# *7* Compaction, Pressure, and Groundwater Flow

## About compaction-driven flow

As sediment is deposited in a basin, it adds to the load on underlying sediments, causing them to compact. Compaction increases pressure on the pore fluid, driving it stratigraphically upward and toward basin margins. In various studies (see reviews by Bethke, 1985, 1986; Harrison and Summa, 1991), geologists have inferred that groundwater flow regimes resulting from sediment compaction play roles in processes of petroleum migration, ore genesis, sediment diagenesis, and the origin of sedimentary brines.

The effect of compaction on fluid pressure is in many cases small. Where compaction is rapid and sediment permeability is low, however, pressure considerably in excess of hydrostatic develops in response to sediment loading. Such pressures, known as overpressures and geopressures, are of interest because of the practical difficulty and danger they pose during drilling and because of their inferred roles in geologic processes such as faulting, structural deformation, and localizing petroleum reservoirs. As discussed in **Chapter 2**, overpressured sediments tend to be less compacted than sediments at normal pressure because the pore fluid relieves the sediment framework of much of the weight of overlying sediments.

Two other processes affect fluid pressure in compacting basins. As sediments accumulate, underlying sediments are buried more deeply. Temperature increases along the geothermal gradient, causing the pore fluid to expand thermally. This thermal contribution, which can also arise when heat flow varies in time, has been termed aquathermal pressuring. Second, the sedimentation surface is a moving boundary. Since we judge pressure relative to a hydrostatic gradient extending downward from the surface, the moving boundary operating in the absence of compaction and thermal expansion works to produce subnormal pressure. This chapter shows how to use Basin2 to model compaction, fluid pressure, and groundwater flow as sediments accumulate in sedimentary basins.

## Rate of groundwater flow

Basin2 calculates the rate at which groundwater flows through the subsurface using Darcy's law, written for a fluid of varying density. Darcy's law gives the fluid's specific discharge, which is the volume of fluid crossing a unit area per unit time. Typical units of discharge are $cm^3/cm^2$ yr, or simply cm/yr.

At any point, the components $q_x$ and $q_z$ of specific discharge along the $x$ and $z$ directions are given

$$q_x = -\frac{k_x}{\mu}\left(\frac{\partial P}{\partial x} - \rho g \frac{\partial z}{\partial x}\right)$$

and

$$q_z = -\frac{k_z}{\mu}\left(\frac{\partial P}{\partial z} - \rho g\right) \qquad (7.1)$$

Here, $k_x$ and $k_z$ are permeability in each direction, $\mu$ is fluid viscosity, $P$ is pressure, $\rho$ is fluid density, and $g$ is the acceleration of gravity.

The derivative $\partial z/\partial x$ in 7.1 is the stratigraphic slope. This term arises because, as discussed in **Chapter 1**, the $x$ direction in Basin2 follows stratigraphic time lines rather than the horizontal. The program carries out its calculations in this curvilinear system but reports discharge and velocity in Cartesian coordinates so the user need not be concerned about details of the internal coordinates.

The terms within brackets in 7.1, carried with the negative sign, are the driving forces for groundwater flow along $x$ and $z$. In the study of petroleum migration, these terms are known as the hydrodynamic forces. The driving forces can be expressed in traditional units (e.g., dynes) or in units of pressure over distance (e.g., atm/km or Pa/km).

Even though discharge can carry units of velocity, this variable is in fact a volume flux. The groundwater velocity, which represents the average rate of displacement of water molecules along a flow path, differs in value from the discharge. To understand why, consider groundwater discharging at a rate of 1 $cm^3/cm^2$ yr through a sediment with 20% porosity. The fluid can travel only within the pore space of the rock, so the first water to cross a $cm^2$ plane in the sediment must move forward 5 cm by the end of the year to make room for the remaining discharge. The general relationship between discharge and components of the average velocity $v_x$ and $v_z$ is

$$v_x = \frac{q_x}{\phi} \quad \text{and} \quad v_z = \frac{q_z}{\phi} \qquad (7.2)$$

where $\phi$ is the sediment porosity. From these equations, it is clear that at a given discharge groundwater moves most rapidly through rocks of small porosity.

# Hydraulic potential

Written in the form of **Equation 7.1**, Darcy's law correctly accounts for convective forces that arise from lateral variations in fluid density. When fluid density remains about constant, as is the case when temperature and salinity vary little over the domain or when you set `density` to a constant value, Darcy's law can be written more simply in terms of hydraulic potential. Hydraulic potential, defined by Hubbert (1940), is the mechanical energy content of a unit volume of groundwater. The potential Φ is given

$$\Phi = P - \rho g z \tag{7.3}$$

where $z$ is depth below sea level. The potential function can be thought of as pressure less the hydrostatic contribution of a column of water extending downward from sea level.

Darcy's law is written in terms of hydraulic potential as

$$q_x = -\frac{k_x}{\mu}\left(\frac{\partial \Phi}{\partial x}\right)$$

and

$$q_z = -\frac{k_z}{\mu}\left(\frac{\partial \Phi}{\partial z}\right) \tag{7.4}$$

By these equations, groundwater moves from areas of high to low potential. The rate at which it moves depends on the potential gradient, permeability, and fluid viscosity.

It is convenient to represent the drive for flow by contouring the hydraulic potential function, as can easily be accomplished with B2plot (see **Chapter 15**). Such contours are known as equipotentials (or isopotentials). Fluid crosses equipotentials in the direction of decreasing Φ; the driving force for groundwater flow in any direction is the contour interval divided by the distance between contours.

In the study of petroleum migration, you can compare this force, the hydrodynamic drive $E_H$, to the buoyant force $E_B$ arising from the lesser density of oil compared to water (Hubbert, 1953). Taking the density of water and oil to be constant, the hydrodynamic force

$$E_{H_x} = -\frac{\partial \Phi}{\partial x}$$

and

$$E_{H_z} = -\frac{\partial \Phi}{\partial z} \tag{7.5}$$

is the negative gradient in hydraulic potential. You calculate the buoyant force from

$$E_{B_x} = -(\rho - \rho_o)\, g \frac{\partial z}{\partial x}$$

and

$$E_{B_z} = -(\rho - \rho_o)\, g \qquad (7.6)$$

where $\rho_o$ is the oil density and $\partial z/\partial x$ is the stratigraphic slope.

