3. SUBSIDENCE ANALYSIS

A. GEOHISTORY ANALYSIS

The goal of geohistory analysis is to produce a graphical representation of the vertical movement of a stratigraphic horizon in a sedimentary basin as an indicator of subsidence and uplift history in the basin since the horizon was deposited (Van Hinte, 1978; Fig. 3.1). Several types of stratigraphic data are needed to do a geohistory or subsidence analysis. These data include a stratigraphic column showing the present-day thickness of the stratigraphic units, types of lithologies, ages of horizons, and estimated paleowater depths. Other types of data that are useful, although not necessary, are porosity information for the units and thermal information, if your goal is to determine thermal history of the basin.

In addition, there are several assumptions and uncertainties that are built into this analysis. Most of these problems can be overcome if thick stratigraphic sections of relatively shallow-water deposits are used and only long-term, large-scale changes are studied.

1. Time Scale

The accuracy of the time scale with which you choose to work limits the accuracy of your results. This problem is compounded by using different types of dating techniques in a single study (e.g., radiogenic versus paleontologic age dates or planktonic foraminifer versus ammonite biostratigraphy). In extreme cases, subsidence rates can vary tremendously between time scales. Although there is no simple solution to this problem, at minimum choose the most accurate time scale as possible, and beware of this problem when comparing subsidence curves between basins where very different scales have been used.

2. Paleobathymetric Scale

The accuracy of the paleobathymetric scale you choose is even more problematic than that of the time scale (Fig. 3.2). Problems in calibrating a depth scale reflect differences in indicator species used by different workers and errors generated by assigning absolute depth values to the general relative-depth indicators described by paleoecologic studies. These problems can be reduced if you only work with stratigraphic sequences consisting of shelf-depth deposits and do not work with sediments deposited in deeper water where absolute water depths have greater uncertainties.

Without a doubt, these problems are more complex when working with nonmarine sequences. Estimates of elevation with respect to sea level (our datum) are usually confined to palynologic studies that may have very large uncertainties. There are other ways to try to constrain paleoelevation, including calculating paleoslope on river systems and estimating the distance upstream of the coastline. We can also assume that in order for thick nonmarine sequences to be preserved in the stratigraphic record, they must have been deposited relatively close to sea level. (Although, the question of how close to sea level is still somewhat arbitrary.) The simplest approach to correcting for elevation above sea level is to assign relatively large error bars to the paleoelevation that are still, hopefully, small compared to the stratigraphic thickness of the unit studied.

3. Compaction Corrections

Although you can correct for the effects of sediment compaction (discussed below), most methods are based on empirically derived porosity-depth relationships from a variety of sediment types (Fig. 3.3). Problems that arise are associated with the method used to collect the data, dealing with scatter in the data, the effects of overpressured horizons, cementation and late-stage diagenesis, and sensitive dependance on the exact lithologies involved. All of these problems lead to uncertainties in the final calculated subsidence history.

4. Sea-Level Effects

Changes in sea level can lead to errors in calculating the basin subsidence history because sea level is the datum from which subsidence is determined. If sea level rises, the record will show deepening water depth. This can be interpreted as an increase in basin subsidence and vice versa. There is no sure way of dealing with sea-level changes, because there is no consensus on the magnitude of global sea-level changes through time. The safest approach to sea level is to realize that short-term, small-scale changes in subsidence may represent sea-level changes, and focus your study on the larger scale events (more than a few tens of meters). This can be facilitated by working with relatively thick sections in which small-scale sea-level changes will only have a minor effect.

B. DISPLAYING THE DATA

It is convenient, although not necessary, to display all the data in a simple form in order to facilitate the calculations. Ingle (1980) has an excellent presentation of the key data needed for subsidence analysis (Fig. 3.4). He shows two columns: one in terms of thickness and one in terms of time. Lithology, paleobathymetry (without error bars), facies, and other data are also clearly shown in this figure.

C. CONCEPTS OF GEOHISTORY ANALYSIS

Van Hinte (1978) gives a clear presentation of his concept of geohistory analysis. His paper should be studied for a complete explanation. Following the approach of Mayer (1982), there are basically three steps in constructing a subsidence or geohistory curve.

1. Step One--Sediment Accumulation

First, the sediment accumulation through time is plotted (Fig. 3.5). In our example, time starts at T_0 and proceeds to T_4. At this point we use the present-day thickness of each stratigraphic unit. Because a constant interval of sediment was deposited during each interval of time, the sediment accumulation rate is linear.

2. Step Two--Compaction

In reality, sedimentary units compact after deposition so that the thickness of the interval that is preserved today is smaller than the unit's thickness at the time of deposition. For example, in Figure 3.6 as unit 1 is buried by younger units, it compacts until, at T_4, it reaches its final, present-day thickness. In our example, the original thickness of unit 1 was twice its present thickness. Therefore, if we remove the effects of compaction over time, the resulting sediment accumulation curve (Fig. 3.6) will lie below the curve generated without correcting for compaction. These two curves meet at the start of the graph (at T_0) because no sediment was yet deposited over the basement, and therefore, there is no compaction correction to be made. The curves also join at the present day (at T_4) because the sediments deposited in the last small interval of time have not yet compacted, and so there is no compaction correction to make.

Note that these curves only meet at T_0 because no sediments were deposited prior to this time. If, for example, you were analyzing the post-Paleozoic subsidence of a sedimentary basin and the basin included a significant Paleozoic stratigraphy, then there would be sedimentary rocks deposited prior to T_0. In this case, the compaction-corrected and noncompaction-corrected curves will not intersect at T_0, because the first early Mesozoic unit laid down compacted the underlying Paleozoic sedimentary rocks. Therefore, even when the first sedimentary units of interest were deposited, the basin subsided to some degree due to compaction of the earlier units and the curves will not come together at T_0. Similarly, if the youngest stratigraphic unit included in the subsidence history is much older than present and was overlain by a significant thickness of younger deposits, than that unit has been compacted. In this case, the two curves will not come together at the young end of the curve (in our example at T_4). Therefore, if only the Mesozoic part of a complete Phanerozoic stratigraphic section were being analyzed, the compaction-corrected and noncompaction-corrected curves would not join at either end of the graph.

3. Step three--Paleobathymetry

The final subsidence curve incorporates changes in paleowater depth in addition to the corrections made for compaction history. These changes in paleobathymetry are made because, in our previous examples, we have assumed that the sediments always filled up the basin to sea level, our datum. However, if the sediments did not build up the sea floor to sea level, then the preserved stratigraphic unit is not as thick as it would be if sediments completely filled in the basin to sea level. Therefore, if we only use the preserved thickness of stratigraphic units, we will underestimate the true amount of basin subsidence. If we have evidence of the depth of the sea floor on which the sediments accumulated (Fig. 3.7) from some paleobathymetric indicator, we can directly add this water depth to our sediment thickness to determine the amount of subsidence the basin underwent through time. In this example, the total subsidence history is episodic; however, without the paleobathymetric corrections the subsidence will appear to be linear.

D. EXAMPLE OF GEOHISTORY DIAGRAM CONSTRUCTION

We will now work through an example of generating a subsidence curve for a synthetic stratigraphic section that represents an hypothetical stratigraphic sequence from a passive margin setting such as the U.S. Atlantic margin (Fig. 3.8). Assume that these data are compiled from a measured section, several sections near each other or well information. The paleobathymetric scale is one used by Heller et al. (1982). Note that we have broken the section up into seven stratigraphic units including one unit that is a major unconformity. A long hiatus can be considered a separate unit that represents a significant period of time; although, it is not represented by stratigraphic thickness. The basis for breaking the section into seven units is somewhat arbitrary. A section can be broken up into different units based on (1) amount of age control available, this is important since subsidence is plotted with respect to time; (2) number of major unconformities, if they represent hiatuses that are long enough to matter on the scale of your final graph; (3) significant changes in paleowater depth, assuming that there is sufficient data to show such changes; (4) significant lithologic breaks, which require different compaction corrections; and (5) the purpose of the study, which may require that you to show more detail during certain periods of time.

1. Compaction Correction

From the compiled data we can draw a simple stratigraphic accumulation curve (Fig. 3.9). The next step is to remove the effects of compaction of the section through time. Here, we follow the method of

Van Hinte (1978) showing that the thickness of a unit at the time of deposition (T_O) and any time thereafter is related to the change in porosity of the sediment during burial (Fig. 3.10). In his derivation, Van Hinte points out that the volume of the grains does not change during burial (assuming no significant diagenesis), but that the volume of the pore space decreases during burial. Therefore, the original thickness is related to the present-day thickness as follows:

$$T_O = \frac{(1-\phi_N)\, T_N}{1-\phi_O} \tag{3.1}$$

where ϕ_O is the original porosity at the time of deposition and T_N and ϕ_N are the present-day thickness and porosity of the unit, respectively.

The rate of decrease of porosity during burial for different types of sedimentary rocks can be determined empirically from a variety of studies (e.g., Bond et al., 1983; Fig. 3.3). For simplicity, we have chosen a simple exponential relationship for the change in unit porosity:

$$\phi_N = \phi_O \, exp\,(-c\,z) \tag{3.2}$$

where the porosity today, or at any time during burial, (ϕ_N) is related to its porosity when originally deposited (ϕ_O), its present depth of burial (z) in meters, and a constant for each lithology (c). The ϕ_O and c used for sandstone, shale, and limestone in our examples are based on studies by Sclater and Christie (1980) and are shown in Figure 3.11. Some studies suggest that porosity decrease with depth follows a similar exponential curve regardless of lithology (Fig. 3.12), which makes computer programming simpler. However, for this example we will use the three different rates for lithologies shown in Figure 3.11.

In this approach, we must presuppose that all the changes of porosity with depth are the result of compaction. If early cementation is significant, than much of the change in porosity may not be due to compaction. Consequently, there will be no compaction corrections to be made, or they will be very slight (see early cementation curves versus compaction curves for the lithologies shown in Fig. 3.3). Of course, then a question arises. Is the cement derived from elsewhere in the stratigraphic column, or is it precipitated out of flow in an open circulation system? Another problem may arise over compaction rates for specific rock types. In this study, we consider compaction of only sandstone, limestone, and shale. It is likely that compaction for silty shale is different then shale or silt alone or that lithic sandstones compact differently then quartz arenites. Fortunately, for the most part, compaction corrections are small for stratigraphic sections of significant thickness (i.e., generally less then 10% of the total subsidence history). Therefore, unless details of the subsidence history are required, compaction corrections can be made as either a single lithology (as in Fig. 3.12) or as three or four lithologies (Fig. 3.3). If you are very concerned about compaction corrections, we encourage you to generate error bars by assuming that all the change in porosity is due to early cementation alone, and then assume that it is all due to compaction. This will define an upper and lower limit in the subsidence curve due to uncertainties in compaction correction. Potentially, petrographic work can help reduce the range of uncertainty.

Using equations 3.1 and 3.2, the thickness of the units at successive stages in burial can be restored (Fig. 3.13). For convenience, we calculate the porosity for the middle of each unit, and assume that this value represents the average porosity of the entire unit. Of course, the depth to the middle of each unit changes as burial proceeds. Following Van Hinte (1978), we set up a table that shows the thickness and porosity of each stratigraphic unit during burial. Figure 3.13 shows how the table is set up for the compaction calculation. Data is taken directly from Figure 3.8. If we take the present stratigraphic column (on the left), we can calculate the present porosity of the midpoint in each unit based on the equation and values of Figure 3.11. In the second column from the left, the uppermost sedimentary unit (unit 7) has been removed. This column, then, represents the sedimentary sequence just after unit 6 was deposited. We can calculate what the porosity was at the midpoint in unit 6 was at this time and, from Figure 3.10, calculate what the thickness of the unit was at that time. Next, we can determine the porosity at the midpoint of the underlying unit 5 in this column and, from that porosity, determine its thickness. This process is repeated for each unit down the column to determine the total thickness (ΣT^*) of the stratigraphic section after the deposition of unit 6. We then repeat the entire process for each column across the diagram, starting with the original column. We use the original column and not the previously calculated column in each iteration. This is because the thickness of each unit in the immediately preceding column contains small errors due to not using the "correct" method described below and because porosities are only calculated to two significant figures. These errors become cumulative during the compaction correction. Thus, the next step is to go back to the present stratigraphic column and remove units 6 and 7. Now calculate the porosity and thickness just after unit 5 was deposited and so on. In this way, we restore the compaction (or in the manner we are doing it, decompaction) history of the entire section through time (Fig. 3.14).

