in temperature because of adiabatic compression and expansion, which result in warming during descent and cooling during ascent, respectively. Evidence for changes in the pattern of the mean meridional circulation can be seen by comparing the temperature changes (Fig. 3C) with the positions of the ascending and descending air in the control run (shading in Fig. 3A). In the middle troposphere, the cooling near 40°S is caused by the southward departure of the descending portion of the Hadley cell, and the warming around 55°S is a result of both the arrival into this region of the descending air and the southward movement of the ascending portion of the Ferrel cell. The latter then causes the extra cooling seen near 70°S.

From these results, it is evident that the vertical profiles of temperature change vary strongly with geographical position. The LvL (*2*) radiosonde data have also shown (for the boreal summer) increases at solar maximum in tropospheric temperature at low and mid-latitudes but decreases at some high-latitude stations. The temperature changes estimated by LvL (*3*) for Lihue, Hawaii (22°N, 159°W), averaged over November–December and January–February between extremes of the three solar cycles occurring between 1959 and 1994 were compared with those calculated for a near position (30°N, 145°W) in January in the present work (Fig. 4). Relative to the observational data, the model profile has a very similar structure but is smaller in magnitude. Given that the model has fixed sea surface temperatures (such that the extra solar energy reaching the surface is "lost") and that the specified ozone changes may be too low (*15*), an underestimate of the solar effects is not surprising. The only other station for which LvL show data for the boreal winter is Truk Island (7.5°N, 152°E); at this site, the observations in the troposphere showed much smaller changes in response to the solar cycle, and the model results likewise are not significantly different from zero. In the stratosphere, both the observations and the model show more warming than at Lihue. [LvL (*3*) also concluded that the shape of the temperature change profiles is consistent with changes in vertical motion.]

The model results suggest that increases in stratospheric temperature in response to enhanced solar irradiance result in stronger summer easterly winds, which penetrate into the tropical upper troposphere and force tropospheric circulation patterns poleward. The model shows changes in temperature, zonal wind, and storm track position that are similar to, although generally smaller than, those observed. The solar-induced increase in stratospheric ozone is important in determining the change in lower stratospheric temperatures and thus the subsequent climate response. There is no quasi-biennial oscillation (QBO) in the model (in effect, it is permanently in the easterly phase); hence, the claim by LvL that the QBO plays a role in modulating the impact of solar variability on the winter lower stratosphere cannot be tested. However, if the strength of the zonal wind in the tropical lower stratosphere plays an important role in transmitting the solar effects from stratosphere to troposphere, as suggested by the present results, then it is clear that modulation by the QBO is probable. The results of the model also imply that changes in stratospheric ozone brought about by any other means may have an impact on tropospheric climate.

REFERENCES AND NOTES

1. W. Herschel, *Phil. Trans.* **91**, 265 (1801).
2. K. Labitzke and H. van Loon, *J. Clim.* **5**, 240 (1992).
3. ———, *Meteorol. Z.* **3**, 259 (1994).
4. G. M. Brown and J. I. John, *J. Atmos. Terr. Phys.* **41**, 43 (1979).
5. J. Lean, *Science* **244**, 197 (1989).
6. R. Garcia, S. Solomon, G. R. Roble, D. W. Rusch, *Planet. Space Sci.* **32**, 411 (1984); L. L. Hood, *J. Geophys. Res.* **91**, 5264 (1986); S. Chandra, *Geophys. Res. Lett.* **18**, 837 (1991); T. Y. W. Huang and G. P. Brasseur, *J. Geophys. Res.* **98**, 20413 (1993); E. L. Fleming, S. Chandra, C. H. Jackman, D. B. Considine, A. R. Douglass, *J. Atmos. Terr. Phys.* **57**, 333 (1995).
7. K. Kodera, M. Chiba, K. Shibata, *Geophys. Res. Lett.* **18**, 1209 (1991).
8. D. Rind and N. K. Balachandran, *J. Clim.* **8**, 2080 (1995).
9. R. T. Wetherald and S. Manabe, *J. Atmos. Sci.* **32**, 2044 (1975); E. Nesme-Ribes, E. N. Ferreira, R. Sadourny, H. LeTruet, Z. X. Li, *J. Geophys. Res.* **98**, 18923 (1993); D. Rind and J. Overpeck, *Quat. Sci. Rev.* **12**, 357 (1993).
10. J. D. Haigh, *Nature* **370**, 544 (1994).
11. J.-J. Morcrette, *J. Geophys. Res.* **96**, 9121 (1991).
12. J. Slingo *et al., Q. J. R. Meteorol. Soc.* **120**, 881 (1994).
13. J. Lean, *Rev. Geophys.* **29**, 505 (1991).
14. L. L. Hood and J. P. McCormack, *Geophys. Res. Lett.* **19**, 2309 (1992).
15. L. L. Hood, *J. Geophys. Res.*, in press.
16. I. N. James, *Introduction to Circulating Atmospheres* (Cambridge Univ. Press, Cambridge, 1994).
17. J. A. Laurmann and W. L. Gates, *J. Atmos. Sci.* **34**, 1187 (1977).
18. K. Kodera, *J. Geophys. Res.* **100**, 14077 (1995).
19. L. L. Hood, J. L. Jirikowic, J. P. McCormack, *J. Atmos. Sci.* **50**, 3941 (1993); N. K. Balachandran and D. Rind, *J. Clim.* **8**, 2058 (1995).
20. E. K. M. Chang, *J. Atmos. Sci.* **52**, 2006 (1995).