Many users expect to see groundwater flow at right angles to equipotentials, as is the case in simple flow problems found in textbooks. This is seldom the case with Basin2 results, because permeability is not likely to be isotropic and, in any event, the results are almost always plotted to a vertical exaggeration. The “right angle rule” applies only to isotropic systems drawn without exaggeration.

When fluid density varies, the drive for flow unfortunately can not be represented in terms of a scalar function like $\Phi$; however, because equipotentials are so useful in visualizing the drive for flow, Basin2 calculates an approximate potential function during runs in which fluid density varies. The program determines potential as the difference between the calculated pressure $P$ and pressure falling along a hydrostatic gradient passing through $P = 0$ at sea level. If fluid density is constant, this definition reverts to the original definition in **Equation 7.3**.

When drawing equipotentials in cases of varying fluid density, keep in mind that the equipotentials only approximately represent the drive for flow. You may note in such cases that fluid at some places in the domain migrates along a direction of increasing potential.

## Compaction in one dimension

First we consider the compaction of a sediment column, assuming that only vertical flow takes place. **Input 7.1** shows the Basin2 input file we will use. The first line sets a single column of nodal blocks and tells the program to output results only at the end of the simulation. For simplicity, we assume constant values for the fluid density $\rho$ and the density $\rho_{sm}$ of the saturated medium. Next, the input defines the permeability correlation for rock type `sh` in terms of two variables `AP` and `BP`. We define these values from the command line when we start the simulation. Variable `p_kxkz` sets the anisotropy in permeability, so that the vertical component $k_z$ is one tenth of the lateral value $k_x$.

**Input 7.1** *Shale compaction in one dimension.*

```
nx = 1;  flow = vertical;  plot = final
density = 1 g/cm3;  bulk_density = 2.3 g/cm3

rock sh
    A_perm = $AP;  B_perm = $BP log_darcy;  p_kxkz = 10
end_rock

X(sh) = 1;  reference = uncompacted;  heat_flow = 1 HFU

strat 'Basal unit'
    t_dep = $TIME0 m.y.
    thickness = 100 m

strat 'Shale deposited over run'
    t_dep = 0 yrs
    thickness = 10 km
```

The final lines define two stratigraphic units, each composed entirely of rock type `sh`. Stratigraphic thicknesses are to be entered in terms of uncompacted sediment, and the basal heat flow throughout the simulation is 1 HFU. The first stratigraphic unit is a thin layer present at the start of the simulation; the program requires that at least one unit be present at startup. The second unit is composed of a ten-kilometer pile of uncompacted sediment, which is to be deposited at a uniform rate from `TIME0` m.y. ago to the present.

We begin by considering the effect of permeability on pressure and porosity. Invoke the program by typing

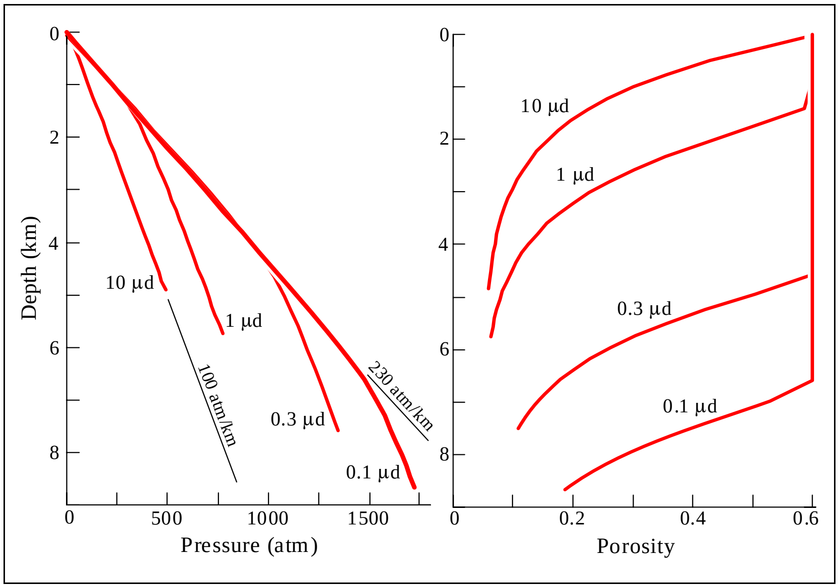
```
b2 Input_7.1 -d AP=0 -d BP=-4 -d TIME0=-10
```

(or start Basin2 from the control panel, setting the variables as shown). This command sets the coefficients *A* and *B* in the permeability correlation (**Equation 2.9**) to zero and −4. Since the anisotropy $k_x/k_z$ in permeability is 10, the program will set the horizontal permeability $k_x$ to a constant value of 100 μdarcy, and the vertical value $k_z$ to 10 μdarcy. Variable `TIME0`, set to −10, is the starting time for the simulation.

Over the course of the simulation, the program added 25 nodes to the column in order to accept sediment, and took about 500 time steps. The model executed on a 400 MHz Pentium II PC in about 15 seconds. We can quickly repeat the calculation using other values for permeability on the command line. In doing so, it is convenient to use the suffix option (`-s`) on the command line to give unique names to the output files, as discussed in **Chapter 6**.

**Figure 7.1**, plotted with B2plot, shows the results of calculations that assumed various values for permeability. Moderate permeability values predict pressures along a hydrostatic gradient of about 100 atm/km. As we decrease permeability, pressures develop significantly in excess of hydrostatic and converge toward the lithostatic gradient of about 230 atm/km. In these runs, the sediments are too impermeable to allow fluid to be expelled rapidly enough to allow normal compaction.

**Figure 7.1** *Profiles versus depth for fluid pressure and porosity after deposition of 10 km of uncompacted shale over 10 m.y. (**Input 7.1**), as predicted by Basin2. Various lines show results of runs assuming differing values for vertical permeability* $k_z$. *Fine lines show pressure gradients of 100 atm/km and 230 atm/km, for reference.*




The differing amounts of compaction observed in these runs explains why the curves in **Figure 7.1** extend from the surface to varying depths. In runs in which compaction proceeds normally, the 10 km pile of uncompacted sediment assumes a thickness of less than 6 km after compaction. When the pile expresses fluid too slowly to compact normally, however, the same amount of sediment assumes thickness greater than 6 km.