2. The "Correct" Method of Decompaction

The technique for decompacting sedimentary strata, outlined above, is only approximately correct. If you plan to write a computer program to do decompaction, you may want to use the more accurate decompaction algorithm discussed below. Figure 3.15 illustrates the decompaction problem. A unit of thickness T_N is buried at a depth of d_N. We want to know the thickness of the unit (T_O) at some earlier time, when the unit was buried only to a depth of d_O. The basic assumptions are that the porosity decreases exponentially with depth (equation 3.2) and that the volume of rock grains within the unit never changes:

$$\int_{d_O}^{d_O+T_O} (1-\phi)\,dz = \int_{d_N}^{d_N+T_N} (1-\phi)\,dz \tag{3.3}$$

where ϕ is the unit porosity and the quantity $(1 - \phi)$ is the volume of rock grains (per unit volume of sediment) at any level within the strata. These two integrals can be evaluated analytically:

$$T_O + \frac{\phi_0}{c} exp(-c\,d_O)\left[exp(-c\,T_O)-1\right] = T_N + \frac{\phi_0}{c} exp(-c\,d_N)\left[exp(-c\,T_N)-1\right] \tag{3.4}$$

There is no way to solve directly for T_O; this is an example of a transcendental equation. The best we can do is to isolate one T_O, guess a value for T_O, and then calculate a new value for T_O. This process is repeated until T_O stops changing from one step to the next. The resulting value is the solution. Isolating T_O gives:

$$T_O = -\frac{\phi_0}{c} exp(-c\,d_O)\left[exp(-c\,T_O) - 1\right] + T_N + \frac{\phi_0}{c} exp(-c\,d_N)\left[exp(-c\,T_N) - 1\right] \qquad (3.5)$$

Let's look at an example of this iteration process using the data given for unit 6. We are dealing with a shale, so use the porosity constants ϕ_0 = 0.5 and $c = 5. \times 10^{-4}\ m^{-1}$. The present thickness and depth of unit 6 are 800 m and 1000 m, respectively. What was the thickness of unit 6 at the end of its deposition ($d_O = 0$)? First, numerically evaluate as much of equation 3.5 as possible:

$$T_O = -1000.\ exp(-5.\ x\ 10^{-4}\ T_O) + 1599.9 \qquad (3.6)$$

where T_O is given in meters. A reasonable first guess for T_O is 800 m (i.e., the present thickness). Put this guess into the right-hand side of the equation and obtain T_O = 929.58 m. Take this value, put it into the right side and obtain T_O = 971.63 m. Continue this process until T_O converges to some value. Our series went as follows: T_O = 800.00, 929.58, 971.63, 984.71, 988.71, 989.94, 990.31, 990.42, 990.46, 990.47, 990.47. Thus, we converge on a solution of T_O = 990.47 m. This value is smaller (by 27 m) than the value of 1017 m obtained by the approximate technique. The difference is less than 3%. If you are trying to make estimates of sea-level variations using backstripped sedimentary sections, you should use the decompaction scheme developed here, particularly if you are dealing with thick units (>100 m). If you are only interested in the tectonic subsidence history and are using thin units (<100 m), save some time by using the approximate technique.

3. Summary

Upon completing the decompaction process, we can draw the sediment accumulation curve for the sequence (Fig. 3.16). This curve has removed the effects of compaction of the section. Because the compaction corrections we are using are quite generalized, the final curve represents only an approximation of the true accumulation curve. Don't forget that any decompaction scheme will overestimate the unit thickness if cementation played a significant role in the lithification of the unit. Nonetheless, even assuming no cementation in our example, you can see that compaction corrections change the magnitude of the curve but do not make a tremendous difference in the general shape of the curve.

4. Total Subsidence Curve

Once compaction corrections have been made, the paleobathymetric estimates for each stratigraphic unit are simply added in, producing the total subsidence curve (Fig. 3.17). Error bars on the water depth should be included. Notice that the subsidence of this site begins rapidly at 75 Ma and then abruptly slows down. The remaining subsidence is relatively slow, with possible uplift at about 20 Ma. However, uplift is not required within the error bars on the paleobathymetry.

There is no record of the subsidence history during the hiatus. There could have been tremendous subsidence and uplift during the time represented by this unconformity. All we know for certain is the amount of net subsidence during this time because we know how much the basin subsided prior to the hiatus, and we know the change in water depth recorded in the sediments that directly overlie the unconformity. The argument is made that if there was no deposition or erosion during the hiatus and the basin had continued to subside during this time, the seafloor would be deeper when the sedimentary record resumed following the unconformity. This is exactly the case in our example. The seafloor was about 100 m deep at the end of unit 3, and it was greater than 500 m deep by the beginning of unit 5. Barring any sea-level change, which is certainly a possibility, the seafloor and, therefore, the basin, at this spot must have continued to subside, on net, during the hiatus.

Possible changes in water depth due to eustatic fluctuations can be determined by comparing this subsidence curve with subsidence curves from other areas and determining if the steps in the subsidence curve of equal magnitude and timing occur in both records. Synchronous steps in unrelated subsidence curves may represent global sea-level changes. This approach was used by Heller et al. (1982) to determine the influence of sea level on regional subsidence curves. Another approach is simply to compare the steps in the subsidence curve with independent sea-level estimates such as those of Vail et al. (1977) or Haq et al. (1987). This approach assumes that the sea-level estimates used are correct.

E. CALCULATING TECTONIC SUBSIDENCE

The total subsidence curve produced in the previous section includes the contributions of all factors that affect the subsidence of the basin: tectonic loads, sediment loads, and sea-level changes (sediment compaction has already been corrected). In studies geared toward interpreting tectonic processes or history and/or the record of sea-level changes, the effects of subsidence caused by loading during sediment deposition need to be removed. The simplest approach for removing the effects of sediment loading, which is especially useful if you have only one stratigraphic section, is by assuming a simple local isostatic model in which the sedimentary units can be removed, and the basin is allowed to isostatically rebound. Be aware that local isostatic backstripping, described below, is not as accurate a technique to interpret basin subsidence in most basins as is regional isostatic (flexural) backstripping (described in Chapter 5, part K). However, isostatic backstripping does have the advantage of being relatively simple to do and, more importantly, requires far less data than a two- or three-dimensional flexural analysis. In addition, if stratigraphic thicknesses do not change over broad regions (i.e., more than a couple of times the flexural wavelength) then isostatic backstripping produces very similar results as flexural analyses (e.g., Steckler and Watts, 1978).

1. Backstripping

Backstripping--as discussed by Watts and Ryan (176), Steckler and Watts (1978) and Watts (1981)--is a fairly straightforward application of an isostatic balance applied to sediment loading. A review of the derivation and application of the isostatic equation (Fig. 3.18) suggests that a when a water-filled basin is filled in by sediment, it subsides to about 2.3 times its original depth, depending on the densities that are used. A backstripping equation that does the inverse problem can be likewise derived (Fig. 3.19) and predicts the depth of the basin if the sedimentary rocks were removed:

$$Z_i = S^*\left(\frac{\rho_a - \rho_s}{\rho_a - \rho_w}\right) + Wd_i \tag{3.7}$$

where Z_i is the depth to the tracked horizon relative to sea level (i.e., the amount of tectonic subsidence); ρ_a is the density of the asthenosphere; ρ_s is the density of the sediment column; ρ_w is the density of water; and Wd_i is the water depth for unit i. The total thickness of sediment column under the top of unit i, corrected for subsequent compaction is:

$$S^* = \sum_{j=1}^{i} T_j^* \tag{3.8}$$

which we have already calculated in the bottom row of Figure 3.14. Because all densities are used solely as a ratio of their contrasts, their units are not important as long as they are consistent. Be aware that equation 3.7 assumes that the basin is filled with water to sea level at all times. The amount of subsidence (Z_i) for a subaerial basin can be calculated by replacing ρ_w with 0 ($\rho_{air} \approx 0$).

Equation 3.7 is for cases in which the sea-level history is not known *a priori.* If the magnitude of sea level rise is known, then equation 3.7 can be rewritten as:

$$Z_i = S^*\left(\frac{\rho_a - \rho_s}{\rho_a - \rho_w}\right) + Wd_i - \Delta SL_i \frac{\rho_a}{\rho_a - \rho_w} \tag{3.9}$$

where ΔSL_i is positive for a sea level rise and negative for a sea-level fall.

On the right-hand side of equation 3.7, all the terms are known except the density of the sedimentary column, ρ_s. Because ρ_s changes as the thickness of the stratigraphic section changes due to compaction, it must be calculated after each new unit i is deposited. Following Steckler and Watts (1978), ρ_s can be calculated for the column after the deposition of each unit i following the derivation shown in Figure 3.20:

$$\rho_{s_i} = \frac{\sum_{j=1}^{i} \left[\phi_j \rho_w + \left(1 - \phi_j\right) \rho_g\right] T_j^*}{S^*} \tag{3.10}$$

where ρ_g is the grain density and ϕ_j is the porosity of unit j. Values for grain density and the maximum surface porosity are shown in Figure 3.11.

F. EXAMPLE OF BACKSTRIPPING

In the subsidence history we are using as an example, we can now apply equations 3.7 and 3.10. By plotting the amount of tectonic subsidence (Z) on the same graphs that we have plotted the total subsidence history (Fig. 3.21), we can now show the subsidence history of the sedimentary section that is due to forces other then sediment loading assuming local isostasy. Notice that in this example approximately one-half, or more, of the total subsidence is due to sediment loading. In fact, it almost appears that all of the tectonic subsidence took place within the first 10 My, with little happening, tectonically, after that time (the curve is nearly flat). During the last 10 My or so, there may have even been some tectonic uplift of the basin, however, uplift is not required given the magnitude of uncertainty in water depth. Again, the tectonic subsidence history during the time represented by the unconformity is unknown. However, at the end of the hiatus the basin had not subsided, on net, much more than where it was at the start of the hiatus. It would be serendipitous for the basin to have undergone significant vertical changes during the hiatus, ending up at approximately the same elevation as it started. Therefore, it seems unlikely that much happened tectonically during this time. As will be discussed in the rest of this course, these curves can be used to interpret the mechanisms of subsidence, the timing of subsidence events, and the sea-level history.