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Universality Classes of Optimal Channel Networks

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Energy minimization of both homogeneous and heterogeneous river networks shows that, over a range of parameter values, there are only three distinct universality classes. The exponents for all three classes of behavior are calculated.

River networks reflect fractal properties in a power law distribution of various quantities (*1*). The striking generality of Horton's law of stream numbers (*2*) motivated Shreve (*3*) to suggest that channel networks developed in the absence of geologic controls are essentially topologically random. Nevertheless, nonrandom river networks have been consistently observed. Their existence has prompted the development of models (*2, 4, 5*) of drainage network optimization based on the concept of energy minimization and optimal channel networks (OCNs) (*6, 7*). Computer simulations of homogeneous OCNs (*6*) have resulted in optimal networks with a striking similarity to those observed in nature. These results have raised the question as to whether some form of global energy minimization underlies the existence of fractal structures. Here, we solve the OCN for a range of parameters for both homogeneous and heterogeneous basins. Although we do obtain fractals, our exact results for the power law exponents do not agree with either the observational data or the computer simulations. The disagreement between the results of our analytic solution and the computer simulations (*6*) is a result of the fact that the latter were only able to access a set of local minima (which depend-

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ed on the initial conditions and the dynamics used) and not the true global minimum.

In computer models, the river network is defined as a spanning, loopless tree on a two-dimensional lattice of linear size L with one global outlet. Each site of the lattice has exactly one outlet bond to one of its nearest neighbors. There are no restrictions on the number of bonds into a site. The area s_i associated with any bond i is

$$s_i = \sum_j s_j + R_i \tag{1}$$

where the sum is over each bond j that inputs into i, and the rainfall R_i (precipitation) (*8*) at the ith bond is commonly assumed to be independent of i and equal to 1. Then, s_i is equal to the number of sites upstream of i connected by the network. The OCN is obtained by choosing the network that minimizes the energy functional

$$E = \sum_{i=1}^{N} k_i s_i^{\gamma} \tag{2}$$

where N is the total number of bonds, k_i is a quantity (assumed to be independent of i and equal to 1 without loss of generality for a homogeneous basin) related to soil properties such as the erodibility, and γ is an exponent that characterizes the physics of the erosional process. The quantity $k_i s_i^{\gamma}$, a measure of the potential energy dissipated in the ith bond, is a product of the water flow and the elevation difference along the bond in the associated landscape: the water flow is proportional to s_i, and the elevation difference is assumed to have an empirical relationship with s_i given by $s_i^{\gamma-1}$ (*4*).

Here we show that homogeneous basins with $\frac{1}{2} \leq \gamma \leq 1$ fall into two universality classes: the Scheidegger class (*9*) for $\gamma = 1$ and mean-field–like behavior (*10*) for $\frac{1}{2} \leq \gamma < 1$. We then show the effects of disorder or heterogeneity in k_i on the scaling behavior.

Consider a site on the lattice and a subbasin comprising all sites upstream that are connected by the network. This subbasin is typically anisotropic with a longitudinal length $\xi_{\parallel}$ and a transverse length $\xi_{\perp} \sim \xi_{\parallel}^{H}$, where $H \leq 1$ is the Hurst exponent. Thus, the number of sites in the subbasin scales as $\xi_{\parallel}^{1+H}$. Let $p(s, L)$ be the probability density distribution for a given bond to be associated with an area s for a system of linear size L. For the optimal network, assume (*7, 11, 12*) that

$$p(s,L) = s^{-\tau} f(s/s_B) \tag{3}$$

where $f(x)$ is a scaling function that is a nonzero constant for small values of x and zero for large values; s_B sets the characteristic scale below which the distribution of s is algebraic with an exponent τ. Because $\xi_{\parallel} \sim L$

