Final Report

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OBJECTIVES

The objective of this study is to investigate the cause(s) of abnormally high pore fluid pressures in the Uinta Basin, Utah.

PROJECT DESCRIPTION

The Altamont-Bluebell oil-field, in the deepest part of the Uinta Basin, has pore pressure that in some places exceeds 0.8 of lithostatic. There are two prevailing explanations for the high pore pressure: 1) sedimentary loading and 2) the volume expansion and collapse of porosity associated with the conversion of solid kerogen to liquid oil. The purpose of this research is to test these hypotheses of pressure generation using three-dimensional, multiphase numerical simulation. These simulations will allow us to further investigate the migration of oil within the Uinta Basin.

ABSTRACT

High pore fluid pressures, approaching lithostatic, are observed in the deepest sections of the Uinta basin, Utah. We analyzed the cause of the anomalous overpressures with a 3-dimensional, numerical model of the evolution of the basin, including compaction disequilibrium and hydrocarbon generation as possible mechanisms. The numerical model builds the basin through time, coupling the structural, thermal and hydrodynamic evolution, and includes in situ hydrocarbon generation and migration. We used the evolution model to evaluate overpressure mechanisms and oil migration patterns for different possible conceptual models of the geologic history.

Model results suggest that observed overpressures in the Uinta basin are probably caused by ongoing oil generation in strata with specific conditions of permeability, relative permeability, TOC content, and oil viscosity. We conducted a sensitivity analysis that suggests that for oil generation to cause overpressures, the necessary conditions are: oil viscosity is ~ 0.05 cP or higher, intrinsic permeability is $\sim 5 \times 10^{-18}$ m² or lower, and source rock TOC values are ~ 0.5 % or higher.

We also analyzed hydrocarbon migration patterns in the basin and how they are affected by the basin's structural history. Oil migration patterns produced by the model are consistent with published oil production maps: oil moves from the deep Altamont source rocks towards Redwash, the eastern Douglas Creek Arch area, and southward towards the Sunnyside tar-sands and Book Cliffs. Peak oil generation occurs from the time of maximum burial in the mid-Tertiary (~ 35 to ~ 30 Ma). Most differential uplift of

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the basin's flanks probably occurs well after this time, and most oil migration to the basin's southern and eastern flanks occurs prior to uplift of these flanks. Model results show that if the basin's flanks are uplifted too soon, reduced temperatures increase viscosity enough to inhibit migration of oil to the southern and eastern flanks.

INTRODUCTION

The condition whereby pore fluid pressure greatly exceeds hydrostatic pressure is known as overpressure and is prevalent in the deeper, low-permeability sections of sedimentary basins throughout the world. Economic interest in sedimentary basin overpressures exists because zones of overpressure may indicate the presence of oil or gas (Timko and Fertl, 1971; Spencer, 1987). Producing overpressured oil or gas can be difficult, however, because overpressured formations are often very difficult or dangerous to drill (Thomeer and Bottema, 1961; Bethke, 1986).

Overpressures occur after the amount of in situ fluid completely fills available pore space, and fluid expulsion during further compaction or fluid volume increase is not rapid enough to maintain hydrostatic pressure. Fluid volume may increase if new fluid is generated in situ by geochemical processes, for example, by clay dewatering or hydrocarbon generation. Alternatively, fluid volume may increase due to a reduction in its density. "Aquathermal pressuring" is a common overpressure mechanism caused by decreases in water density accompanying increases in temperature (Daines, 1982). "Compaction disequilibrium," another common overpressure mechanism, refers to both abnormal compaction trends and overpressures resulting because the rate of compaction exceeds the rate at which in situ fluids escape a compacting medium (Magara, 1975). The history of overpressure research to date has recently been reviewed by Neuzil (1995).

Overpressures observed in the northern half of the Uinta basin approach lithostatic pressure (Fouch et al, 1992), and are thought to result from either hydrocarbon generation (Spencer, 1987; Sweeney et al, 1987; Bredehoeft et al, 1994) or from compaction disequilibrium (Sweeney, written communication, 1992). Although it is difficult to prove or disprove any overpressure mechanism, strong inferences can be made from numerical modeling studies. Sweeney et al (1987) concluded that oil generation is a feasible overpressure mechanism in the Uinta basin because their modeled oil generation rate vs. depth curves mimic the shape of observed overpressure-depth curves. Bredehoeft et al (1994) arrived at a similar conclusion based on a 3-dimensional numerical model of the hydrodynamics of the present-day basin. The study of Bredehoeft et al (1994) is limited because their model included only single phase (water) flow, and compaction disequilibrium was not addressed. No studies that test compaction disequilibrium together with hydrocarbon generation as possible overpressure mechanisms in the Uinta basin have been published.

In fact, few published sedimentary basin studies consider both compaction disequilibrium and oil generation as potential overpressure mechanisms. This deficiency occurs primarily because structural evolution and associated compaction processes are difficult to model. Numerical models necessary for these types of analyses are complex and difficult to construct, and require considering an unwieldy number of physical properties and interrelated parameters. Additionally, for a numerical model to simulate

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compaction, it must necessarily be dynamic, i.e., the model domain must deform corresponding to compaction, which is coupled to the structural and hydrodynamic history via effective stress.

One example of a dynamic model is discussed in Burrus et al (1992), who analyzed compaction processes, hydrocarbon generation, expulsion and migration along a cross-section of the Northern Viking Graben of the North Sea. Burrus et al (1992) concluded that compaction disequilibrium was the primary cause of overpressures in the Northern Viking Graben. Although their study was very thorough, a caveat of the Burrus et al (1992) analysis is that it was only 2-dimensional. This study of the Uinta basin began as a 2-dimensional study of a cross-section of the basin. We concluded, however, that the model could not properly reconstruct the complex thermal and hydrogeologic history of the basin because it was not clear how surrounding parts of the basin affected the part being modeled (McPherson and Bredehoeft, 1992). We thus expanded this study to a 3-dimensional analysis.

The primary goal of this study is to understand better high pore fluid pressures and regional scale hydrocarbon migration patterns in the Uinta basin, Utah. Implicit in this objective is the need to understand the whole basin system, the individual geological processes that comprise the system, and how the processes interact. To achieve this goal we constructed 3-dimensional, numerical models of the evolution of the basin with special consideration of compaction disequilibrium and hydrocarbon generation as possible overpressure mechanisms. We developed these models not to reproduce observed data, but rather to understand the complicated basin system.

GEOLOGIC SETTING

The Uinta basin of northern Utah (Figure 1) marks the northernmost edge of the Colorado Plateau. The basin is a structural down-fold that developed contemporaneously with the Laramide orogeny of latest Cretaceous and early Tertiary age (75 to 50 Ma); in shape (Figure 1), and structurally it is asymmetrical to the north (Figure 2). Sediment deposition was continuous through the mid-Tertiary. In map view, the basin is elliptical with its major axis almost due east-west, which is unique among the Laramide intermontane basins. The deepest part of the basin is along the northern edge where it is bounded by a large basement thrust fault that dips northward (Campbell, 1978).

The Uinta basin is bounded on the north by the Uinta Mountains, on the west by the Wasatch uplift, on the east by the Douglas Creek arch, and on the south by the Uncompahgre uplift and the San Rafael Swell. These adjacent highlands are sources of Tertiary sediments, in addition to other sources several hundred km south of the basin such as the San Juan Mountains (Hintze, 1980; Pitman et al, 1982).

Tertiary sediments approach 6 km thickness in the northern part of the basin, and strata along the southern flank of the basin dip between 3 and 6 degrees north (Figure 2; location of profile shown on Figure 1). Cross-section N-S is used to illustrate many model results. The location of profile N-S (Figure 1) was chosen because the profile crosses both the Altamont oil field and the Sunnyside Tar Sands, both important hydrocarbon bearing areas to which we compared oil migration results.

This study focuses primarily on the Tertiary Green River Formation. Figure 2(a) shows a generalized stratigraphic column of the Green River Formation, which is an Eocene lake deposit consisting of intercalated sandstones, shales, mudstones, and carbonates, especially dolomite. The Green River Formation is also common to many other Laramide basins, including the Bridger basin, Washakie basin, and the Piceance Creek basin. Within the Uinta basin, the Green River Formation consists of three primary depositional facies (Fouch, 1975; Ryder et al, 1976): open lacustrine, marginal lacustrine, and alluvial. These three facies comprise a lacustrine depositional system associated with Lake Uinta, a lake thought to cover much of the Uinta basin, Piceance Creek basin and much of the Wasatch Plateau during late Cretaceous to middle Eocene time. In this system the open lacustrine facies is the most distal facies and consists primarily of mud-supported carbonates and other organic-rich sediments that are associated with lake-bottom deposits. The marginal lacustrine facies consists primarily of claystones, sandstones and carbonates associated with lake-margin deltaic and interdeltic environments. Strata in both open lacustrine and marginal lacustrine facies are source rocks for oil and gas in the basin, especially oil shale. Finally, the alluvial facies is the most proximal in the system and consists primarily of channel sandstones, fine-layered siltstones and very fine grained sandstones and claystones. Ryder et al (1976) provide excellent conceptual schematic diagrams of the depositional system and describes facies and stratigraphy in great detail. Additionally, Fouch et al (1992) provide updated maps of facies for several isochronous horizons in the Uinta basin. The geologic history of the Uinta basin is discussed in the following section.

CONCEPTUAL MODEL OF UINTA BASIN GEOLOGIC HISTORY

A numerical model of a basin's history is meaningful only if the conceptual model of the history is realistic. The structural history of the Uinta basin is relatively simple, and can be subdivided into three main parts:

- (i) Subsidence and deposition occurred contemporaneous with the Laramide orogeny and uplift of the Uinta mountains (from ~70 to ~80 Ma until ~40 to ~30 Ma). The structural history of the basin after it achieved maximum depth (maximum deposited thickness) is not definitively known because subsequent uplift and erosion have eliminated most evidence of the surface of maximum aggradation (maximum original topography). A consequence is that the time at which maximum depth was reached is uncertain. However, several studies (listed in Table 1) have estimated the time of maximum depth for various parts of the basin.
- (ii) A period of uplift/erosion of the basin's flanks occurred, in particular the southern periphery of the basin was uplifted to form the "bowl" shape of the basin. It is the bowl-type structure and topography within and around the basin that control the hydrodynamic and thermal regimes. Thus, the timing (e.g., time of initiation and duration) of this phase of the basin's history is critical. A structure-contour map of the Flagstaff member of the Green River Formation (Figure 3) illustrates that the southern periphery of the basin has been uplifted, forming the ~5° dip of strata south of the basin axis (see also Figure 2, structural cross-section N-S). A cross-section extending from the northern side

of the Uinta basin (Vernal, Utah) through the San Rafael Swell (Hintze, 1980) suggests that this uplift is tied to the Laramide uplift of the San Rafael Swell.

(iii) Relatively rapid uplift of all strata (uplift of the entire Colorado Plateau) occurred beginning ~10 Ma until the present. Many researchers (Fouch et al, 1992; Anders et al, 1992; Johnson and Nuccio, 1993) agree that major uplift of the entire Colorado Plateau began ~10 Ma and is continuing today.

We have strong confidence in parts (i) and (iii) of our conceptual model of the basin's history. The most uncertain aspect of the history model is part (ii), the timing (i.e., starting time, duration, rate) of the uplift phase that forms the primary structure of the basin, but the range of possibilities lies between two end-member cases: (1) all differential uplift and erosion of the basin's southern periphery is coincident with uplift of the entire region during just the past 10 My -- what we will refer to as the "late uplift" history model; (2) all uplift and associated erosion began during the Laramide orogeny and continued through the Oligocene -- what we will refer to as the "early uplift" history model. Oligocene is chosen as the earliest possible ending time of uplift because primary deposition of the Uinta and Duchesne River formations is thought to have ceased at that time; in other words, it is possible that uplift occurred only until the surface layer finished deposition.

Although both conceptual models are geologically possible, we believe the early uplift model is probably the most geologically reasonable of the two cases. Various thermal history indicators (e.g., fission track, $^{40}\text{Ar}/^{39}\text{Ar}$, vitrinite reflectance analysis) could be used to better constrain the timing of uplift and erosion for at least parts of the basin. A primary objective of this study, however, is to analyze how the full range of possible structural histories affect hydrodynamic aspects, thermal aspects, overpressures and oil migration. Thus, these two end-member cases form a basis for the model simulations: simulating the end-member possible structural histories provides the end-members of possible hydrodynamic histories and oil migration patterns. Finally, the model simulations provide valuable insight about overpressure development for the Altamont field and surrounding area.

NUMERICAL MODEL OF BASIN EVOLUTION

Integrated Finite Difference Simulator

In the initial phase of this study (McPherson and Bredehoeft, 1992), we concluded that overpressures in the Uinta basin could not be analyzed properly with a 1-dimensional or 2-dimensional model. A priority, therefore, was to develop a 3-dimensional model that reconstructs the basin through time, and includes hydrodynamic, thermal, and hydrocarbon aspects. Rather than build a 3-dimensional model from scratch, we modified TOUGH2 (Pruess, 1991), a public domain geothermal reservoir simulator developed at the Lawrence Berkeley Laboratory, for our purposes. TOUGH2 was specifically designed to simulate coupled heat and multiphase fluid transport in porous and fractured media. We examined and tested several simulators before using TOUGH2, and concluded that TOUGH2 offered many advantages for this study. First and most important, it is a 3-dimensional simulator. Second, it is fairly well known, is currently used by many workers, and has been tested and compared with several other geothermal reservoir simulators (Pruess and Wang,

1984; Moridis and Pruess, 1992; Oldenburg and Pruess, 1994). Finally, TOUGH2 uses an integrated finite difference method (Narasimhan and Witherspoon, 1976), the advantages of which we detail below.

