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ENERGY BALANCE MODEL, SURFACE

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Introduction

The climate of the Earth system is constantly changing, and scientists are interested in understanding how and why the climate of the Earth system has changed in the past, what has shaped the climate of the current Earth system, and how the climate of the Earth system might change in the future. While it is possible to collect information from geological records about the climates of the past, and to monitor the Earth system today to determine what the climate is now, these data do not allow us to understand how the characteristics and processes in the Earth system produced the climates of the past and the present. In addition, we can only speculate about the climates of the future.

In order to develop a better understanding of how the Earth system has worked in the past and how it might evolve into the future, a hierarchy of computer models has been developed as tools to study the Earth system. A computer model of the Earth system is a mathematical formulation of how that system works implemented in a computer program. Into each model is incorporated all current knowledge of how parts of the Earth system interact, after which, using data that describe the present climate or a climate of the past as a starting point, the model is employed to simulate the climate of the Earth under a wide range of conditions and assumptions.

If all of the current understanding of the climate system on all spatial and time scales were included in a computer model it would be too complex and costly to run. As a result even the most complicated models are relatively simple representations of the real Earth system. There are basically three ways that models simplify the Earth system. First, all models at some level use empirical relationships to represent complex physical processes. Second, models use and compute values of different physical quantities on a grid. The physical spacing of that grid defines the resolution of the model. Models average the physical quantities over the area of a grid box, which in the simplest models could be over an entire spatial dimension. Third, models represent processes that occur on scales smaller than the grid spacing of the model with formulas that are based on processes that occur on the larger scales. This type of formulation is called a parameterization. With all of these simplifications models cannot exactly simulate the climate system.

The earliest models were also the simplest ones, both in the number of processes included and in the spatial resolution of the model, since computer resources available to run them were limited. As computing power increased, models were able to include more of the processes that occur in the Earth system and to increase the spatial resolution to finer grids.

The simplest models average over all horizontal and vertical scales to obtain a single value for the temperature of the Earth system. Only the most basic physics acting on the largest scales is incorporated into these models, and the models can be used only in the broadest sense to understand how this physics shapes climate. The results cannot be assigned to any particular place on the Earth but only to the system as a whole.

As models become more complex their resolution increases, so that variations in latitude, in longitude, and in the vertical can be examined. With the increase in resolution comes an increase in the number of characteristics and physical processes included in the model. While this increase in complexity increases the model's ability to simulate the climate system and its change over time, there are still many characteristics and processes acting on scales smaller than the resolution of the model that are either completely omitted or are parameterized in simple ways. As a result these models have a limited ability to simulate or predict climate. However, their simplicity allows scientists to examine how the components and processes of the Earth system that are included interact, and to study how changes in those components and processes might change the climate.

The most complex models are the general circulation models. These models, developed originally for the atmosphere and now including the atmosphere, ocean, biosphere, and cryosphere, have the highest spatial resolution and demand large amounts of computer resources, though the resolution is still too coarse, for example, to resolve individual clouds. As a result many of the smaller-scale processes important to shaping the climate of the Earth system are parameterized only simply in these models.

The Surface Energy Balance

The Earth system operates close to an energy balance. This means that an equal amount of energy comes into the Earth system and goes out of it, and as a result the temperature of the entire system over a long period of time is relatively constant. However, within the Earth system there are variations over time and over space. Some of these are the result of regular cycles such as the seasonal cycle, the El Niño Southern Oscillation (ENSO), and glacial/interglacial variations, or regular changes in location such as the steady decrease in temperature from the Equator to the poles. Some are the result of random variations called natural variability, which produce, for example, the day-to-day variations we see in our weather, variations in the strength of yearly monsoons, variations in the number and the location of landfall of hurricanes, and the somewhat irregular intervals of ENSO. Some of the variations in time and space are the result of changes in surface conditions such as whether the surface is land or water, or covered by snow and ice covered. These changes in surface conditions produce changes in the surface energy balance. The changes in these surface conditions affect the amount of energy retained by the Earth system and how it is distributed within that system.