$$s_B \sim L^{1+H} \tag{4}$$

and because the maximum area of the basin is L^2, $H \leq 1$. The mean area $\langle s \rangle$ is equal to the average distance to the outlet from the sites, which is shortest for directed networks (*13*). Generally, $\langle s \rangle$ scales as L^{d_l}, where the fractal dimension of the stream $d_l \geq 1$, with the equality holding for directed networks. Thus

$$\langle s \rangle = \int_1^{\infty} ds\, s\, p(s,L) \sim L^{d_l} \tag{5}$$

and therefore

$$d_l = (1 + H)(2 - \tau) \tag{6}$$

Hack's law (*14*), relating the length l of the longest stream in the drainage region (measured from any site to the edge of the subbasin) to the drainage area s of the basin (the number of upstream sites), $l \sim s^{h}$, holds with

$$h = \frac{d_l}{(1 + H)} \tag{7}$$

because the number of sites in the subbasin scales as $\xi_{\parallel}^{1+H}$ and $l \sim \xi_{\parallel}$. Thus, for directed, self-affine river basins, the wandering of the river characterized by the Hurst exponent uniquely determines several of the other exponents.

The Scheidegger model (*9, 10*) is obtained for homogeneous basins on choosing $k_i = 1$ for all i and $\gamma = 1$ in Eq. 2. Minimization of $\Sigma_i s_i$ over the ensemble of all networks (directed and undirected) leads to all directed networks being degenerate and optimal because $(1/L^2)(\Sigma_i s_i) = \langle s \rangle$, which is a minimum and the same for all directed networks. Because the path from a given point to the outlet executes a random walk, $H = \frac{1}{2}$ and $d_l = 1$ (*13*) for the Scheidegger model.

For the case $\frac{1}{2} \leq \gamma < 1$, the minimum of the energy functional in Eq. 2 scales as

$$\text{Min } E \sim L^{1+2\gamma} \tag{8}$$

because Min E scales with a power greater than or equal to $1 + 2\gamma$ and separately with a power less than or equal to $1 + 2\gamma$. The lower bound is obtained by dividing the sum in Eq. 2 into the sum over rows of sites in the direction transverse to the flow and the sum over sites within the rows and by using the inequality $\Sigma X_i^{\gamma} > (\Sigma X_i)^{\gamma}$ for $X_i \geq 0$. The upper bound results from an explicit construction of classes of self-similar networks that satisfy the scaling $E \sim L^{1+2\gamma}$. For the optimal spanning tree, from Eq. 2

$$\text{Min } E \sim L^{2} \langle s^{\gamma} \rangle \tag{9}$$

Combining with Eq. 8 the scaling form of Eqs. 3 and 6, one obtains $H(1 - \gamma) = d_l - \gamma$, and thus, for $\frac{1}{2} \leq \gamma < 1$ in the homogeneous case, $H \geq 1$ (which is consistent with $H \leq 1$ only when $H = 1$), $d_l = 1$, and $\tau = \frac{3}{2}$.

Natural river basins are heterogeneous. The simplest heterogeneity that one may consider is random precipitation with R_i no longer uniform as in Eq. 1. Sites with large s_i have contributions from the R_j of all the upstream sites. Thus, small fluctuations in R_i would be expected to average out to yield the homogeneous result. Indeed, an explicit analysis (*10*) within the framework of the Scheidegger model shows that the universality class is unchanged upon addition of this kind of heterogeneity.

Now, consider the case in which erosion is random, that is, the values of k_i are random and uncorrelated. In this case, we determine the nature of the optimal spanning tree (each site in a spanning tree has a unique path to the global outlet) for which the quantity $\Sigma_i k_i s_i$ ($\gamma = 1$ in Eq. 2) is a minimum. For any spanning tree, $\Sigma_i k_i s_i$ can be exactly rewritten (*15*) as $\Sigma_m C_m$ with $C_m = \Sigma_j k_j$. Here, C_m physically represents the cost of the path from site m to the global outlet, the sum over j runs over all bonds on the path, and k_j is the cost of traversing bond j. The summation over m is a sum over the paths originating from each of the sites. Indeed, the factor s_i in $\Sigma_i k_i s_i$ arises from the number of paths passing through a given bond i. The optimal spanning tree, for which $\Sigma_m C_m$ is the lowest, is obtained if C_m is a minimum for each m. The minimization of all the individual C_m's can be effected simultaneously, and the resulting optimal spanning tree is simply the union of all the optimal paths from each of the sites to the global outlet. As an implication of optimality, when two paths intersect they overlap exactly the rest of the way to the outlet. Also, the path from a site along another path coincides exactly with the subpath of the original path. Thus, the bond randomness provides a

Table 1. Summary of exponents for models studied and for river basins (*1*). The random erosion model for $\frac{1}{2} \leq \gamma < 1$ yields the same exponents as the homogeneous model (mean field).