We developed separate routines that modify the integrated finite difference mesh, its parameterization and its boundary conditions according to the geologic history imposed. These routines modify the finite difference mesh during the simulation: a layer of finite difference cells is added to simulate addition of a new formation, and those cells are thickened to simulate sedimentation. Note that sediments are always deposited horizontally. Similarly, a layer of nodes is thinned to simulate erosion, and that layer of cells is removed when the entire formation is eroded. Cells are moved with respect to each other to simulate differential uplift or subsidence. With this dynamic mesh the transient effects of sedimentation, erosion, and structural changes (uplift and subsidence) on the hydrogeological and thermal aspects of the basin are explicitly considered. The integrated finite difference method is particularly appropriate for modeling with a dynamic mesh. According to Anderson and Woessner (1992), the finite element method requires that each element be constructed so that its aspect ratio (the ratio of maximum to minimum element dimensions) is close to unity, at least for isotropic materials. Aspect ratios can be higher than 1, but a value exceeding 5 should be avoided (Anderson and Woessner, 1992); otherwise, numerical instabilities or errors may result. The finite difference method is not as prone to aspect ratio problems, but finite difference grids must be rectangular. The integrated finite difference method offers the mesh flexibility of the finite element method without the aspect ratio constraints, making the method particularly useful for modeling active sedimentation. A layer begins as a thin veneer of sediment and thus nodes representing it can have very large aspect ratios at the initial stage of deposition. Additionally, the integrated finite difference method offers the numerical robustness (at high nodal aspect ratios) of the finite difference method without the regular, rectangular grid spacing constraints. This property is beneficial for simulation of later stages of the basin's history when differential uplift and flexure cause the basin structure to have variable curvature.

Algorithm

A flow chart summarizing the model algorithm is detailed in Figure 4. The algorithm is based on the original TOUGH2 model; major revisions are indicated by a larger, bold font. These additions include (reference portions labeled (a) through (g) on Figure 4):

- a. At the time in the basin's history in which sedimentation of a new formation begins, a new layer(s) of finite difference nodes is added to the model domain mesh. Each new layer is assigned an initial thickness of 1 m (to minimize numerical instability associated with extreme aspect ratios (e.g., $< 10^4$), and thickening of the layer begins as the time that deposition would have accumulated 1 m of sediment.
- b. The amount of oil generated from kerogen, as determined for the previous time step, is injected at the source cells to simulate oil expulsion.
- c. The amount of water that compaction expels from pore space within each individual cell, as determined from the previous time step, is injected to account for changes in fluid mass balance associated with compaction.

The processes outlined in d. through f., including changes in the model mesh and estimation of oil generation rates, are performed after each successful (converged) time step:

- d. Changes in effective stress and resulting compaction (collapse of pore space) are determined and assigned for each cell.
- e. Changes in permeability as dictated by porosity changes are assigned for each cell.
- f. Structural changes, including thickening of layers due to sedimentation, thinning or removal of layers by erosion, and layer curvature caused by differential uplift/subsidence are performed.
- g. Integration of the time-temperature history up through the current successful time step and resulting oil generation rate and cumulative amount of oil generated for each oil source cell is computed.

Model aspects (a) through (g) are discussed in greater detail in subsequent sections. For a complete description of the original TOUGH2 algorithm, including the governing equations and solution method, see Pruess and Narasimhan (1985), Pruess (1988), and Pruess (1991).

Treatment of Porosity Reduction Due to Compaction

During sedimentation, increased overburden causes the sedimentary section to compact and lithify. Compaction causes porosity reduction and subsidence, and it squeezes out pore fluids. If sedimentation rates are relatively high and the deposited sediment is highly compactible, pore fluid pressures often increase, in turn causing a minor increase in porosity (Hanshaw and Bredehoeft, 1968).

Many numerical models of basins use a depth-porosity correlation to govern porosity evolution. A depth-porosity correlation implies that porosity monotonically decreases during the basin's history. However, porosity changes are caused by changes in effective stress, which tracks changes in the depositional history and the hydrodynamic history. Effective stress is a function of both overburden and fluid pressure, and thus porosity evolution is not necessarily monotonic.

The magnitude of porosity change for a given change in effective stress depends on the compressibility of the medium. Compressibility, α , is defined as

$$\alpha = -\frac{\partial e}{\partial \sigma_e(1+e_o)} \quad (1)$$

where e is void ratio, e_o is the initial void ratio of the sediment at deposition and σ_e is the effective stress. Void ratio is expressed in terms of porosity as

$$e = \phi/(1-\phi), \quad (2)$$

and effective stress is defined as the difference between the total stress and fluid pressure, or

$$\sigma_e = \sigma_t - p. \quad (3)$$

The model addresses both elastic and inelastic changes in porosity. The solid line shown on Figure 5 is a plot of porosity versus effective stress and represents the functional relationship used to assign porosity in the basin evolution model. The relationship is calibrated with measured porosity data discussed in the next section and also shown on Figure 5. The solid curve is termed the "virgin loading curve" and represents non-elastic or permanent changes in porosity. The dashed curves are examples of "unloading/reloading curves" (unloading curves and reloading curves are typically slightly different, but in this study we approximate them as the same for convenience (Domenico and Mifflin, 1965) and represent elastic changes in porosity. During sedimentation the rock column thickens, effective stress increases (except in the case of high fluid pressure and an effective stress decrease) and in turn model porosity decreases along the virgin curve. If effective stress decreases, an elastic increase in model porosity occurs that tracks the slope of an unloading curve. When effective stress subsequently increases, model porosity decreases along the reloading curve until it reaches the virgin curve, at which point model porosity continues to decrease along the virgin curve. Thus, two values of compressibility can be defined: α_v is the compressibility dictating the virgin curve and inelastic porosity changes, and α_r is the compressibility dictating the reloading curve and elastic porosity changes. Measured porosity data, discussed in the following section, were used to calibrate approximate values of compressibility $\alpha_v = 10^{-7} \text{ Pa}^{-1}$ and $\alpha_r = 10^{-9} \text{ Pa}^{-1}$ for all strata, which are within the range of virgin and reloading compressibilities for most sedimentary lithologies (Domenico and Mifflin, 1965).

A compaction term, derived by Bredehoeft and Cooley (1983), was added to the governing equations of the numerical model to account for changes in fluid mass balance associated with porosity changes. The continuity equation for flow through a differential volume element, $\Delta x \Delta y \Delta z$, can be stated as

$$\nabla \cdot \rho \mathbf{v} = \frac{\partial(\Delta M)}{\partial t} / \Delta x \Delta y \Delta z, \quad (4)$$

where ρ is the density of the fluid, $\mathbf{v}$ is the fluid flux, and ΔM is the fluid mass within the volume element (Jacob, 1950). An implicit assumption behind equation (4) is that the position and size of the differential volume element is stationary. Incidentally, equation (4) is the basic governing equation solved in TOUGH2, minus any source term (see equation (1) of Pruess, 1987).

The mass of fluid in the volume element is

$$\Delta M = \rho \phi \Delta x \Delta y \Delta z \quad (5)$$

where ϕ is porosity. For the sake of mathematical simplification, we assume that changes in the lateral dimensions are negligible, but that the z-dimension may change (e.g., due to compaction). Under this assumption the rate of change of the fluid mass through time may be stated as

$$\frac{\partial(\Delta M)}{\partial t} = \left[\phi(\Delta z) \left(\frac{\partial \rho}{\partial t} \right) + \rho(\Delta z) \left(\frac{\partial \phi}{\partial t} \right) + \rho \phi \left(\frac{\partial(\Delta z)}{\partial t} \right) \right] \Delta x \Delta y. \quad (6)$$

Following Jacob (1950), if it is assumed that the change in volume of solids is negligible compared to the change in porosity, then

$$\frac{\partial(\Delta z)}{\partial t} = -\alpha \Delta z \left(\frac{\partial \sigma_e}{\partial t} \right), \quad (7)$$

where α is vertical compressibility and σ_e is effective stress. Additionally, the change in porosity is

$$\frac{\partial \phi}{\partial t} = -\alpha(1 - \phi) \left(\frac{\partial \sigma_e}{\partial t} \right), \quad (8)$$

and the change in density is

$$\frac{\partial \rho}{\partial t} = \beta \rho \left(\frac{\partial p}{\partial t} \right), \quad (9)$$

where β is the compressibility of the fluid and p is the pore pressure. Combining equations (6) through (9) and introducing the effective stress law,

$$d\sigma_e = d\sigma_t - dp, \quad (10)$$

yields the following form of the flow equation:

$$-\nabla \cdot \rho \mathbf{v} = \phi \rho \left(\beta + \frac{\alpha}{\phi} \right) \frac{\partial p}{\partial t} - \rho \alpha \frac{\partial \sigma_t}{\partial t} \quad (11)$$

The vertical matrix compressibility, α , in the above equations is either α_r or α_v as appropriate. This equation, minus the last term on the right-hand-side, is equivalent to equation (4), the basic flow equation solved by TOUGH2:

$$-\nabla \cdot \rho \mathbf{v} = \phi \rho \left(\beta + \frac{\alpha}{\phi} \right) \frac{\partial p}{\partial t}. \quad (12)$$

Thus, the last term on the right-hand-side of equation (11), $-\rho \alpha \frac{\partial \sigma_t}{\partial t}$, is a source term that must be included in the flow equation to account for compaction at a fixed depth. Finally, compaction, uplift and subsidence require that the depth, z , of the volume element change through time. The condition of a changing z requires that equation (11) becomes

$$-\nabla \cdot \rho \mathbf{v} = \phi \rho \left(\beta + \frac{\alpha}{\phi} \right) \left(\frac{Dp}{Dt} \right) - \rho \alpha \left(\frac{D\sigma_t}{Dt} \right), \quad (13)$$

where D indicates the total derivative (N.B. changes in the x - and y -dimensions are still considered negligible). This expression can be rewritten as

$$-\nabla \cdot \rho \mathbf{v} = \phi \rho \left(\beta + \frac{\alpha}{\phi} \right) \left[\frac{\partial p}{\partial t} + \frac{\partial p}{\partial z} \frac{\partial z}{\partial t} \right] - \rho \alpha \left[\frac{\partial \sigma_t}{\partial t} + \frac{\partial \sigma_t}{\partial z} \frac{\partial z}{\partial t} \right], \quad (14)$$

where $\frac{\partial z}{\partial t}$ is the rate of subsidence, uplift, erosion and/or sedimentation. Separating the basic flow equation from equation (14), the remaining source terms are:

$$\phi \rho \left(\beta + \frac{\alpha}{\phi} \right) \left(\frac{\partial p}{\partial z} \frac{\partial z}{\partial t} \right) - \rho \alpha \left[\frac{\partial \sigma_t}{\partial t} + \left(\frac{\partial \sigma_t}{\partial z} \frac{\partial z}{\partial t} \right) \right]. \quad (15)$$

The terms in expression (15) are the source/sink terms we added to the flow equation to account for changes in fluid mass balance due to compaction in a moving medium.

Hydrocarbon Aspects

Additional routines that we developed calculate oil generation and expulsion in situ in the model, such that hydrocarbon generation and migration histories can be constructed. Oil generation from potential source rocks is kinetically controlled by an Arrhenius-type rate expression, in which heat is the driving force:

$$\frac{-dK}{dt} = \frac{dO}{dt} = A \exp\left(\frac{-E}{RT}\right) K, \quad (16)$$

where K is the concentration of kerogen, O is the amount of oil converted from K , A is a frequency factor (s^{-1}), E is the activation energy ($kcal\ mol^{-1}$), and R is the gas constant, $1.987\ kcal\ mol^{-1}\ K^{-1}$. We explicitly coupled the algorithm of Sweeney (1990) to the thermal history in the basin evolution model. Sweeney's (1990) method is to calculate oil generation by integrating equation (16) over the thermal history. We adapted the algorithm to integrate over the time-temperature history during the model simulation, such that the effects on subsurface temperature of changing porosity, permeability, groundwater circulation, variable uplift, erosion, surface temperature, and oil generation can be explicitly considered. Following oil generation and expulsion, oil migration is also simulated in the model. The particular version of TOUGH2 modified for this study included flow of a third phase consisting of water-insoluble oil. Another version of the TOUGH2 code was recently released that includes water-soluble NAPLs and hydrocarbon gases (called M²NOTS, described in detail in Adenekan et al, 1993), but at the time of release this study was committed to the original TOUGH2.

IMPLEMENTING THE CONCEPTUAL MODEL IN THE NUMERICAL MODEL

Basin Structure and Stratigraphy

The first step is to determine the present 3-dimensional geometry of the basin, including assessing which individual formations or groups of formations are appropriate to serve as individual model layers, and obtaining or constructing structure contour maps and isopach maps of those layers. Figure 2 (structural-stratigraphic cross-section N-S) shows the breakdown of the eight individual stratigraphic layers serving as discrete model layers

for the analysis. Structure contour maps and isopach maps of the eight layers were developed by Fouch (1975) and Bredehoeft et al (1994), and the 3-dimensional finite difference grid was constructed from these maps.

Stratigraphic Projection

The second step is to complete or "fill out" the stratigraphic section by projecting the stratigraphic thicknesses of eroded section, assuming that all sediments were deposited horizontally. Figure 6 is a schematic drawing illustrating the concept of projecting formation thicknesses. Because the evidence of the surface of maximum aggradation has been destroyed by erosion, the original maximum depth and original thickness of the top unit, the Duchesne River Formation, is uncertain (for example, see layer 1 on Figure 6). Several studies have attempted to estimate the unknown amount of deposition and erosion for various parts of the Uinta basin; Table 1 summarizes these estimates for the local area of the Altamont oil field. The estimates listed in Table 1 vary a great deal because of the lack of a definitive surface of maximum aggradation and because of inherent uncertainties associated with the methods used.

An estimate of the amount of unknown deposition and subsequent erosion is necessary for the model. We took a reasonable but conservative approach, estimating the amount of strata by projecting all layers beyond their erosional surface (Figure 6), which provides an estimate of the minimum original surface prior to uplift and erosion. After projecting the Duchesne River Formation (the surface layer), we concluded that this method was reasonable because the maximum surface elevation of the projected surface layer did not exceed 3000 m, the estimated elevation of maximum aggradation (Johnson and Nuccio, 1993).