Researchers who have sought to simulate the climate of the Earth system and to understand how and why it changes over time have used models, which in all but the simplest globally averaged cases have taken into account the surface energy balance. While more recent work has included the complexity of the biosphere in the surface energy balance, the surface energy balance is most simply described by the following equation:

$$(1 - P_e)\times(1 - \alpha)\times F_{sw} + F_{ir} + F_{lw} + F_l + F_s + F_{cond} = 0 \qquad [1]$$

Here P_e is the fraction of the solar radiation that is not reflected that penetrates the surface, α the surface albedo (the fraction of radiation reflected), F_{sw} the short-wave radiation available at the surface, F_{ir} the long-wave radiation from the surface to the atmosphere, F_{lw} the long-wave radiation from the atmosphere to the surface, F_l the latent heat flux, F_s the sensible heat flux, and F_{cond} the conductive flux from below the surface

If the sum of the energy fluxes in eqn [1] does not equal zero, the imbalance of energy results in a change in temperature defined by

$$(\rho c)\,\partial T/\partial t \qquad [2]$$

where T is the temperature of the surface, t time, ρ density, and c the specific heat.

F_{sw} is the incoming short-wave radiation available at the surface. This radiation is also referred to as solar radiation as most short-wave radiation in the Earth system originates from the Sun. Most solar radiation entering the top of the Earth's atmosphere is transmitted through the atmosphere to the surface or to the top of clouds. At that point it is either reflected back through the atmosphere to space, absorbed at the surface or in the cloud where it heats the surface or cloud, or penetrates through the surface or edge of the cloud to be absorbed below.

The albedo, α, determines how much of the short-wave radiation that reaches the surface gets reflected back to space. α is expressed as a fraction ranging from 0 (no radiation reflected) to 1 (all reflected). So in eqn [1] $(1 - \alpha)$ is the fraction of the short-wave radiation, not reflected, and $(1 - \alpha) \times F_{sw}$ the amount of the available short-wave radiation that is not reflected.

P_e determines how much of the short-wave radiation not reflected back to space is transmitted through the surface. P_e is also expressed as a fraction ranging from 0 (no available short-wave radiation not reflected penetrates the surface) to 1 (all available short-wave radiation not reflected penetrates the surface). So in eqn [1] $(1 - P_e)$ is the fraction of the available short-wave radiation that does not penetrate the surface and $(1 - P_e)\times(1 - \alpha)\times F_{sw}$ is the amount of the short-wave radiation not reflected that does not penetrate the surface, or, in other words, the amount of short-wave radiation absorbed at the surface. This energy flux is always directed toward the surface, representing a gain by the surface, and is never negative.

Every object radiates energy at a wavelength proportional to the fourth power of its temperature in kelvin (K). The temperatures of the Earth's atmosphere and surface are in a range where the wavelength of the radiation they emit is in the infrared part of the electromagnetic spectrum. Since the wavelength of this radiation is longer than that coming from the Sun, it is commonly called long-wave radiation.

F_{ir} is the long-wave radiation coming from the surface to the atmosphere. It can be represented in the surface energy balance equation as

$$F_{ir} = \varepsilon\sigma T_s^4$$

where ε is the emissivity which ranges from 0 to 1, σ the Stefan–Boltzman constant, and T_s the surface temperature.

F_{lw} is the long-wave radiation coming from the atmosphere to the surface. The representation of F_{lw} in the surface energy balance equation is mainly determined by the temperature of the air near the surface, though it is also affected by the atmospheric humidity and cloudiness.