Exponent	Scheidegger $\gamma = 1$	Mean field $\frac{1}{2} \leq \gamma < 1$	Random erosion $\gamma = 1$	River basins
H	$\frac{1}{2}$	1	$\frac{2}{3}$	0.67 to 0.92
τ	$\frac{4}{3}$	$\frac{3}{2}$	$\frac{7}{5}$	1.40 to 1.46
h	$\frac{2}{3}$	$\frac{1}{2}$	$\frac{3}{5}$	0.52 to 0.60

natural mechanism of aggregation.

The problem of the optimal path, for which C_m is a minimum, has been solved exactly in two dimensions in the context of domain walls in random ferromagnets and directed polymers in a random medium (*16*). The Hurst exponent $H = 2/3$. For directed, self-affine river basins, the values of τ and h can readily be deduced to be $7/5$ and $3/5$, respectively. These values are robust and do not change even if the minimization of the energy functional includes both directed and undirected networks (*16*).

For $1/2 \leq \gamma < 1$, heterogeneities in the erosional properties are irrelevant, and the exponent values are the same as their homogeneous counterparts. Our proof relies on first observing that Min $E \leq L^{1+2\gamma}$ (Min $E \leq \Sigma_i k_i s_i^\gamma$ for the tree for which $\Sigma_i s_i^\gamma$ is a minimum, but $\Sigma_i k_i s_i^\gamma \leq k_i^{\max} \Sigma_i s_i^\gamma \approx k_i^{\max} L^{1+2\gamma}$, where $k_i^{\max}$ is the largest of the k_i values) and then using Eq. 9 in conjunction with Eq. 3 to show that $H \geq 1$. Because $H > 1$ is not physically meaningful, the Hurst exponent remains unchanged at $H = 1$.

We have thus shown that OCNs with $1/2 \leq \gamma \leq 1$ show three classes of behavior (Table 1). Our results indicate that the OCN, in its present form, does not describe the behavior of river basins. Rinaldo and co-workers (*6*) have carried out numerical studies of the $\gamma = 1/2$ case. Their work, which was restricted to the statistics of local minima (and not the global minimum, as in our analysis), yielded exponents different from our results but in good accord with observational data.

