taylor, 1995).

The second scaling problem in surface hydrology is that different processes require a different minimum scale in order to occur. Surface runoff generation in nature involves two distinct processes: the occurrence of overland flow ("Hortonian" runoff) when precipitation rate exceeds infiltration rate, and precipitation on saturated regions. Hortonian overland flow is a point process, whereas saturated overland flow arises when the water table rises to the land surface as a result of the accumulated upslope subsurface runoff. As such, it requires a minimum upslope catchment area before it can begin. However, Entekhabi and Eagleson (1989) were able to incorporate these two separate mechanisms and their changing relative importance at the model grid scale in their parameterization of runoff generation. As noted above, this parameterization accounts for subgrid-scale variation in precipitation intensity and antecedent soil moisture, but assumes a horizontal surface. Wood and Lakshmi (1993) suggest that it is possible to account for topographic effects on the relative importance of Hortonian and saturated overland flow through the use of a statistically distributed model based on elevation. Thus, Entekhabi and Eagleson's (1989) scheme could be applied within each of several elevation intervals, with appropriate parameter values in each interval.

The third scaling problem is that the dominant process can change as the scale under consideration changes. Thus, matrix flow - the flow of water through the pores within the bulk of a soil sample - gives way to preferential flow in certain concentrated regions when viewed at a larger scale (Blöschl and Sivapalan, 1995). Under these conditions, the appropriate hydraulic conductivity at any given depth is not a simple linear average of the horizontally varying hydraulic conductivity (Binley and Beven, 1989). This constitutes an alternative to distributed models – the use a lumped model with *effective* parameters whose values are not simple arithmetic averages of the sub-grid parameter values. A problem with this approach is that there are parameters (such as soil hydraulic conductivity), where no single effective value works for all soil moisture conditions (Blöschl and Sivapalan, 1995). Furthermore, the appropriate spatially averaged or effective values of quantities such as soil density, porosity, and hydraulic conductivity depend on the scale under consideration. Scaling laws based on the use of fractals have been developed that appear to do well in predicting this scale-dependence for scales ranging from individual grains of sand up to small field plots (Crawford, 1994).

2.1.2   *Surface-air fluxes of heat, trace gases, and momentum*

In global-scale climate models, the vertical fluxes of heat, water vapour, and momentum between the land surface and atmosphere are usually computed using relationships that were developed at the scale of a few square meters to tens of square meters. These relationships are applied to the entire grid cell, without modification, using the grid-average parameter values as inputs. A number of researchers have compared the grid-scale fluxes computed by this lumped approach with fluxes computed using a statistically distributed model. Wood and Lakshmi (1993) concluded that use of grid-average leaf area index (LAI), precipitation, and initial soil moisture is valid for computing areal mean evaporation over the Amazonian rainforest for a case involving realistic spatial variability in these parameters. Analysis of observed data collected from the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) by Sellers et al. (1995) indicates that, in scaling from 30 x 30 m cells to an area of 2 x 15 km, areal mean LAI, canopy resistance, and soil moisture can be used in computing areal mean evapotranspiration. They note that the evaporation-soil moisture relationship is close to linear at soil moisture contents greater than 50% of field capacity (so that simple aggregation of variability above this threshold should work), but that the spatial variability in soil moisture decreases as the soil dries, so that the nonlinearity at relatively low soil moisture is not important at this field site. In contrast, Bonan et al. (1993) and Li and Avissar (1994) concluded that subgrid variabilities in LAI, stomatal resistance, and soil wetness can be important for the larger scale vertical sensible heat and moisture fluxes. Bonan et al. (1993) note that the true statistical distribution of these parameters is unknown, so the importance of subgrid-scale variability for global climate simulations cannot yet be assessed. Arola and

Lettenmaier (1996) compared simulations of areal mean snow depth and the vertical sensible and latent heat fluxes in the mountainous terrain of Montana (USA) as computed using coarse- and fine-resolution grids. They found that the coarse-grid model could closely replicate the fine grid results if it was integrated using a probability distribution function based on altitude bands, but otherwise did poorly. The main reason the coarse-grid model did poorly without using a probability distribution is that, in the high-resolution model, snow persisted at the highest elevations - something that could not be captured by the coarse-resolution model.

Liston (1995) used a high-resolution (50 m horizontal grid cells) atmospheric model covering a horizontal 10 km x 10 km domain to compute the energy available for melting snow as a function of the fractional area covered by snow and of the snow patch size. He found that performing separate energy balance calculations for snow-free and snow-covered conditions, and area-weighting the results, gave very accurate areal-mean energy fluxes. This is because the large spatial variation in the energy available for snow melt between the upwind and downwind sides of a snow patch (due to horizontal advection) largely cancels out over a large enough area. This procedure, which is equivalent to lumping all the snow-free and snow-covered areas together and treating them separately, is what is used in AGCMs. Thus, based on this study and that of Arola and Lettenmaier (1996), it appears that a probability distribution function is needed to account for the effects of elevation but not the effects of advection when computing snowmelt. However, in order to correctly partition the snowmelt between changes in snow depth without reducing the extent of snow cover, and changes in snow extent, a spatial distribution in melt rates or in pre-existing snow depths should be assumed. Similar procedures are required in modelling the melting of sea ice, as will be discussed later.

Not all researchers have used statistically distributed models in climate modelling applications. Seth et al. (1994) used a mixture of deterministically and statistically distributed models. They subdivided GCM grid cells into subgrid cells in order to specify the variability of vegetation type within the GCM grid cells (a deterministically distributed model), then randomly applied the GCM grid cell precipitation to the subgrid cells with a statistical distribution of intensities. In previous approaches combining spatial variability of vegetation type and rainfall, all of the points with a given vegetation type were combined into a single subgrid cell, whereas in the Seth et al. (1994) scheme, several subgrid cells could have the same type of vegetation but different amounts of soil moisture. In agreement with some of the studies cited above, Seth et al. (1994) find that allowing for sub-GCM-grid scale variability can significantly alter the mean energy fluxes (by up to 30%) and runoff (by up to 50%). Giorgi (1997a,b) has pursued this hybrid deterministic-statistical approach further.

As in the treatment of surface hydrology, an alternative to the use of deterministically or statistically distributed models is to apply the point process model to the entire grid cell, but using effective parameter values (the lumped

approach). An important parameter in the computation of the vertical momentum and sensible and latent heat fluxes is the aerodynamic resistance, which depends on the roughness of the surface (rougher surfaces generate more turbulence, so the "resistance" to vertical transfers is smaller). Considerable attention has been devoted to how to best compute effective values of the resistance or surface roughness (e.g.: Mason, 1988; Claussen, 1991; Chenbouni et al., 1995; Raupach and Finnigan, 1995). The important point of general interest is that a simple linear weighting in general does not give the correct large-scale parameter values. A second alternative is to develop a new model structure that is applicable at the larger scale, an approach adopted by Wetzel and Chang (1987). They developed a parameterization to account for regional variability in soil moisture when computing the evaporation flux. This lead to a relationship between mean soil moisture and evaporation which is different than in the original model.

As noted above, the surface roughness is a key parameter that enters into the computation of the vertical fluxes of momentum and heat. The effective roughness height needed in order to compute the vertical flux of momentum between the surface and atmosphere at a scale of 10 km can be an order of magnitude larger than the local value. This is due to the existence of drag on dispersed obstacles covering a small fraction of the grid cell. However, the effective roughness height for heat (and water vapour) tends to decrease with increasing scale since heat transfer is not concentrated on obstacles. Thus, the way in which the effective roughness varies with scale is different for the momentum and heat fluxes. This complicates the effective-parameter approach to the upscaling problem. Beljaars and Holstag (1991) present data on the observed change in the effective roughness heights in going from point to 10 km scales for one particular terrain type; however, the variation in roughness heights with scale, and in particular, the variation when increasing to the GCM grid-cell scale (several 100 km) may depend on the degree of landscape heterogeneity at different scales. The essence of the scaling problem in this case is that the surface looks different at different scales.