These calculations are unrealistic in several ways: the sedimentation rate of 1 mm/yr remains constant over 10 m.y.; the sediment is homogeneous, so there are no facies changes; and permeability does not change as the sediment compacts. To address the latter concern, we repeat the calculation for various sedimentation rates, assuming that the horizontal permeability of the shale follows the trend

$$\log k_x \text{ (darcy)} = 8\phi - 8 \qquad (7.7)$$

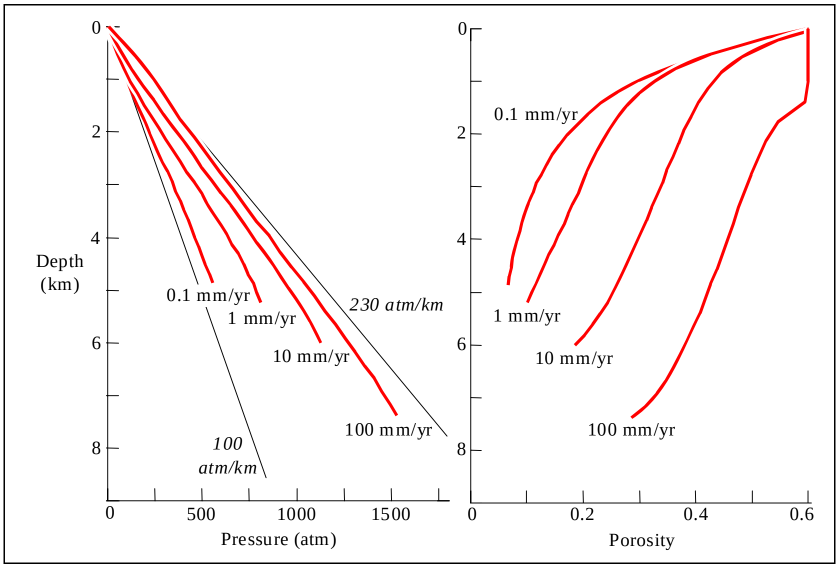
taken from the data of Neglia (1979).

We invoke the program with the command

```
b2 Input_7.1 -d AP=8 -d BP=-8 -d TIME0=-10
```

and then repeat the calculation, varying the sedimentation rate by changing the value of variable `TIME0`. The calculation results (**Figure 7.2**) show that as we increase the sedimentation rate, fluid pressure increases and approaches the lithostatic gradient. As before, porosity is preferentially preserved in overpressured sediments.

**Figure 7.2** *Profiles for fluid pressure and porosity at the end of Basin2 simulations in which 10 km of shale was deposited at differing rates. Sedimentation rates are expressed in terms of uncompacted sediment. The calculations assume that permeability follows the correlation* $\log k_x = 8\phi - 8$ *and that the permeability anisotropy* $k_x/k_z$ *is 10.*




An interesting further exercise is to have Basin2 calculate the bulk density $\rho_{sm}$ from the sediment porosity, instead of assigning the variable a constant value. To accomplish this, set

```
bulk_density = variable
```

in the second line of **Input 7.1**. In this case, the program predicts that, as permeability and sedimentation rates increase, the fluid pressure converges to values greater than hydrostatic but somewhat less than the 230 atm/km gradient seen in the previous results.

This result occurs because the program calculates a value for the bulk density of overpressured sediments (which are undercompacted) that is considerably less than 2.3 g/cm$^3$. As such, lithostatic pressures in these runs fall to the left of the 230 atm/km gradient commonly taken to define lithostatic. Interestingly, fluid pressures measured in compacting shale basins seldom plot above a gradient of about 200 atm/km.

## Compaction of the Niger Delta

In this section, we show how Basin2 can be used to model the compaction of an actual basin, the Niger Delta. **Input 7.2** was compiled by Charles Norris from references listed in **Appendix 5**. The input describes a cross section running 432 km landward from left to right. We set the finite difference grid to contain 30 columns of nodal blocks, each with a target thickness of 800 m of uncompacted sediment. The *y*-thickness of the cross section, set by variables `y_LHS` and `y_RHS`, increases from right to left to account for the seaward increase in the delta's width. Distributed across the cross section are 14 wells, the first 6 km from the left, and the last 426 km from the left.