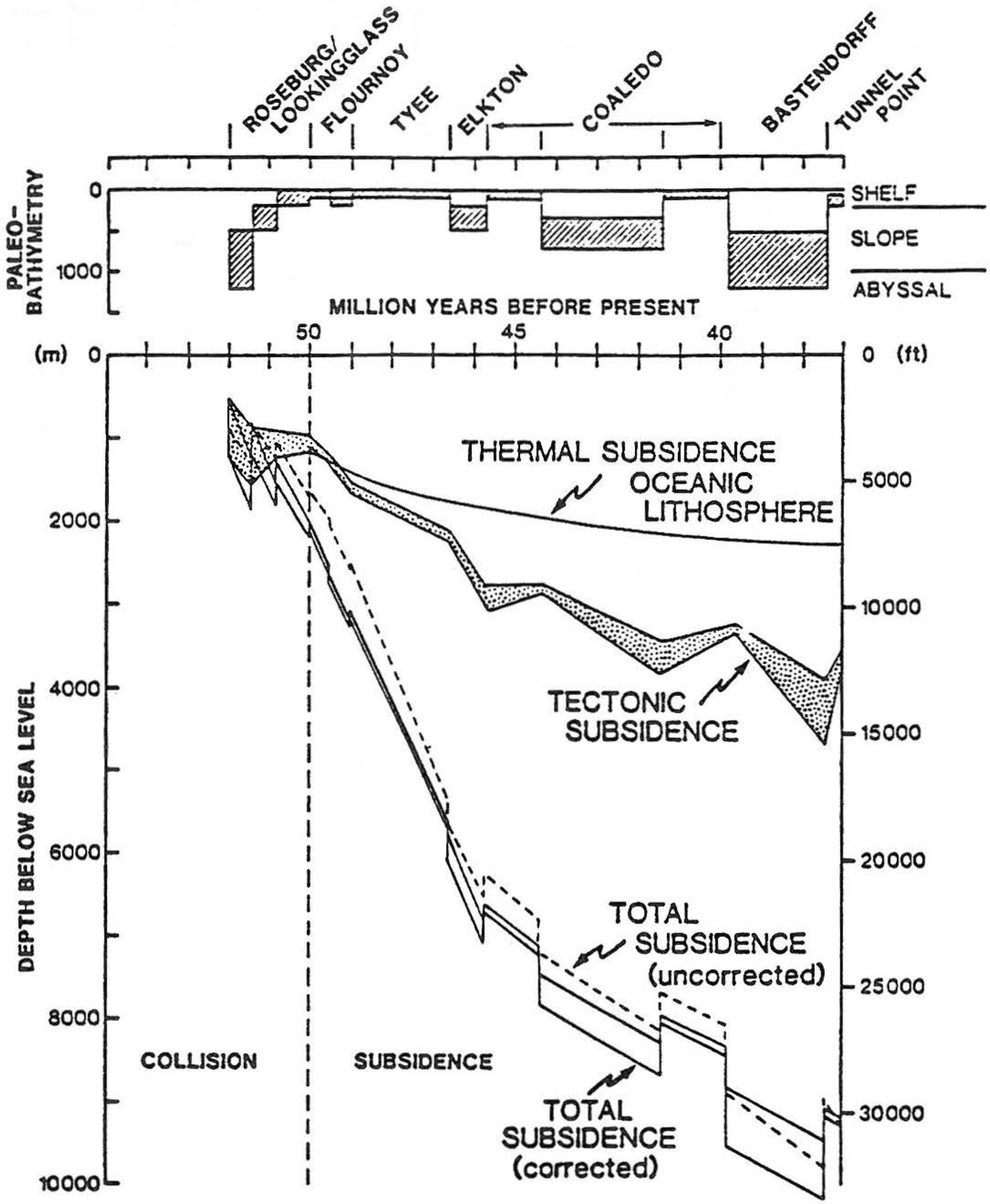


Figure 3.1 Subsidence history of the Oregon Coast Range during Eocene time. From Heller (1983).

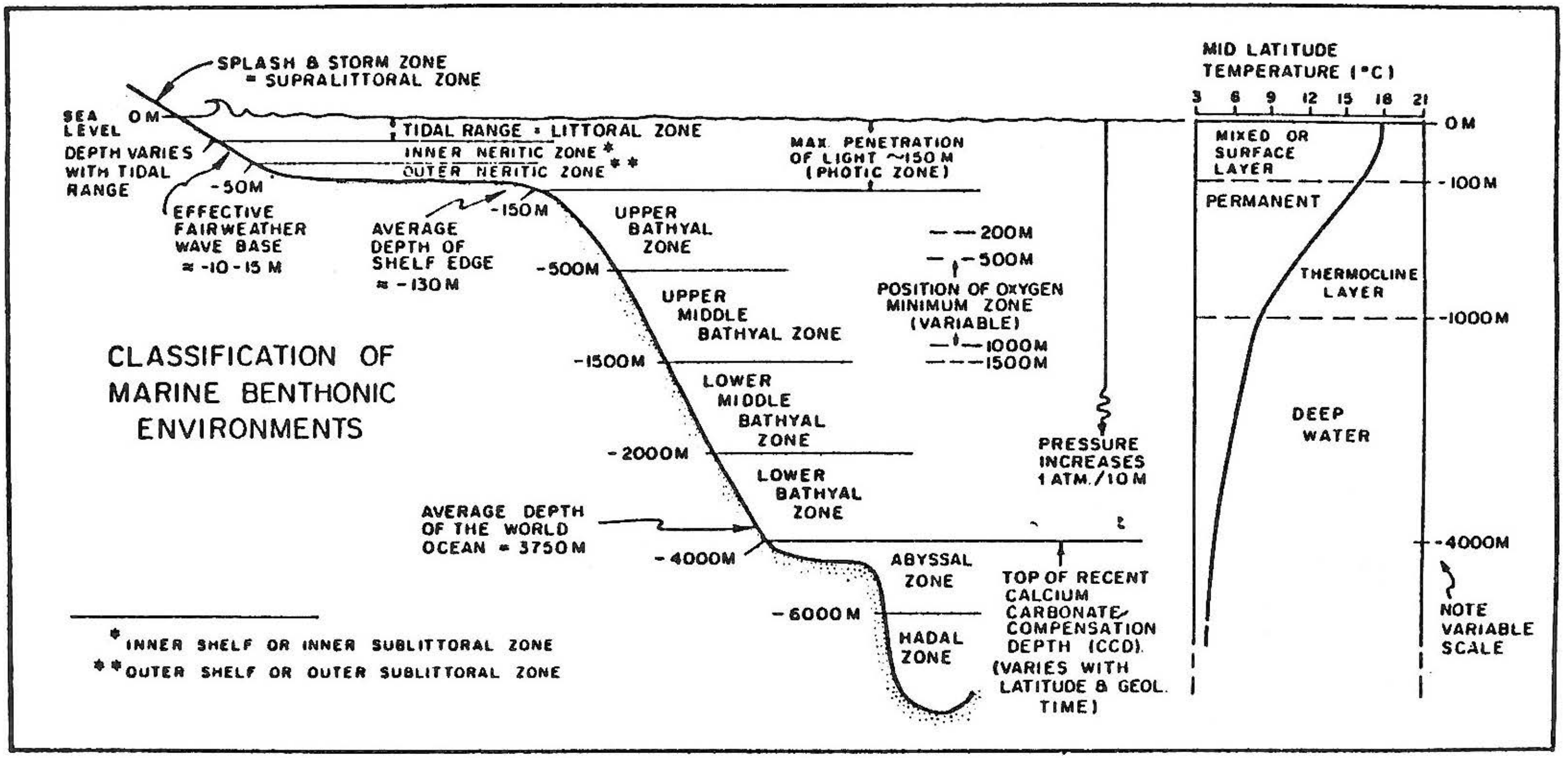


Figure 3.2 Example of a paleobathymetric scale from Ingle (1980).

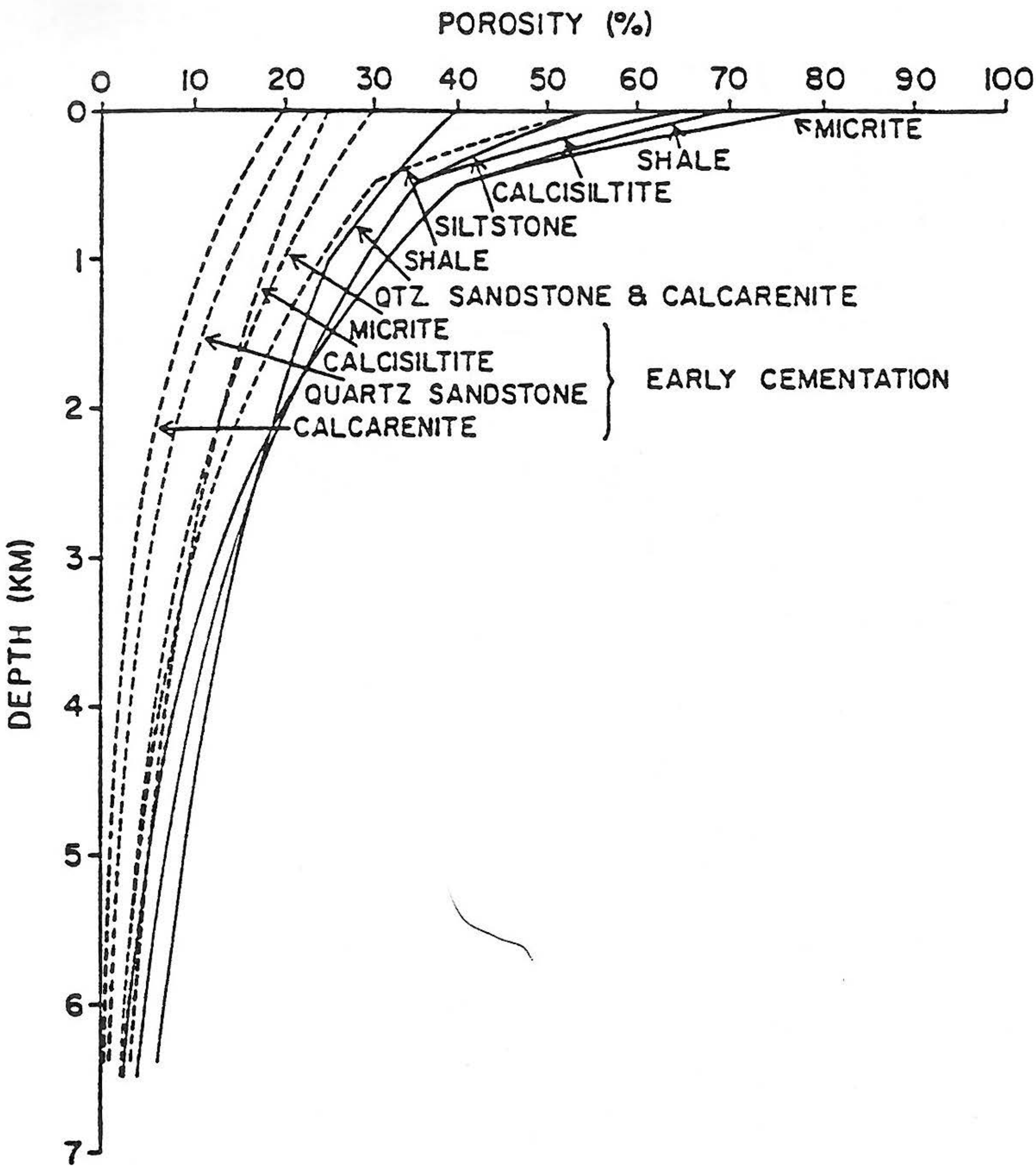


Figure 3.3 Porosity versus depth for a variety of lithologies, assuming there was either early cementation (dashed lines) or that all changes in porosity were due to compaction (solid lines). From Bond et al. (1983).