Since F_{ir} and F_{lw} are both long-wave radiative fluxes, one of which is directed away from the surface (F_{lw}) and one toward the surface (F_{lw}), they partially cancel. In regions or at times when the surface is warmer than the atmosphere, such as over land during the day, when the surface is heated by incoming solar radiation, and over ocean in the high polar latitudes, where the water surface is usually warmer than the air, the net long-wave radiative flux is directed from the surface to the atmosphere. However, when the atmosphere is warmer than the surface, as over sea ice during the Arctic winter and over land surfaces at night during the winter, the net long-wave radiative flux is directed from the atmosphere to the surface. Thus the net long-wave radiative flux can be directed to the surface or to the atmosphere, depending on conditions.

The latent heat flux, F_l, is the exchange of energy between the surface and the atmosphere that occurs when water is evaporated from or condenses onto the surface. When water is evaporated, the energy absorbed by it causes it to change state from a liquid to a gas (water vapor) rather than to change the temperature. The gas then mixes with the rest of the atmosphere, carrying the latent heat with it. When the gas changes back into a liquid (condensation), such as when clouds form, that heat is released into the environment where the water condenses. As a result the energy has been moved from the surface, where the water was evaporated, to higher in the atmosphere, where the water condenses. The strength of this latent heat flux is dependent mostly on the relative amounts of water at the surface and in the air just above the surface. In general, the water, and thus the latent heat, moves from where there is more water to where there is less. In most circumstances the surface contains more water than the atmosphere, so the latent heat flux is generally from the surface to the atmosphere, representing a loss to the surface and a gain by the atmosphere.

In early energy balance models the latent heat flux was computed using a bulk aerodynamic formula of the form

$$F_1 = c_{dl}\upsilon L(a_s - a_a)$$

where c_{dl} is the drag coefficient for latent heat flux (typical value 1.32×10^{-3}), υ the wind speed, L the latent heat of vaporization, and a_s and a_a the absolute humidities of the surface and the air at the surface respectively. Recent models also account for the impact of vegetation on the latent heat flux.

The sensible heat flux, F_s, is the exchange of energy between the surface and the atmosphere that results from the temperature difference between the surface and the atmosphere. The bigger the difference in the temperature between the surface and the atmosphere the larger is the flux of energy. If the surface is warmer than the atmosphere then the flux is from the surface to the atmosphere. If the atmosphere is warmer than the surface then the flux is from the atmosphere to the surface. Thus this flux can provide either a gain or a loss by the surface.

In early energy balance models the sensible heat flux was computed using a bulk aerodynamic formula of the form

$$F_s = c_{ds}\upsilon\rho C_p(T_s - T_a)$$

where c_{ds} is the drag coefficient for sensible heat flux (typical value 1.41×10^{-3}), υ the wind speed, ρ the surface air density, C_p the specific heat of dry air, and T_s and T_a the surface and surface air temperatures.

The latent heat flux and the sensible heat flux are turbulent energy fluxes. This is because these exchanges of energy between the surface and the atmosphere are affected by the magnitude of the wind speed. As the wind speed increases the energy flux also increases.

F_{cond} is the conductive flux from beneath the surface. This flux varies, depending on the material the surface is made of and the environmental conditions. If temperature decreases with depth, as it generally does on land near the surface, the conductive flux is directed from the surface to beneath the surface, representing an energy loss by the surface. However, if the temperature increases with depth, as in the case of sea ice, which is floating on warmer ocean water, the conductive flux is directed toward the surface, representing an energy gain by the surface.

When the amount of energy coming to the surface equals the amount of energy leaving it, the surface is said to be in energy balance and the temperature remains constant. However, if more energy is coming into (leaving) the surface than is leaving (coming into) it then there is a positive (negative) energy imbalance and the temperature increases (decreases) in order to restore the balance.

In order to study the state of the Earth system and how it varies over time, computer models of the Earth system, which includes the surface energy balance equation discussed above, are used to identify which of the energy fluxes are important in establishing the state of the Earth's environment and how the Earth system responds to changes in the various energy fluxes.