**Input 7.2** *Niger Delta basin.*

```
title 'Development of the Niger Delta basin'
nx = 30; delta_z = 800 m
y_LHS = 10 cm; y_RHS = 1 cm
press_increase = 5; temp_increase = 5

rock ss
    A_perm = 15.5; B_perm = -5 log_darcy; p_kxkz = 2.5
rock sh
    A_perm = 8;    B_perm = -8 log_darcy; p_kxkz = 10
end_rock

width = 432 km
x_well(km)     6   56   76  121  161  186  216  \
             251  276  306  336  366  396  426

heat_flow = 1.5 HFU; relief = off
X_average = geometric, Z_average = harmonic

strat 'pre-Eocene'
    t_dep = -55 m.y.
    thickness  = 10 m; X(sh) = 1.0
    column  water_depth(km)
    w(1:8)      2.60
    w(9)        2.56
    w(10)       2.37
    w(11)       2.04
    w(12)       1.61
    w(13)       1.14
    w(14)       0.70

strat 'Early Eocene'
    t_dep = -49.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1:4)      0.0            2.600         0.0    1.0
    w(5)        0.064          2.536          "      "
    w(6)        0.239          2.361          "      "
    w(7)        0.574          2.026          "      "
    w(8)        1.081          1.519          "      "
    w(9)        1.435          1.128          "      "
    w(10)       1.688          0.682          "      "
    w(11)       1.722          0.317          "      "
    w(12)       1.722          0.079          "      "
    w(13)       1.700          0.0           0.10   0.90
    w(14)       0.800         -0.009         0.0    1.0

strat 'Lower Middle Eocene'
    t_dep = -46.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1:3)      0.0            2.600         0.0    1.0
    w(4)        0.010          2.590          "      "
    w(5)        0.171          2.365          "      "
    w(6)        0.265          2.096          "      "
    w(7)        0.346          1.679          "      "
    w(8)        0.383          1.136          "      "
    w(9)        0.400          0.759          "      "
    w(10)       0.450          0.375          "      "
    w(11)       0.550          0.111          "      "
    w(12)       0.800          0.002         0.10   0.90
    w(13)       1.000         -0.007         0.82   0.18
```

```
    w(14)       0.284         -0.022         0.22   0.78

strat 'Upper Middle Eocene'
    t_dep = -42.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1:2)      0.0           2.600         0.0    1.0
    w(3)        0.002         2.598          "      "
    w(4)        0.135         2.455          "      "
    w(5)        0.320         2.045          "      "
    w(6)        0.401         1.694          "      "
    w(7)        0.450         1.229          "      "
    w(8)        0.475         0.703          "      "
    w(9)        0.550         0.386          "      "
    w(10)       0.700         0.118          "      "
    w(11)       1.400         0.003         0.05   0.95
    w(12)       1.200        -0.007         0.72   0.28
    w(13)       0.491        -0.021         1.0    0.0
    w(14)       0.113        -0.049         0.50   0.50

strat 'Late Eocene'
    t_dep = -39.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1:2)      0.0           2.600         0.0    1.0
    w(3)        0.044         2.554          "      "
    w(4)        0.151         2.304          "      "
    w(5)        0.240         1.805          "      "
    w(6)        0.269         1.425          "      "
    w(7)        0.271         0.958          "      "
    w(8)        0.305         0.474          "      "
    w(9)        0.425         0.213          "      "
    w(10)       0.850         0.031          "      "
    w(11)       1.400        -0.003         0.5    0.5
    w(12)       0.620        -0.014         1.0    0.0
    w(13)       0.254        -0.035          "      "
    w(14)       0.027        -0.075          "      "

strat 'Early Oligocene'
    t_dep = -31.75 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.0           2.600         0.0    1.0
    w(2)        0.005         2.595          "      "
    w(3)        0.082         2.472          "      "
    w(4)        0.165         2.138          "      "
    w(5)        0.224         1.581          "      "
    w(6)        0.235         1.190          "      "
    w(7)        0.250         0.738          "      "
    w(8)        0.500         0.306          "      "
    w(9)        1.000         0.102          "      "
    w(10)       2.300         0.001         0.02   0.98
    w(11)       1.300        -0.008         0.95   0.05
    w(12)       0.526        -0.023         1.0    0.0
    w(13)       0.077        -0.052          "      "
    w(14)       0.026        -0.100          "      "

strat 'Late Oligocene'
    t_dep = -24.75 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.0           2.600         0.0    1.0
    w(2)        0.033         2.562          "      "
    w(3)        0.102         2.370          "      "
    w(4)        0.166         1.973          "      "
    w(5)        0.202         1.378          "      "
```

```
    w(6)        0.225          0.989          "       "
    w(7)        0.400          0.562          "       "
    w(8)        0.660          0.187          "       "
    w(9)        1.911          0.038          "       "
    w(10)       1.290         -0.002         0.55   0.45
    w(11)       0.620         -0.013         1.0    0.0
    w(12)       0.340         -0.034          "       "
    w(13)       0.020         -0.072          "       "
    w(14)       0.0           -0.100          "       "

strat 'Early Miocene'
    t_dep = -19 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.0            2.600         0.0    1.0
    w(2)        0.171          2.392          "       "
    w(3)        0.299          2.070          "       "
    w(4)        0.400          1.573          "       "
    w(5)        0.428          0.951          "       "
    w(6)        0.450          0.594          "       "
    w(7)        0.700          0.253          "       "
    w(8)        1.600          0.029          "       "
    w(9)        1.921         -0.002         0.47   0.53
    w(10)       1.304         -0.012         0.80   0.20
    w(11)       0.500         -0.031         1.0    0.0
    w(12)       0.140         -0.068          "       "
    w(13)       0.028         -0.100          "       "
    w(14)       0.0           -0.100          "       "

strat 'Lower Middle Miocene'
    t_dep = -14.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.061          2.539         0.0    1.0
    w(2)        0.296          2.096          "       "
    w(3)        0.391          1.679          "       "
    w(4)        0.437          1.136          "       "
    w(5)        0.396          0.555          "       "
    w(6)        0.800          0.272          "       "
    w(7)        1.600          0.057          "       "
    w(8)        2.507         -0.002         0.55   0.45
    w(9)        1.520         -0.011         0.80   0.20
    w(10)       0.760         -0.030         1.0    0.0
    w(11)       0.213         -0.064          "       "
    w(12)       0.032         -0.100          "       "
    w(13:14)    0.0           -0.100          "       "

strat 'Upper Middle Micoene'
    t_dep = -11.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.118          2.421         0.0    1.0
    w(2)        0.248          1.848          "       "
    w(3)        0.285          1.394          "       "
    w(4)        0.282          0.854          "       "
    w(5)        0.500          0.337          "       "
    w(6)        1.060          0.121          "       "
    w(7)        1.500          0.003         0.19   0.81
    w(8)        1.500         -0.008         0.80   0.20
    w(9)        0.740         -0.021         1.0    0.0
    w(10)       0.550         -0.048          "       "
    w(11)       0.035         -0.100          "       "
    w(12:14)    0.0           -0.100          "       "
```

```
strat 'Lower Late Miocene'
    t_dep = -8.75 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.147          2.273         0.0    1.0
    w(2)        0.237          1.611          "      "
    w(3)        0.251          1.143          "      "
    w(4)        0.600          0.627          "      "
    w(5)        1.200          0.187          "      "
    w(6)        1.900          0.038         0.31   0.69
    w(7)        1.400         -0.002         0.50   0.50
    w(8)        0.700         -0.016         1.0    0.0
    w(9)        0.550         -0.034          "      "
    w(10)       0.180         -0.072          "      "
    w(11:14)    0.0           -0.100          "      "

strat 'Upper Late Miocene'
    t_dep = -6.25 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.328          1.945         0.0    1.0
    w(2)        0.425          1.190          "      "
    w(3)        0.850          0.738          "      "
    w(4)        1.700          0.306          "      "
    w(5)        1.800          0.031         0.63   0.37
    w(6)        2.016         -0.002         0.50   0.50
    w(7)        0.900         -0.012         1.0    0.0
    w(8)        0.450         -0.035          "      "
    w(9)        0.225         -0.067          "      "
    w(10)       0.028         -0.100          "      "
    w(11:14)    0.0           -0.100          "      "

strat 'Pliocene'
    t_dep = -2.5 m.y.
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        0.600          1.378         0.0    1.0
    w(2)        0.850          0.627          "      "
    w(3)        1.514          0.277          "      "
    w(4)        2.190          0.038         0.33   0.67
    w(5)        2.095         -0.005         0.50   0.50
    w(6)        0.450         -0.016         1.0    0.0
    w(7)        0.600         -0.038          "      "
    w(8)        0.150         -0.091          "      "
    w(9)        0.033         -0.100          "      "
    w(10:14)    0.0           -0.100          "      "

strat 'Pleistocene'
    t_dep = 0 yrs
    column  thickness(km)  water_depth(km)  X(ss)  X(sh)
    w(1)        3.000          1.600         0.0    1.0
    w(2)        3.035          1.113          "      "
    w(3)        2.500          0.696         0.12   0.88
    w(4)        0.864          0.277         0.85   0.15
    w(5)        0.260          0.021         1.0    0.0
    w(6)        0.240         -0.002          "      "
    w(7)        0.509         -0.013          "      "
    w(8)        0.009         -0.038          "      "
    w(9)        0.0           -0.072          "      "
    w(10:14)    0.0           -0.100          "      "
```