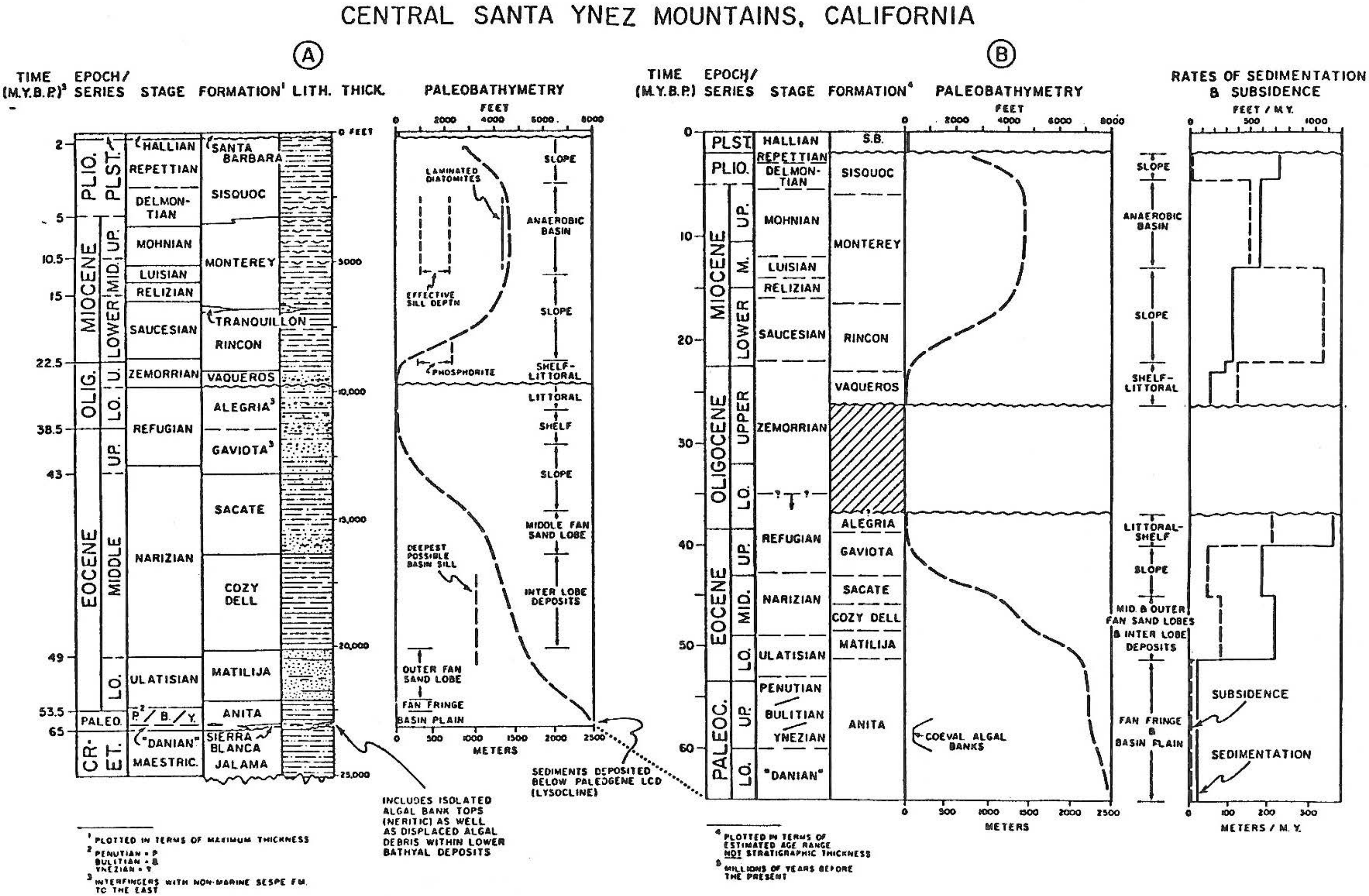


Figure 3.4 Display of data that could be used for a subsidence analysis. From Ingle (1980).

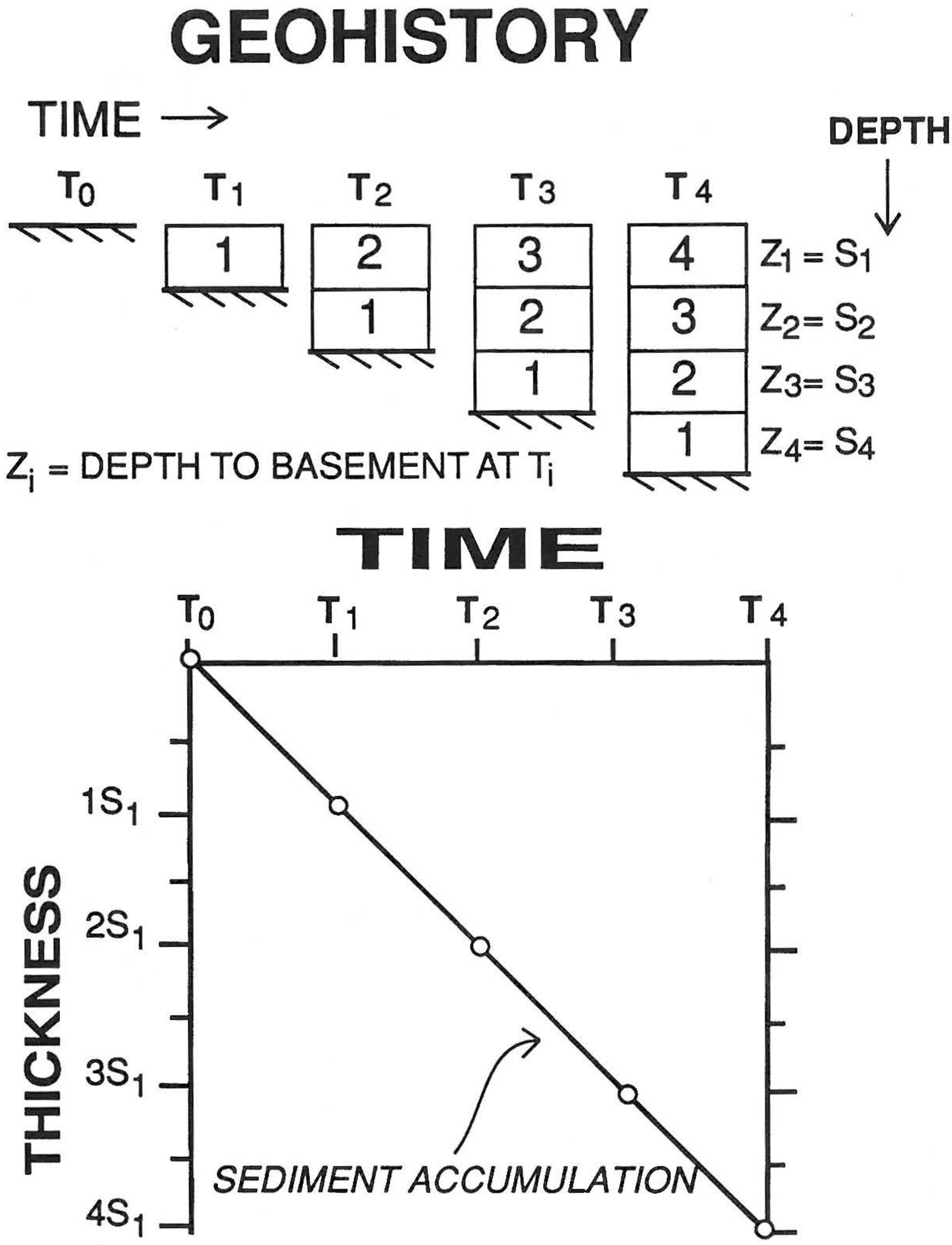


Figure 3.5 Using modern sediment thickness alone in geohistory analysis.

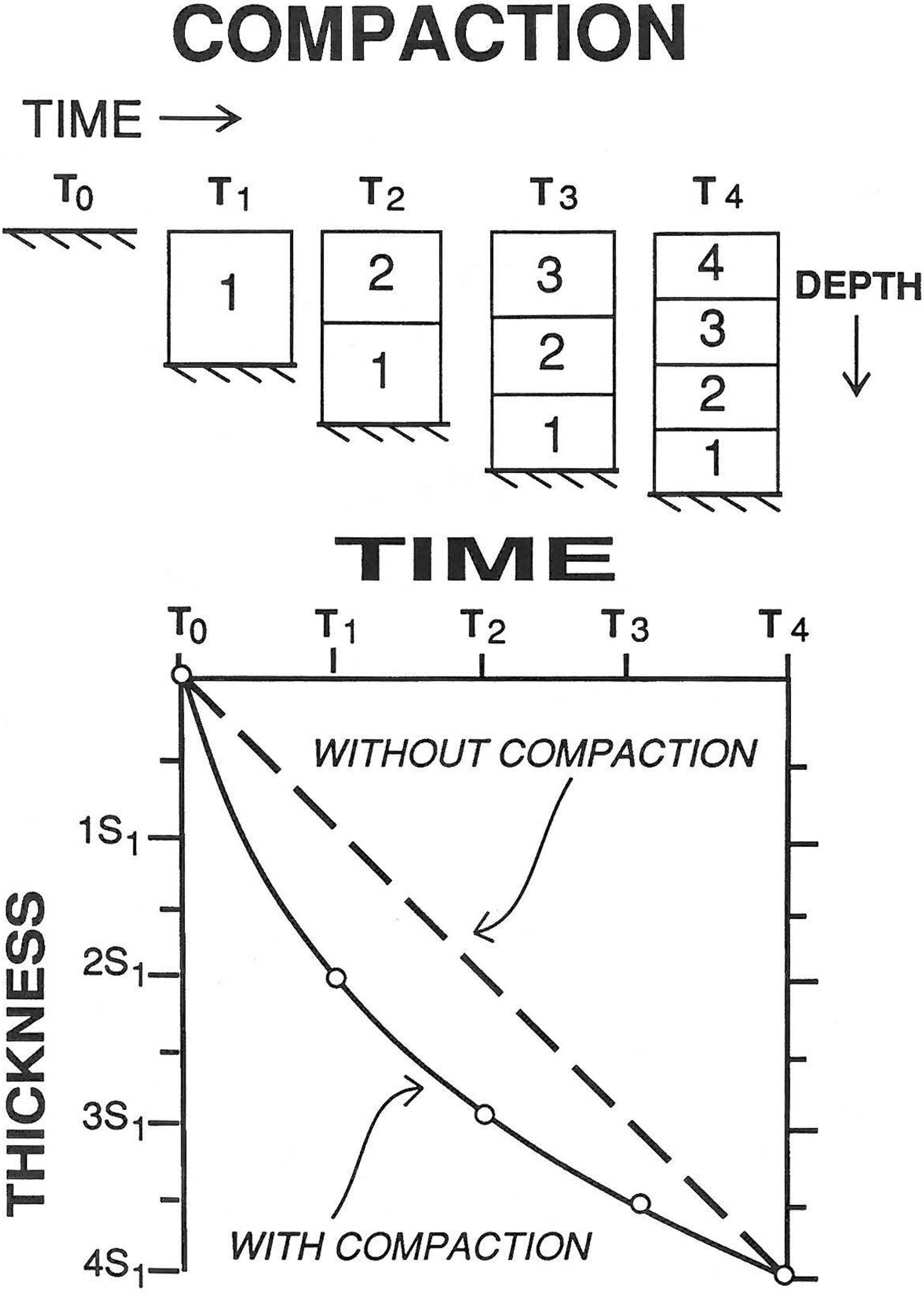


Figure 3.6 Removing the effects of compaction in geohistory analysis.