2.1.3   Free-air vertical heat fluxes and interactions with clouds

The preceding section dealt with the computation of the vertical fluxes of sensible and latent heat immediately adjacent to the land surface at the climate model grid scale, when there is spatial heterogeneity within the grid cell. The atmospheric motions that give rise to these vertical fluxes are turbulent in nature, and thus involve random motions that have no preferred horizontal spatial structure. As one moves a few tens to hundred of meters above the surface, however, organized air motions on a scale of one to several tens of kilometers can arise. These are referred to as *mesoscale* motions, and are intermediate in scale between turbulence and those motions that can be resolved by GCMs (Avissar and Pielke, 1989). These motions are particularly dependent on surface heterogeneity, as variations in surface temperature can produce regions of concentrated uplift (convection), separated by regions of subsidence (sinking

motion). The main surface characteristics that can cause variations in surface temperature are albedo and soil moisture content.

The importance of mesoscale circulation depends on the typical spatial extent of surface heterogeneities, as well as on the larger scale atmospheric conditions (such as stability and wind velocity). In particular, a minimum spatial extent of about 10 km for contrasting surface types is required to set up mesoscale motions; below this scale, turbulent motions are dominant (Chen and Avissar, 1994). Given a knowledge of the length scale of subgrid variability within GCM grid cells, and of larger scale atmospheric conditions, it might be possible to parameterize the effects of mesoscale motions, as suggested by Avissar (1995). Lynn et al. (1995) and Zeng and Pielke (1995) present initial attempts at developing parameterizations of the effects of mesoscale motions on the vertical sensible and latent heat fluxes. Unlike the case of turbulent heat fluxes, where grid-mean surface albedos can be used (Li and Avissar, 1994), the spatial variability of surface albedo must be taken into account when attempting to parameterize the vertical fluxes due to mesoscale motions (Pielke et al. 1993).

Interactions with clouds complicate the picture further. Even if the net vertical mass flux due to mesoscale motions is zero at the GCM grid-scale, there will be a net effect of such motions on the radiative heating and cooling if clouds form in regions of rising air motions. Since precipitation in rising columns is not cancelled by processes in sinking columns, the effect on the moisture field is also non-zero. Indeed, once localized clouds begin to form, the mesoscale and larger scale motions could be profoundly altered. Avissar and Liu (1996) found, in comparing heterogenous and homogenous domains covering an area of about 1000 km^2 that mesoscale motions induced by surface heterogeneities can cause areal-mean precipitation to be up to four times greater than for the homogenous case.

Interactions with clouds and the generation of precipitation were not considered in the parameterizations of the effects of mesoscale motions developed by Lynn et al. (1995) and Zeng and Pielke (1995). Wetzel and Boone (1995) proposed a parameterization for the effect of surface heterogeneity on non-precipitating cumulus clouds, using either deterministically or stochastically distributed surface patches coupled to a single atmospheric layer that covers all the patches. However, their scheme does not include the effects of mesoscale motions. Thus, although one may be able to successfully aggregate surface heterogeneity in computing grid-cell mean sensible and latent heat fluxes, a much more difficult - and as yet unresolved - scaling problem remains when it comes to interaction between the surface and clouds. It might be that the effect of mesoscale motions is simply to change the distribution of precipitation but not the total precipitation within a GCM grid cell, but the conditions under which this is the case remain to be defined.

A second scaling problem for clouds arises due to the existence of spatial variability within atmospheric grid boxes, involving humidity. At a given point, condensation requires a relative humidity of 100%. Upscaling to the grid-cell is

accommodated in atmospheric models by setting the threshold for cloud formation at something less than 100% relative humidity, to account for the fact that parts of the cell can be saturated even when the mean relative humidity is less then 100%. Subgrid-scale variations in relative humidity might also be important to the grid-average radiative properties of aerosols (Haywood et al., 1997; 1998; Ghan and Easter, 1998).

A third scaling problem related to clouds involves accounting for the effect of ensembles or collections of cumulus clouds within a single grid cell. A cumulus cloud modifies the air around it through detrainment of water vapour and liquid water at the top of the cloud, and through the induced subsidence of air adjacent to the cloud. The mass flux and vertical extent of each cloud, however, depend on the large-scale atmospheric conditions. At any given time there will be a collection of cumulus clouds of varying thickness and hence varying heights of detrainment. The presence of each cumulus cloud in an ensemble influences the occurrence and characteristics of all the other clouds through its effect on the large-scale atmospheric conditions. The net effect is therefore not given by the direct effect of some "average" cumulus cloud in the ensemble. The essence of this scaling problem is (a) the existence of a spectrum of clouds of differing characteristics, and (b) the existence of feedback between the clouds and the large-scale environment. In spite of the complexity of this scaling problem, Arakawa and Shubert (1974) developed a parameterization that accounts for the mutual interactions of an ensemble of cumulus clouds. The key to this upscaling parameterization is the disparity in time scales between that of cumulus convection (1 hour) and that of the large-scale driving atmospheric conditions (days). This disparity implies that the ensemble of cumulus clouds must stay nearly in balance with the large-scale variables. Pan and Randall (1998) developed this approach further by allowing for an adjustable degree of disequilibrium between the cloud ensemble and the large-scale environment. The degree of disequilibrium serves as a new parameter that is introduced by the process of upscaling.

A fourth scaling problem with respect to clouds arises from the fact that cloud albedo (α_c) and emissivity (ε_c) depend nonlinearly on the vertically integrated cloud liquid or ice water content (LWC) (Stephens, 1987; Ramaswamy and Chen, 1993), and LWC can vary substantially within a GCM grid cell. Taylor and Ghan (1992) point out that one cannot parameterize the grid-mean α_c in terms of the average LWC because the average LWC is dominated by thick clouds, while thin clouds contribute relatively more to α_c. Cloud feedbacks on climatic change depend on how LWC changes with climate and on the rate of change of α_c and ε_c with LWC. Because of the nonlinearities involved, computation of the net cloud feedback using the mean cloud LWC and using a probability distribution of LWCs will yield different results, even for the same grid-mean change in LWC, if the within-grid cell variation in LWC is sufficiently large. The importance of these nonlinearities is demonstrated by a

sensitivity experiment performed by Senior and Mitchell (1996). They changed the assumed shape of the statistical distribution of cloud liquid water path in the grid cells of an AGCM in such a way as to slightly improve the simulation of present-day radiative fields, and in so-doing, increased the global mean model response to a CO_2 doubling from 3.4 K to 5.4 K.

Considine et al. (1997) developed a parameterization for the spatial variability of LWC in marine boundary layer clouds. The development of parameterizations for the spatial variability of LWC (or other cloud properties) requires some *hypothesis* about the mechanisms causing spatial variation. Considine et al. (1997) assumed that the spatial variation in LWC is related to the spatial variation in the relative humidity of air rising below the cloud base, which in turn leads to a spatial variation in the height of the cloud base and in the vertically integrated LWC if one makes the additional assumption that the cloud top height is spatially uniform and fixed by an overlying inversion. Cloud fraction can be deduced from the integral of the LWC probability distribution function. Considine et al. (1997) specified the diurnal variation in the mean height of the cloud base and used a time-invariant standard deviation of cloud base height, and in this way were able to replicate the observed diurnal variation in mean cloud fraction and in the LWC distribution. Since models have been built which can simulate the observed diurnal variation in the spatially averaged cloud base height, the possibility of successfully linking these two predictive components exists. It will still be necessary, however, to determine how the cloud top height and the standard deviation of the cloud base height should vary with model-predicted, grid cell-scale vertical heat fluxes.

The fifth, and final, scaling problem involving clouds arises from the occurrence of partial cloud cover within a horizontal grid cell. The simplest approach for dealing with partial cloud cover is to compute radiative fluxes separately for cloud-free and cloud-covered cases, and to use a simple areal weighting to give grid-mean fluxes. Apart from inaccuracies arising due to radiation from cloud sides, this approach is inapplicable when – as is usually the case – there are clouds in several model layers, with varying degrees of overlap. The alternative is to scale the cloud optical thickness in each layer in some way according to the areal fraction. Chou et al. (1998) provide a recent analysis of this approach, which amounts to finding an *effective* optical depth for each cloud layer. The appropriate optical depth is different for diffuse and direct beam radiation, and depends on other conditions (such as solar zenith angle). These limitations to the effective-parameter approach to scaling are similar to the limitations encountered in other fields (e.g.: surface hydrology), as discussed earlier.