It is possible, however, that some unknown amount of additional deposition and subsequent erosion occurred beyond this minimum surface. For example, in the schematic drawing in Figure 6, the top surface (dashed) is an example of an unknown amount of additional strata that is entirely removed for which no evidence remains. During each simulation of the late uplift model, we added enough additional Duchesne River Formation at each column of the model to bring the maximum surface elevation of the projected layer to 3000 m, the estimated elevation of maximum aggradation. No additional Duchesne River Formation was added during early uplift model simulations because uplift causes erosion to occur simultaneous with deposition, and thus any additional Duchesne River Formation is removed when it is added.

It is clear that this approach is somewhat conservative because the 3000 m maximum aggradation elevation may be an underestimate. However, a conservative approach is justified because the purpose of this study is to understand the basin system better (not to reproduce the basin history exactly), and a geologically reasonable conceptualization of the basin's history is necessary. Sophisticated approaches, such as fission-track annealing (Naesar, 1979), could be used to estimate unknown amounts of deposition and erosion, but such an effort would be a separate study.

Backstripping

The third step is to estimate sedimentation rates for each layer at all points in the basin, including the projected sedimentary layers. We used a backstripping method (described in Miall, 1990) for estimating sedimentation rates at each column of the model grid. Backstripping requires use of an empirically calibrated effective stress-porosity function, described in the following data section, assumes that hydrostatic conditions were maintained through the geologic history, and neglects time-dependent effects such as overpressures. Thus, in areas of the basin where extreme overpressures (or underpressures) occur in the basin history, the structure is not reproduced exactly. However, the error in modeled formation contact elevation caused by abnormal pressures was never greater than 30 m. This error is acceptable given that the estimated accuracy of the structure contour maps, constructed by interpolation, is 50 m or less except at wells used for interpolation.

Uplift and Erosion

The fourth step is to estimate timing and rates of uplift and erosion at each column of the model grid. The amount of uplift at each column in the basin model grid is simply the difference in elevation between the beds at their maximum depth and their present structural elevation. Uplift rates are assumed to be linear and constant and equal to the amount of uplift divided by the duration of uplift, where the amount and duration are specified by the conceptual model. Erosion is addressed in a simple manner. At the time of maximum burial, all layers are assumed below the present day surface elevation in a relative coordinate system. During periods of uplift, as model strata cross the present-day surface they are stripped off or eroded. Numerically, erosion is simulated by reducing the thickness of a node as it intersects the specified erosional surface. Sweeney et al (1987) and Fouch et al (1992) estimate the time of maximum depth for parts of the basin to occur at approximately 31 to 32 Ma. Anders et al (1992) and Johnson and Nuccio (1993) estimate the time of maximum depth for parts of the basin to occur at ~37 to ~38 Ma.

Boundary Conditions

Specific boundary conditions were applied during the model simulations, including:

- (i) A uniform, constant basal heat flow was applied at the basal boundary. We assigned the basal heat flow to be 65 mW m^{-2} , the value necessary to fix the present day average surface heat flow in the northern third of the basin at 57 mW m^{-2} , the average surface heat flow value for that region estimated by Chapman et al (1984). Note that we set the basal heat flow value constant through time for the entire basin; we found no justification for varying the value in the past from the present-day value, nor did we find reason to vary basal heat flow geographically. However, oil generation and migration results are very sensitive to this parameter. If basal heat flow is assumed to be 10 mW m^{-2} less, and all other variables are held constant, the time of peak oil generation will occur as much as 5 My later, the magnitude of peak oil generation will be reduced by as much as an order of magnitude, the viscosity of oil throughout the section will be increased, and thus migration will be stifled. Likewise, if basal heat flow is assumed to be 10 mW m^{-2} greater, the time of peak oil generation will occur sooner and the magnitude increased. The value of heat flow chosen for the history of the basin also has a tremendous influence on the development of overpressures due to hydrocarbon generation. However, both occurrence

time and magnitude of peak hydrocarbon generation also depend on the hydrocarbon kinetic parameters assumed; a different set of kinetic parameters can result in much different hydrocarbon generation peak times and magnitudes for the same basal heat flow.

(ii) A specified surface temperature history (temperature versus time), discussed in a subsequent section, was used to govern the surface boundary temperature. Model simulations using a constant surface temperature (11° C) through time were run and compared to the variable surface temperature history model results.

(iii) Constant head, equal to (surface topography - 100 m), was assigned at the surface of the domain throughout the history simulation i.e. during deposition and erosion, head at the surface was assigned the current topographic surface value minus 100 m. This approach is justified because the observed value of hydraulic head in most intracontinental groundwater systems is within 100 m of the surface and is stable with time (Hubbert and Rubey, 1959).

(iv) No-flow boundaries were assigned at the sides and bottom of the model domain. We assumed that the permeability of shale layers below the Cretaceous North Horn Formation (e.g., the Mancos shale) provide a low permeability barrier that justifies a no-flow boundary for purposes of the model simulation. Because portions of the Cretaceous section crop out in most areas of the basin, no-flow boundaries were assigned to the entire base/sides of the model except at the Uinta mountain boundary, where the Uinta south flank fault also warrants a no-flow boundary.

The data used for boundary conditions and model parameterization are discussed in the following section.

DATA

An important aspect of using a numerical model to analyze a groundwater flow system is model parameterization, or the assignment of petrophysical flow properties. The petrophysical flow properties of concern include porosity, permeability, relative permeability, and capillary pressure. We collected 196 individual plug samples from Uinta basin core for the purpose of measuring permeability and porosity. The bulk mineralogy of the samples was also assessed for the purpose of analyzing the relationship between mineralogy and permeability. Described below is how we used these data to govern permeability and porosity development in the forward model, estimate a compaction trend, and backstrip the sedimentary section. We sampled core only from the Altamont/Bluebell oil fields, believing this area underwent a minimal amount of uplift and erosion and thus that the porosity-depth data best represent the actual compaction trend. Table 2 lists the name, location, and number of samples of each well.

The data reported here comprise the first substantial, public domain suite of core-plug measurements of permeability, porosity and bulk mineralogy for northern Uinta basin strata. The 196 samples cover a large portion of the Tertiary Green River Formation, from above the Middle Marker to the deepest Flagstaff member (see cross-section, Figure 2). Virtually no core-plug petrophysical data for the Altamont/Bluebell area were available prior to this study; Pitman et al (1982) report petrophysical data for some relatively shallow samples south of the Bluebell field.

Bulk Mineralogy Data by XRD Analysis

We measured bulk mineralogy to help interpret permeability trends; specifically, to analyze the relationship between permeability and its two dominant controlling variables -- porosity and mineral composition (especially clay content). X-ray diffraction methods are used routinely to identify detrital and authigenic minerals in sedimentary rocks. For our study, measurements were made on a Rigaku X-ray diffractometer, using the full range of intersection angles for initial reconnaissance of the mineral assemblage of the 196 samples (for a general description of the use of XRD methods for mineralogy, and associated terminology, see Moore and Reynolds, 1989). After assessing what principal minerals we wished to identify, we narrowed the ranges of XRD intersection angle specifically to evaluate the presence and relative abundance of quartz, feldspar, calcite, dolomite, ankerite and clay minerals. For example, most clay minerals produce a significant diffraction intensity peak at the intersection angle of approximately 19.8° . This diagnostic peak is referred to as the composite clay mineral peak (Parry, personal communication, 1995). It usually occurs very close to intersection angle 19.8° , but may be off by as much as $\pm 0.06^\circ$. To evaluate the samples for presence and abundance of clay, we looked for an intensity peak at $\sim 19.8^\circ$ by scanning each sample from intersection angle 19° to 21° .

The diffraction patterns for each sample provided information about the presence or absence of specific minerals. In general, quartz is common to all 196 samples, and it is the major component of nearly all samples. Petroleum companies preferentially core reservoir strata; the samples reflect this bias. Feldspar, calcite, dolomite, ankerite and clay minerals are also common components, but with widely varying abundances. We found no systematic changes in mineral abundances with depth, except a general, very slight decrease in calcite content with increasing depth.

Although X-ray diffraction patterns provide an easy method for determining the presence or absence of specific minerals, quantitative determination of relative mineral abundances is a more difficult task. The difficulty in estimating relative mineral abundance occurs because the intensity of any given peak is influenced not only by its own abundance, but also by the absolute abundance, crystallinity, and orientation of all other minerals in the specimen (Fisher and Underwood, 1995). In other words, many minerals have diagnostic peaks that are affected by the presence of other minerals. Depending on the relative abundance of other interfering minerals, a mineral's diagnostic peak may shift in position (angle), intensity, or peak width. If minerals' diffraction patterns did not interfere with each other, then the intensity of a mineral's diagnostic peak would be proportional to the mineral's abundance. However, interference among minerals requires that peak intensity weighting factors, determined from mixes of mineral standards, must be applied. Cook et al (1975) derived peak intensity weighting factors from 50:50 mixtures of two mineral standards. However, this approach is flawed because the error associated with the single weighting factor for each mineral pair increases as the mixture deviates from the ideal 50:50 blend (Fisher and Underwood, 1995). Single best fit weighting factors can be determined for each individual mineral in a known mix using laboratory blends of mineral standards (Underwood et al, 1993). Weighting factors for every mineral pair in a multicomponent system also can be determined empirically (Moore, 1968). The accuracy of this method is improved if each mineral pair is measured over its entire range of possible mixtures (Fisher and Underwood, 1995).

We adapted the method of Moore (1968), but calculated weighting factors only for specific minerals of interest that we know to exist from thin section analyses of our core samples (Morris, personal communication, 1995), and only factors for each mineral compared to quartz. In other words, instead of calculating weighting factors for every possible combination of mineral pairs, we calculated weighting factors for the following pairs of minerals: (1) calcite-quartz, (2) dolomite-quartz, (3) potassium feldspar-quartz, (4) plagioclase-quartz, and (5) clay-quartz (using composite clay-peak of montmorillonite standard). We chose these pairs because we were primarily interested in the relative abundances of quartz, feldspar, clay, calcite and dolomite, and their effects on porosity and permeability. Additionally, we assume that quartz is the most significant interference mineral for each and neglect interference by minerals other than quartz.

Figure 7, a plot of calcite/quartz abundance versus peak intensity, is an example of our weighting-factor determination. We measured the calcite peak intensity at the intersection angle (2θ) of 29.4° for 6 blends of calcite(%):quartz(%): 5:85, 11:89, 16:84, 20:80, 24:76, 30:70, 100:0. A regression fit of these data yields a weighting factor of an exponential form:

$$C = 3 \exp(I \times 3.7 \times 10^{-4}) \quad (17)$$

where C is the abundance of calcite (%*) corrected for the presence of quartz and I is the raw intensity of the calcite peak (counts per second). Corrected abundance is expressed as %* instead of % to emphasize the explicit caveat that the results are semiquantitative relative abundance, not absolute abundance. Correcting the raw X-ray diffraction data with empirical weighting factors provided semiquantitative estimates of relative abundance of quartz, clay, calcite and dolomite in each sample. The effects of these minerals on porosity and permeability are discussed below.

Detectable Microfractures

The extreme reduction in pressure from in situ to laboratory conditions commonly either creates microfractures or opens existing microfractures in plug samples taken from depth. Porosity and permeability measurements of such samples are too high to be representative of in situ conditions, unless -- as was not done in this study -- the measurements are made at in situ pressures. Therefore, we qualitatively assessed the presence of microfractures by spraying each sample with a volatile, quick-drying solvent. The solvent on the surface of a sample would dry first, and any solvent that seeped into microfractures that intersect the surface would dry last, leaving the fractures visible to the naked eye for a few seconds. We noted whether microfractures were present in each sample, and tried to rank qualitatively the intensity of the fractures. Of the 196 samples, 73 contained detectable microfractures. We discuss the effects of microfractures on porosity and permeability in the following sections.

Porosity Data

We measured the intergranular, connected porosity of every sample with a very simple procedure:

- 1) evacuate the samples of all pore water from connected pore-space by storing the samples in a bell jar connected to a vacuum pump (keep under vacuum for at least 24 hours)
- 2) measure the "dry" mass of each sample
- 3) evacuate in vacuum bell jar again (to rid samples of any moisture acquired during mass measurement and to prepare for next step)
- 4) while under vacuum, flood samples with water to fill connected pore space
- 5) measure the "wet" mass of each sample
- 6) measure bulk volume of each sample by immersing in graduated cylinder with water; the volume of water displaced by the sample indicates the bulk volume of the sample

The difference between the "wet" and "dry" mass is equal to the volume of water in the connected pore space; this volume divided by the bulk volume of the plug sample is equal to the connected porosity. The dry mass divided by the matrix volume (bulk volume minus pore volume) of a plug sample is equal to the grain density of the sample.

Thin sections (Morris, personal communication, 1994) of the sampled core show that some core plugs contained detectable amounts of "dead oil." We did not treat these samples to remove this oil prior to porosity determination, and presence of this oil would cause some porosity underestimation for these samples. We consider this bias to be present but minor in the overall data population.

Measured values of intergranular, connected porosity are plotted against depth in Figure 5. Distinguishing between microfractured (open diamonds) and nonfractured samples (solid circles) illustrates that microfracture porosity is significant. Porosity varies from less than 0.02 % to about 10 %, and averages 2.1%. Porosity exhibits a weak inverse correlation with clay content, but does not seem to correlate with abundance of other minerals. Augmenting this data set (Figure 5) are porosity data for the Pariette Bench area from Pitman et al (1982; open squares).