An Energy Balance Model

Figure 1 is a schematic diagram of an energy balance model that incorporates a full surface energy balance. The diagram shows one latitude zone between latitudes ϕ and $\phi + \Delta\phi$, and indicates that the model treats four distinct regions of the Earth system, including air over land, land, air over ocean, and ocean. An energy balance is computed for each of these regions. If there is an energy surplus (deficit) in a region then the temperature is computed to rise (fall) to restore energy balance.

In an energy balance model the main parameter to be computed is the temperature. In eqns [1] and [2] the temperature under discussion is the temperature of the surface. In the energy balance model shown schematically in **Figure 1** the focus is the temperature of each of the regions. The temperature of each region is determined by summing the energy crossing each of the boundaries of the region. So the net increase or decrease of energy in the region of the air over land (for example) is determined by the following:

$$\begin{array}{lclc} & \begin{array}{l}\text{Energy exchange}\\ \text{between the land}\\ \text{and air}\\ \text{A}\end{array} & + & \begin{array}{l}\text{Energy exchange}\\ \text{between the air}\\ \text{space}\\ \text{B}\end{array} \\ + & \begin{array}{l}\text{Energy exchange}\\ \text{between this zone}\\ \text{and zone to south}\\ \text{C}\end{array} & + & \begin{array}{l}\text{Energy exchange}\\ \text{between this zone}\\ \text{and zone to north}\\ \text{D}\end{array} \\ + & \begin{array}{l}\text{Energy exchange}\\ \text{between this zone}\\ \text{and air over ocean}\\ \text{E}\end{array} & = & \begin{array}{l}\text{Net change in}\\ \text{energy content}\\ \text{of zone}\\ \text{F}\end{array} \end{array} \quad [3]$$

where the term is positive if the flux is into the zone and negative if the flux is out of the zone.

In a more complex model, which has more resolution in the east–west direction, rather than the simple land–sea resolution depicted in **Figure 1**, eqn [3] would have to include two terms for the exchange of energy within a latitude zone, one for the exchange to the east and one for the exchange to the west. In this model, with only two latitudinal zones resolved, we need consider only the exchange of energy between them.

Term A in eqn [3] is the surface energy balance explicitly stated in eqn [1]. If an energy balance occurs at the surface then term A is 0, as shown in

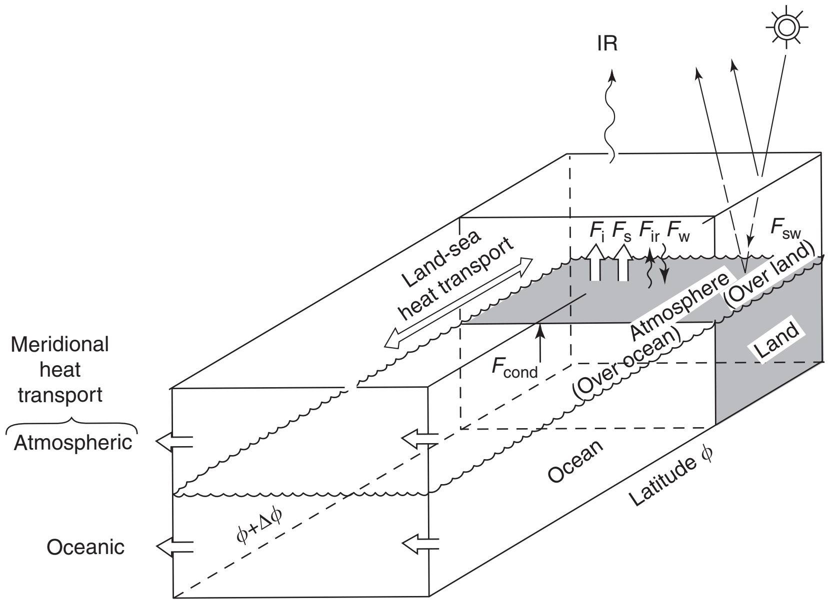

Figure 1 Schematic diagram of one latitude zone of a simple energy balance model that employs a full surface energy balance. (Adapted with permission from Ledley TS (1988) A coupled energy balance climate–sea ice model: impact of sea ice and leads on climate. *Journal of Geophysical Research* 93: 15919–15932. ©1988 by the American Geophysical Union.)