2.1.4   *Simulating the impact of widespread deforestation*
In assessments of the climatic impact of deforestation using climate models, the deforestation has been applied to entire model grid cells. Deforestation in reality proceeds as a growing patchwork of deforested areas, with secondary regrowth

on abandoned patches (O'Brien, 1996; 1999, this issue). As noted in the review by Pielke and Avissar (1990), the juxtaposition of transpiring vegetation next to dry, bare land can generate circulations as strong as sea breezes. As noted above, the development of mesoscale motions requires a minimum spatial scale for surface heterogeneities of about 10 km. Hence, the local and any possible large-scale effects of deforestation on atmospheric circulation could very well depend on the small-scale structure of deforestation. Since this structure is likely to change as deforestation changes, the local and possible global effects of deforestation could very well change in an abrupt manner.

2.2   ECOLOGICAL MODELLING

Upscaling issues arise in the computation of carbon exchange between the land surface and atmosphere under present atmospheric conditions, in computing the response of the coupled water and carbon fluxes to higher atmospheric CO_2 concentrations, in simulating the response of terrestrial ecosystems to disturbances, and in simulating plant-animal-environment interactions on land and in the sea and hence in projecting their response or assessing their vulnerability to climatic change. Each of these areas is discussed below.

2.2.1   Net photosynthesis
The net flux of carbon between plants and the atmosphere, which depends on photosynthesis and respiration, is closely associated with the vertical fluxes of heat and moisture. A number of land surface models have been developed which simulate these coupled fluxes in an internally consistent manner (Bonan, 1995; Hunt et al. 1996; Sellers et al. 1996). In the long run, such land surface models will permit computation of feedbacks between climatic change and the terrestrial component of the carbon cycle. At present, such models have been used to evaluate the present-day global distribution of net photosynthesis using a lumped approach, namely, applying leaf or canopy-scale relationships to grid squares 1° x 1° in size or larger (Hunt et al. 1996; Zhang et al. 1996).

Pierce and Running (1995) assessed the errors incurred in using the lumped approach to estimate average net primary productivity (NPP) over an area of 110 km x 110 km in western Montana (about 1° x 1°). The effects of averaging the smaller scale variation in climate, topography, leaf area index, and soil water holding capacity were determined by comparing the lumped results with results obtained with a distributed model. The lumped approach gave areal-mean NPP that differed by 15-30% from the distributed approach, depending on the season.

Computation of rates of photosynthesis also requires summing over vertical heterogeneities and nonlinearities. The rate of photosynthesis depends primarily on the available light and the photosynthetic capacity, the latter varying roughly linearly with the leaf nitrogen content. If photosynthetic capacity is assumed to be uniformly distributed through the canopy, then, as the incident light at the top of the canopy increases, the leaves at the top of the canopy will become saturated

while leaves lower down are net yet saturated. That is, with increasing light, the rate of photosynthesis at the top of the canopy will begin to level off while continuing to increase in the lower canopy. As a result, canopy-scale photosynthesis saturates more slowly than leaf-level photosynthesis at the top of the canopy.

If, however, the photosynthetic capacity varies vertically in direct proportion to the average available light, then photosynthesis will saturate at the same rate at all levels in the canopy. Thus, the canopy-scale curve of photosynthesis versus light will have the same shape as the leaf-scale curve. There is abundant evidence that this is indeed the case, and Sellers et al. (1992) used this to develop a method for scaling leaf-scale processes (photosynthesis, dark respiration, stomatal conductance) to the canopy scale. In this scheme, the canopy-scale processes are computed as the product of (i) leaf-scale biophysical parameters (such as maximum catalytic capacity) and the available light at the top of the canopy; (ii) a factor that involves the vertical structure of the light penetration (which can be computed from the specified vertical profile of leaf area); and (iii) factors that depend on environmental variables within the canopy (temperature, relative humidity, CO_2 concentration), all of which are assumed to be vertically uniform. This upscaling scheme is upscaling by analytic integration, and is possible because nonlinearities that are otherwise present have been removed by assuming that environmental variables (other than light) are constant within the canopy. The bulk scheme gives canopy-scale process rates that are within a few percent of an exact calculation involving 80 layers, if the exact scheme uses vertically uniform environmental variables (other than light). When vertical variation in these variables is allowed, the results are still generally within 10% of the bulk results. Kruijt et al. (1997) performed further tests of this upscaling technique and found that it gives accurate estimates of canopy-scale photosynthesis compared to a detailed, vertically-resolved model. This upscaling technique has been incorporated in a land-surface module for use in AGCMs (Berry et al., 1997).

To summarize, it appears that a distributed model is required for horizontal upscaling, at least in mountainous terrain, but that canopy-scale parameters can be defined for vertical upscaling using a leaf-scale process model. The vertical upscaling technique is the only example discussed in this paper of upscaling by analytic integration. In common with other upscaling techniques, it hinges on some hypothesis concerning the behavior of the system in question. In this case, the hypothesis is that the vertical distribution of nitrogen is fully acclimated to the average availability of light or, equivalently, that the plant has optimized the vertical distribution of a key limiting nutrient. As in the Arakawa and Shubert (1974) parameterization of the effects of an ensemble of clouds, a disparity in time scales is also critical to the upscaling. In the canopy case, the time scale for variation in light intensity (hours) is much shorter than that for distribution of nitrogen (many days or weeks), so that the nitrogen distribution can be assumed to be adjusted to the long-term light distribution, in the same way that the

ensemble of clouds is assumed to be adjusted to the slowly-varying, large-scale forcing.

2.2.2   Computing the distribution of light within a forest canopy

In the preceding section, the method of Sellers et al. (1992) for scaling photosynthesis from the leaf to the canopy was discussed. This method requires, as input, the vertical variation in the flux of solar radiation. Various approaches have been used to compute this term in canopies. The greatest complication arises from sub-canopy scale structure in the vegetation canopy, so that relationships that assume a very large number of consistently oriented scattering and absorbing elements (e.g.: Beer's Law) are not applicable. As in the computation of field-scale soil hydrological properties or grid-scale cloud properties, description of the canopy structure in terms of fractals facilitates the scaling process (Knyazikhin et al., 1998). However, this is a scaling problem with a twist: the problem is not one of applying a relationship applicable at a single point (i.e.: a leaf) to a large scale (i.e: a canopy), but rather, one of taking a relationship that is valid in a statistical sense only for a very large number of elements and applying it to cases where there is spatial structure in the scattering elements.

2.2.3   Physiological response of plants to higher atmospheric CO_2

In Section 2.2.1 I noted the need for upscaling in the computation of the coupled heat, moisture, and carbon fluxes using current land surface modules. Upscaling is required across both horizontal and vertical heterogeneity. The fluxes of both water vapour and carbon are, however, directly affected by the atmospheric CO_2 concentration, and the need to correctly incorporate this effect at the model grid scale introduces yet further upscaling problems.

As discussed in the preceding section, vertical upscaling can be performed analytically if (i) the vertical distribution of leaf area (which affects the vertical distribution of available solar radiation) is specified, (ii) the vertical distribution of nitrogen in the canopy is adjusted to the average distribution of light, and (iii) all other environmental and plant variables are vertically uniform. This scheme works very well in predicting canopy-scale process rates for present-day conditions, but it does not automatically follow that the response to changes in atmospheric CO_2 concentration or temperature will be accurately projected. For one, temperatures can vary by up to 4 K within a canopy, so different parts of the canopy will be at different points relative to the optimum temperature and might therefore respond differently to a uniform change in temperature. Reynolds et al. (1992) investigated the importance of these and other effects using a multi-layer canopy model of scrub oak - in effect, a deterministically distributed model. Not surprisingly, a single layer model using average leaf properties did very poorly in simulating absolute canopy-scale rates of photosynthesis and evapotranspiration

for either present or doubled CO_2. However, the relative response of an entire plant to a CO_2 doubling differed little between distributed and lumped models.

The second way in which scaling of the photosynthetic response can be problematic is through *feedback* between the plant and the surrounding environment. This can arise in at least two ways. First, if higher CO_2 concentration leads to a greater leaf area in the upper canopy layers, this will reduce the availability of light in lower canopy layers. The response of the entire canopy will not be equal to the scaled response of an individual leaf, except possibly by chance. It is not clear, however, that higher CO_2 will lead to a greater LAI (Körner, 1996). Second, feedback between the leaf and canopy scales has been shown to be important in modulating the response of evapotranspiration, at the scale of the canopy, to increasing CO_2. At the scale of the leaf, higher CO_2 tends to reduce evapotranspiration because it leads to partial closure of the stomata. However, in ecosystems such as closed forests, where the vegetation strongly influences the air humidity next to the plant, the effect of stomatal closure in all the leaves of the canopy is to reduce the canopy humidity, thereby tending to drive evapotranspiration rates back up (Jarvis and McNaughton, 1986). This feedback will be accounted for in models that explicitly represent the temperature and water vapour pressure within the canopy air and allow for a variable coupling with the overlying atmosphere. The feedback between the leaf and surrounding environment can extend to still-larger scales through the effect of warmer surface temperatures (resulting from less evapotranspiration) on the development of the planetary boundary layer and on mesoscale circulations (Field et al., 1995). This feedback is harder to incorporate, as discussed in Section 2.1.3.