The model accounts for two types of basin sediments, a sandstone `ss` and a shale `sh`. Except for the basal unit which is pure shale, the 14 stratigraphic units grade from sandstone landward to shale seaward.

Moving upward through the stratigraphic column, the sandstone progrades seaward across the cross section. Thus, the distribution of rock types (x values) reflects in a general way the typical distribution of sedimentary facies in a delta.

Since we are mainly interested in compaction, the input suppresses topographic relief, eliminating the groundwater flow that would arise from the elevation of the subaerial portion of the cross section. Overall permeability for each unit, as discussed in **Chapter 4**, is determined in the *x* direction by geometric averaging, and by harmonic averaging along *z*.

The model takes about 900 times steps and runs in about 90 seconds on a 400-MHz Pentium II PC. **Figure 7.3** shows at three points in geologic time the distribution of sandy facies in the basin along with the calculated pressure distribution and flow regime.

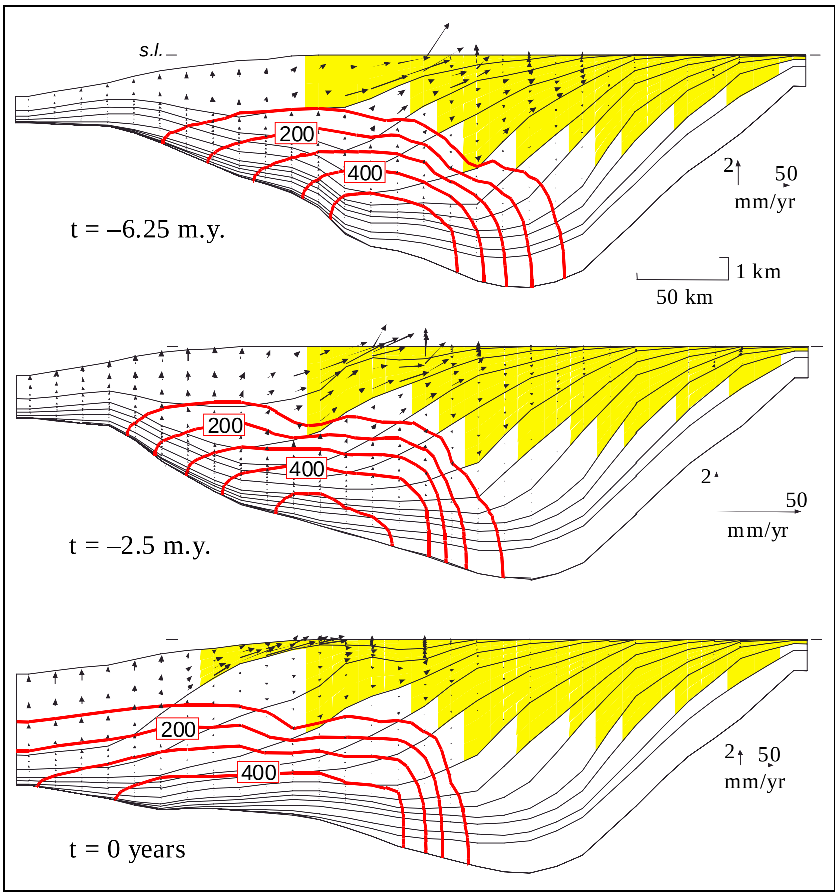



**Figure 7.3** *Distribution of overpressure (contour lines, in atm) and distribution of sandy facies (shaded areas) at three points in geologic time in the Niger Delta basin, as calculated by Basin2 and rendered using B2plot. Arrows represent velocities* $v_x$, $v_z$ *(see scales) of fluid expelled by sediment compaction.*

# *8* Groundwater Flow due to Topographic Relief

## About topography-driven flow

**Chapter 7** introduced the mathematical model that Basin2 uses to solve for the pressure distribution in a sedimentary basin and presented examples of groundwater flow driven by sediment compaction. In this chapter we consider the role of topographic relief in driving fluid flow through the subsurface.

As discussed in **Chapter 7**, gradients in hydraulic potential (or, when fluid density varies, gradients in pressure and elevation) drive groundwater flow. To see why flow occurs when the surface of a basin is exposed to topographic relief, consider the definition of hydraulic potential (**Equation 7.3**). Basin2 assumes that sediment is saturated with fluid everywhere in the basin, so the water table lies at the land surface. Since pressure $P$ along the water table is atmospheric, term $\rho gz$ in **Equation 7.3** requires that hydraulic potential increase with increasing elevation above sea level (i.e., with decreasing depth $z$ below sea level). Here, $\rho$ is fluid density, $g$ is the acceleration of gravity, and $z$ is depth below sea level.

A topographic slope, therefore, causes hydraulic potential to vary across the basin surface, giving rise to groundwater flow. No such effect occurs where the basin surface lies below sea level. Hydraulic potential is the same everywhere in a standing body of water because $\rho gz$ balances hydrostatic pressure. A subsea portion of the basin surface, therefore, defines an equipotential along which there is no drive for fluid flow.