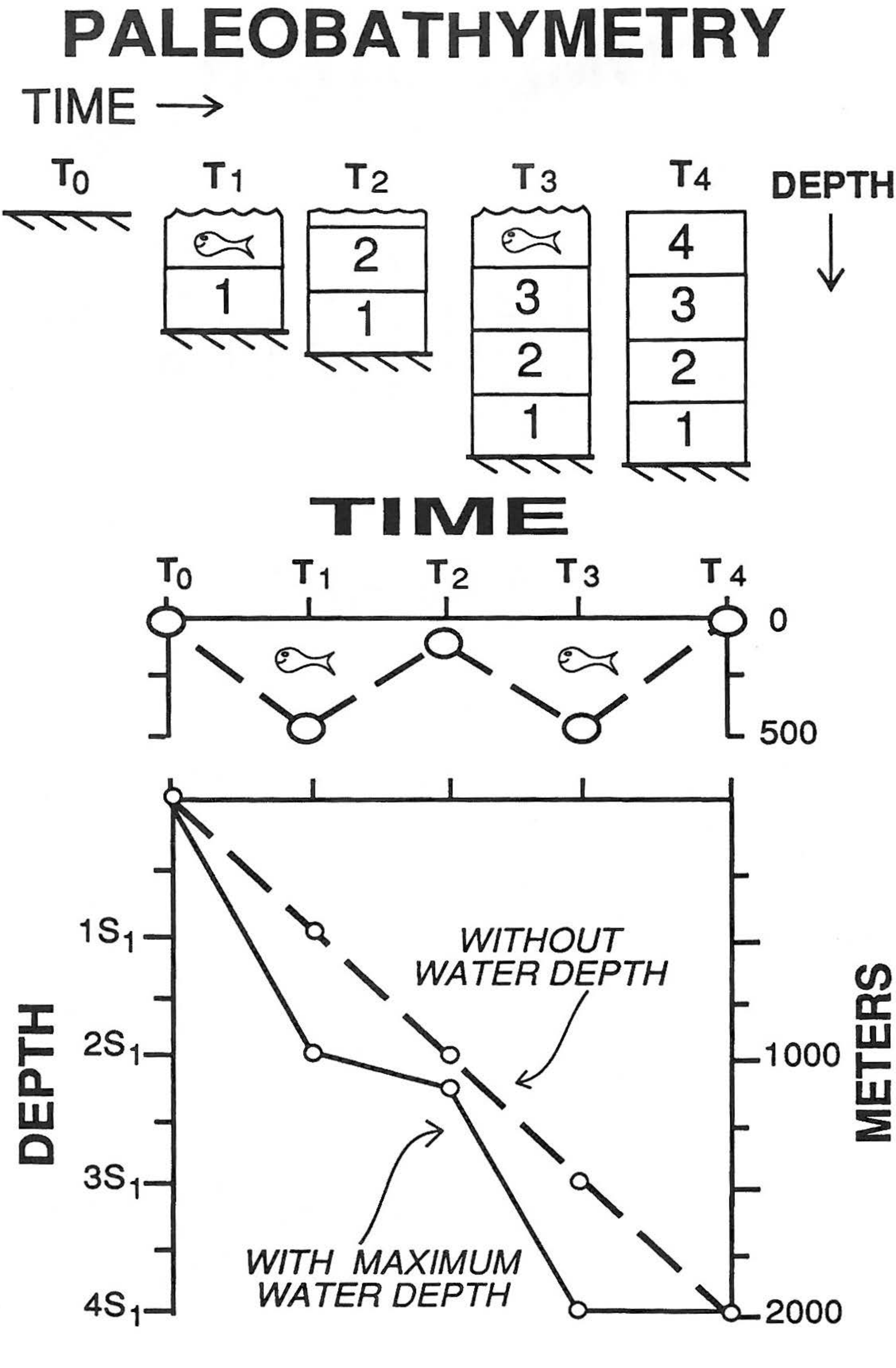


Figure 3.7 Complete geohistory analysis incorporating paleobathymetry.

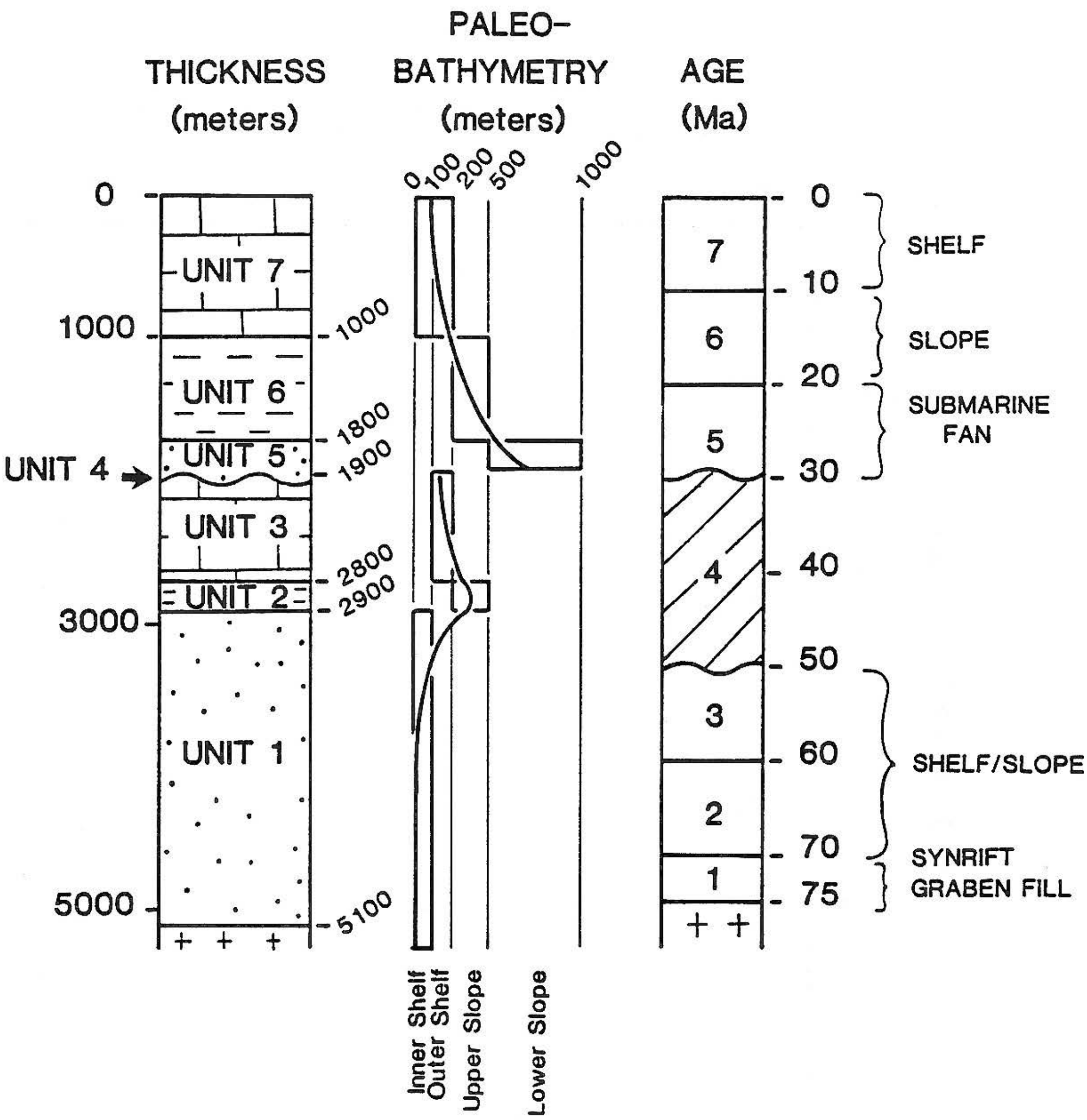


Figure 3.8 Data to be used in worked example of subsidence analysis.

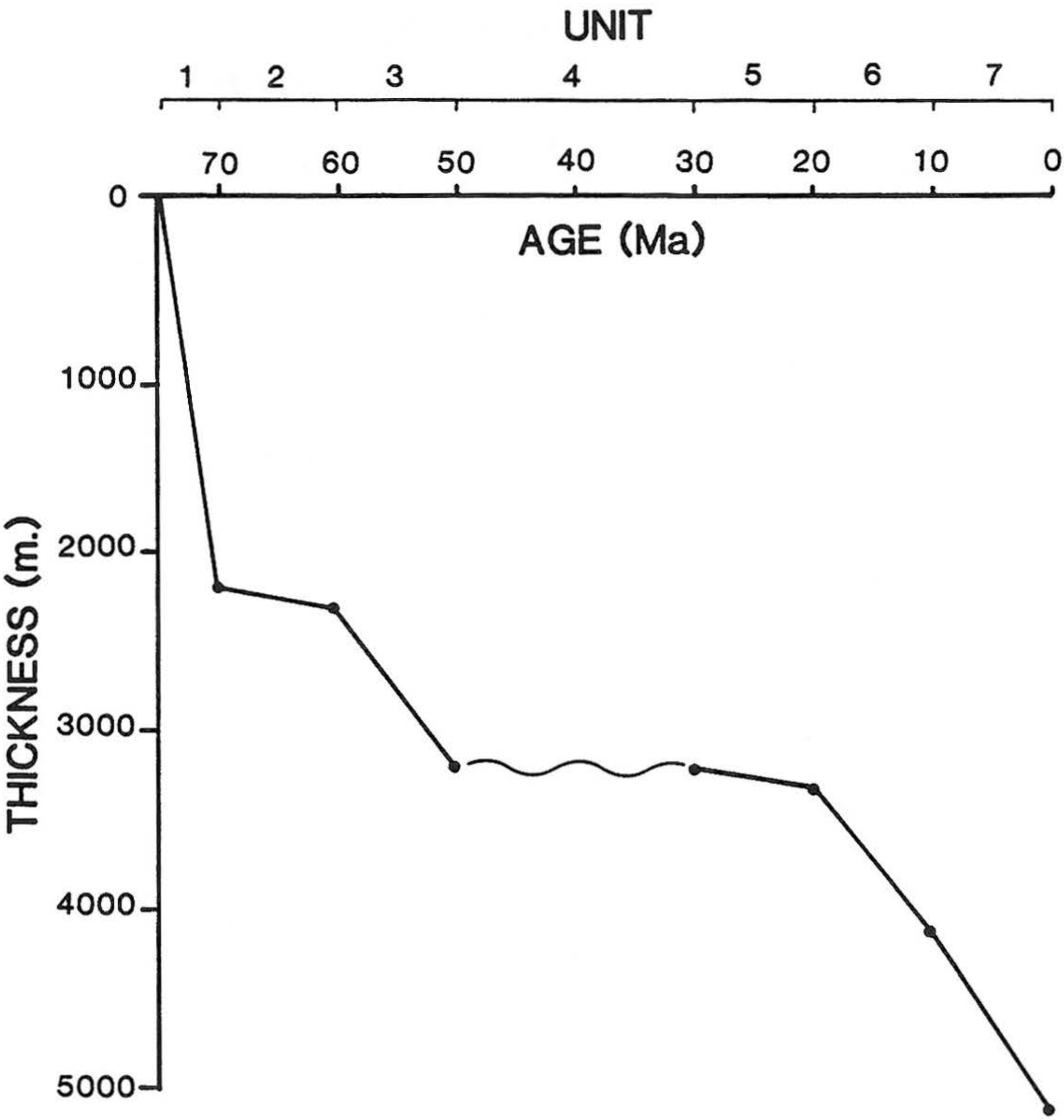
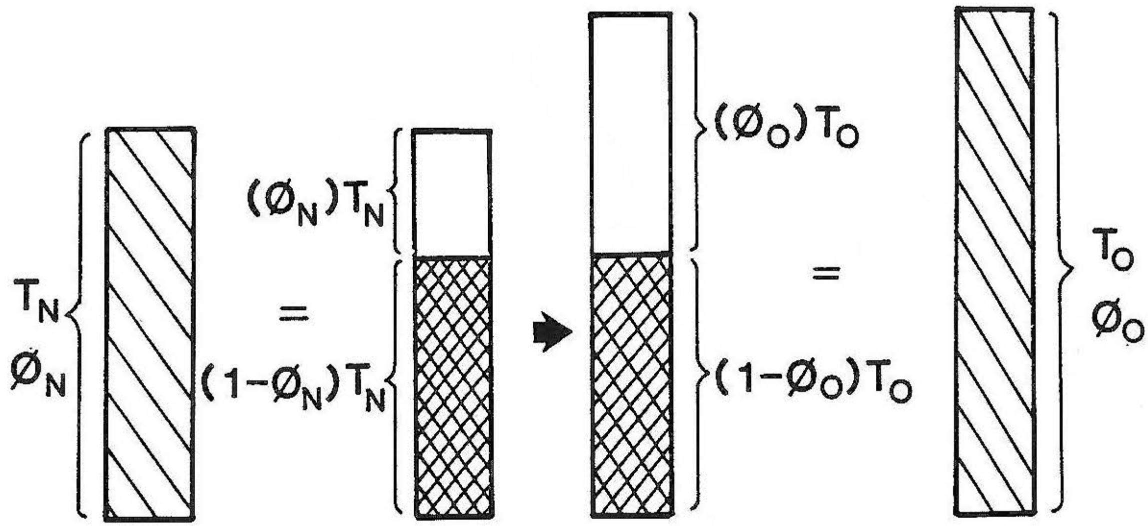