Many authors suggest that compaction trends are exponential in nature (e.g., Sclater and Christie, 1980), with the equation describing a trend being of the form

$$\phi = \phi_0 * e^{-z/h}, \quad (18)$$

where ϕ_0 is the porosity at the top or surface of the sediment column, and h is the depth at which porosity becomes ϕ_0 / e . We estimated an exponential compaction trend by fitting a regression line (solid line in Figure 5). Only the unfractured samples were used to estimate the compaction trend because fracture porosity is significant. Compaction is expressed by equation (18) with $\phi_0=0.4$ and $h=1250$ m, or

$$\phi = 0.4 * \exp(-z/1250.) \quad (19)$$

where ϕ is fractional porosity and z is depth in m. The value of h , 1250 m, is lower than most values estimated in the literature for sandstone/shale sections, which suggests that some additional deposition and subsequent erosion was possible. Rather than try to align the compaction trend with trends published for other sedimentary sections, we used this fitted expression to backstrip the section. As the modeling results will show, this approach is acceptable for our purposes of trying to understand the system. Note that this curve is identical to the "virgin loading curve" described previously, and as such is used to assign primary porosity in the basin evolution model.

Permeability Data

Matrix permeability of the 196 samples was measured by mini-permeameter; measurements were contracted to a local laboratory (Terra Tek, Inc.) specializing in permeability measurements. Distinguishing between natural microfractures and those caused or enlarged by coring and associated pressure decrease is extremely difficult; the permeability data for samples with observed microfractures are not plotted or used in the analysis. Permeabilities of the remaining 123 apparently unfractured samples are plotted against their associated porosities in Figure 8. Although sample bias is towards sandstone, correlating bulk mineralogy to porosity revealed that neither porosity nor permeability depends strongly on relative proportions of quartz, feldspar, calcite or dolomite. In contrast, the presence of even a small proportion of clay (e.g., 3 to 5% or greater) reduces porosity somewhat and reduces permeability very significantly: solid circles on Figure 8 represent samples with little or no clay and solid triangles represent permeability of samples with significant (> 5%) clay. Other minerals do not seem to affect permeability significantly. Permeability data from Pitman, et al (1982; open squares on Figure 8) were included to augment the data-set. The solid line is a least-squares regression fit (with exponential-form equation) through the non-clay data and the data of Pitman et al (1982). The equation for the function is

$$k = 10^{-18} * \phi^{(1.5 + 0.6 * \ln \phi)} \quad (20)$$

where k is matrix permeability in m^2 and ϕ is fractional porosity. This function was used to assign permeability in the basin evolution model.

Other properties or boundary conditions of the basin evolution model that need to be considered are fractures and fracture generation, sedimentary facies, hydrocarbon data such as total organic content, generation kinetic parameters, oil density and viscosity, and thermal data such as thermal conductivity, surface temperature, and basal heat flux.

Fractures and Fracture Generation in the Model

Throughout the Altamont/Bluebell fields are vertical to nearly vertical, mineralized, open fractures that enhance oil production (Lucas and Drexler, 1976). Extensive subsurface systems of fractures are thought to exist, and significant regional systems of fractures are evident at the surface. The most prominent fracture systems (order of tens of km long) observable at the surface are: (1) the east-west trending Duchesne fault and

fracture zone on the south-central flank of the basin parallel to the Uinta Mountains, and (2) a series of gilsonite veins in the southeastern Uinta basin striking west-northwest roughly parallel to the Uncompahgre uplift (Chidsey and Laine, 1992). Surface fracture systems are thought to be caused by changes in the regional tectonic stress regime (Narr and Currie, 1982), whereas subsurface fracture systems local to the Altamont/Bluebell fields are thought to be hydraulic fractures caused by overpressures (Spencer, 1987; Sweeney et al, 1987). This general knowledge of the character and distribution of fractures is helpful for interpreting the hydrodynamic regime because of the effects of fractures on permeability.

Permeability is a scale-dependent parameter. The permeability data used in this study were measured of core samples and are explicitly valid only for the core scale. Well-scale to regional-scale permeabilities -- measured or inferred -- can be very much higher than core-scale permeabilities measured of the same strata (McPherson and Chapman, 1996; Deming, 1993; Willett and Chapman, 1987; Neuzil, 1986). In the Uinta basin specifically, Bredehoeft et al (1994) published 78 well-scale permeabilities inferred from drill stem test analyses that cover much of the depth section, and at any given depth these data span two to six orders of magnitude. The variability is attributed to fractures which the core scale measurements do not sample. It is not appropriate to assume that the core-scale permeabilities sufficiently describe the well-scale and basin-scale permeabilities. We addressed this problem by inducing fractures during the model simulation: if the fluid pressure in a cell exceeds the least principal stress, the vertical and horizontal permeability of that cell is increased by a specified amount. We arbitrarily assigned a single order of magnitude increase of permeability for a single fracture; more than one order of magnitude permeability increase will occur if a single cell fractures more than once. In most model simulations, lateral fracture networks of several km develop and thus the regional-scale permeability is increased. No fracture annealing (e.g., calcite precipitation within fractures) is built into the model. However, permeability of fractured cells continue to change with porosity at the same rate as unfractured cells, but at an order of magnitude higher value (see dashed line on Figure 8).

Sedimentary Facies

The facies architecture of the Uinta basin is detailed in Fouch (1975), Ryder et al (1976), and Fouch et al (1992). We digitized the facies maps of Fouch et al (1992) and assigned each cell in the 3-dimensional model grid a facies designation. These facies designations determined initial total organic carbon (TOC), hydrocarbon kinetic parameters, relative permeability functions, and capillary pressure functions, as described in subsequent sections. Porosity and permeability were not assigned based on facies designation. One porosity/effective stress function (Figure 5) and one permeability/porosity function (Figure 8) were used for all strata; the data do not indicate significant differences between facies.

Relative Permeability

The relative permeability of an individual fluid phase is the effective permeability expressed as a fraction of the absolute or intrinsic permeability of that phase, or

$$k_r = k_e / k \quad (21)$$

where k_r is relative permeability, k_e is the effective permeability, and k is the absolute permeability. Effective permeability is always lower or equal to absolute permeability because different fluid phases tend to interfere with one another as they flow.

Some workers that specifically examine relative permeability issues are Brooks and Corey (1964), Van Genuchten (1980), Demond and Roberts (1987), Parker et al (1987), and Okui and Waples (1993). In this study, we used a formulation published by Parker et al (1987) in which relative permeability is expressed in terms of capillary pressure curve parameters derived by Van Genuchten (1980). Figure 9 shows the relative permeability curves we applied to the different facies in the model. We used different formulations for different facies because the relative permeability relationships between hydrocarbon fluids and water in typical reservoir strata are different from those for typical source strata such as oil shale (Okui and Waples, 1993). Relative permeability curves for sandstones and other typical reservoir strata are usually measured experimentally; numerous "typical" relative permeability curves for sandstone are published in the literature. However, experimental relative permeability data for shales and other low permeability source rocks are not available. Burrus et al (1992), Okui and Waples (1993), and Wendebourg (1994) proposed different relative permeability functions for source rocks (than those for reservoir strata) to improve hydrocarbon expulsion aspects in their modeling studies.

Capillary Pressures

Oil in the subsurface is acted on by many forces: buoyancy causes upward movement, hydrodynamic drive and overpressures cause migration away from the source, and capillary pressures cause resistance to migration. Oil drops immersed in water try to assume the smallest surface area possible, ideally a sphere (Tissot and Welte, 1984). If a pore throat is smaller than the droplet of oil, the interfacial tension between the oil and water resists the deformation necessary to squeeze the droplet through the pore throat. This interfacial tension is capillary pressure. Capillary pressures that develop between two immiscible fluid phases, e.g., oil and water, that come in contact in a porous medium are proportional to the interfacial surface tension between the two fluids (Berg, 1975):

$$p_c = p_w - p_o = 2\lambda/r \quad (22)$$

where p_c is capillary pressure, p_w is water pressure, p_o is oil pressure, λ is surface tension, and r is the radius of the capillary or pore throat. Some papers that discuss capillary pressure issues are Brooks and Corey (1964), Berg (1975), Van Genuchten (1980), and Parker et al (1987). We adapted the formulation of Parker et al (1987) for this study.

The capillary pressure function we assigned to alluvial facies, the lower curve on Figure 10, is the Parker et al (1987) capillary pressure formulation with experimental parameters for sandstone published by Essaid et al (1993). Burrus et al (1992) estimated the capillary pressures for oil-water interfaces in source shales with permeability $\sim 10^{-21} \text{ m}^2$ to be ~ 100 bars (10 MPa). Thus, the capillary pressure formulation we used for lacustrine facies is the identical alluvial facies function, but scaled to a maximum of 10 MPa (top curve on Figure 10).

Hydrocarbon Generation Data

The Green River Formation is a particularly rich hydrocarbon source, due to its lacustrine origin. Green River Formation open lacustrine facies are primarily type I kerogen. Type I kerogen is lipid- and hydrogen-rich, almost always occurs in lacustrine environments, and primarily generates oil during catagenesis (Fouch et al, 1992; Tissot et al, 1978). The Green River Formation of the Uinta basin is commonly cited as a typical source rock of Type I kerogen produced in a lacustrine environment. The marginal lacustrine facies rocks contain types I, II and III kerogen. Type II kerogen also primarily generates oil, whereas type III kerogens generate mainly methane gas. The alluvial facies contains minor amounts of type III kerogen. In this study we neglect the hydrocarbon potential of alluvial strata. Source rocks are prevalent throughout the open and marginal lacustrine facies, and are as much as 4500 m thick at the depositional center of the basin (Anders and Gerrild, 1984). The amount of oil generated from these source rocks depends primarily on the type of organic matter, the initial total organic content (TOC), and the thermal history.

The source rock considered in this study is termed "Green River oil shale" or "black shale" (Ruble, written communication, 1994; Sweeney et al, 1987; Tissot et al, 1978), a bituminous-rich carbonate mudstone that occurs in the open and marginal lacustrine facies. We considered only Tertiary Green River Formation source rocks, and only type I kerogen. Types II and III kerogens of the marginal lacustrine facies were neglected. Different sets of kinetic parameters for the type I black shale have been determined by Campbell et al (1978), Burnham and Singleton (1983), Sweeney et al (1987), and most recently Ruble (written communication, 1995). All sets of activation energies and frequency factors (see Sweeney, 1990, for an explanation of these terms and the associated Arrhenius formulations) are similar and produce consistent results. We used the most recent set by Ruble (1995, written communication), which consists of a frequency factor and a distribution of activation energies (Table 3).

The initial TOC distribution is among the least constrained variables. Anders and Gerrild (1984) analyzed over 90 samples from among six wells in the northern Uinta basin. Their analyses revealed that thermally immature samples above the Mahogany oil shale bed (Figure 1) have TOC values averaging 6% but ranging as high as 21 %. Below the middle marker, mature lacustrine facies average about 1.6%. Anders and Gerrild (1984) state that the amount of oil that can be generated from their immature kerogen samples is 44% of the total organic carbon, if only kerogen is considered. Anders and Gerrild (1984) also found that open lacustrine source rocks are generally twice as rich as marginal lacustrine source rocks. On the basis of these results, we assigned all open lacustrine rocks an initial TOC of 1.6 mass %, and all marginal lacustrine rocks half that much. Further, only 44 mass % of the TOC is convertible material.

This parameterization of initial TOC is a gross simplification. TOC values vary a great deal among samples collected: Ruble (written communication, 1995) recently collected and assayed several samples with as much as 55 mass % TOC, some samples discussed in Fouch et al (1992) are as much as 60% TOC, and some exposed upper Paleocene rocks in the southwest part of the basin contain algal coals and kerogenous limestone with TOC values as high as 75% (Fouch and Hanley, 1977). Little is known about the actual distribution of rocks with sufficient TOC to generate significant oil. On

one hand, by assigning all open lacustrine facies rocks with sufficient TOC to generate significant oil, it is possible that we overestimated the distribution of possible source rocks and the potential amount of oil generated. On the other hand, most major oil accumulations of the world originated in source rocks with TOC > 2.5 mass % (Palciauskas, 1991); thus because the assigned TOC (1.6 mass % for open lacustrine) is relatively low, it is possible that we underpredicted the potential amount of oil generated. Although the TOC parameterization is simplified, we assume that it is reasonable.

Density and Viscosity of Oil

Oil density has a great effect on migration patterns. In general, the difference in density between oil and water manifests itself through buoyancy effects; buoyancy tends to move oil upward because it is less dense than water. Similarly, Athy (1930) introduced the concept that large bodies of oil may move upward due to buoyancy, but higher surface energies per unit volume cause resistance to flow in smaller bodies of oil. Layers of lower permeability and capillary pressures are barriers to this upward movement.

The density of oil at the time it is generated is also important because the volume increase associated with conversion of solid kerogen to liquid oil influences overpressure development. Following oil generation, oil density will change during basin evolution due to changes in chemical composition, caused in turn by changes in temperature and pressure, biodegradation, and depletion of saturated hydrocarbons (Tissot and Welte, 1984).

Figure 11 is a plot of oil density vs. depth for 1070 Uinta basin wells. Assuming that all API gravity data were measured at conditions of atmospheric pressure and 60° F (15.5° C), we translated API gravity to SI units using the following formula (Tissot and Welte, 1984):

$$\delta = 141.5/(API + 131.5), \quad (23)$$

where δ is density in kg m^{-3} (at 60° F and atmospheric pressure) and API is API gravity. The plotted data are corrected for in situ conditions of temperature and pressure using the corrections of Standing (1952). Data scatter are due to differences in temperature, pressure, chemical composition and dissolved hydrocarbon gases. The solid line through the data is the relationship used to assign the density of initially generated oil. The line was fit by least squares regression based on a second-order polynomial, and is mathematically expressed as

$$\delta = 967 - 0.078 * z + 9.6 * 10^{-6} * z^2. \quad (24)$$

We made an effort to describe initial oil density based on observed data, but were unsuccessful. This initial density formulation is a compromise because it does not account for changes in density caused by changes in paleotemperatures and paleopressures. Furthermore, it is impossible to reconcile the effects of past temperature and pressure because of data scatter. Plotting the data in map view (by formation) provides no better basis for assigning initial density.