In this chapter we present a few examples to illustrate how Basin2 can be used to model the role of topography in driving fluid flow. You set topographic relief with the keyword `water_depth`. Negative values of `water_depth` represent the elevation of the basin surface above sea level. By assigning values of `water_depth` well by well, you can specify topographic relief in any configuration along the basin cross section. As noted in **Chapter 4**, you can adjust the position of sea level, and hence the amount of topographic relief, with the `eustat` keyword.

## Flow in homogeneous basins

We start with a simple case: a homogeneous basin in which elevation varies linearly from left to right. **Input 8.1** shows the Basin2 input file used for this run. We set the program in steady-state mode, and by default leave both the left and right boundaries closed to flow. Since temperature is assumed constant and porosity does not change, topographic relief is the only force driving flow. Densities of fluid and rock (bulk density) are assumed constant for simplicity.

**Input 8.1** *Two-layer basin.*

```
run = steady;  start = 0 yrs
nx = 20;  delta_z = 200 m
temperature = 20 C
fluid_density = 1 g/cm3;  bulk_density = 2.3 g/cm3

x_well(km)   0   50

strat 'Unit 1'
   t_dep = -1 m.y.;  X(rk1) = 100%
   thickness = 1 km;  water_depth = 0

strat 'Unit 2'
   t_dep = 0 yrs;  X(rk2) = 100%
   column  thickness(km)  water_depth(km)
   w(1)          2             -1
   w(2)          1              0

rock rk1
   bpor = 0; A_perm = 0; B_perm = $BP1 log_darcy; p_kxkz = 10
rock rk2
   bpor = 0; A_perm = 0; B_perm = $BP2 log_darcy; p_kxkz = 10
end_rock
```

There are two stratigraphic formations. The bottom formation has a constant thickness, but in the top formation we set different values of `water_depth` and `thickness` at the left and right of the section. The `water_depth` values are negative or zero, so the whole basin surface lies above sea level.

Each formation is made up of a single rock type, either `rk1` or `rk2`. Porosity and permeability for each rock type are constant (`bpor = 0` and `A_perm = 0`); `p_kxkz` is set so that vertical permeability is one-tenth of the lateral value. The parameter `B_perm` is defined for rocks `rk1` and `rk2`, respectively, in terms of the variables `BP1` and `BP2`. Permeability in this two-layer basin can then be defined as either homogeneous or heterogeneous, depending on the values you set on the command line for these variables.

We begin by calculating the flow pattern when permeability is the same in both layers. Start Basin2 by typing

```
b2 Input_8.1 -d BP1=-3 -d BP2=-3
```

**Figure 8.1** shows the calculated potential distribution and flow pattern. Water recharges across the basin surface at high elevations and discharges at low elevations. Except near the left and right bounds, the equipotentials are almost vertical, and flow paths indicated by the velocity vectors follow lines almost parallel to the basin's stratigraphy. Velocities are slightly larger along the shortest pathlines (those near the basin surface).

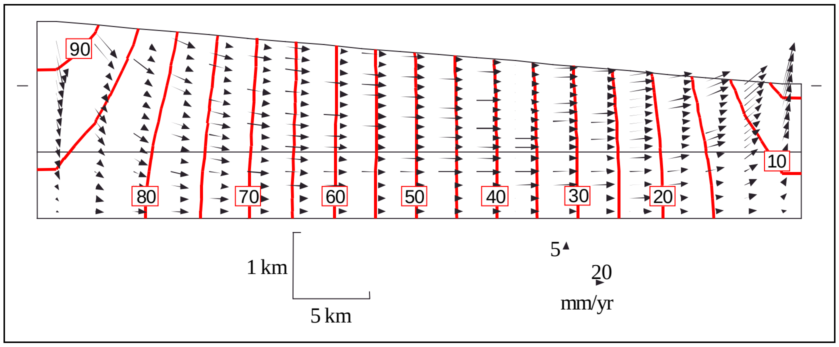



**Figure 8.1** *Basin cross section showing equipotentials and flow velocities in a homogeneous basin (**Input 8.1** with* BP1 = −3 *and* BP2 = −3*), as predicted by Basin2 at steady state.*

Next we consider the more interesting case (**Input 8.2**) in which topography is uneven instead of linear. This case differs from the previous example in that there is a single stratigraphic unit for which the values of `water_depth` and `thickness` at a number of wells across the basin define an irregular land surface. The drop in elevation between the left and right boundaries is the same as in **Input 8.1**, but now there are two valleys perched on the main slope. As before, the whole basin is above sea level.

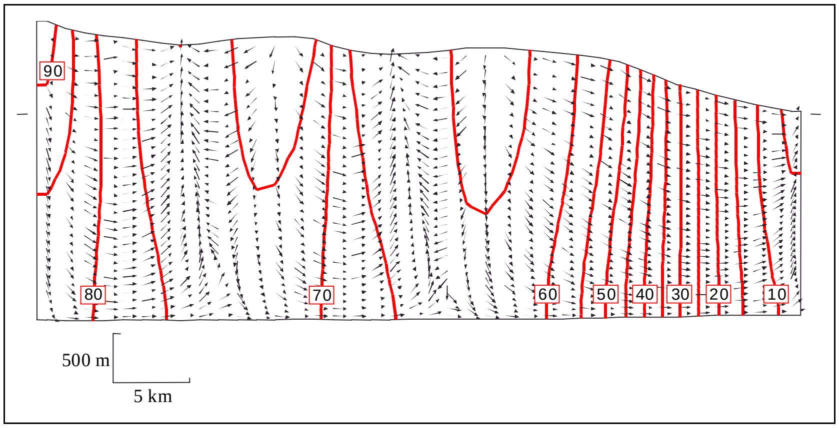
**Input 8.2** *Homogeneous basin with uneven topographic relief.*

```
run = steady_state; start = 0 yrs
nx = 40;  delta_z = 100 m
temperature = 20 C
fluid_density = 1 g/cm3;  bulk_density = 2.3 g/cm3

x_well(km)    0  1.5    3  4.5    6  7.5    9 10.5   12 13.5 \
             15 16.5   18   19   21   23   25   26   27   28 \
             30   34   38   42   46   50
```

```
strat 'Lone unit'
    t_dep = 0 yrs;  X(sh) = 100%
    column  thickness(km)  water_depth(km)
    w(1)        3.00          -1.00
    w(2)        2.90          -0.90
    w(3)        2.84          -0.84
    w(4)        2.80          -0.80
    w(5)        2.78          -0.78
    w(6)        2.74          -0.74
    w(7)        2.71          -0.71
    w(8)        2.72          -0.72
    w(9)        2.75          -0.75
    w(10)       2.78          -0.78
    w(11)       2.79          -0.79
    w(12)       2.80          -0.80
    w(13)       2.78          -0.78
    w(14)       2.72          -0.72
    w(15)       2.64          -0.64
    w(16)       2.62          -0.62
    w(17)       2.63          -0.63
    w(18)       2.64          -0.64
    w(19)       2.66          -0.66
    w(20)       2.68          -0.68
    w(21)       2.69          -0.69
    w(22)       2.64          -0.64
    w(23)       2.55          -0.55
    w(24)       2.30          -0.30
    w(25)       2.13          -0.13
    w(26)       2.00           0.00
```