eqn [1]. If there is no surface energy balance then term A is non-zero and it contributes to the net change in energy of the region. In recent modeling work, term A has been developed to include the influence of vegetation on the surface energy balance. Vegetation, with its leaves and root systems, introduces a much more complex picture of the surface energy balance, which when included in models results in a better simulation of the climate of the Earth system.

Term B in eqn [3] is the net change in energy at the top of the atmosphere. Since the atmosphere is relatively transparent to solar radiation, most solar radiation reaches the surface and is represented as the first term in eqn [1]. The only other significant transfer of energy between the atmosphere and space is of infrared radiation (long-wave radiation). Since the atmosphere is warmer than space, the flux of infrared radiation is from the atmosphere to space, representing a loss by the atmosphere (see the IR term in **Figure 1**).

Terms C and D in eqn [3] represent the meridional heat transport between the zone under consideration and those to the north and south. Since any flux across the pole represents a flux out of the zone under consideration and then back into it, this flux is generally set to zero. In the simplest models this meridional heat transport is computed on the basis of the temperature gradient between the zone under consideration and those to the north and south, and the assumed diffusion coefficients. In some cases there may be more than one adjacent zone to the north or south, i.e., air over land in one zone may be adjacent to both air over land and air over water to the north or south, depending on the amount of land and ocean in each zone. In that case the fluxes from both of these zones must be included in terms C and D.

Term E in eqn [3] represents the zonal heat transport between the air over land in the zone under consideration and the air over ocean in the same zone. This can be computed from the temperature gradient between the air over land and sea and a diffusion coefficient.

Term F in eqn [3] represents the net change in energy in the zone. It can be represented in general by eqn [2]; however, the temperature now is the temperature of the air over land and the density and heat capacity are those for the air over land.

The temperature of the air over land is computed as follows:

$$\begin{aligned}\text{Net change in the energy in the zone} &= \Delta E \\ &= (\rho c)\partial T/\partial t \quad [4]\end{aligned}$$

By choosing a particular time step, Δt, over which to apply eqn [4] one can solve for the new temperature:

$$T(t+1) = \frac{T(t) + \Delta E^{*}\Delta t}{\rho c} \qquad [5]$$

The temperature computed here is the temperature representative of the whole layer of air. In order to determine the surface air temperature, which is generally desirable in order to compute the surface energy fluxes, another equation must be applied that relates the temperature computed in eqn [5] to the surface air temperature.

In using eqn [5] to compute the new temperature one must be careful that the energy balance equation is valid over the chosen time step size. If the equation is not valid over the chosen time step size then the numerical result of the calculation may not be physically realistic.

The other three regions have similar energy balance equations applied to them. The equation for the temperature of the air over ocean is different only in the values of the variables, constants, and coefficients, which are changed so that they represent the character of the ocean surface rather than the land surface.

The energy balance applied to the land includes terms A, B, and F in eqn [3]. Term A represents the energy exchange between the land and the air over land. Term B is altered to represent the exchange of energy between the land surface layer and deeper layers within the Earth (not shown in **Figure 1**). This energy flux is called the geothermal flux. Term F represents the effect of any imbalance in energy on the temperature of the land. In general the resolution of energy balance models is too coarse for the horizontal exchanges of energy between land in the zone and adjacent land or water to have a significant impact on the temperature of the land in the zone.

The energy balance applied to the ocean includes all the terms applied to the land with the addition of the meridional heat transport between adjacent ocean regions. The currents in the ocean carry a significant amount of energy both meridionally and zonally, and thus, while the ocean currents are not included explicitly in the model, their impact on the energy balance must be included.