Once kerogen is converted to liquid oil and assigned an initial density, the following simple relationships were used to dictate changes in density (via changes in volume) due to changing pressure and temperature (Eremenko, 1991), respectively:

$$\Delta V = -C(V_o)\Delta p + \alpha(V_o)\Delta T \quad (25)$$

where ΔV is the change in oil volume, V_o is the initial volume, Δp is the change in pressure, and ΔT is the change in temperature. C is crude oil compressibility, and in the model is assigned a value of 10^{-9} Pa^{-1} . Note that for oils containing high amounts of dissolved gas, C approaches $1.5 * 10^{-8} \text{ Pa}^{-1}$. α is the coefficient of thermal expansion for crude oil, and in the model it is assigned a value of $0.001 \text{ }^\circ\text{C}^{-1}$. For high amounts of dissolved gas, α approaches $0.0025 \text{ }^\circ\text{C}^{-1}$.

The model does not include gas generation. Although Altamont/Bluebell are dominantly oil fields, some gas has been produced from the area and many gas fields exist around the southern half of the basin. Spencer (1987) states that the underlying Cretaceous at Altamont/Bluebell may be gas-bearing, although data supporting this conclusion are limited.

Oil viscosity will also change during basin evolution due to changes in temperature and pressure, biodegradation, and depletion of saturated hydrocarbons (Tissot and Welte, 1984). In general, crude oil viscosity decreases with increasing temperature and increases with increasing pressure, if the only effect of the pressure is compression of the liquid. Beal (1946), Standing (1952), and Chew and Connally (1959) provide detailed discussions of hydrocarbon fluid viscosity and related issues.

We obtained several assays for crude oils produced from Altamont (Frank Mueller, Phillips Petroleum Company, written communication, 1994) which listed crude oil viscosities ranging from 0.6 cP to 6 cP. Not having enough data to constrain oil viscosity for model simulations, we used the correlations of Beal (1946) to assign oil viscosity corresponding to oil density at in situ temperature and pressure. All model simulations were conducted with the assumption of no dissolved gas; however, dissolved gas can lower oil viscosity significantly.

Thermal Data

Additional data needed to parameterize the thermal properties and boundary conditions through time are the thermal conductivity distribution, the background heat flow, and the estimated surface temperature history.

Maps of lithologic thermal conductivity were constructed by Chapman et al (1984) for the Uinta, Green River and Wasatch Formations, based on facies maps, stratigraphy, and 852 thermal conductivity measurements. We digitized their maps and assigned each cell in each model layer accordingly. The maps display the matrix thermal conductivity, the thermal conductivity of the solid component only, measured at laboratory conditions (20°C). A correction is applied for in situ temperature (Touloukian et al, 1970),

$$k_t = k_s * (1 / (1 + 0.0005 * T)) \quad (26)$$

where k_s is uncorrected matrix conductivity, T is in situ temperature, and k_t is the corrected thermal conductivity. A correction is also applied for in situ porosity (Robertson and Peck, 1974),

$$k_r = k_w^\phi k_t^{(1-\phi)}, \quad (27)$$

where ϕ is porosity, and k_w is the thermal conductivity of pore water (Touloukian et al, 1970), or

$$k_w = a + bT + cT^2 \quad (28)$$

(where, $a = .564842$, $b = .00187829$, and $c = -.000007231084$ if $T < 137^\circ \text{C}$, otherwise $a = .60200489$, $b = .00130975$, and $c = -.00000514$).

We assigned the background heat flow for the entire basin equal to 57 mW m^{-2} , the average surface heat flow of 94 values determined by Chapman et al (1984). Sweeney et al (1987) and Fouch et al (1992) both assume that the average background heat flow has been constant through Tertiary time to the present. We followed this assumption and assigned the background heat flow constant through time in all model simulations. Surface temperature, however, was varied through time.

Surface temperature is often treated as a constant in thermal history models. Most papers involving thermal models of the Uinta basin do not describe how surface temperature was specified. Heat flow and thermal maturity are very sensitive to surface temperature, and the surface temperature history must be constructed a priori.

Physiognomic data from fossil plants make it possible to reconstruct paleoclimates. Tanai and Huzioka (1967), Wolfe and Hopkins (1967), Addicott (1969), Devereux (1967) and Wolfe (1978) all document a significant warming episode during the Miocene that apparently affected a great proportion of the earth. Other major events in the Tertiary climatic history include Miocene Alpine glaciation in Alaska (Denton and Armstrong, 1969), and a rather large, rapid (1 to 2 My) reduction in mean global surface temperature that occurred at the end of the Eocene (Wolfe, 1978). The mean annual surface temperature history selected for the Uinta basin, shown in Figure 12, is adapted from a surface temperature history determined for the Pacific Northwest (Wolfe, 1978, Figure 1). Because the present mean annual surface temperature of the Uinta basin (NOAA, 1989) is approximately 3°C lower than the present mean annual surface temperature of the Pacific Northwest region (Wolfe, 1978), we applied a uniform reduction of 3°C to the Wolfe (1978) temperature history before assigning it as the Uinta basin's surface temperature history.

ANALYSIS AND RESULTS

Present-Day Surface Heat Flow

Heat flux in every cell is calculated by multiplying the temperature gradient across the cell by its thermal conductivity. Surface heat flow in the model, mapped in Figure 13, is the value of heat flux determined for the uppermost cell in each column of the model grid. The model predicts surface heat flow ranging between 45 and 70 mW m^{-2} , with an average of 65 mW m^{-2} for the entire basin. The average surface heat flow in the northern third of the basin is 57 mW m^{-2} , a condition we tried to meet as discussed in the previous boundary conditions section. Surface heat flow in the southern two thirds of the basin is surprisingly homogeneous, ranging only between 65 and 70 mW m^{-2} .

A cross-sectional display provides insight for interpreting the surface heat flow distribution. Figure 14 plots cross-section N-S (Figures 1, 2) with corresponding model fluid flow vectors and surface heat flow (indicated by solid line plotted above the cross-section). Depressed surface heat flow in the northern part of the basin corresponds to groundwater recharge, indicated by the direction and magnitude of fluid flow vectors. Higher topography at the northern periphery drives groundwater deep in the basin, where it picks up heat by advection, and the heated groundwater subsequently flows toward the center. Thus, model surface heat flow reaches its highest value in the elevation trough at the center of the basin. These model results are consistent with the work of Willett and Chapman (1987).

Probably the most interesting aspect of the surface heat flow distribution is the continued trend of high surface heat flow on the southern side of the basin. Fluid flux vectors on Figure 14 suggest that the groundwater recharge rates are highest on the southern edge of the basin; thus, corresponding depressed surface heat flow in the southern portion would seem to be the logical result. However, surface heat flow results do not reflect advective depression by recharge along the southern periphery. An explanation of this result is found by examining the conductive aspects of the thermal regime.

To isolate the conductive aspects of the surface heat flow distribution, we ran a simulation with the permeability of all strata limited to 10^{-21} m^2 . Surface heat flow results for this model varied from 63 to 70 mW m^{-2} with variations caused only by the heterogeneous thermal conductivity distribution. Also plotted above cross-section N-S on Figure 14 is the conduction-only surface heat flow profile (dashed line). Comparison of the advection plus conduction heat flow profile (solid line) to the conduction-only profile suggests that heat flow variations due to thermal conductivity are local in nature (because conductive heat flow is primarily vertical). Advective effects, however, are manifested on a larger scale. Comparing the two profiles also shows that cooling by advection dominates the northern third of the basin (Figures 13 and 14), whereas the southern half of the basin is clearly dominated by conductive heat flow: the two heat flow profiles track each other within 2 to 3 mW m^{-2} in the southern half. The reason why advection is minor in the southern half of the basin is that the permeability of sediments close to the surface is relatively low. Recall that these surficial sediments were buried deep early in the basin's history and subsequently uplifted and denuded. During the burial phase of the history, these sediments became compacted and permeability was irreversibly reduced. This conclusion is corroborated by the magnitude of the flow vectors close to the surface along the southern edge of the basin (Figure 14). Recharge rates are fairly high, as illustrated by the large magnitude vertical flow vectors at the surface, but immediately below the surface the flow vector magnitudes are nominal, indicating minor flow rates. In comparison, flow vector magnitudes as much as 3 km deep in the northern half of the basin are relatively high (Figure 14), providing an explanation for depressed heat flow by advection.

Comparison of Surface Heat Flow Model Results to Previous Studies

Although the amount of actual heat flow data is limited, we compared model heat flow results to what was available. Heat flow data of Chapman et al (1984; Figure 10 of that paper) are superimposed on Figure 13 for comparison. The heat flow data of

Chapman et al (1984) consist of 97 values, all located in the northern third of the basin. All values were estimated using Bullard's (1939) thermal resistance method with BHT (bottom hole temperature) data and measured thermal conductivity data. The estimated heat flow data of Chapman et al (1984) range from 39 mW m^{-2} to 72 mW m^{-2} , with an average of 57 mW m^{-2} . Chapman et al (1984; page 5) state that they "attach no significance to short-range fluctuations of order $\pm 5 \text{ mW m}^{-2}$, but believe the smooth variations are real." Based on this assumption, we arbitrarily grouped their data (by location) and the mean of heat flow in each grouping was computed and plotted (at the average locations of the individual groups) on Figure 13. Also plotted on Figure 13 are two heat flow values measured using more conventional techniques, shown in brackets, as determined by Sass et al (1971; southernmost value) and Reiter et al (1979; Red Wash location). The distribution of model surface heat flow results is remarkably consistent with the averaged observations: lowest heat flow is along the northern periphery of the basin, and the highest gradient for this decrease is near Altamont. At most specific locations of the averaged heat flow observations, model surface heat flow values fall within 3 mW m^{-2} of the observed average, including the conventional measurements (bracketed values on Figure 13). In each of these cases, the discrepancy is less than the standard deviation associated with the average observed heat flow. The only exception is the southernmost average value of 59 mW m^{-2} (standard deviation $\pm 7 \text{ mW m}^{-2}$) which is 9 mW m^{-2} lower than the model result at that location. Comparison of the actual values of estimated heat flow (i.e., not the averaged values) to the model results is less satisfactory: the five circles on the heat flow profile plotted above the cross-section in Figure 14 correspond to the estimated heat flow values (Chapman et al, 1984) coincident with profile N-S. Only one of the values falls on the modeled surface heat flow profile, but the trend of decreasing heat flow values to the north is consistent. The model parameterization could have been adjusted to match heat flow model results to the observations. However, the aim of the study was not to replicate the system, but rather to improve understanding of it.

Chapman et al (1984) interpreted their data as showing a trend of decreasing heat flow from the center of the basin to the Uinta Mountains, and they discussed two possible mechanisms that could cause the trend: (1) lateral refraction of heat into the Precambrian quartzite (very high thermal conductivity) which comprises much of the Uinta mountains, or (2) advective heat transport accompanying groundwater circulation. Willett and Chapman (1987) tested these hypotheses by constructing a 2-dimensional finite element model of a cross-section within the northern third of the basin. Their model, which was deliberately very simple in design, suggested that most lateral variation in the thermal field could be caused by a regional, topographically driven groundwater flow system. Our model results agree with their suggestion.

In summary, comparison of model surface heat flow results with observed heat flow data (Figures 13 and 14) suggests that advective heat transport associated with local recharge/discharge and regional groundwater flow sufficiently explain large variations in the thermal field, at least to the extent observed in the data of Chapman et al (1984). Advective cooling associated with groundwater recharge is predominant on the northern side of the basin. A somewhat unexpected result is the primarily conductive surface heat flow distribution on the southern side of the basin, interpreted to be caused by lack of

advection within the compacted, low permeability sedimentary section close to the surface on that side of the basin.

Subsurface Temperature History

Probably the most important factor governing the subsurface temperature history is the structural/depth history. Figure 15 is the burial history for strata at the northernmost edge of the Altamont oil field (location indicated by solid triangle on Figure 1), corresponding to our late uplift model; for the sake of discussion, we will focus on the Altamont area because it is the site of highest observed overpressures. Recall that all deposition of Tertiary strata occurs from the beginning of the model simulation at 56 until 32 Ma, followed by a period of no deposition or erosion until 10 Ma, with all uplift and erosion within the basin occurring from 10 My until present day. The temperature history of the northernmost Altamont strata associated with this burial history is shown in Figure 16. Temperature at each horizon increases as depth increases, and when deposition ceases at 32 Ma, temperature continues to increase at a slower rate due to conductive equilibration of the sedimentary column; for a detailed discussion of this type of temperature equilibration following deposition, see Chapman and Furlong (1992). As uplift and erosion of the sedimentary section occurs during the last 10 My of the history, temperature decreases as depth decreases. Secondary effects of thermal conductivity structure, erosion, sedimentation, and advection are explicitly treated in the model, but the effects of each are not always easy to distinguish from each other in temperature space.

The depth and temperature histories of the northernmost Altamont strata associated with the alternative structural history, the "early uplift" model, are shown in Figures 17 and 18, respectively. Recall that the main difference with this model is that all differential uplift of the basin occurs from the early Tertiary, i.e., onset of the model simulation at 56 Ma, until deposition ceases approximately 32 Ma (Figure 17). Two primary differences between the contrasting temperature histories are illustrated by comparing Figures 18 and 16. First, the maximum temperature reached at depth in the early uplift model (Figure 18) is 8° C lower than that of the late uplift model (Figure 16) because the maximum depth reached is shallower (compare Figures 15 and 17). This 8° disparity is consistent with the observed thermal regime of the area: given that the difference between the maximum depth reached in the two models is 352 m (0.35 km), the corresponding average temperature gradient in the modeled sedimentary column is $(8^{\circ})/(0.35 \text{ km})$, or $23^{\circ} \text{C km}^{-1}$, which is close to the observed average of $25^{\circ} \text{C km}^{-1}$ (Chapman et al, 1984). The second major difference between the temperature histories associated with the alternative structural histories is that during the last 10 my of the basin's history, temperatures in the late uplift model decrease due to depth decrease. In contrast, no such temperature reduction occurs in the early uplift model because depth does not change.