Figure 3.9 Stratigraphic accumulation over time for example data.



$$\text{since } (1-\O_N)T_N = (1-\O_O)T_O$$

$$T_O = \frac{(1-\O_N)T_N}{1-\O_O}$$

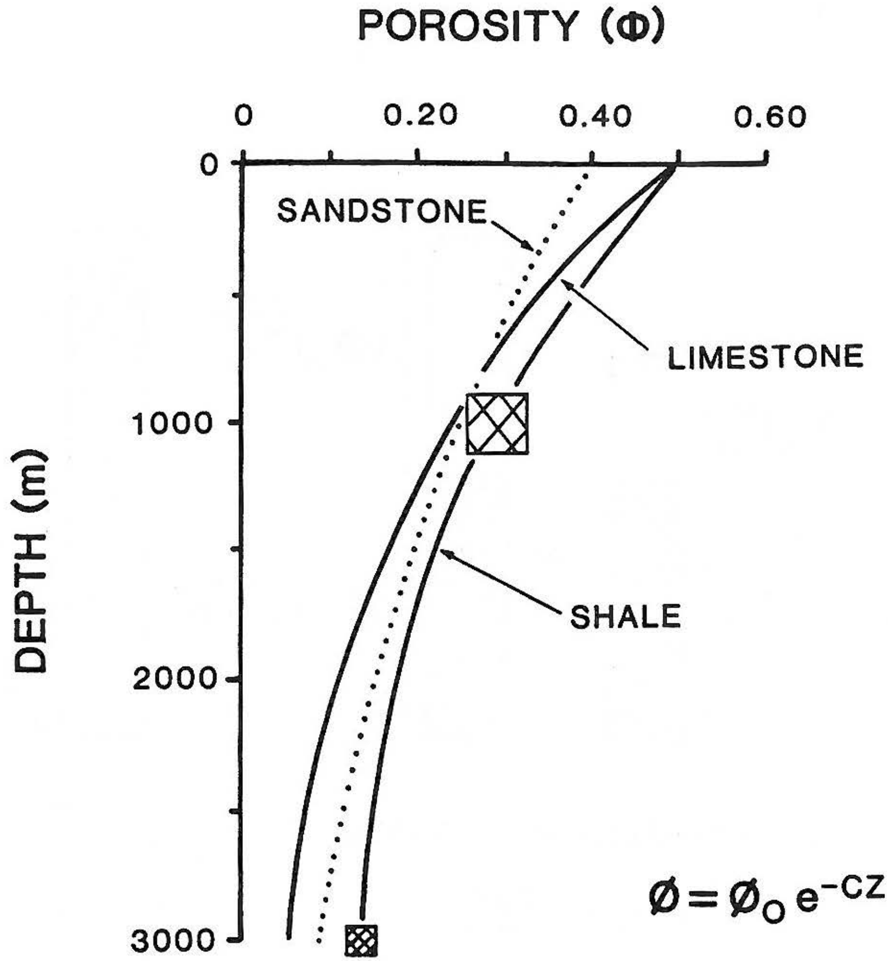
EG. $T_N = 100\text{ m}$ $\quad \O_O = 0.50$

$\O_N = 0.20$

$$(1-\O_N)T_N = 0.8(100) = 80 \text{ m of sediment}$$

$$T_O = \frac{(1-\O_N)T_N}{1-\O_O} = \frac{80}{0.5} = 160 \text{ m.}$$

Figure 3.10 Theory and example of compaction correction following the approach of Van Hinte (1978).



	SHALE	SANDSTONE	LIMESTONE
Φ_O	0.5	0.4	0.5
C m^{-1}	5.0×10^{-4}	3.0×10^{-4}	7.0×10^{-4}
ρ_g	2.72	2.65	2.71

Figure 3.11 Idealized porosity versus depth curves for different lithologies.

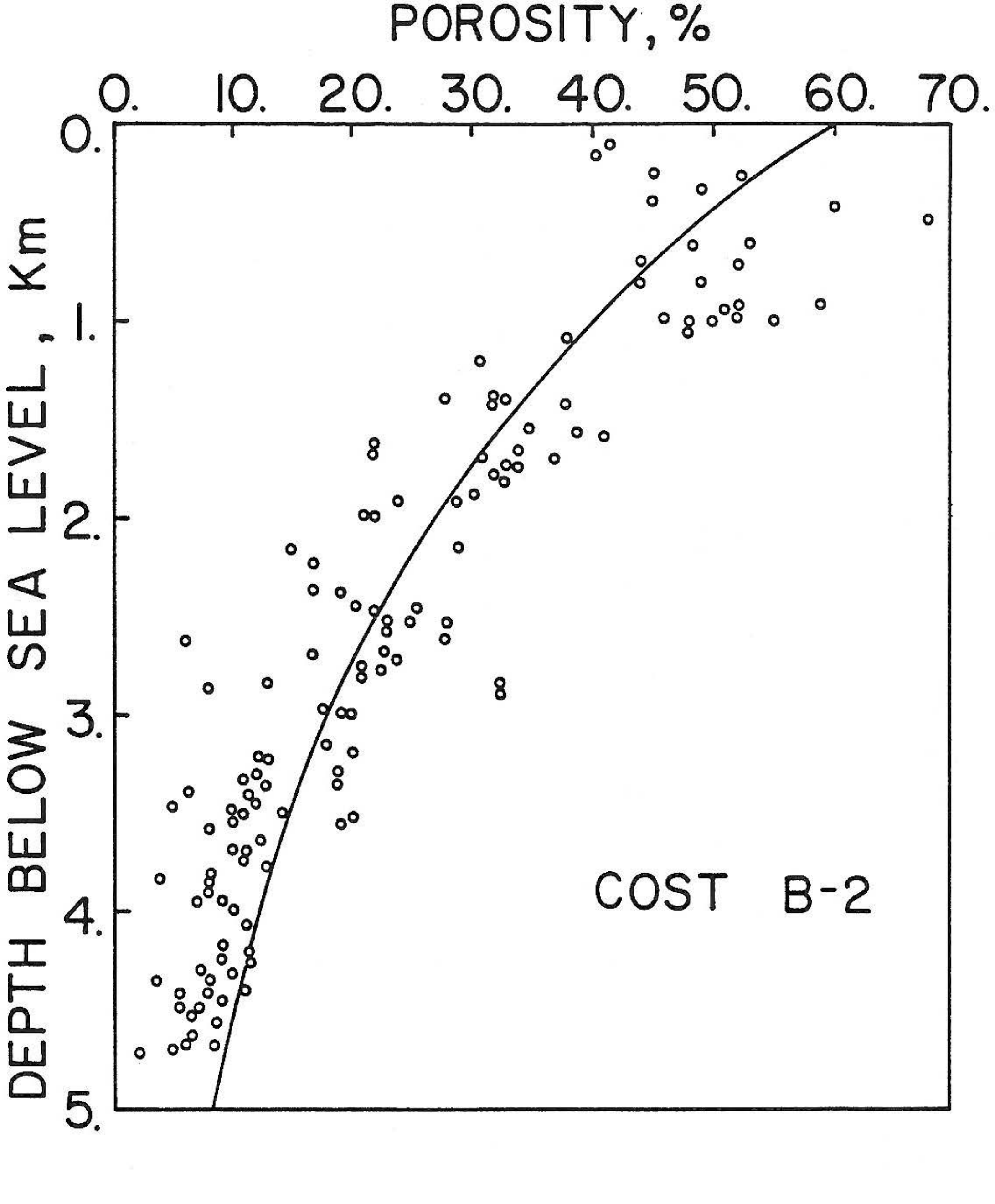


Figure 3.12 Porosity versus depth data from the COST B-2 well off the New Jersey coast. Modified form Steckler and Watts (1978).

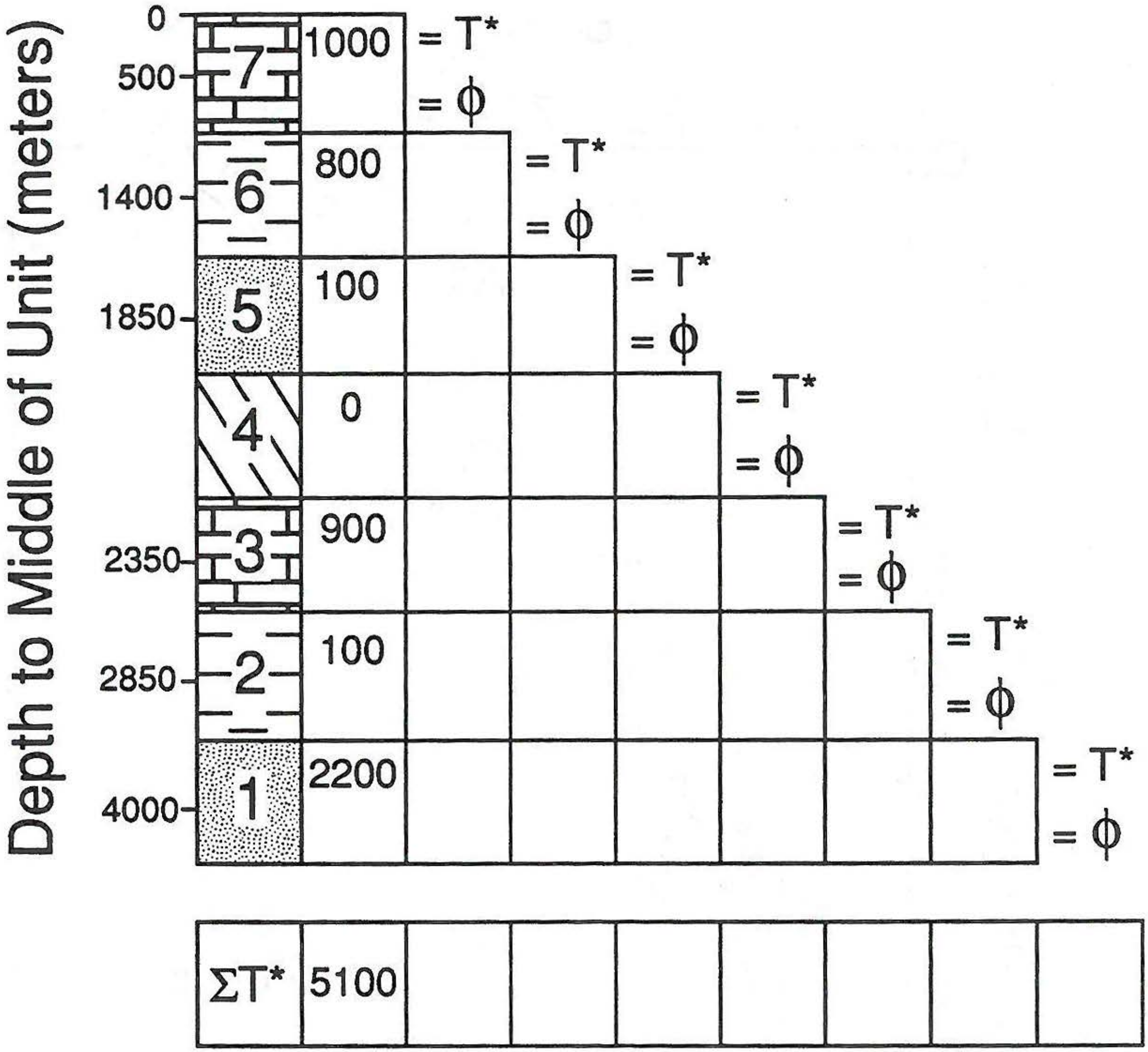
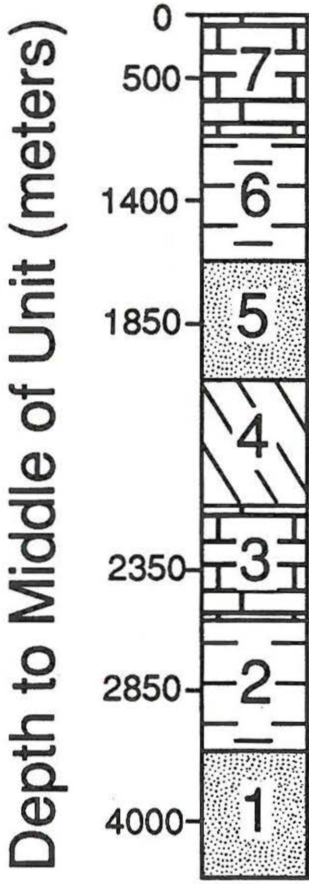


Figure 3.13 Table set up for making compaction corrections for the example problem.