Energy Balance Models as Tools

Energy balance models have been valuable tools in the study of the climate of the Earth system; however, they are dramatic simplifications of the Earth atmosphere system that either exclude or else represent in only a simple form the real physics and biogeochemical processes that occur. One example of this simplification is that the dynamics of the atmosphere, i.e., the processes that produce the high- and low-pressure systems in midlatitudes, monsoons, hurricanes, and tornadoes that are key in producing the exchanges of energy between different latitudinal zones and between regions over land and water, are represented only with respect to how they effect energy distribution. They are not included in a realistic way. Thus any results from energy balance models must be viewed with these simplifications in mind.

However, energy balance models have several advantages. The most important of these are (1) they are quick, allowing multiple experiments, and (2) it is relatively easy to analyze model results so that the causes of a change in the simulated system can be traced from the imposed perturbation to the resulting change. These advantages allow the study of particular processes in the Earth system and of how the Earth system responds to various changes. The results of these studies can then be used to guide the development of experiments in more complex models which are more costly to run and more complex to analyze, but do include the dynamics of the atmosphere and ocean, a much higher spatial resolution, and many other physical processes of the Earth system.

An example of how energy balance models can be used to investigate the relative importance of processes that contribute to climate change is an investigation using a version of the energy balance model described earlier that was conducted to examine how small variations in the minimum amount of open water in the sea-ice-covered polar oceans during the winter affect the maximum amount of open water during the summer and the seasonal cycle of surface air temperatures.

The first step in this study is to compare the climate simulated with the energy balance model to observations of the current climate to assure that the model is able to produce a reasonable simulation of the present climate. This includes comparisons of the mean annual seasonal cycle of simulated and observed surface air and surface temperatures, sea ice thickness, area of open ocean, and energy fluxes. The energy balance model used for this comparison specifies a minimum lead fraction (the minimum fraction of open water in sea ice, a lead being a crack in the sea ice that exposes ocean water to the atmosphere) of 1.1%. This means that for the winter over the polar oceans, when the heat loss would produce ice growth in any area of open water, the model determines the amount of ice that would grow and then mechanically open 1.1% of the ocean area to be ice-free, as would occur as the result of wind stresses and ocean currents in the real word. The climate simulated by this version of the energy balance model is called the control case.

Table 1 Mean annual sea ice thickness and maximum percentage of open water or period of ice-free conditions at 75° N and 75° S for various cases of specified minimum lead fraction

		75° N	*75° S*
Minimum lead fraction = 0%	Ice thickness (m)	1.42 m	2.02 m
Maximum % open water or period of ice-free conditions		0%	0%
Minimum lead fraction = 1.1%	Ice thickness (m)	1.30 m	2.10 m
Maximum % open water or period of ice-free conditions		8 weeks	1.7%
Minimum lead fraction = 2.2%	Ice thickness (m)	1.31 m	2.29 m
Maximum % open water or period of ice-free conditions		9 weeks	3.4%
Minimum lead fraction = 4.3%	Ice thickness (m)	1.36 m	2.94 m
Maximum % open water or period of ice-free conditions		10 weeks	6.3%

Adapted with permission from Ledley TS (1988) A coupled energy balance climate – sea ice model: impact of sea ice and leads on climate. *Journal of Geophysical Research* 93: 15919–15932. © 1988 American Geophysical Union.

In the study the effect of changing this minimum area of open ocean in the winter is investigated. **Table 1** shows the mean annual sea ice thickness and the maximum percent of open water or period of ice-free conditions during the summer for various specifications of the minimum lead fraction. **Figure 2** shows the seasonal cycle of surface air temperature zonally averaged over ocean, sea ice, and land as a function of latitude for the control case (minimum lead fraction during the winter of 1.1%) and the changes in the seasonal cycle of surface air temperature when the minimum lead fraction is reduced to 0%, increased to 2.2%, increased to 4.3%, and increased to 100% (meaning no sea ice is allowed to form).