The third way in which scaling can be problematic is as a result of feedback or interactions between different plant components and with light. The initial stimulation of photosynthesis at the scale of the leaf leads to changes in the allocation of carbon and nitrogen to roots and shoots, which then feed back to the leaves and alter the photosynthetic response at the scale of the plant by changing the availability of nutrients. The response is different still at the scale of the canopy, due to feedback between plants and the availability of light. Reynolds et al. (1993) investigated these effects by linking a detailed mechanistic plant photosynthesis model to a stand-level ecosystem model. They found, for the one case that they investigated, the following longterm increases in the rate of photosynthesis following a doubling of atmospheric CO_2: 18% at the leaf level, 16% at the plant level, and 20% at the stand level. Thus, feedbacks altered the plant-level or stand-level photosynthetic response by only 10% compared to the longterm leaf-level response in this case. Interestingly, growth is most stimulated by higher CO_2 under high-nutrient conditions at the plant level, but under low-nutrient conditions at the stand level. This is because, under low nutrient conditions, the stand is more open and has more light, so it is more responsive to higher CO_2.

A fourth consideration that gives rise to a scaling problem is that an atmospheric CO_2 increase will not have the same relative effect on photosynthesis or water use in all species. Consequently, the interactions among species will be altered, thereby precluding a simple upscaling.

2.2.4   Response of forests to climatic change

The response of land plants to climatic change has been assessed using a variety of different approaches, and prominent among these different approaches has been the use of "gap" models which simulate the competition between different plant species within a small (about 10 m^2) patch (see Shugart and Smith, 1996, and other papers in that special issue of Climatic Change). All except the very most recent gap models assume the immediate availability of seed stock for all the species that could potentially grow in a given region. If these models were applied simply to evaluate the response to isolated disturbances in a restricted region, this would be an adequate assumption. However, when modelling continental-scale ecological responses to continental-scale climatic changes, this assumption is inappropriate because of the lag that would occur in reality between the time when the climate becomes appropriate for the growth of a species not currently found nearby, and the arrival of seeds from the closest initial occurrence of the species in question. Thus, a scaling problem arises due to the increasing importance of temporal lags as the spatial scale over which a response must occur increases.

As discussed by Bugmann et al. (this issue), most gap models disregard disturbances other than the mortality of individual trees. In extrapolating to the landscape scale, it is assumed that the forest consists of a series of patches of different ages that develop independently of one another. Interactions between patches through fire, windthrow, or insect infestations, are neglected. Such interactions can be important to forest dynamics and to changes in forests at both large and small scales.

2.2.5   Distribution of terrestrial and marine plants and animals

A number of upscaling issues arise in terrestrial ecology pertaining to the distribution of plants and animals. Among these issues are (i) the changing importance of different controlling variables at different scales; (ii) the dynamics of heterogeneous landscapes, in particular, the role of interactions between adjacent landscape units and of landscape connectivity; and (iii) the relationship between disturbances and the large-scale ecosystem structure.

Ecological patterns depend on an interplay between large-scale and small-scale processes. Large-scale variations tend to be more controlled by factors such as environmental stress, dispersal, and productivity, while small-scale patterns tend to be controlled by competition and predation (Menge and Olson, 1990). Because the relative importance of different variables in explaining species distributions changes with scale, observations at a small scale might not correctly identify the dominant processes that generate the large-scale pattern (Root and

Schneider, 1995). A particularly striking example is provided by Neilson and Wullstein (1983), who found that oak seedling mortality in the American southwest increases with decreasing precipitation at the local scale, but at the regional scale is lowest in dry latitudes. Peterson (1999, this issue) provides another example of how the occurrence of different processes at different scales causes both local and regional ecosystem behavior to differ from that expected based on small-scale studies alone. The example involves the interaction of forest succession and spruce budworm outbreaks; at the local scale, a given path of forest development might be expected. However, as this development occurs over a large enough area, conditions conducive to an outbreak of spruce budworm arise, that alters the forest development at both regional and local scales. These local-scale impacts cannot be predicted from a consideration of processes at the local scale. This example is a case where prediction at the local scale must take into account local processes and their modulation by larger scale variables.

A concrete example of how an ecological model that spans scales can be developed is provided by Rothschild and Osborn's (1988) study of predator-prey interactions in marine plankton communities. A key parameter in this ecosystem is the rate of contact between predator and prey species. This depends not only on species densities, but also on turbulence and its characteristic length scales. Large-scale patterns in plankton communities depend, in part, on the large-scale variation in the nature of small-scale turbulence. As suggested by Rothschild and Osborn (1988), it should be possible to parameterize the large- scale variation in predator-prey contact rates in terms of the large-scale variation in the parameters (such as wind velocity) which determine small-scale turbulence.

In the plankton example given above, the fine-scale detail does not need to be explicitly represented, although its statistical characteristics must be determined. A general problem in linking across scales is to determine the extent to which fine-scale detail can be neglected. Some detail is just noise, but in other cases, emergent phenomena can arise from the collective behaviour of small-scale processes (Levin, 1992).

Another example of a cross-scale model is the Everglades model of DeAngelis et al. (1998). In this case, models of individual predator organisms are linked to models of the statistical distribution of prey species within a spatially varying landscape. Predator species (birds) are modelled at the level of individuals, rather than at the aggregated population level (as for prey species) because of the dependence of the predator species on patchy resources and their ability to travel long distances in foraging flights. The dynamics of such organisms, in a heterogeneous and changing environment, could not be captured if modelled at the population level.

Spatial heterogeneity combined with dispersal alters the dynamics of species interactions in a number of ways (Levin, 1976). First, different combinations of species can be favored in local regions and maintained elsewhere, thereby increasing overall species richness. Second, it allows nondominant competitors to

survive. Third, dispersal allows exploitation of resources that are locally unreliable but regionally dependable. Since dispersal can be episodic and opportunistic (by capitalizing on ephemeral resources), there is a time as well as a spatial component to the dynamics and hence to the upscaling problem.

Thus, the response of animal species to climatic change will surely involve interactions between adjacent landscape units, as well as direct biotic-abiotic relationships at a fine-scale. This in turn implies that the heterogeneity of the landscape is important to the response, as also argued by Pickett and Cadenasso (1995). The input-response relationship is therefore likely to be different than that expected based on small-scale considerations alone. Landscape heterogeneity is also important since the conditions necessary for the survival of a given species might be able to persist somewhere within a grid cell even after the mean conditions are no longer suitable.

Landscape connectivity - a particular attribute of spatial heterogeneity - is also crucial to species and their response to climatic change. Even in the absence of climatic change, connectivity is critical inasmuch as the survival of populations depends on the rate of local extinctions (within patches) and the ease of movement between patches (Turner, 1989). The importance of connectivity will be amplified when rapid shifts in climatic zones occur (Schwartz, 1992). Thus, knowing only the proportion of different landscape types within a GCM size grid cell is not sufficient for predicting impacts.

The large-scale impact of climatic change will likely depend on the spatial (and temporal) scales of disturbances and on the interaction between disturbances and the small-scale structure of landscape variability. Turner et al. (1993) show how the spatial and temporal scales of disturbance influence the large-scale statistical properties of a landscape (for example, the proportions of the landscape in different successional stages). Conversely, the effect of a change in the disturbance regime, and how individual disturbances propagate, depends on the spatial arrangement of patches that are susceptible or resistant to disturbances (Turner et al., 1989). Landscape heterogeneity can enhance or retard the spread of disturbances. Thus, the impact of changes in large-scale climatic parameters, which alter the disturbance regime, will depend on subgrid-scale landscape variability.

Landscape heterogeneity may further complicate the large-scale response to changes in the disturbance regime when multiple disturbances occur. However, the interactive effect of multiple disturbances across large landscapes is unknown, as past research has emphasized the study of single disturbances in small areas (Turner, 1989). Miller and Urban (1999) have developed a spatially explicit forest gap model that simulates between forest pattern, climatic controls, and fire.