Our model simulations also suggest that surface temperature is a critical boundary condition. The model simulations that produced the two temperature histories shown in Figures 16 and 18 were both executed with a constant surface temperature of 11° C throughout the basin's history to isolate the effects of the structural history on subsurface temperature. Note that 11° C is the current average surface temperature of the region

today (NOAA, 1989). Figure 19 is a plot of the subsurface temperature histories corresponding to the early uplift model (see Figure 17 for the depth history), but with the surface temperature boundary condition specified by the variable temperature history shown in Figure 12 (adapted from Wolfe, 1978). The greatest effects of the variable surface temperature history on subsurface temperatures occur after 35 Ma, where disparities as much as 8° C exist between the contrasting temperature histories. The differences occur because a greater surface temperature causes the subsurface temperature to be greater. Subsurface temperature corresponding to the variable temperature history is always greater than subsurface temperatures in the constant surface temperature model, simply because the lowest temperature in the variable surface temperature history is 11° C.

We paid special attention to the subsurface temperature history because it is the most important factor in determining the timing, rates and amounts of hydrocarbon generation. The timing of hydrocarbon generation during the basin's structural history may dictate hydrocarbon migration patterns. More importantly, the timing and amounts of hydrocarbon generation may influence overpressure development, and thus the hydrodynamic history, as the following section details.

Overpressures

Overpressures Caused by Oil Generation

Among the most interesting aspects of the Uinta basin are the significant overpressures observed in the northern portion of the basin. Mapped in Figure 20 is an isometric projection of observed hydraulic head at the stratigraphic equivalent to the Flagstaff member of the Green River Formation, illustrating the extremely high fluid pressures in the Altamont area compared to the surrounding ambient head map. Plotted on Figure 21 is a pressure-depth profile in the Altamont oil field. Maximum observed pressures in this particular area of the field are 65 MPa, or 27.5 MPa overpressure ("overpressure" is explicitly defined as the amount of pressure above hydrostatic fluid pressure), at ~4 km depth (~ -1.5 km below sea level), equivalent to ~5.1 km hydraulic head. The maximum observed fluid pressure in the Altamont area, as well as the entire basin, is equivalent to 6050 m head, about 3 km higher than the maximum topography (~3 km). Highest anomalous pressures are observed within the Green River Formation and are thought to be associated with oil generation (Spencer, 1987). Fluid pressures decrease to normal in the North Horn Formation, below the oil production interval of the Green River Formation. Substantial overpressuring, caused by generation of natural gas, is speculated to be present in the Cretaceous Mesaverde Group below the Altamont field (Spencer, 1987; Fouch et al, 1992).

Our simulation results suggest that observed overpressures in the Uinta basin are probably caused by ongoing oil generation. In simulations of both the early uplift model and late uplift model, overpressures comparable to the observed overpressure distribution result. Overpressure development depends on conditions of permeability, relative permeability, TOC content, and oil viscosity. The most important factor, however, is the subsurface temperature history.

The correlation between temperature history and overpressures is illustrated by Figure 22, which shows the modeled temperature history (a), oil generation rate history (b), and overpressure history (c) for the Flagstaff member of the Green River Formation at

the northernmost edge of the Altamont field. These results correspond to the late uplift model (constant surface temperature boundary condition). During the depositional phase from 56 to 32 Ma, temperature in the Flagstaff increases as depth increases (Figure 22(a)). Temperature reaches the "oil window," or temperature range in which kerogen to oil conversion is significant, at about 40 Ma (Figure 22(b)). Overpressure, (plotted in Figure 22(c)), or the amount of pressure above the hydrostatic condition, becomes significant at ~37 Ma, shortly after oil conversion rates become significant. The chemical reaction that converts kerogen to oil approaches its peak at 28 Ma, and the peak in overpressures is coincident (see vertical reference line on Figure 22). The discontinuity, or sudden drop, in pressure (crossed by vertical reference line in Figure 22(c)) occurred due to an induced fracture at ~27.5 Ma; when fluid pressure exceeds the value of least principal stress, a vertical fracture is simulated by increasing vertical permeability one order of magnitude (reference Figure 8, the permeability data plot, and associated discussion). During the last 25 My of the basin's history, the temperature of the Flagstaff remains in the "oil window," yet the oil generation rate decreases to close to zero by present day. The rate reduction is attributable to the chemical reaction reducing available kerogen for conversion. We were led to infer that overpressures are associated with oil generation because overpressure dissipation coincides with oil generation rate reduction, much as overpressure increase is coincident with increases in oil generation rates, and a sensitivity analysis verified the conclusion. Overpressures dissipate but do not reduce to zero because permeability of the Flagstaff is fairly low and thus pressure bleeds off slowly. The rate of dissipation suggests that another 10 to 15 My would be required for overpressures to disappear completely. For this particular model (late uplift model with constant surface temperature), resulting overpressures are comparable to observed overpressures. Figure 23 is a map of hydraulic head results for the Flagstaff member, corresponding to the late uplift model at present day, e.g., final result. Comparing this map to the map of observed hydraulic head (Figure 20) illustrates that the model qualitatively reproduces observed overpressures, including peak overpressure approaching 6 km head in the northernmost part of the basin. The model overpredicts overpressures west of the Altamont field, suggesting that we may have overestimated the TOC distribution in that area.

Whereas the late uplift model qualitatively reproduces observed overpressures, the alternative end-member early uplift model does not reproduce as close of a match, albeit we consider the early uplift model to be more reasonable in a geologic context. The early uplift model with a constant surface temperature boundary condition (11° C) overpredicts pressures. Comparing the final overpressure value shown in Figure 24(a), the fluid pressure histories produced by this model for strata at the northernmost edge of the Altamont field, to the maximum observed overpressure, 27.5 MPa, shown in the observed fluid pressure profile (Figure 21) illustrates that this model overpredicts fluid pressure by about 15 MPa. However, the same model using the variable surface temperature history (Figure 12) underpredicts the final values of overpressure by about 12.5 MPa (refer to Figure 24(b), the fluid pressure histories for the early uplift model with variable surface temperature history).

Closer examination of both fluid pressure histories (Figures 24(a) and 24(b)) suggests that a tradeoff between induced fracturing and oil generation must occur for the

final value of overpressure to match the observed value. For example, the early uplift model with variable surface temperature history (Figure 24(b)) underpredicts the final value of overpressure. However, this occurs because initial overpressures are higher in this model and, consequently, many more fractures are induced, creating higher permeability than in the same model with constant surface temperature (Figure 24(a)). In both cases, four of the nine model layers are significantly overpressured during much of the modeled basin history, but higher overpressures are maintained longer in those formations that are less fractured because the permeability in those formations is lower. Note that in both models the North Horn Formation is overpressured, albeit it is not a source rock. The proximity of the North Horn to the highly pressured Flagstaff stratigraphically above, combined with its low permeability, causes the modeled overpressures in that formation. Of the three layers of Green River Formation that are overpressured, only the Flagstaff member is open lacustrine; the other two are marginal lacustrine and thus have a lower TOC assigned.

We could have adjusted model parameters to force a match between the modeled and observed overpressures in the early uplift model. The system is very nonunique, and any number of parameters could have been adjusted to achieve a match. However, the model simulations were performed with the objective of understanding the system, not to replicate it. Given that either structural history, combined with reasonable model parameters of TOC, oil viscosity, etc., qualitatively emulates the observed overpressure distribution, we performed a sensitivity analysis to isolate threshold values of certain parameters with respect to overpressure development.

Sensitivity Analysis

The two end-member structural histories illuminated the effects of the temperature history on overpressure development. A remaining issue was to elucidate how individual parameters, including permeability, relative permeability, porosity, capillary pressure, TOC, oil density, and oil viscosity affected overpressure development. We designed a sensitivity analysis that would isolate how these individual parameters affect overpressure development. Our study revealed that the individual parameters that affect overpressures the most are permeability, TOC and oil viscosity. Porosity and relative permeability are also significant, but the porosity effect manifests itself through associated permeability and relative permeability is a scaling factor of intrinsic permeability. For each of the three parameters, we ran a series of model simulations in which the individual parameter within a single finite difference cell was assigned a fixed value for the duration of the history simulation whereas all other parameters in all other cells vary as described previously. In all simulations, the late uplift model with the constant surface temperature condition was used, and the cell in which we fixed either permeability, TOC value, or oil viscosity corresponded to the Flagstaff member of the Green River Formation at the northernmost edge of the Altamont oil field and the deepest section of the basin.

Among the three parameters, permeability is probably the most important because permeability is a limiting factor that determines whether overpressures will occur due to any mechanism. Figure 25(a) displays the results for the permeability sensitivity analysis. Each point on Figure 25(a) represents the maximum overpressure reached in a single finite

difference cell in the Flagstaff member during a model simulation in which the cell was assigned a constant value of permeability; we tested the range of permeability from 10^{-19} m^2 to $5 \times 10^{-18} \text{ m}^2$, as shown by the abscissa on Figure 25(a). The results show that for values of permeability less than 10^{-18} m^2 , overpressures are insensitive to permeability, i.e., if permeability is lower than 10^{-18} m^2 , overpressures always occur, at least for these conditions of temperature history, TOC, porosity, etc. We refer to this lower limit of sensitivity the "saturation" value of permeability. Below the saturation value of permeability, the maximum overpressure reached in the cell was $\sim 51 \text{ MPa}$, the value of the least principal stress. In every simulation in which the fixed permeability was lower than the saturation permeability and fluid pressure reached the least principal stress, the cell of interest fractured and the pressure dissipated some amount.

For values of permeability greater than 10^{-18} m^2 , the maximum overpressure reached was less for each increase in fixed permeability (Figure 25(a)), and at $5 \times 10^{-18} \text{ m}^2$, no appreciable overpressure developed. Because overpressures do not occur for fixed permeability values above $5 \times 10^{-18} \text{ m}^2$, we refer to this upper limit of sensitivity as the "threshold" value of permeability. Note also that for values of permeability greater than 10^{-18} m^2 , the maximum overpressure reached was less than the least principal stress, and thus the cell of interest did not fracture. In all model simulations except the sensitivity analysis models, the minimum permeability for any cell in the model was $\sim 10^{-18} \text{ m}^2$, as indicated by Figure 8.

The sensitivity simulations also suggest that if intrinsic (absolute) permeability is increased an order of magnitude above the functional relationship we assigned, model overpressures are significantly reduced (all other variables kept the same). Similarly, if permeability is reduced an order of magnitude, overpressures increase tremendously. In all model simulations except the sensitivity analyses, we assigned all strata the same intrinsic permeability formulation (Figure 8). A valid alternative, however, is to assign alluvial strata a somewhat higher permeability (e.g., an order of magnitude higher) than what is assigned lacustrine strata; alluvial facies rocks like sandstone typically have higher permeability values than lacustrine facies rocks such as shale. Results for models with this alternative parameterization resulted in slightly diminished overpressures in lacustrine facies and no overpressure development in alluvial facies, provided other variables favor overpressure development. We used a permeability anisotropy ratio of $K_v = K_h/1000$, as the permeability/porosity data suggest (Figure 8). Fine stratification of clay within the Green River Formation implies that the maximum vertical permeability may be only as high as the maximum observed permeability of clay-rich strata, not considering fracture permeability. In a simulation in which the vertical permeability in the model was kept equal to the maximum observed permeability of clay-rich samples, $\sim 10^{-17} \text{ m}^2$, overpressures did not develop at all. Similarly, a simulation in which intrinsic permeability was isotropic ($K_v = K_h$) resulted in no overpressure development.

Similar results were obtained for the TOC sensitivity simulations. Figure 25(b) plots the maximum overpressure reached for fixed TOC ranging from 0.4 % to 1.7 %. The saturation value of source rock TOC is about 1.2% and the threshold value is about 0.5%, i.e., for values of TOC greater than 1.4 %, the maximum overpressure reached met or exceeded the least principal stress and for values of TOC less than 0.5% no appreciable overpressures developed. Note that the maximum TOC assigned in all model simulations

(except these particular sensitivity models) was 1.6%, which is slightly higher than the saturation value, and is the average value of TOC measured of Green River source rock samples, upper Paleocene to middle Eocene lacustrine facies below the middle marker zone (Tissot et al, 1978; Anders and Gerrild, 1984; Anders et al, 1992; Fouch et al, 1992). The sensitivity study also suggested that TOC can be too high and permeability too low for results to be consistent. If assigned TOC is much higher than 5%, or permeability assigned much less than 10^{-18} m^2 , then overpressures grow much more rapidly and climb much higher. Fracture density is increased to the extent that overpressures dissipate rapidly and are not maintained as long because of the induced high fracture permeability. Also important to consider is that we assigned the average value of observed TOC to open lacustrine source strata in the model. However, as much as 60 mass % and as little as no organic content have been measured for Green River samples, and a definitive distribution of TOC is unknown. Therefore, estimation of the uncertainty in model overpressure response associated with the 3-dimensional distribution of TOC would be very difficult to assess.

The results of the oil viscosity sensitivity models are similar in character to those of TOC and permeability. A range of 0.0 cP up to 10 cP was tested, but the saturation value was just less than 0.5 cP and the threshold value was about 0.05 cP. Therefore, the range of viscosity shown on Figure 25(c), a plot of maximum overpressure vs. oil viscosity, is from 0.0 cP to 0.7 cP. The maximum viscosity for the deep Flagstaff dictated by the Chew and Connally (1959) correlation approaches 2 cP, well above the saturation value of 0.45 cP (Figure 25(c)).

The typical values of permeability, TOC and oil viscosity in the deep Flagstaff at Altamont, as determined by rock and oil samples, approach or exceed the parameter saturation values determined by the sensitivity study, thus explaining why the model qualitatively reproduces observed overpressures. Table 4 summarizes the saturation and threshold values of the permeability, TOC and viscosity determined in the sensitivity study. In summary, the conditions necessary for high overpressures to develop are oil viscosity ~ 0.05 cP or higher, intrinsic permeability $5 \times 10^{-18} \text{ m}^2$ or lower, and TOC values of ~ 0.5 % or higher.