Unit	Depth to Middle of Unit (m)								
7	500	1000							= T*
		35%							= ϕ
6	1400	800	1017						= T*
		25%	41%						= ϕ
5	1850	100	108	126					= T*
		23%	29%	39%					= ϕ
4		0	0	0	0				= T*
									= ϕ
3	2350	900	976	1209	1209	1209			= T*
		10%	17%	33%	33%	33%			= ϕ
2	2850	100	106	117	120	120	172		= T*
		12%	17%	25%	27%	27%	49%		= ϕ
1	4000	2200	2278	2361	2390	2390	2652	2652	= T*
		12%	15%	18%	19%	19%	27%	27%	= ϕ

ΣT*	5100	4485	3813	3719	3719	2824	2652	0

Figure 3.14 Completed table for making compaction corrections for the example problem.

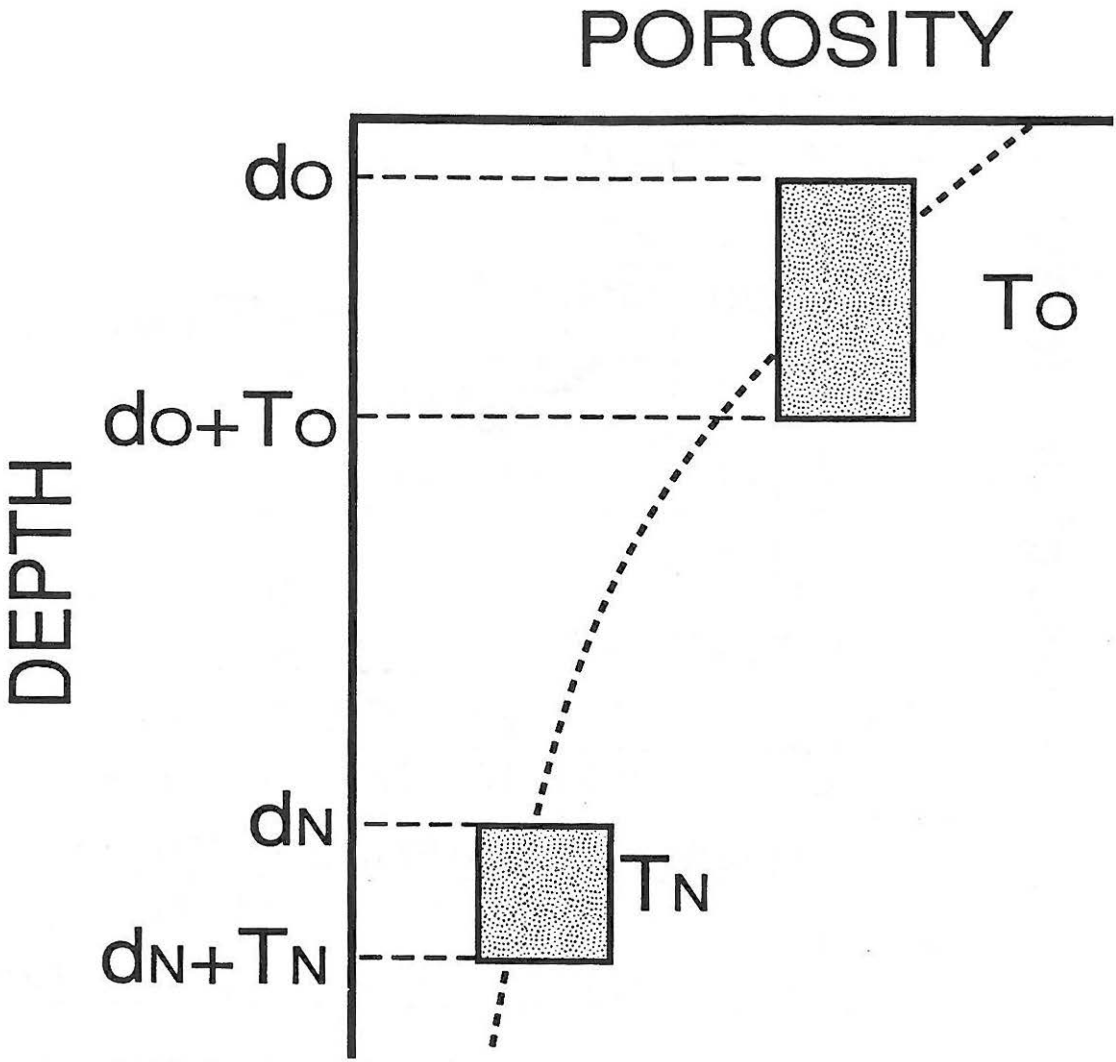


Figure 3.15 Example of the "correct" method for making compaction corrections.

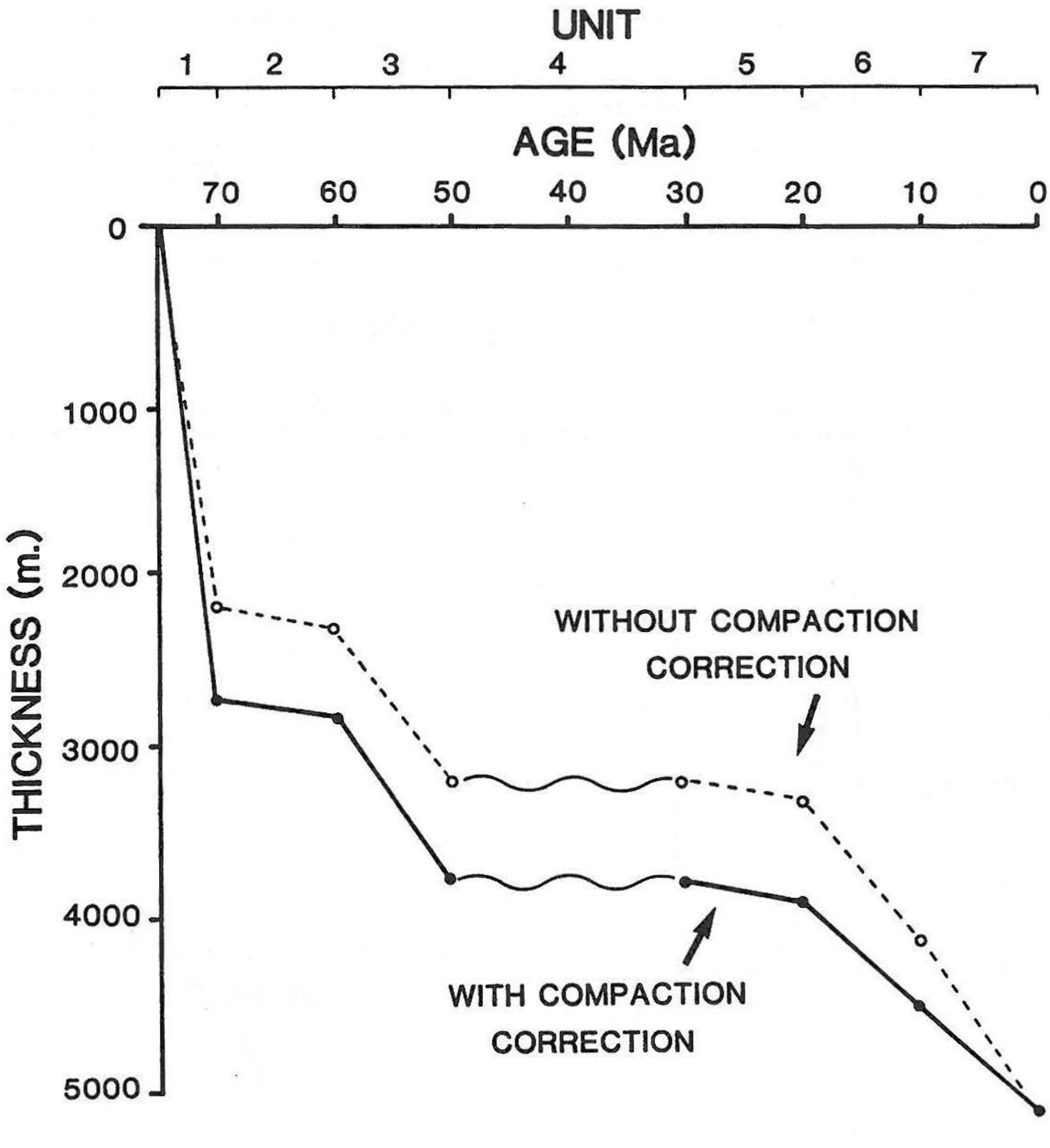


Figure 3.16 Subsidence curve corrected for compaction for example data.