Table 1 shows that when the minimum lead fraction is increased from the 1.1% control case, increasing the amount of open water during the winter, the mean annual thickness of the sea ice and the maximum area of open water or period of ice free conditions during the summer both increase. The increase in sea ice thickness occurs because the increase in the area of open ocean during the winter causes an increase in the amount of heat lost by the ocean and thus increases the production of sea ice. When leads in the sea ice are completely eliminated (minimum lead fraction = 0%) the sea ice thickness at 75° S decreases owing to the decrease in the ice growth rate. However, at 75° N the mean annual sea ice thickness is increased because of a decrease in the melting of sea ice during the summer.

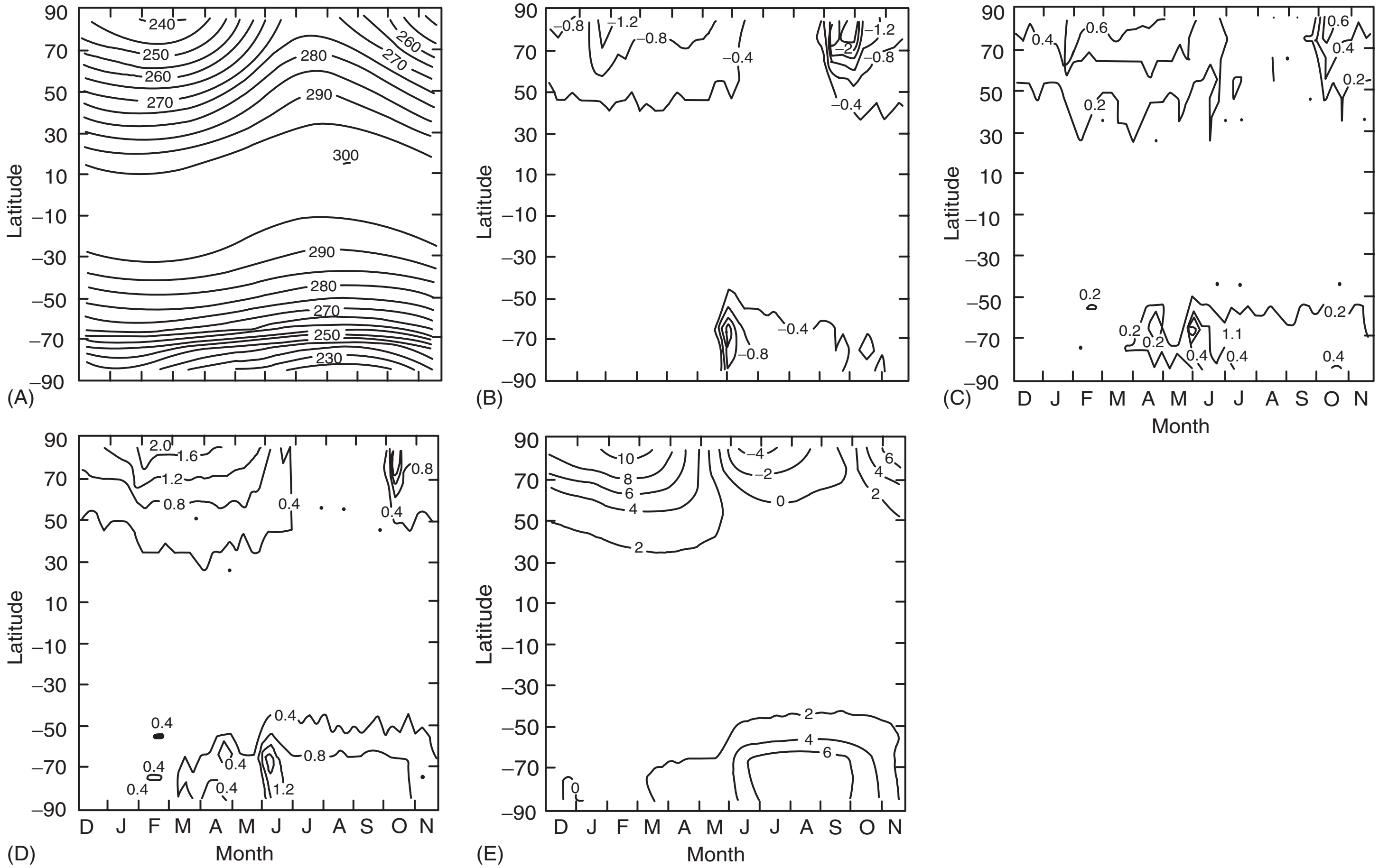

Figure 2 (A) The seasonal cycle of surface air temperature zonally averaged over the ocean, sea ice, and land, as a function of latitude for the control case, in degrees Kelvin. (B–D) The seasonal cycle of the change in the surface air temperature zonally averaged over the ocean, sea ice, and land, as a function of latitude from the control case when (B) no leads are specified, (C) when the minimum lead fraction is equal to 2.2%, and (D) when the minimum lead fraction is equal to 4.3%. (Adapted with permission from Ledley TS (1988) A coupled energy balance climate–sea ice model: impact of sea ice and leads on climate. *Journal of Geophysical Research* 93: 15919–15932. ©1988 by the American Geophysical Union.)

The study goes on to examine the impact of those changes in sea ice thickness and the relative amounts of open water on surface air temperatures through the year. **Figure 2** shows that the impact on the surface air temperatures is rather small during the summer; however, during the winter, when there is a large difference between the surface temperature of ocean water and of sea ice, the impact of small changes in the area of open ocean on surface air temperature is large. When areas of open ocean are eliminated, during the winter, the atmosphere is cut off from a heat source, namely the relatively warm ocean, and surface air temperatures drop by between 0.4 K and 1.2 K during the winter. When areas of open ocean are increased during the winter, the atmosphere is in contact with an expanded heat source, and surface air temperatures increase by up to 2.0 K when sea ice is allowed to form, and much more when it is not.