**Figure 8.2** shows the steady-state flow pattern calculated by Basin2. The potential distribution differs considerably from that in the first example. The upper half of the basin is divided into three sub-basins, and the two points of high elevation act as groundwater divides. A regional flow system, in which all water flows from left to right, appears only at depth. The influence of the water table configuration on the upper half of the basin is felt also in the lower half, where equipotentials are not spaced as evenly as those in the previous example. Tóth (1963) and Freeze and Witherspoon (1967) show further examples of how the water table configuration and permeability distribution affect basin flow regimes.

**Figure 8.2** *Basin cross section showing equipotentials and flow pattern in a basin with uneven topography (**Input 8.2**), as predicted by Basin2 at steady state. Arrows show the direction but not velocity of groundwater flow.*




## Flow in heterogeneous basins

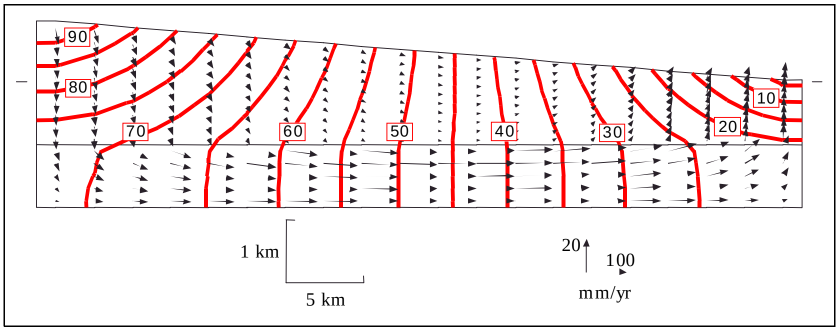
We now consider flow in basins composed of stratigraphic layers of differing permeability. In the first example, we take a two-layer basin in which an aquitard overlies an aquifer. In the input (**Input 8.1**), we set the aquifer to be an order of magnitude more permeable than the aquitard. Invoke Basin2 by typing

```
b2 Input_8.1 -d BP1=-2 -d BP2=-3
```

Here, the values of `BP1` and `BP2` are the log permeabilities (in log darcys), respectively, of the aquifer (type `rk1`) and aquitard (type `rk2`). **Figure 8.3** shows the equipotentials and fluid velocities calculated by Basin2.

The results displayed in **Figure 8.3** differ from those shown in **Figure 8.1**. Flow in the top layer is now mainly vertical, whereas in the more permeable bottom layer it is mostly horizontal. Equipotentials are far from vertical in the top layer, in agreement with the direction of flow indicated by the flow velocity vectors. Note that if the approximately 10:1 vertical distortion of the plot were eliminated, equipotentials near the left and right boundaries in the top layer would be almost horizontal. Most water in the basin flows through the aquifer from left to right; it is only because the left and right boundaries are closed that water migrates vertically through the aquitard.

**Figure 8.3** *Cross section showing equipotentials and flow velocities in a two-layer basin (**Input 8.1** with* BP1 = –2 *and* BP2 = –3, *as predicted by Basin2 at steady state.*

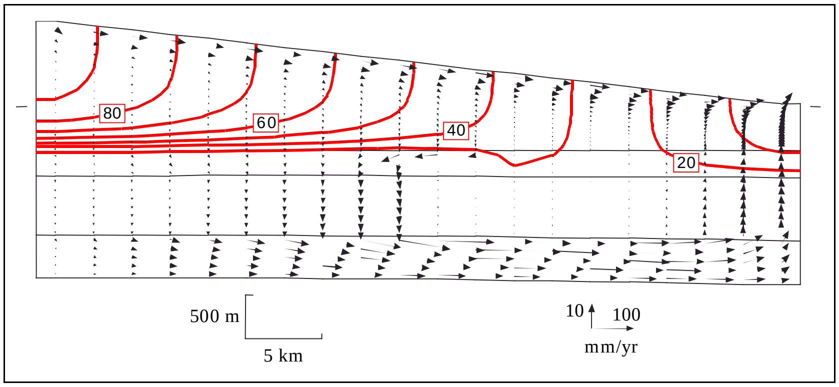


With Basin2 you can predict the flow pattern in a basin having a rock composition that varies not only with depth but also from well to well. Consider the data in **Input 8.3**. The basin is composed of four layers. The lowermost layer is a sandstone aquifer that runs through the basin. The second layer from the bottom and the uppermost layer are homogeneous aquitards made up of shale and carbonate, respectively. The third layer from the bottom is heterogeneous: the left half is composed of sandstone and the right half of shale. The left half of this layer is therefore a second aquifer. **Figure 8.4** shows equipotentials and velocity vectors calculated by Basin2.