2.3 IMPACT OF CLIMATIC CHANGE ON AGRICULTURE

Computer simulation models of the growth of crops have been developed based on climatic and soil data and observed crop growth at individual experimental sites. These models are then used to simulate mean growth rates over entire AGCM grid cells, using grid cell average climatic and soils data as inputs. As discussed by Russell and van Gardingen (1977), there is considerable spatial variability in growing conditions within AGCM-scale gird cells, and the experimental conditions used to calibrate crop models are often not representative of the average conditions encountered in the field. The general applicability of simulated crop yields, and changes in yields resulting from changing climate, is therefore an open question.

Easterling et al. (1998) used crop production models to simulate maize and wheat yields in the U.S. Great Plains during the period 1984-1992. They used observed climatic variations during this period and observed soil characteristics as inputs, and compared the simulated yields with observed yields. Simulations were first performed by applying the point process model to 2.8° x 2.8° GCM grid cells with areally averaged soil and climatic data as input. Simulations were then repeated with the original cells subdivided into progressively smaller cells (down to 0.5° x 0.5°with either disaggregation of climatic data only, disaggregation of soil data only, or disaggregation of climatic and soil data. Representation of climatic data at progressively finer scales improved the agreement between model and observations down to a scale of 1° x 1°, but no further improvement occurred for the 0.5° x 0.5° simulation. Disaggregation of soil data alone had little effect on simulated yields. Thus, use of a deterministically distributed model with regard to climatic data improved the areal-mean yields, but only up to a certain point.

2.4 PHYSICAL OCEANOGRAPHY AND SEA ICE

Physical oceanography poses a number of difficult scaling problems. First, the global-scale ocean circulation depends on mixing processes occurring at a scale of 1 m. This was first shown by Bryan (1987), who demonstrated, using a 3-D ocean GCM, that the intensity of the thermohaline overturning circulation depends on the value of the subgrid-scale diffusion coefficient. This coefficient, which is a prescribed parameter in ocean GCMs, is meant to represent the effect of vertical mixing processes that have a typical scale of 1 m. Diffusion is also used to represent the effects of horizontal mixing, but in this case the mixing involves eddies with a spatial scale of about 50 km. In both cases, the diffusion parameterization is an implicit upscaling that assumes the existence of an emergent property. However, the diffusion parameterization does not work well in the case of horizontal mixing. Rather, the only solution to the upscaling problem in this case seems to be to explicitly represent the subgrid-scale variability by increasing the model resolution. This has already been done with

global-scale oceanic models having a grid resolution of $1/2°$ x $1/2°$ (Semtner and Chervin, 1992).

The modelling of sea ice involves two components: thermodynamics, dealing with energy flows, the freezing of sea water, and the melting of ice; and dynamics, dealing with forces and motions. With regard to the thermodynamics, the air-sea heat flux and vertical rate of ice growth in winter are a very strong function of ice thickness for thicknesses less than 1 m (Maykut, 1978). Thin ice is continually created as ice motions expose open water. Consequently, the heat flux from thin ice covering a few percent of the grid area can dominate the grid-average heat flux. Correct modelling of grid-mean fluxes requires performing calculations over a statistical distribution of ice thicknesses (i.e.: the usual statistically distributed model). Harvey (1988) applied this approach in modelling sea ice in a zonally averaged climate model, but it has yet to be used in an AOGCM (where, admittedly, the need is somewhat smaller). It is also important to allow for a statistical distribution of ice thicknesses when computing vertical melting and its effect on ice extent. Where the ice is thin enough, a given vertical melting will lead to the complete disappearance of sea ice, whereas there could be no change in ice extent if the ice thickness is assumed to be the same everywhere and exceeds the average thickness of ice melted. This issue has also been treated by Harvey (1988).

Two distinct scaling issues arise in the treatment of sea ice dynamics: (i) how the dynamic behaviour of sea ice changes with scale, and (ii) determination of the scale at which atmospheric forcing of sea ice motion is most directly applicable. Overland et al. (1995) recognize five spatial scales for sea ice, which are summarized in Table 1. The most important property relevant to the dynamics of sea ice - that is, pertaining to its motion - is the relationship between the force or *stress* exerted on the ice, and its rate of deformation or *strain*. The smallest scale, ranging from 1 cm to 1 m, is the scale at which properties such as compressive strength and elasticity can be directly measured. At the 1-100 m and 1 km scales, *effective properties* need to be applied. These properties depend on the statistical characteristics of the smaller scale features, but do not require explicit representation of each individual element (in much the same way that atmospheric diffusion coefficients bypass the need for explicit representation of each turbulent eddy). At larger scales, *emergent properties* arise. In this case, the aggregate behaves in ways that could not be predicted from a consideration of the individual components alone. The components in this case are individual ice floes, and their collective behaviour depends on floe-floe interactions. In particular, discontinuities in velocity at the boundaries between floes are important to the regional ice flow. Another factor that conditions the behaviour of ice as the spatial scale increases is that the strength of ice at a given scale depends on the weakest element at the next smaller scale. As a result, the ice behaves like a granular medium at the 0.1-1 km scale, while at a regional scale it behaves like a viscous fluid.

Table I

Spatial and temporal scales in sea ice, based on Overland et al. (1995)

Spatial Scale	Temporal Scale, Days	Selected Features
100 km	100-500	pack ice
10 km	10-100	floes assemblage, polyna
1 km	1-10	floes, lead, ridge
1-100 m	< 1	thermal cracks, melt ponds, macrocracks
1 cm - 1 m		crystals, brine pockets, microcracks

The second scaling issue involves relating atmospheric forcing to the response of sea ice. As discussed by Overland et al. (1995), the local velocity of sea ice cannot be directly related to the local atmospheric shear stress. Rather, the valid linkage is between regional atmospheric forcing and regional sea ice deformation. However, atmospheric forcing varies much more rapidly in time than the sea ice response, so the history of atmospheric forcing must also be taken into account.

A consideration of hierarchy theory indicates the existence of a fundamental problem when it comes to parameterizing sea ice dynamics at the scale of global climate model grid cells. Under hierarchy theory, properties at level n-1 can, with proper averaging and consideration of interactions, be used to mechanistically explain phenomena at level n but not at higher levels (O'Neill, 1988). Since the GCM grid-cell scale (> 100 km) is two levels removed from the ice floe scale (Table 1), care is needed in applying ice floe effective properties to climate models. Sea ice is self-similar, meaning that the geometric patterns of floes within assemblages of floes are similar to the patterns of assemblages within the regional sea ice. This self-similarity can be quantified through the fractal dimension. Overland et al. (1995) suggest that the fractal dimension could be a predictive property of sea ice models, and used in upscaling effective properties from the ice floe to climate model grid scales.

2.5 ATMOSPHERIC CHEMISTRY

Scaling problems arise in the atmospheric chemistry of compounds that have a very short lifespan in the atmosphere, which results in strong spatial variations in their concentration, and where the reaction chemistry depends highly nonlinearly on concentration. Since most compounds have a rather long atmospheric lifespan once they reach the stratosphere (several months to years), concentration variations with longitude are not particularly strong and two dimensional (latitude-height) models are adequate. However, significant upscaling problems

can exist in the case of the troposphere. The main difficulty involves NO_x gases (NO and NO_2), which have highly concentrated emission sources and have a lifespan of only a few days. The impact of emissions of carbon monoxide (CO) on ozone (O_3) depends on the NO concentration: in NO-rich environments, O_3 is produced, while in NO-poor environments, O_3 is destroyed (Crutzen, 1988). NO also affects the CH_4 oxidation pathway, with net production of O_3 in NO-rich environments and possible O_3 destruction in NO-poor environments. Both kinds of environments are encountered at present. Even fairly high resolution models (60 km x 60 km) cannot adequately represent the range of concentrations encountered in nature, so that significant errors in the projected impact of emission changes can still occur.