Overpressures Caused by Compaction Disequilibrium

Other mechanisms of Green River Formation overpressures have been proposed, including compaction disequilibrium associated with sedimentary loading (Sweeney, written communication, 1993) and excessive dewatering of clays associated with transformation of smectite to illite within low permeability strata. In this paper we demonstrate that ongoing oil generation sufficiently explains observed overpressures in the Tertiary Green River Formation. We did not consider effects of clay diagenesis (albeit some illite and smectite are present) or conversion of oil to gas as potential overpressure mechanisms.

The model results suggest that compaction disequilibrium probably did cause overpressures during periods of high sedimentation, but that these overpressures dissipated relatively rapidly. Plotted on Figure 26 is a portion of the fluid pressure histories corresponding to the late uplift model described previously. When the model was run with only one of either oil generation or compaction inactive, significant overpressures

were not induced (lower curve on Figure 26) during the period between 39 and 32 Ma. However, if both oil generation and compaction were active, some overpressuring was induced (upper curve on Figure 26). However, the presence of oil was necessary for compaction disequilibrium to cause overpressures; the oil generation rate during this particular period of the basin history was not high enough to produce overpressures, but the presence of oil reduced the relative permeability of water such that compaction disequilibrium became significant. Once deposition of the Duchesne River Formation ceases at 32 Ma, compaction ceases to be a significant overpressure mechanism. Simulation results suggest that overpressures due to compaction disequilibrium would probably be much higher and be maintained for significant geologic time beyond the cessation of deposition if the absolute permeability were much lower, e.g., less than 10^{-20} m². Alternatively, if the assigned compressibility of the strata is increased, overpressures by compaction would be more significant. However, a shift in compressibility shifts the modeled porosity trend to be inconsistent with the observed porosity trend.

Although either structural history can reasonably explain the observed overpressure distribution, the hydrodynamic and oil migration histories corresponding to the contrasting structural histories illuminate some constraints of the system and its parameters, as discussed in the following section.

Hydrodynamic Evolution and Oil Migration

Late Uplift Model Versus Early Uplift Model

The hydrodynamic history and oil migration patterns corresponding to the early uplift model are starkly different from those corresponding to the late uplift model. In terms of the hydrodynamic history, the timing of development of the basin's "bowl" shape structure affects the history of the surface topographic profile, directly influencing the hydrodynamic evolution. Regarding oil migration patterns, the previous section about overpressures demonstrated that the subsurface temperature histories for the contrasting models are significantly different, and therefore the oil generation rate histories are different. More importantly, the timing of differential uplift of the basin's periphery versus the timing of greatest oil generation influences migration patterns. In the early uplift model, the highest rates of oil generation occur during the formation of the "bowl" shape of the basin, or differential uplift of the basin's periphery. In contrast, the highest rates of oil generation occur before the "bowl" structure of the basin forms in the late uplift model, which affects where and how far oil can migrate.

These concepts are easily demonstrated using a panel of structural cross-sections along profile N-S constructed by the numerical model at selected time steps, corresponding to both the late uplift model (Figure 27, upper sequence of plots) and the early uplift model (Figure 27, lower sequence of plots). Overlain on each cross-section are fluid flow vectors, illustrating that the paleohydrogeologic regime evolves with changing structure and material properties. Also superimposed on the cross-sections are contours of oil saturation, illustrating the timing of oil generation and patterns of oil migration in the north-south direction.

The paleohydrologic regime depends greatly on the structural and topographic histories. In the early uplift model (Figure 27, lower sequence of plots), from 56 Ma until Uinta and Duchesne River formations are deposited in full, recharge occurs on the

southern side of the basin with flow primarily to the north. By the end of the Laramide Orogeny ~32 Ma, the flow patterns mimic what is thought to be the present day flow system. In contrast, the late uplift model (Figure 27, upper sequence of plots) has a very different history. From early in the history until uplift begins 10 Ma, most recharge is on the northern side of the basin with flow primarily from north to south. Uplift and erosion rates are very rapid during the final 10 My and the flow system assumes its present form during that time. At the final time (present day), both models converge to essentially the same flow system. However, oil migration patterns at the final time are starkly different. At 32 Ma, both the late uplift and early uplift models have produced significant oil within the Altamont oil field at the northern side of the basin. Much more oil forms in the early uplift model than in stratigraphically equivalent portions of the late uplift model by 32 Ma because more of the deep Green River section in the early uplift model is in the "oil window" longer prior to 32 Ma.

The difference in the amounts of oil generated between the two models is more easily understood by referring to Figure 28, plots of the temperature and oil generation rate histories for the Flagstaff member along cross-section N-S and at the center of the Altamont field (location indicated by solid circle on Figure 1). In the early uplift model, prior to ~32 Ma, the Flagstaff member is warmer than the same in the late uplift model, by as much as 20°C (Figure 28(a)). Advection is responsible for the lower temperatures in the late uplift model: topographic drive of groundwater from the northern side of the basin towards the southern side advectively carries heat away from the oil generation zone within Altamont. Additionally, in the early uplift model prior to 32 Ma, topography on the south drives groundwater from the south towards Altamont, with some heat transported to the area by advection. The net result is that prior to 32 Ma, the northern half of the basin is cooled by advection in the late uplift model and, in contrast, is warmed by advection in the early uplift model. In turn, the Flagstaff oil window is reached much sooner in the early uplift model: the oil generation rate becomes significant at ~40 Ma, and peaks at 35 Ma. In comparison, the oil generation rate of the Flagstaff in the late uplift model does not become significant until after 35 Ma, explaining why so much less oil has been generated in the late uplift model. The Flagstaff oil generation rate in the late uplift model does not peak until after 10 Ma, during the uplift phase. Temperature increases slightly during the last 10 My because uplift of the southern periphery of the basin changes the groundwater flow system to a primarily south to north flow system, which promotes some slight advection of heat towards the Altamont area (refer to the flow vectors on the last three of the upper sequence of plots in Figure 27). No significant oil migration has occurred by 32 Ma, i.e., the oil saturation contours shown in both models at 32 Ma reflect oil in place that has not been expelled from source rocks.

The fluid flow and oil migration results corresponding to the last 10 My of the basin's history are illustrated in the final three plots (upper and lower sequences) of Figure 27. In the late uplift model (Figure 27, upper sequence of plots), much more oil has been generated and significant migration from the source rocks in the northern half to the southern half of the basin has occurred prior to the last 10 My of the basin's history. The lacustrine facies source rocks in the Green River formation do not extend the full length of the profile. The southern extent of the source rocks is about 2/3 the length of the cross-section from the northern edge of the profile (N) towards the southern edge (S).

Thus, it is clear that oil migration has occurred out of these source rocks into the southern third of the basin, approaching Sunnyside and its tar sands. Buoyancy drives oil upward from the source rocks as well as laterally. Oil does not migrate directly to the surface primarily because of interlayered shales with tight permeability and high capillary pressures.

In contrast to the late uplift model results, no significant migration out of source beds occurred in the early uplift model during the last 10 My (Figure 27, lower sequence of plots). The primary reason migration is stifled is that viscosity is too high for oil to migrate beyond the Altamont area. Strata in the center of the basin never get deeper than 4 km, whereas equivalent strata in the late uplift model get as deep as 5 km, implying a temperature difference of as much as 25° to 30° C between equivalent strata of the two models. Consequently, oil viscosity in the middle of the basin for the early uplift model is 0.5 to 1 cP higher than oil viscosity in equivalent strata of the late uplift model at the same time (as estimated by the Chew and Connally (1959) correlation we employed in the model). The higher viscosity is an effective barrier to migration from the deeper rocks of Altamont past the center of the basin in the early uplift model. Another reason why oil is not found in the center of the basin for the early uplift model is that source rocks in the middle of the basin never reach the oil window, and thus no oil was generated or is in place in the center of cross-section N-S for this model.

Map views of hydrodynamic and oil migration results for the Flagstaff member of the Green River formation are also very useful for contrasting the results of the end-member structural histories. Figure 29 is a map of flow vectors and oil saturation contours corresponding to the late uplift model at 10 Ma. Water flow vectors are oriented north south, reflecting the relatively flat structure with higher topography at the northern side driving flow southward. Oil saturation indicated by the darkest shade approaches 0.9. The northern edge of this heavily oil-saturated zone also demarcates the northern boundary of alluvial facies within the Flagstaff member of the Green River formation. Note also that all strata north of the highly oil-saturated zone are open lacustrine and marginal lacustrine, all strata west of the highly oil-saturated zone are marginal lacustrine, and all strata south of the highly oil-saturated area are also alluvial facies; for a detailed mapping of facies environments at several stratigraphic levels within the Green River Formation, see Fouch (1975) and Fouch et al (1992). The facies mapping is important because of the relative permeability relationships applied to the different facies (Figure 9). In both open and marginal lacustrine, the oil saturation can never exceed ~20 % because water becomes immobile at 24 % oil saturation (Figure 9(b)) and therefore cannot be forced out of the pore space. In contrast, water stays mobile (i.e., relative permeability with respect to water > 0.) until oil saturation approaches 100% in the alluvial facies relative permeability formulation (Figure 9(a)). For these reasons oil saturation in alluvial facies strata exceeds 80% whereas oil saturation in the lacustrine facies never exceeds 20% (Figure 29). Also, these constraints on water mobility explain why water flow vectors are of such small magnitude in lacustrine areas with oil saturation approaching 20%, but are of relatively high magnitude in nearly oil-saturated alluvial facies. The North Horn Formation possesses no source strata (TOC=0%); thus, oil is not sourced from it. Rather, most oil migrated from the open lacustrine source rocks to the north of the alluvial facies environment. A small portion of the oil that migrated into the alluvial facies is

sourced from the marginal lacustrine facies located west of the alluvial facies. Recall that we assigned marginal lacustrine facies to have 30% of the TOC relative to the open lacustrine facies.

Much as the cross-sectional display of results (Figure 27) showed, the map views clearly demonstrate how different are the hydrodynamic and oil migration results for the two competing structural histories. Figure 30 is a map view of water flow vectors and oil migration patterns for the Flagstaff corresponding to the early uplift model at 10 Ma. Whereas the hydrodynamic regime was primarily north to south flow in the late uplift model at 10 Ma, the fluid flow vectors in the early uplift model at 10 My are nearly identical to the present day flow system results: flow is generally from the topographically higher periphery of the basin towards the center of the basin, especially the area of lowest surface elevation southwest of Redwash. Flow vectors are of highest magnitude coming from the southern side of the basin. In models without in situ oil generation/migration, flow vectors coming from the northern edge, where the topography of the Uinta mountains is an effect, are greater. However, relative permeability of water in the presence of oil limits the water flux rate, as explained above. Figure 30 also evidences the limited migration out of the Altamont area, due to the "viscosity barrier" discussed previously.

An important conclusion to draw from these results is that the late uplift model provides a reasonable mechanism for hydrocarbons to reach the south side of the basin and form the Sunnyside tar sands. In contrast, the limits on oil generation and viscosity imposed by the temperature history of the early uplift model do not promote oil migration to Sunnyside. The early uplift model is probably the more geologically reasonable of the two models -- it is more likely that the uplift of the periphery of the basin is coincident with other Laramide structural events (e.g., San Rafael uplift) and not associated with regional uplift of the Colorado Plateau during the last 10 My. Thus, the fact that the less reasonable model produces more reasonable migration results is somewhat paradoxical. The likely explanation may consist of at least two reasons: (1) the viscosity correlation (Chew and Connally, 1959), even though consistent with the few viscosity measurements we obtained, is somewhat unreasonable and is inhibiting migration too much; (2) neither proposed conceptual model of the structural history is correct, and the actual structural history lies somewhere between them. Recall that overpressure results for both models are reasonably consistent with the observed overpressure distribution. The viscosity correlation that we used imposes oil viscosity values in the deepest part of the basin (e.g., 1.5 cP) that are still significantly above the "threshold" viscosity value for overpressures (0.45 cP) determined in the sensitivity analysis discussed previously. Therefore, we could probably scale down the viscosity correlation which would not affect overpressure results but would probably have a significant effect on migration patterns for the early uplift model.

Static Model

Migration results for the early uplift model are similar to those of an early stage of this study when we had modeled the basin with a static mesh, prior to developing the evolution model (McPherson and Bredehoeft, 1992). In this early study, we developed an a priori temperature and oil generation history, and modeled oil migration patterns for the

present-day structure. Results of this model also indicated little to no migration away from the Altamont source rocks, with migration local to Altamont occurring solely due to overpressure gradients. The groundwater flow system (flow towards Altamont) and viscosity were also barriers to oil migration. These results suggest that a static model is inadequate for proper migration analysis, except to study current trends.

SUMMARY AND CONCLUSIONS

The primary goal of this study was to understand better high pore fluid pressures and regional scale hydrocarbon migration patterns in the Uinta basin, Utah. To achieve this goal we developed a 3-dimensional, numerical model of the evolution of the basin with special consideration of compaction disequilibrium and hydrocarbon generation as possible overpressure mechanisms. The model simulates the structural evolution of the basin through time, coupling the thermal and hydrodynamic evolution, and including in situ hydrocarbon generation and migration.

Because the structural history of the basin is not definitively known, the evolution of the Uinta basin was modeled using two "end-member" extremes of possible structural histories. Depositional histories are the same for both cases, but the history of differential uplift and erosion of the basin's periphery, that which forms the "bowl structure" of the basin, is different for the two cases. In the "late uplift model," all uplift and erosion occurs in the last 10 My of the basin's history, what is thought to be a period of uplift of the entire Colorado Plateau. In the "early uplift model," all uplift and erosion occurs during the Laramide orogeny through mid-Tertiary time, when the Uinta Mountains formed, the San Rafael to the south of the basin was uplifted, and sedimentation from these and other sources filled the trough of the basin. Results of neither end-member case is entirely consistent with observed overpressures and oil migration patterns, but the model with all uplift during the last 10 My produces migration trends much more consistent with observed oil production maps. However, overpressures produced by both models are generally consistent with observed present-day overpressures. Depending on the specific parameterization used, both models are consistent with regards to induced fractures in the northern Uinta basin, which is well known for its high fracture density.