The use of energy balance models in this kind of study permit further investigation to identify which energy fluxes contribute to the simulated changes in surface air temperature and under what conditions each energy flux is the most important.

See also

Air–Sea Interaction: Momentum, Heat and Vapor Fluxes; Sea Surface Temperature. **Boundary Layers:** Modeling and Parameterization. **Coupled Ocean–Atmosphere Models**. **General Circulation:** Models. **Land–Atmosphere Interactions:** Overview. **Mesoscale Meteorology:** Models. **Radiation (Solar)**. **Reflectance and Albedo, Surface**. **Teleconnections**.

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EVOLUTION OF ATMOSPHERIC OXYGEN

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Introduction

Abundant free oxygen in the atmosphere distinguishes our planet from all others in the solar system. Earth's oxygen-rich atmosphere is a direct result of life. The current atmosphere contains (by volume) 78.09% N_2, 20.95% O_2, 0.93% Ar, 0.036% CO_2, and additional trace gases. Apart from argon, all of the quantitatively important gases are at least in part biologically controlled, but oxygen in particular has no significant abiotic source. Diatomic oxygen is generated by oxygenic photosynthesis, the biological process in which water molecules are split using the energy of sunlight. Today, green plants, single-celled phytoplankton (free-floating organisms in the ocean), including cyanobacteria (chlorophyll-containing bacteria) all perform oxygenic photosynthesis. Of these, cyanobacteria are the most numerous, with $\sim 10^{27}$ in the oceans, and probably their ancient ancestors were just as plentiful. However, geological differences between the ancient and modern Earth show that there was insufficient O_2 in the early atmosphere to leave traces of oxidation that today are ubiquitous, such as the reddening of exposed iron-rich rocks. The