REFERENCES AND NOTES

1. I. Rodriguez-Iturbe, R. L. Bras, E. Ijjasz-Vasquez, D. G. Tarboton, *Water Resour. Res.* **28**, 988 (1992); D. G. Tarboton, R. L. Bras, I. Rodriguez-Iturbe, *ibid.* **24**, 1317 (1988); *ibid.* **26**, 2243 (1990); P. La Barbera and R. Rosso, *ibid.* **25**, 735 (1989); S. P. Breyer and R. S. Snow, *Geomorphology* **5**, 143 (1992); D. R. Montgomery and W. E. Dietrich, *Science* **255**, 826 (1992); D. G. Tarboton, R. L. Bras, I. Rodriguez-Iturbe, *Water Resour. Res.* **25**, 2037 (1989).
2. R. E. Horton, *Geol. Soc. Am. Bull.* **56**, 275 (1945); *Eos* **13**, 350 (1932).
3. R. L. Shreve, *J. Geol.* **74**, 17 (1966); *ibid.* **75**, 178 (1967); *ibid.* **77**, 397 (1969).
4. L. B. Leopold and W. B. Langbein, *U.S. Geol. Surv. Prof. Pap. 500-A* (1962).
5. A. D. Howard, *Water Resour. Res.* **26**, 2107 (1990); *ibid.* **7**, 863 (1971); C. T. Yang, *ibid.*, p. 311; P. S. Stevens, *Patterns in Nature* (Little-Brown, Boston, MA, 1974).
6. I. Rodriguez-Iturbe, A. Rinaldo, R. Rigon, R. L. Bras, E. Ijjasz-Vasquez, *Water Resour. Res.* **28**, 1095 (1992); *Geophys. Res. Lett.* **19**, 889 (1992); A. Rinaldo, I. Rodriguez-Iturbe, R. Rigon, E. Ijjasz-Vasquez, R. L. Bras, *Phys. Rev. Lett.* **70**, 822 (1993); I. Rodriguez-Iturbe and A. Rinaldo, *Fractal River Basins: Chance and Self-Organization* (Cambridge Univ. Press, Cambridge, 1996); see also P. Yam, *Sci. Am.* **271**, 26 (November 1994).
7. T. Sun, P. Meakin, T. Jøssang, *Phys. Rev. E* **49**, 4865 (1994); *ibid.* **51**, 5353 (1995); *Water Resour. Res.* **30**, 2599 (1994); P. Meakin, J. Feder, T. Jøssang, *Physica A* **176**, 409 (1991).
8. The effects of evaporation of the water are captured by a negative contribution to R_i. It can be shown that randomness in R_i does not change the universality class.
9. A. E. Scheidegger, *Bull. Int. Assoc. Sci. Hydrol.* **12**, 15 (1967).
10. H. Takayasu, M. Takayasu, A. Provata, G. Huber, *J. Stat. Phys.* **65**, 725 (1991).
11. A. Maritan, A. Rinaldo, R. Rigon, I. Rodriguez-Iturbe, A. Giacometti, *Phys. Rev. E* **53**, 1501 (1996).
12. Our finite-size scaling arguments and the notion of universality are similar to those in critical phenomena. See, for example, M. E. Fisher, *Rev. Mod. Phys.* **46**, 597 (1974).
13. Directed networks are ones in which there is no local flow with a component away from the dominant flow direction. Generally, the mean area $\langle s \rangle$ scales as L^{d_l}, where the fractal dimension of the stream $d_l \geq 1$. The fractal situation $d_l > 1$ corresponds to $H = 1$, whereas a self-affine geometry has $d_l = 1$ and $H \leq 1$. All of our results hold for arbitrary networks without any restriction on their directedness.
14. J. T. Hack, *U.S. Geol. Surv. Prof. Paper 294-B* (1957). The principal cause of the anomalous exponent of Hack is a matter of debate. Basin elongation occurs in the Scheidegger model (*9, 10*), but this model is yet untested with large sets of data. Alternately, computer simulations (*7*) suggest that basin shapes have a similarity across scales.
15. This substitution is a generalization of the observation that the mean area is equal to the average distance to the outlet [B. M. Troutman and M. R. Karlinger, *Water Resour. Res.* **28**, 563 (1992), and references therein].
16. D. A. Huse and C. L. Henley, *Phys. Rev. Lett.* **54**, 7708 (1985); M. Kardar, *ibid.* **55**, 2923 (1985); D. A. Huse, C. L. Henley, D. S. Fisher, *ibid.*, p. 2924.
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Predatory Dinosaurs from the Sahara and Late Cretaceous Faunal Differentiation

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Late Cretaceous (Cenomanian) fossils discovered in the Kem Kem region of Morocco include large predatory dinosaurs that inhabited Africa as it drifted into geographic isolation. One, represented by a skull approximately 1.6 meters in length, is an advanced allosauroid referable to the African genus *Carcharodontosaurus*. Another, represented by a partial skeleton with slender proportions, is a new basal coelurosaur closely resembling the Egyptian genus *Bahariasaurus*. Comparisons with Cretaceous theropods from other continents reveal a previously unrecognized global radiation of carcharodontosaurid predators. Substantial geographic differentiation of dinosaurian faunas in response to continental drift appears to have arisen abruptly at the beginning of the Late Cretaceous.

Major continental areas became increasingly isolated during the Late Cretaceous as non-avian dinosaurs underwent their final radiation. The influence of continental fragmentation on dinosaur evolution during this interval has remained uncertain because of uneven sampling of the fossil record. Late Cretaceous dinosaurian remains have been recovered principally from Asia and western North America and consist largely of coelurosaurian predators and ornithischian herbivores (*1*). Although dinosaur bones of Late Cretaceous age have been reported from all major southern land areas including Antarctica (*2–12*), reasonably complete skeletons have been recovered only from South America (*13*), where abelisaurid predators and titanosaurian herbivores flourished.

On continental Africa, the most complete remains of Late Cretaceous dinosaurs were discovered in Egypt and include the lower jaw and vertebrae of an unusual fin-backed theropod, *Spinosaurus* (*3*); skull fragments and bones of another large theropod, *Carcharodontosaurus* (*4*); isolated bones of a third predator, *Bahariasaurus* (*5*); and bones of a large titanosaurian sauropod, *Aegyptosaurus* (*6*). In beds of similar age in Morocco, several bones of an enigmatic sauropod *Rebbachisaurus* were discovered (*7*). Phylogenetic interpretation of these remains (*14–17*) has been difficult because many of the bones are fragmentary and because the Egyptian collection was destroyed during World War II (*18*).

We describe here new vertebrate remains from Late Cretaceous beds in the

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