The actual structural history lies somewhere between these two end-member cases. For oil to migrate to the southern and eastern flanks of the basin, areas of known oil production, differential uplift of the basin's flanks must occur after oil has migrated to these areas. Thus, the model does not narrow the timing and duration of differential uplift, but it does suggest that most uplift and erosion occurs well beyond early Tertiary time. Peak oil generation occurs from the time of maximum burial in the mid-Tertiary (~35 to ~30 Ma), and most differential uplift probably occurs well after this time. Otherwise, oil viscosity increases too much to permit migration of oil to the southern and eastern flanks. If the periphery of the basin is uplifted too soon, topographic drive of groundwater towards the oil source rocks in the northern basin may also stifle migration.

A summary list of key results includes:

- 1) The actual structural history lies somewhere between the proposed end-member conceptual models of the structural history, the "early uplift model" and the "late uplift model." It is likely that some, if not most, of the uplift and erosion that formed the

"bowl structure" of the basin occurred during the Laramide orogeny, and some may be coincident with regional uplift of the Colorado Plateau beginning 10 Ma.

- 2) Model results suggest that observed overpressures in the Uinta basin are probably caused by ongoing oil generation in strata with specific conditions of permeability, relative permeability, TOC content, and oil viscosity.
- 3) A sensitivity analysis elucidated "threshold" values of certain parameters -- values of the parameters that must be met, exceeded or not exceeded -- for overpressures to develop. Most or all of these specific threshold values must be coincident, depending on the structural history imposed, for overpressures to be significant: oil viscosity is 0.05 cP or higher, intrinsic permeability $5 \times 10^{-18} \text{ m}^2$ or lower, TOC values of 0.5 % or higher.
- 4) The sensitivity analysis also elucidated "saturation" values of viscosity, TOC and permeability. Results show that if oil viscosity is greater than 0.45 cP, permeability is less than 10^{-18} m^2 , or TOC is greater than 1.2%, fluid pressures always met or exceeded the least principal stress. If all three parameters approach or exceed their saturation value, overpressures always developed. Otherwise, if one or two of these parameters were far below their saturation value, then the other parameter(s) must exceed the saturation value for overpressures to be favored.
- 5) Typical values of permeability, TOC and oil viscosity in the deep Flagstaff at Altamont, as determined by rock and oil samples, approach or exceed the parameter saturation values determined by the sensitivity study, thus explaining why the model qualitatively reproduces observed overpressures
- 4) Most differential uplift and erosion of the basin's flanks probably occurs well after mid-Tertiary time. Otherwise, oil migration to the periphery of the basin is inhibited by reduced oil viscosity and fluid flow towards the area of source rocks.
- 5) Compaction disequilibrium occurs, but is relatively minor.
- 6) The basin's hydrodynamic history is directly coupled to the structural history. The two end-member possibilities of the structural history provided extremely different hydrodynamic histories and oil migration histories.
- 7) Oil generation and migration histories are controlled by the 3-dimensional temperature history, which depends on the structural history and boundary conditions imposed.

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Figure 1. Location map of Uinta basin, Utah. Locations of profile N-S is indicated on the map, as are most oil and gas production fields.

Figure 2. Generalized stratigraphy and structure of the Uinta basin. (a) Generalized stratigraphic column of the Tertiary formations. (b) Structural cross-section N-S (location of profile N-S shown on Figure 1.), vertical exaggeration = 20. The lines on the cross-section demarcate the hydrostratigraphic units used in the numerical model. See Fouch (1975) and Ryder et al (1976) for an explanation of stratigraphic terms.

Figure 3. Structure-contour map of the Flagstaff Member of the Green River Formation. Values of elevation are in m. The contour interval is 500 m.

Figure 4. Flow chart detailing numerical algorithm of basin evolution model.

Figure 5. Uinta basin porosity data plotted against depth and effective stress. Solid line represents compaction trend used to backstrip sedimentary section as well as the functional relationship used to govern forward porosity development in the basin evolution model. Open diamonds represent porosity values measured of samples showing microfractures. Open squares represent porosity data from Pitman et al (1982) for the Pariette Bench area. The remaining solid circles represent porosity-depth data for unfractured samples from the Altamont/Bluebell area.

Figure 6. Schematic drawing of a basin cross-section illustrating method of projecting stratigraphic thickness. Dotted lines represent the boundaries of the projected part of the layer of interest. Layer 1 in this drawing represents a layer for which estimating the thickness cannot be done by structural inference alone.

Figure 7. Calcite/quartz abundance versus XRD calcite peak intensity, an example of determining XRD weighting-factors.

Figure 8. Permeability of Uinta basin samples, plotted against porosity. Solid circles represent permeabilities from unfractured samples with no clay present. The solid triangles represent permeability from unfractured samples with clay present, illustrating how any clay can significantly reduce permeability. The open squares are data from Pitman et al (1982). The solid line is a least-squares regression fit (with exponential-form equation) through the no-clay data and the data of Pitman et al (1982). The dashed line is the permeability/porosity relationship tracked by a fractured cell (discussed in the text).

Figure 9. Relative permeability curves for oil and water. (a) Relative permeability formulation applied for oil and water in alluvial facies (e.g., sandstone). (b) Relative permeability formulation applied for oil and water in lacustrine facies (e.g., shale).

Figure 10. Capillary pressure curves assigned to alluvial and lacustrine strata.

Figure 11. Oil density data, corrected for in situ pressure and temperature conditions (we used corrections from Beal, 1946), for 1070 Uinta basin wells. The solid line is the relationship used to assign initial density of oil expelled from source strata.

Figure 12. Surface temperature history assigned to Uinta basin. Adapted from Wolfe (1978).

Figure 13. Modeled surface heat flow in the Uinta basin. Also posted are measured values from various studies (Chapman et al, 1984; Reiter et al, 1979; Sass et al, 1971).

Figure 14. Structural cross-section N-S with corresponding model fluid flow vectors and surface heat flow results (solid line on top plot). The dashed line on the heat flow plot represents heat flow results for a conduction-only simulation and the solid circles represent measurements made by Chapman et al (1984).

Figure 15. Burial history for strata at the northernmost edge of the Altamont oilfield, late uplift model. Key: UDR=Uinta and Duchesne River Formations, TG=Top of Green River Formation, MH=Mahogany Oil Shale, MM=Middle Marker of Green River Formation, TCM=Top of Carbonate Marker within Green River Formation, PE=Paleocene-Eocene Boundary, FS=Flagstaff member of Green River Formation, NH=North Horn Formation (Cretaceous).

Figure 16. Temperature history for strata at the northernmost edge of the Altamont oil field, late uplift model. Key: UDR=Uinta and Duchesne River Formations, TG=Top of Green River Formation, MH=Mahogany Oil Shale, MM=Middle Marker of Green River Formation, TCM=Top of Carbonate Marker within Green River Formation, PE=Paleocene-Eocene Boundary, FS=Flagstaff member of Green River Formation, NH=North Horn Formation (Cretaceous).

Figure 17. Burial history for strata at the northernmost edge of the Altamont oil field, early uplift model. Key: UDR=Uinta and Duchesne River Formations, TG=Top of Green River Formation, MH=Mahogany Oil Shale, MM=Middle Marker of Green River Formation, TCM=Top of Carbonate Marker within Green River Formation, PE=Paleocene-Eocene Boundary, FS=Flagstaff member of Green River Formation, NH=North Horn Formation (Cretaceous).

Figure 18. Temperature history for strata at the northernmost edge of the Altamont oil field, early uplift model (constant surface temperature boundary condition). Key: UDR=Uinta and Duchesne River Formations, TG=Top of Green River Formation, MH=Mahogany Oil Shale, MM=Middle Marker of Green River Formation, TCM=Top of Carbonate Marker within Green River Formation, PE=Paleocene-Eocene Boundary, FS=Flagstaff member of Green River Formation, NH=North Horn Formation (Cretaceous).

Figure 19. Temperature history for strata at the northernmost edge of the Altamont oil field, early uplift model (variable surface temperature history boundary condition). Key: UDR=Uinta and Duchesne River Formations, TG=Top of Green River Formation, MH=Mahogany Oil Shale, MM=Middle Marker of Green River Formation, TCM=Top of Carbonate Marker within Green River Formation, PE=Paleocene-Eocene Boundary, FS=Flagstaff member of Green River Formation, NH=North Horn Formation (Cretaceous).

Figure 20. Isometric projection of observed hydraulic head in the Flagstaff member of the Green River Formation within the Uinta basin.

Figure 21. Observed fluid pressure-depth profile in Altamont field (adapted from Bredehoeft et al, 1994). Shown for reference is a hydrostatic pressure profile (freshwater; dashed line).

Figure 22. Correlation between temperature history and overpressures for the Flagstaff member of the Green River Formation at the northernmost edge of the Altamont field: (a) modeled temperature history, (b) oil generation rate history, and (c) overpressure history.

Figure 23. Isometric projection of modeled hydraulic head in the Flagstaff member of the Green River Formation within the Uinta basin (present day).

Figure 24. Overpressure histories for strata in northernmost Altamont field, early uplift model, with different surface temperature boundary conditions: (a) pressure histories for early uplift model with constant surface temperature boundary condition; (b) pressure histories for early uplift model with variable surface temperature history boundary

Figure 25. Results of overpressure sensitivity analysis: maximum overpressure versus (a) fixed permeability, (b) initial TOC content, (c) fixed oil viscosity.

Figure 26. Selected window of the fluid pressure history for the Flagstaff member of the Green River Formation, illustrating that compaction disequilibrium contributes to modeled overpressures.

Figure 27. Hydrodynamic evolution and oil migration results. The sequence of graphs are plots of structural cross-section N-S with superimposed flow vectors and contours of oil saturation, at selected time steps. The plots show the structural evolution, hydrodynamic evolution, and oil migration through time. The upper sequence shows results for the late uplift model; the lower sequence of plots shows results for the early uplift model.

Figure 27 (continued).

Figure 27 (continued).

Figure 27 (continued).

Figure 28. Comparison of (a) temperature histories and (b) oil generation rate histories of the early and late uplift models, corresponding to the Flagstaff member along cross-section N-S and at the center of the Altamont field (location indicated by solid circle on Figure 1). The solid lines in both plots represent the early uplift model results and the dashed lines represent the late uplift model results. The vertical axis of plot (b) is normalized generation rate: before plotting, all generation rates were divided by the maximum rate, such that the results are normalized to unity.

Figure 29. Map of flow vectors and oil saturation for the Flagstaff member of the Green River Formation, corresponding to results of the late uplift model at the 10 Ma time step. S_o indicates oil saturation, expressed as fraction of total fluid present. Note that the patterns representing different oil saturations are slightly different from those used for Figure 27, the display of results in cross-section.

Figure 30. Map of flow vectors and oil saturation for the Flagstaff member of the Green River Formation, corresponding to results of the early uplift model at the 10 Ma time step. See Figure 29 and accompanying caption for the legend associated with this figure.

Table 1. Published estimates of time of maximum burial and of eroded overburden for area of Altamont field.

Reference	Time of Max. Depth	Meters Eroded	Method
Tissot, et al 1978	not specified	1780	unknown
Narr and Currie, 1982	not specified	1968	fluid inclusion
Sweeney, et al 1987	31 to 32 Ma	1796	inference from shale compaction data
Anders, et al, 1992	37 to 38 Ma	976	comparing modeled to observed maturity
Johnson and Nuccio, 1993	37 to 38 Ma	1890-335	extrapolating vitrinite profiles

Table 2. Oil/gas wells sampled for petrophysical measurements.

Well Name	Location (Township, Range, Section)	# Samples	Depth Range (m)
Chevron #1 Chasel Unit Flying Diamond	1 S, 1 W, 18	40	3240-3420
Chevron 1-33-3 Blanchard	1 S, 2 W, 3	65	2722-3781
Chevron 1-24 L. Boren Unit	1 S, 2 W, 19	4	3701-3716
Bow Valley 2-19 A1E DR Long	1 S, 1 E, 19	12	2793-2962
Chevron #1 Walker Hiko Bell	1 S, 2 W, 12	14	3195-3227
Bow Valley 2-22AI Ute Tribal	1 S, 1 E, 22	12	3746-3781
Chevron 9-4B1 Ute Tribal	2 S, 1 W, 4	33	3241-3534 3856-3891
Chevron 3-17-A2 Lamiq Urruty Unit	1 S, 2 W, 17	9	4142-4168
Chevron 4-19 A2 Ute Tribal	1 S, 2 W, 19	11	4214-4250

Table 3. Kinetic parameters for Green River black shale in the Uinta basin.

Mass %	Activation Energy (kcal/mol)
0.001	38
0.001	39
0.001	40
0.1	41
0.3	42
0.29	43
1.16	44
0.19	45
5.73	46
0.001	47
14.81	48
24.81	49
34.46	50
3.72	51
11.81	52
0.001	53
0.001	54
0.11	55
1.29	56
0.001	57
0.001	58
1.22	59
Frequency Factor	$3.61 * 10^{12} \text{ s}^{-1}$

Table 4. Threshold and saturation values of source rock and hydrocarbon (oil) properties for which overpressures developed.

Rock/Hydrocarbon Property	Threshold Value	Saturation Value
Intrinsic Permeability	$5 \times 10^{-18} \text{ m}^2$	10^{-18} m^2 (or lower)
Total Organic Carbon	0.5%	1.2% (or higher)
Hydrocarbon (oil) Viscosity	0.05 cP	0.45 cP (or higher)