There are at least two distinct scale thresholds in modelling tropospheric chemistry: scaling from the scale of point emission sources to model grid cells, and scaling from the regional to the global scales. In order to address the second scale transition, one must run global-scale models, but this necessitates such coarse grids that the first upscaling transition cannot be explicitly modelled. The regional-to-global transition can be avoided altogether by computing the impact of changes in the emissions of a variety of pollutants for representative chemical conditions (e.g.: clean continental, polluted continental, clean maritime, and polluted maritime), without trying to integrate globally, as in Thompson et al. (1990). This amounts to a refusal to perform upscaling, but still yields policy-relevant information. With regard to the first scale transition, Calbo et al. (1998) developed parameterizations for the "effective" emissions from urban areas as functions of the true emissions and the local meteorology. The difference between true and effective emissions accounts for nonlinear effects in transformation and removal processes when emissions are confined to urban areas rather than artificially spread over the entire AGCM grid cell. Poppe et al. (1998) have also addressed this problem. They point out that artificial dilution of ozone precursors to an AGCM grid cell usually but not always causes the calculated production of ozone to be overestimated.

2.6   SCALING IN THE SOCIAL SCIENCES AS APPLIED TO GLOBAL CHANGE

According to Young (1994), the problem of scale has not been as prominent in the social sciences as it has been in the natural sciences. Nevertheless, a number of scaling issues that are critical to understanding global environmental change have emerged with regard to the driving forces of population growth and changes in land use, and in the economic assessment of the costs and benefits of abatement of greenhouse gas emissions. These issues are outlined below.

2.6.1 Land use and deforestation

The scaling issues that arise in the study of land use changes pertain to the scale from which the driving forces for land use change operate. This is in contrast to the scaling issues discussed up to this point, which pertain to processes and the integration of these processes over successively larger scales. Land use decisions in the developing world involve a mix of indigenous or traditional (pre-capitalist) rationales, and modern (capitalist) rationales (Meyer et al., 1992). Studies of land use change at the small scale have focused on individual households and indigenous agricultural knowledge, while regional assessments have examined broad socioeconomic conditions, in some cases mediated by local factors. In many cases the forces driving land use changes originate outside the area being studied. As in the explanation of species distributions in terrestrial ecology, there is a need to nest explanations derived at different scales of analysis.

2.6.2 Demography

There are two distinct streams of demography that are relevant to research into global environmental change. The first is applied demography, which is concerned with the derivation of statistical relationships between demographic variables (for example, between income and education). The scaling problem that arises is that census data are already aggregated spatially to some extent. This aggregation in many cases masks some of the variation that is needed in order to derive useable correlations. Conversely, the boundaries of the aggregated units, for which data are reported, might not be ideal for the particular purposes at hand. Thus, considerable effort has to be devoted to dealing with problems arising from data aggregation or unsuitable boundaries, as discussed by Gutmann (1999, this issue).

The second demographic stream is theoretical demography, which includes the projection of future demographic trends and conditions - one of the key driving factors in future environmental change. Of particular importance are projections of future population and its age structure, and projections of potential future migrations. Both population growth and migration occur in response to a variety of local- and global-scale driving forces.

2.6.3 Economics

The main concerns of economics with regard to global environmental change pertain to (1) estimating the costs of actions taken to prevent or minimize environmental changes, and (2) estimating the costs of environmental changes.

The costs of greenhouse gas emission abatement can be assessed at the project, sectoral, or macro-economic scale. If costs are first estimated at the project level, then there is a need to scale up these costs by aggregating over the whole range of projects that could be undertaken in a sector or economy. This is the classical "bottom-up" approach, and entails comparing the lifecycle costs of currently-used technologies and new, more efficient technologies that could be used in their place. These cost assessments depend on the prevailing (or

projected) prices of energy, technology, and labour. The costs estimated for individual firms or projects cannot, however, be simply aggregated. This is because the implementation of efficiency measures by a single firm cannot noticeably affect the price of energy or of other factors, but if a large number of individual firms (and consumers) implement energy efficiency (or fuel switching) measures, this can be expected to noticeably influence prices. This in turn will alter the ultimate reductions in emissions of greenhouse gases that are achieved. Thus, a scaling problem arises because of feedback between the small and large scales. This scaling problem is analogous to that of extrapolating the response of evapotranspiration (to a change in atmospheric CO_2) from an individual leaf to a forest canopy.

At the other end of the scale spectrum is the "top-down" approach, in which economic correlations derived at the national scale are used in models of a national or the global economy. Beaver (1993) reviews many such models, some of which do not distinguish between different fossil fuels or which do not distinguish between more than three or four sectors in the economy. Top-down models do not involve upscaling in the sense of having to link processes or relationships derived at a small scale and applied at a larger scale. Rather, they are based on direct observations at the larger scale. However, when top-down models are used in a predictive mode under entirely different circumstances than in the past (as indeed they are), there is an implicit upscaling. This is because the large-scale response to a change in price involves the integrated effect of a large number of small-scale actions, and this integration is assumed when large-scale price-response relationships are used. Potential error arises in that a different set of detailed response options could come into play in the future. For example, there could be a greater role of non-price induced improvements in efficiency or fuel switching due to partial removal of barriers to energy-efficient investments by government action. The preferred solution to what can now be recognized as an upscaling problem is to provide enough detail that processes (response options) at the next lower level in the economic model hierarchy can be explicitly represented. I shall return to this point in Section 3.2.

   The problem here is analogous to the prediction of the vegetative response to climatic change based on models that use large-scale correlations that obscure the real controlling factors (see Bugmann and Martin, 1995; and Loehle and LeBlanc, 1996). As previously noted, a particular question that arises in upscaling is to know when fine-scale details can be safely neglected. This case might provide a concrete example where loss of detail in the upscaling process is likely to be important to the results.

   On the other hand, some of the potential costs of greenhouse gas abatement can be assessed only at scales beyond some minimum scale of analysis. These include costs related to shifts in investment patterns and changes in the rate of growth in the productivity of labour and capital (many abatement measures entail changes in the productivity of energy, which imply changes in the productivity of other factors of production). Another scale-related issue is the question of

whether there are economies or diseconomies with increasing scale – a question that is discussed by Green (1999, this issue).

Apart from anticipated cost, another important consideration in the development of policies to reduce greenhouse gas emissions is risk. Scaling issues arise here as well. Risks related to individual investments will counterbalance to some extent at the scale of a country, but not at the scale of individual investors. This occurs for two reasons. First, the conditions which adversely affect the economics of one investment might enhance the attractiveness of an alternative investment. For example, an increase in the price of natural gas that renders fuel switching from coal to natural gas less attractive will make renewable forms of energy more attractive in places where natural gas but not coal is an option. Second, not all projects in a portfolio are likely to fail (or succeed) at the same time.

With regard to international policy instruments for greenhouse gas emission abatement, the concept of "joint implementation" or "JI" currently enjoys wide support. Under JI, countries or corporations with comparatively high emission reduction costs invest in what are supposed to be cheaper options in developing or former East-bloc countries, and receive at least partial credit for the emission reductions so achieved. JI has been critically reviewed by Harvey and Bush (1997) and Bush and Harvey (1997). The scaling-related assumption behind JI is that if individual investments are cost-effective from the viewpoint of all the individual private sector investors, then the sum of all such investments will be cost-effective for society as a whole. This assumption is unlikely to be valid because of societal costs and benefits that are not reflected in private sector costs and benefits. As Jackson (1995) has pointed out, "allowing JI to be pursued on an ad-hoc project by project basis by a heterogeneous community of private investors - far from promoting cost-effectiveness - may run the risk of hampering ... least cost paths to global warming abatement both nationally and globally". A concrete example is provided by Harvey and Bush (1997): total societal costs could be smaller if domestic greenhouse gas emissions are reduced through adoption of measures that decrease electricity use rather than being offset by investments in JI projects. An electric utility, however, might prefer the JI project because that course of action will not reduce its revenues.

Rothman (1999, this issue) raises a number of scaling issues related to the damage cost of projected climatic change: (i) The value of a given natural asset (e.g.: a lake) depends on the scale of the analysis, as this affects the availability of substitutes (e.g.: other lakes) and hence the attached value. (ii) The impact of the loss of a forest or of a decline in agriculture depends on whether one takes a local, national, or global perspective due to the potential for adjustments or multiplier effects at larger scales. (iii) The value of natural assets for society as a whole is not simply the sum of individual valuations within society, due to society's much longer life expectancy. This last point is another example where temporal considerations enter into the determination of the proper spatial

upscaling (temporal upscaling is another problem in its own right, and is also discussed by Rothman (1999, this issue)).