**Input 8.3** *Four-layer basin.*

```
run = steady_state;  start = 0 yrs
nx  = 20;  delta_z = 100 m
temperature = 20 C
fluid_density = 1 g/cm3;  bulk_density = 2.3 g/cm3

width = 50 km
x_well(km)   0   25   30   50

strat 'Unit 1'
   t_dep = -3 m.y.;  X(ss) = 100%
   thickness = 500 m

strat 'Unit 2'
   t_dep = -2 m.y.;  X(cn) = 100%
   thickness = 700 m

strat 'Unit 3'
   t_dep = -1 m.y.
   thickness = 300 m
   column   X(ss)  X(sh)
   w(1:2)     1      0
   w(3:4)     0      1

strat 'Unit 4'
   t_dep = 0 yrs;  X(sh) = 100%
   column  thickness(km)  water_depth(km)
   w(1)        1.5            -1.0
   w(2)        1.0            -0.5
   w(3)        0.9            -0.4
   w(4)        0.5             0.0
```

**Figure 8.4** *Cross section showing equipotentials and flow velocities in a four-layer basin (**Input 8.3**), as predicted by Basin2 at steady state.*




The equipotentials in **Figure 8.4** are spaced most closely together near the left and right boundaries, indicating that potential gradients are greatest there. The vertical gradients drive water into the top aquifer in the left half of the basin and out of the bottom aquifer near the right boundary. Flow is channeled along a path of high permeability, following the sandstone aquifers. Although water is drawn into the top aquifer, significant horizontal flow does not develop within it because the permeable part of the layer spans only part of the basin; therefore, water has to travel vertically downward to the bottom aquifer. In the top layer, hydraulic potential has a more constant gradient, and flow paths near the surface are similar to those of the homogeneous basin of the first example. Because of the different permeabilities in underlying formations, however, a smaller fraction of the total discharge in this example passes near the surface.

# *10* Heat Transfer and Thermal Convection

## About heat transfer

Basin2 can calculate basin thermal regimes accounting for groundwater flow, variation in thermal conductivity of basin sediments, and changing heat flows and surface temperature. Determining the effects of these factors on the subsurface temperature regime and on the thermal history of a basin are essential steps in attempting to model the thermal maturation of organic matter in basin strata (see **Chapter 11**).

Basin2 can also calculate the effect of buoyant forces in driving fluid convection. Buoyant forces arise whenever fluid density, and hence temperature, vary laterally. For this reason, variation in heat flow, surface temperature, or thermal conductivity along the cross section can cause fluid convection. Convection can also arise spontaneously in an aquifer heated from below, even if initially there is no lateral temperature gradient, depending on the aquifer's permeability, thickness, and thermal conductivity, and the thermal gradient across it. Combarnous and Bories (1975) give a thorough description of the origin and nature of thermal convection.

## Rate of heat transfer

Basin2 calculates heat transfer by conduction through sediments and pore fluid and, if specified (see **Chapter 6**), by fluid advection. The program does not account for radiative heat transfer, because at submetamorphic temperatures sediments and rocks are opaque to infrared radiation.

Fourier's law gives the rate of heat conduction $q_H$

$$q_{H_x} = -K_x \frac{\partial T}{\partial x}$$

and

$$q_{H_z} = -K_z \frac{\partial T}{\partial z} \tag{10.1}$$

from the thermal conductivity $K_x$ and $K_z$ and the temperature gradient along $x$ and $z$. Here, $q_H$ might be given in cal/cm$^2$ sec or J/cm$^2$ sec, and the thermal conductivity in cal/cm$^2$ sec °C or J/cm$^2$ sec °C.

The rate of advective transport $q_A$, in cal/cm$^2$ sec or J/cm$^2$ sec, is the rate at which migrating fluid carries thermal energy across a unit plane. This quantity is the product of the mass flux and the fluid's enthalpy $h_w$

$$q_{A_x} = \rho q_x h_w$$

and

$$q_{A_z} = \rho q_z h_w \tag{10.2}$$

where $\rho$ is fluid density, $q_x$ and $q_z$ are specific discharge along $x$ and $z$, and $h_w$ is fluid enthalpy, as discussed in **Chapter 3**.

## Thermal convection

To illustrate how we use Basin2 to model heat transfer, we calculate at steady state the effect of topographic relief on thermal convection. We set up a cross section 20 km wide and about 2 km deep (see **Input 10.1**). The section is composed of a basal formation overlain by a wedge of sediment that extends from the left to the center of the section. We set heat flow to be 4 HFU at the left of the cross section, 2 HFU at its center, and 1 HFU along the right side. The variation in heat flow creates a lateral temperature gradient that drives convection.

**Input 10.1** *Effect of topography on thermal convection.*

```
run = steady; start = 0 yrs
nx = 30;  delta_z = 100 m
temperature = full; surface_temp = 20 C

rock ss
    A_perm = 0;  B_perm = –2 log_darcy;  p_kxkz = 10
    phi0 = 30%; phi1 = 0; bpor = 0
end_rock

X(ss) = 1
x_well(km)  0 10 20

strat 'Basal unit'
    t_dep = –1 m.y.
    thickness = 2 km

strat 'Topography'
    t_dep = 0 yrs
    column    water_depth(m)  thickness(m)  heat_flow
    w(1)        $TOPO            $THICK           4
    w(2)            0                 0           2
    w(3)            0                 0           1
```

The wedge and underlying formation are composed of a single rock type. We set porosity to a constant value of 30%, and lateral and vertical permeability, respectively, to $10^{-2}$ and $10^{-3}$ darcy.

Variables TOPO and THICK control the elevation and thickness of the wedge. We set TOPO to a negative value and THICK to a positive number of the same magnitude. For example, we might initialize a run with the command

```
b2 -i B2in_convect.txt -d TOPO=-50 -d THICK=50
```

In this way, the base of the wedge and the top of the underlying formation lie along the horizontal.

We set a steady-state run, and in doing so implicitly assume that with time the flow regime becomes invariant. This assumption is not necessarily valid, because natural flow regimes in which fluids convect actively sometimes will oscillate rather than assuming a steady flow pattern. For this reason, Basin2 may have difficulty converging to solutions for certain problems posed at steady-state.

**Figure 10.1** shows the results of the calculation made by setting zero, 50 m, and 100 m of topographic relief. When no relief is set, a plume of upwelling fluid develops along the left side of the section, above the point of highest heat flow. The effect of 50 m of relief is to diminish the plume and produce a second plume toward the right side of the section, where water driven by the topography discharges. With 100 m of relief, a single plume develops in the right half of the cross section, even though the basal heat flow is smaller there than along the left half.

**Figure 10.1** *Effects of topography on thermal convection, calculated assuming (top) no relief, and (middle) 50 m and (bottom) 100 m of relief from left side to center of cross section. Contours are isotherms labeled in °C. Arrows show velocities* $v_x$, $v_z$ *(see scale) of fluid migration driven by topographic relief and buoyant forces.*