2.6.4   Political Science

Young (1994) identifies two scales at which political scientists have analyzed human-environment interactions. One is at the scale of small, stateless societies, while the other is at the international scale of nation states. A topic of enquiry in stateless societies are *common pool resources (CPRs)*, which are resources that are not owned by specific individuals and whose use by others cannot be excluded. Political scientists have found that the "tragedy of the commons" (Hardin, 1968) often does not occur in societies that lack a strong or any central control, and considerable attention has been devoted to explaining how self-interested actors are able to use resources sustainably in the absence of an overarching authority (e.g.: McCay and Acheson, 1987; Bromley et al. 1992). McKean (1996) gives an illuminating example of CPR arrangements governing the management of natural resources by groups of villages in seventeenth century Japan. She describes a nested control structure, where different but overlapping subsets of groups of villages have varying degrees of control over their common land resource, depending on their susceptibility to harm from resource degradation and their ability to enforce CPR arrangements. This presents a tantalizing analog to the problem of devising schemes for the international regulation of global resources by nation states, for which an overarching authority is also absent, if the insights gained from this small-scale analysis can be applied to the global scale.

At the international scale, political scientists have found that relations often do not conform to the non-cooperative logic of the prisoner's dilemma (Young, 1994). Political scientists working at this scale try to understand the basis for sustained cooperation for a range of analytically distinct situations. Young (1994) indicates that there is a need to closely compare the local and international streams of analysis, and he presents some initial conclusions concerning the applicability and/or roles of pre-negotiation, monitoring, and transparency at the local and international scales.

Another, not entirely independent, scaling issue in political science involves game theory. Here the scaling issue is that the outcome of individual decision-making when a collection of players is involved can be quite different from the decision that a given player would make in the absence of other players. That is, the decision of the group is not the average of the decisions that each player would make if acting alone. This failure of simple scaling arises because of the prevalence of strategic decision making when a group is involved, that is, trying to anticipate the response of the other players to the decision made by a given player.

Finally, an upscaling problem arises in the aggregation of individual voter preferences in the formulation of national-level policy responses to global warming, as discussed by Sprinz (1999, this issue).

3. Synthesis

The preceding discussion is summarized in Table 2, which classifies upscaling problems in global change research according to the underlying fundamental cause, and in Table 3, which summarizes the solutions adopted. I shall briefly elaborate upon these two tables below.

3.1 CAUSES OF UPSCALING PROBLEMS

A common reason for an upscaling problem is the existence of spatial heterogeneity combined with nonlinearities in the relevant processes. This problem arises in surface hydrology, the computation of surface-air fluxes of heat and water vapour, in plant physiological processes, in cloud dynamics, in atmospheric chemistry, and in ecological processes related to landscape connectivity and fire. This is the case that was considered by King (1991) and discussed at the beginning of Section 2. However, there are a number of other, conceptually distinct, reasons why upscaling problems arise.

First, it has been widely observed in marine and terrestrial ecology that different processes are primarily responsible for producing the spatial distribution of plants and animals at different scales. The same is also true with regard to human land use patterns, and for the driving factors for population growth and migration. This implies that correlations derived at one scale might not be applicable at a larger scale or to changes through time. It also implies that the simplification to a model (such as exclusion of certain processes or interactions) that is acceptable at one scale may not be acceptable at a different scale (this of course is well known in fluid dynamics). Closely related to this dependence of the dominant process on the scale of observation is the fact that sometimes a minimum spatial scale is required before a given process can operate.

Second, feedbacks can occur between the small-scale components of a system and the larger scale. This has the net effect of altering the relationship between large-scale driving factors and the aggregate response of the system. Thus, the response of transpiration from a leaf and ultimately from an entire forest canopy to changes in atmospheric CO_2 concentration is strongly modulated by the effect of transpiration on the relative humidity within a forest canopy. Similarly, the effect of a change in energy prices (through a carbon tax, for example) on energy use at the scale of an individual firm and ultimately for an entire national economy is modulated by the feedbacks of energy demand on the price of energy.

Another conceptually distinct cause of an upscaling problem is the development of emergent properties. Emergent properties arise from the mutual interaction of small-scale components among themselves, whereas the feedback causation (discussed above) involves interaction between small-scale components and larger scale variables. The most striking example of the development of emergent properties is in sea ice, where the mutual interaction of

individual ice flows imparts properties (such as viscous behaviour) at the large scale which are not found in any of the constituent components. It is tempting to draw an analogy between upscaling in sea ice dynamics and upscaling in political science - the response of a collection of individual nation states (or ice floes) to an "external" situation or forcing factor such as global warming (or atmospheric

Table II

A classification of upscaling problems based on the underlying fundamental cause

Fundamental Cause	Examples
Spatial variability+process nonlinearity	Surface hydrology, formation of clouds and precipitation, photosynthetic response to higher atmospheric CO_2, natural disturbances
Minimum spatial scale required for certain process	Runoff generation
Different processes dominate at different scales	Terrestrial and marine ecology
Feedbacks between scales	Transpiration response to higher CO_2 Economic costs of greenhouse gas emission abatement
Development of emergent properties	Sea ice
Edge effects	Sea ice, terrestrial ecology
Temporal lag dependent on spatial scale change	Response of ecosystem species composition to climatic change
Collective response with differential effects	Ecosystem response to higher CO_2 Societal valuation of impacts

stress in the case of ice floes) may be quite different from that which would occur if each nation state (or ice floe) acted in isolation.

The emergent properties that arise with increasing scale in sea ice depend largely on interactions at the edges between ice floes. Edge effects have also been identified in this review as being important to the dynamics of terrestrial ecosystems at larger scales. The edge effects in this case occur between landscape patches with different characteristics. The larger scale ecological characteristics (such as overall species diversity) also depend on dispersal from one patch to surrounding patches, so that both spatial heterogeneity and the characteristics (such as speed) of the dispersal process are critical to the system

statistical properties and dynamics at the larger scale. To the extent that the ecosystem properties at the larger scale are different from those at the patch scale and depend on the mutual interaction between different patches, this provides another example of the occurrence of emergent properties as we scale up.

An upscaling problem can also arise when there is a temporal lag in the response of a system to a perturbation, if this lag increases the larger the spatial scale over which the adjustment to the perturbation must occur. A clear example is in the response of forest species composition to large-scale climatic change, when the climatic change is so large that the climate becomes suitable for species whose seedlings are not currently available at a given site. Correct modelling of the time-dependent response at a given site *requires* scaling up to a large spatial scale so that the gradual dispersal of species into new regions can be modelled.

Finally, aggregating environmental values and the costs of climatic change across individuals to the scale of a society presents upscaling problems for yet different reasons. A key issue here is to determine how to weight different values and costs when aggregating to the larger scale; the choice of weighting scheme contains implicit assumptions concerning the distribution of income, and depends on ethical judgements - something that is not always explicitly acknowledged. The only thing that one can say with confidence is that the least defensible weighting is a uniform weighting, yet this is exactly what has been done in many cases. One can draw an analogy to the computation of effective parameter values for use in lumped models of physical systems; the correct effective value is usually anything but a uniform weighting of the individual values, and the correct weighting often changes with the conditions.

3.2   SOLUTIONS TO UPSCALING PROBLEMS

The solution adopted when upscaling is required depends in part on the underlying reason for the particular scaling problem in question. Table 3 outlines the solutions that have been adopted to the scaling problems that were identified above.

The first solution, which is not always a solution, is to ignore the problem. This is done, for example, when point process models are applied to an entire GCM grid cell without modification, or when estimated costs of climatic change are simply summed over all members of a society. In some cases this can be an acceptable approach, if the underlying nonlinearities are weak or if they fortuitously cancel. The task in this case is to determine the conditions under which the upscaling problem can be ignored.

The second solution is to apply the point process model to the entire grid cell, but to adjust certain critical thresholds. This is referred to here as implicit upscaling. An example is to allow stratiform clouds to form at a relative humidity of less than 100%.

Table III

Solutions to upscaling problems that arise in modelling of physical and biological processes

Solution	Examples
Ignore	Surface hydrology, surface-air fluxes
Adjust critical thresholds	Formation of clouds at RH < 100%
Use a distributed model	Surface hydrology, surface-air fluxes
Use a lumped model with effective parameter values	Surface hydrology, surface-air fluxes, treatment of partial cloud cover with overlap, atmospheric chemistry (effective emissions)
Parameterize interactions between distinct patches	Effects of mesoscale motions in the atmosphere
Parameterize small-scale details	Diffusion approximation
Create a new model to integrate effects of next smallest scale	Cumulus cloud ensembles Ecophysiological models
Greatly increase model resolution	Ocean GCMs
Refuse	Tropospheric chemistry (regional to global transition)

The third solution is to use a distributed point process model, whereby processes are computed for a number of distinct patches within a region, and the results summed. This is an acceptable approach for dealing with scaling problems that arise due to spatial heterogeneity combined with process nonlinearity, but is not valid when there are interactions between adjacent patches.

The fourth method of upscaling is to apply the process model to the entire heterogenous domain but to use effective parameter values that account for the heterogeneity. A problem with this approach is that the relationship between the distribution of real parameter values and the effective parameter value can depend on the state of the system, and thus can change over time. The implicit upscaling approach can be thought of as a variant of this approach, since the adjusted threshold serves as an effective threshold. In some instances where the effective-parameter approach has been used, considerable effort has been devoted to developing parmeterizations of the difference between the true and effective parameter values.

Where interactions between patches are important, it is necessary to either directly parameterize these effects, or to create a whole new model which incorporates these interactions and their effects. The parameterization approach to representing inter-patch interactions can be used when the details of small-scale interactions do not matter. The simplest example is the diffusion parameterization to represent the effects of turbulent mixing or the random dispersal of plants and animals. The diffusion approximation works at scales where the unpredictability of specific events cancels out, and the overall statistical properties can be relied upon. A more complicated example is the parameterization of the effect of organized mesoscale motions on vertical heat fluxes. This is a case involving interactions between adjacent land surface patches, inasmuch as rising motion over one patch affects the tendency for rising or sinking motion over adjacent patches. In the case of cumulus cloud ensembles, in which feedbacks between the large-scale environment and the ensemble occur, but which also involve strong interactions between the members of the ensemble, rather complicated parameterizations have been developed which bear little resemblance to earlier parameterizations that considered only a single cloud interacting with its environment. Models of species diversity and the spread of disturbances also explicitly consider interactions between adjacent patches.

When spatial upscaling involves the integration of different components from one level within a system hierarchy to a higher level, the preferred approach is to link mechanistic models of the individual components, as discussed by Reynolds et al. (1993). For example, a model of plant growth would link modules involving shoot biomass, root biomass, and carbon and nitrogen substrate pools. The submodules themselves should be based on phenomenological relationships (i.e.: relationships based on a fundamental understanding of the processes involved, rather than being purely empirical) and parameterized with data collected at that level. The model would then be validated against data collected at the scale of interest and constrained by data at larger scales. Models that include mechanisms across a wide range of levels should be avoided because, as discussed by Reynolds et al. (1993), they tend to be very complex, unstable, and difficult to verify and alter.

These principles can also be applied to the question of estimating the cost of greenhouse gas emission reduction measures at the scale of a national economy. As in models of sea ice dynamics or in the response of plants to higher atmospheric CO_2 concentration, a model hierarchy can be identified, where different levels in the hierarchy roughly correspond to different spatial scales. Here, the lowest level in the hierarchy consists of models (and measurements) of the energy use by specific technologies (e.g.: motors and commercial chillers). The next level up consists of models of individual buildings or industrial processes, where the integrated effect of individual technologies on the energy use at the scale of the building or industrial plant can be assessed. These provide estimates of cost-effective energy saving potential at the scale of the individual firm, the scale at which bottom-up assessments typically begin. The next level up

consists of models of an entire sector or of a national economy. Based on the principle in hierarchy theory that assessments at level n should be based on the integration of models from level n-1 (O'Neill, 1988, and Section 2.3, above), national level assessments should be based on the integration of models at the next lower level, which involves individual energy end uses and energy supply choices. That is, considerable bottom-up detail needs to be built into top-down models if they are to be credible, and this is now being done to an increasing extent.

The next-to-last solution to the upscaling problem that has been considered here is to simply run a model at a fine enough resolution that the important processes can be explicitly represented. This approach has been used with some success in ocean GCMs to account for the effects of mesoscale (50 km) eddies on the large-scale flow. This has also been tried in models of tropospheric chemistry, but does not entirely work because strong variations in the concentrations of important chemical species can occur inside grid cells as small as 50 km x 50 km. An alternative in this case is to compute changes for different representative chemical regions but without attempting a global integration - what I have termed "Refuse" in Table 3.

3.3   UPSCALING AND PARAMETERIZATION

The processes of upscaling and of parameterizing subgrid-scale processes in physical or biophysical models are intimately connected. As noted in the introduction to Section 2, some forms of parameterization represent an implicit upscaling, the best example being the use of a threshold for condensation of water vapor that is less than the true, physical threshold. In other instances, upscaling is achieved through the use of effective parameter values that are explicitly parameterized in terms of larger scale variables and the known spatial structure of the subgrid-scale variability in surface boundary conditions. Parameterization thus becomes a means for carrying out upscaling. At the same time, and as discussed extensively in this paper, there are many methods of upscaling that do not rely on parameterization of the difference between point and grid-scale relationships between variables. Thus, parameterization can be a form of upscaling, or upscaling can explicitly rely on parameterization, or upscaling can be done without parameterization.

However, not all parameterizations of subgrid-scale processes represent upscaling. Upscaling involves taking an observed or theoretical relationship that is applicable at the point scale, and altering the relationship so that it is applicable at a larger scale. Some parameterizations simply represent the *effects* of unresolved processes without trying to represent the processes themselves or the quantitative relationships that govern them. An example would be the inclusion of the effects of updrafts and downdrafts in a parameterization of cumulus clouds (Alexander and Cotton, 1998). Another example is the parameterization of tree growth in terms of growing degree days; no attempt has been made in this case to

represent the relevant biophysics, or to scale up what are thought to be the mechanistic interactions involved in tree growth. In summary, then, not all parameterization is a form of upscaling, and not all methods of upscaling involve parameterization.

4. Research Questions and Needs

As indicated in this review, upscaling is widely required in models used to predict or understand global environmental change. An upscaling problem can arise for a variety of conceptually distinct reasons, and a number of distinct solutions have been applied to this problem. Greater recognition of the existence of upscaling problems by researchers across the spectrum of disciplines involved in global change research should, hopefully, lead to the formulation of better models for purposes of analysis and prediction. At the same time, it should lead to greater appreciation of the weaknesses of current approaches.

For this to happen, a number of specific research needs will have to be addressed. In the physical and biological sciences, these research needs are as follows:

(1) There is a need for information on the spatial variability that exists within 1° x 1° grid cells on a global basis. This is needed for parameters such as soil moisture holding capacity, soil infiltration rate, vegetation type, and leaf area index. In order to adequately characterize the probability distribution functions, information on means, variabilities, and skewness is required. Information is also needed on the spatial covariation among variables.

(2) The conditions under which a lumped approach can be used instead of a distributed model approach need to be thoroughly investigated.

(3) For conditions in which a distributed approach is required, work is needed to determine which variables need to be represented in a distributed manner and which can be safely averaged.

(4) When a statistically distributed model approach is used and the probability distribution functions could themselves change as large-scale conditions change, the ways in which and the extent of such potential changes need to be determined. This is a possible problem in cloud modelling in particular, but is unlikely to be a problem in the modelling of surface hydrology.

In both the biological and social sciences, an important area for further research is to determine ways to properly compare studies that were carried out at different scales. Levin (1992, pg. 1953) raises the possibility that scaling laws can be developed (in ecology, at least) to allow such comparisons. There is scope for considerably more work on upscaling in all disciplines involved in global change research, but the greatest challenges may very well lie in the modelling of land-atmosphere-cloud interactions.

The need for upscaling has mixed implications for our ability to correctly predict changes at the model grid scale and larger. First, in cases such as soil

moisture and precipitation, our ability to correctly predict changes in these variables at the grid scale improves as the simulation of the present conditions improves (see, for example, Meehl and Washington, 1988). Inasmuch as correct upscaling alters the grid-mean simulated values for the present climate, and generally improves the simulation, it will improve our predictability. On the other hand, if the probability distribution functions or parameterized interactions between units, used in the upscaling algorithm, are themselves subject to change as the climate changes, then predictability will worsen if these changes cannot be correctly anticipated.

In closing, it seems reasonable to believe that much of the information needed to improve our upscaling techniques could also help in dealing with the complementary scaling problem, which I have not addressed here - that of downscaling from grid-average output data to specific points within the grid. Downscaling techniques are reviewed in Bass and Brook (1997). In particular, much of the data that will need to be collected in order to construct probability distribution functions for surface and vegetation properties can, if also retained in a spatially distributed form, be used for downscaling.

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