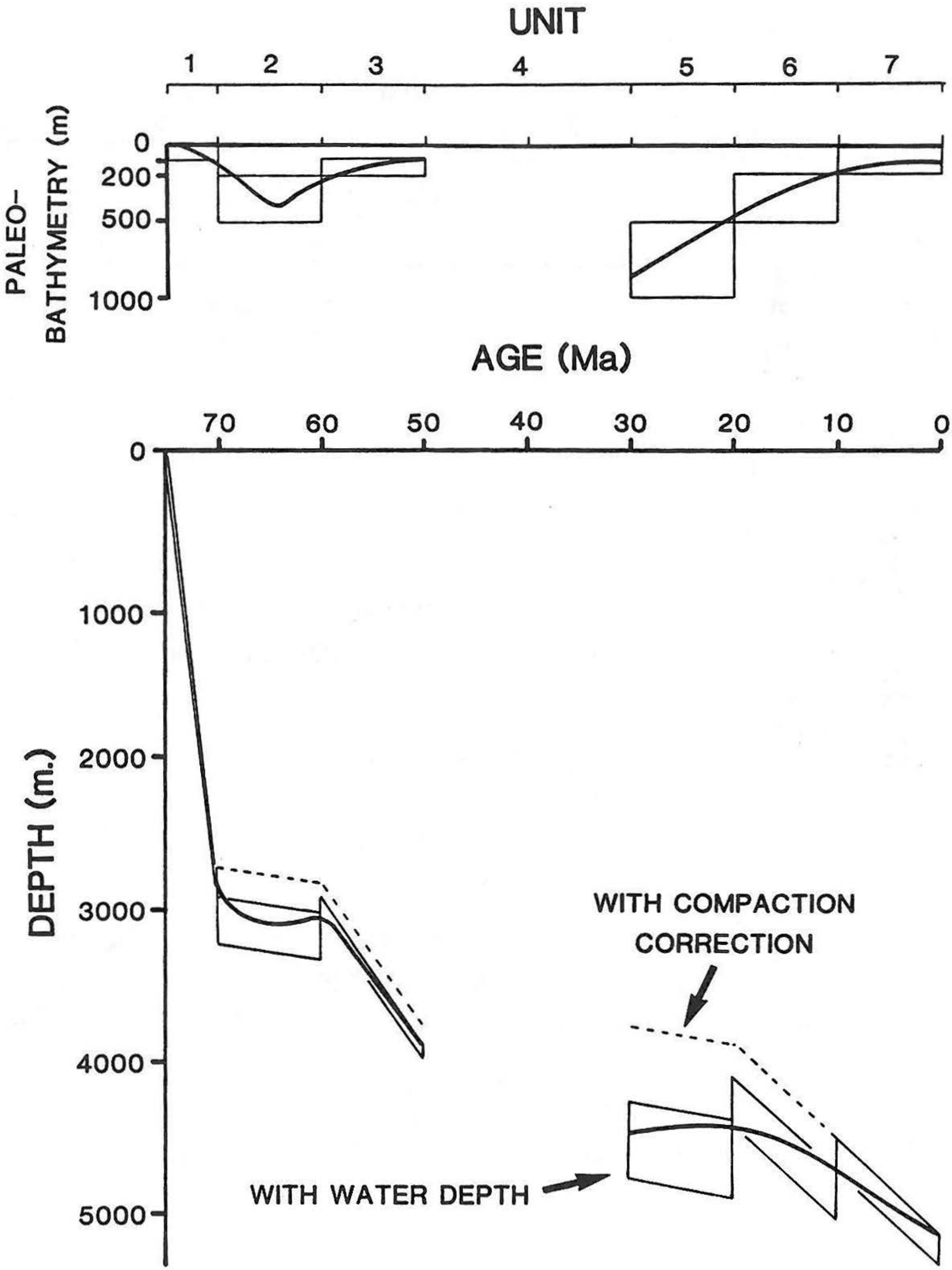
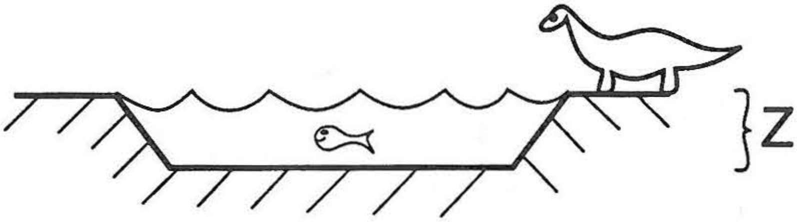
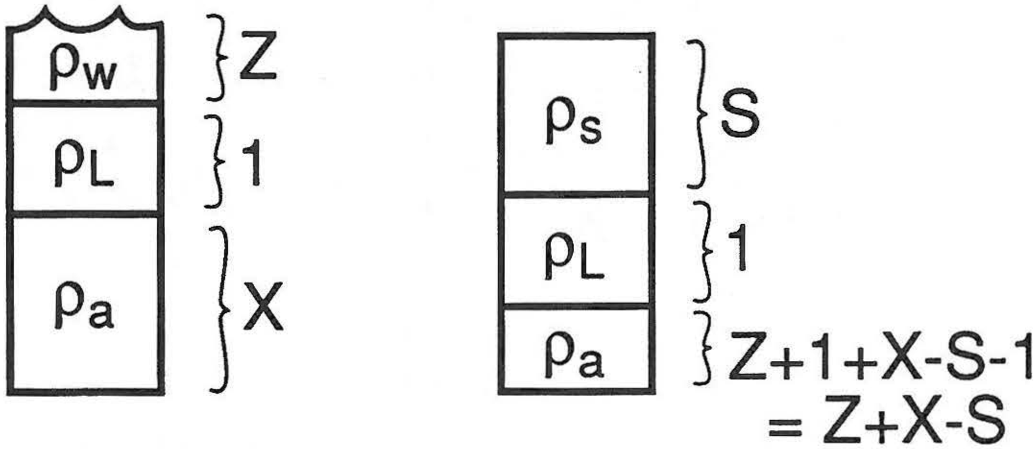


Figure 3.17 Total subsidence curve corrected for compaction and paleobathymetry for example data.





$$\rho_w Z + (1)\,\rho_L + \rho_a X = \rho_s S + (1)\,\rho_L + \rho_a (Z + X - S)$$

$$= \rho_s S + (1)\,\rho_L + \rho_a Z + \rho_a X - \rho_a S$$

SOLVE FOR S:

$$\rho_a S - \rho_s S = (1)\,\rho_L - (1)\,\rho_L + \rho_a Z - \rho_w Z + \rho_a X - \rho_a X$$

CANCEL AND SIMPLIFY:

$$(\rho_a - \rho_s)\, S = \rho_a Z - \rho_w Z$$

$$= (\rho_a - \rho_w)\, Z$$

$$S = \left(\frac{\rho_a - \rho_w}{\rho_a - \rho_s}\right) Z$$

If $\rho_a = 3.3\ \text{g/cm}^3$; $\rho_s = 2.3\ \text{g/cm}^3$; and $\rho_w = 1.0\ \text{g/cm}^3$

then $S = 2.3\ Z$

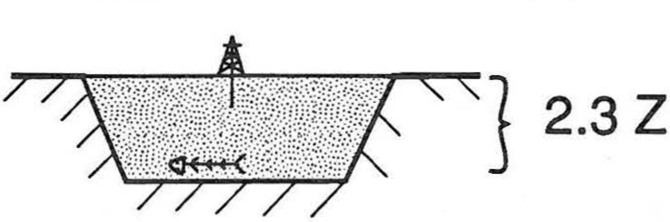
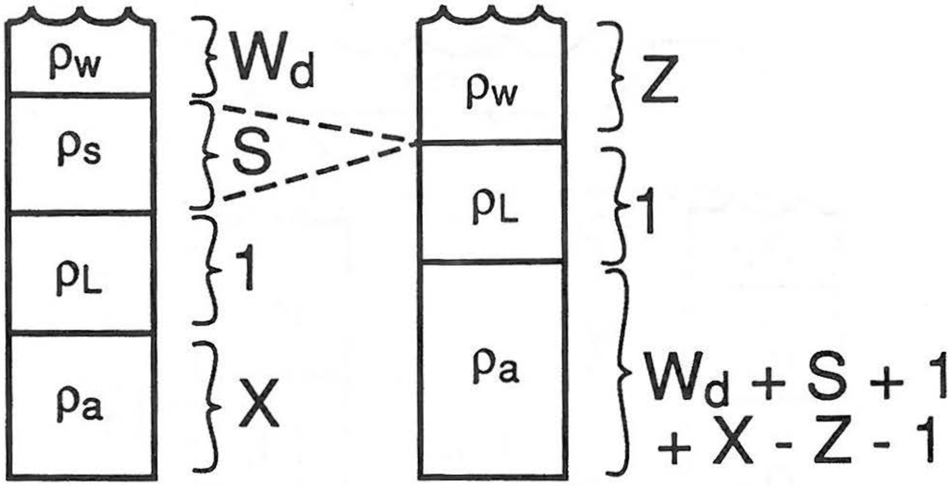


Figure 3.18 Derivation of isostasy equations, example for basin filling.



1. BALANCE COLUMNS:

$$\rho_w W_d + \rho_s S + \rho_L 1 + \rho_a X = \rho_w Z + \rho_L 1 + \rho_a (W_d + S + 1 + X - Z - 1)$$

$$= \rho_w Z + \rho_L 1 + \rho_a W_d + \rho_a S + \rho_a 1 + \rho_a X - \rho_a Z - \rho_a 1$$

$$= \rho_w Z + \rho_a W_d + \rho_a S + \rho_a X - \rho_a Z$$

2. LUMP Z TERMS:

$$\rho_a Z - \rho_w Z = \rho_a W_d - \rho_w W_d + \rho_a S - \rho_s S$$

3. FACTOR:

$$(\rho_a - \rho_w) Z = (\rho_a - \rho_w) W_d + (\rho_a - \rho_s) S$$

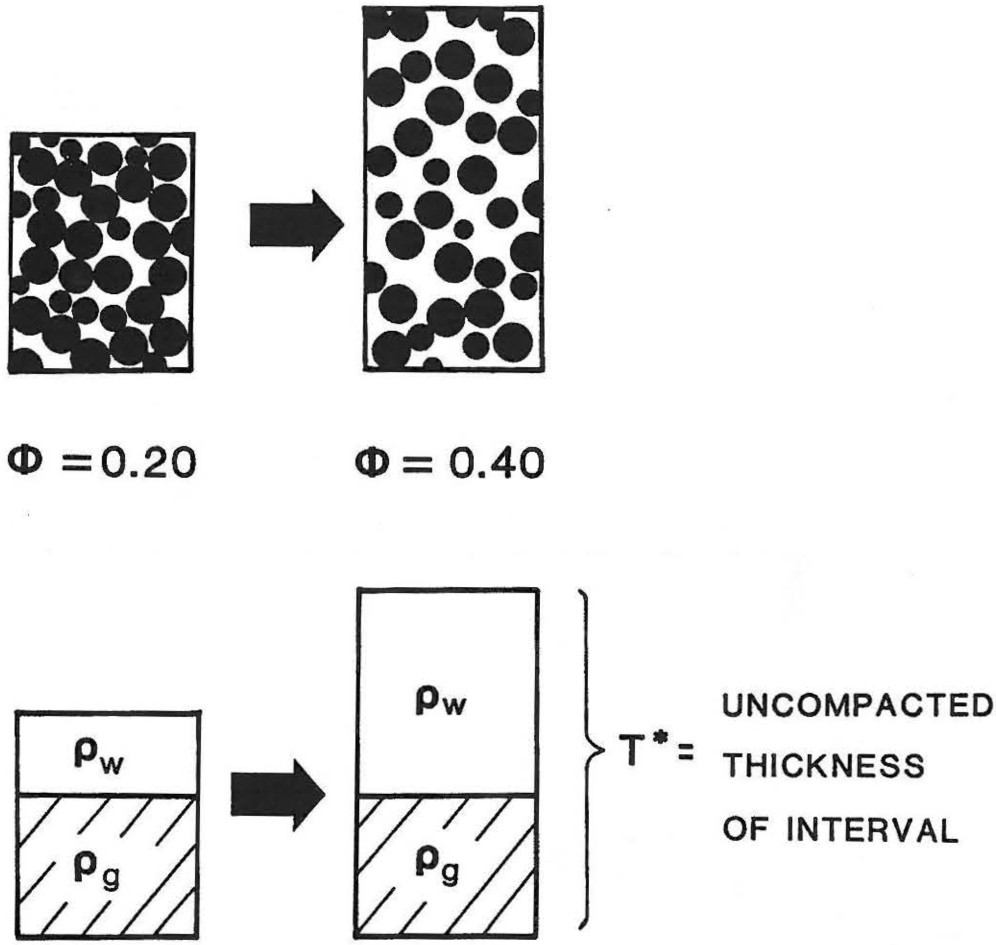
4. DIVIDE THROUGH BY $\rho_a - \rho_w$:

$$Z = \left(\frac{\rho_a - \rho_s}{\rho_a - \rho_w}\right) S + W_d$$

5. IF CHANGE OF SEA LEVEL (ΔSL) IS KNOWN, THEN:

$$Z = \left(\frac{\rho_a - \rho_s}{\rho_a - \rho_w}\right) S + W_d - \Delta SL \frac{\rho_a}{\rho_a - \rho_w}$$

Figure 3.19 Derivation of backstripping equations (after Steckler and Watts, 1978).



$$\rho_{s_i} = \Phi_i \rho_w + (1 - \Phi_i)\rho_g$$

IF WE WANT ρ_s OF AN ENTIRE COLUMN

AT A GIVEN POINT IN TIME (TIME i)

$$\rho_{s_i} = \frac{\Sigma_1^i \left[\Phi_i \rho_w + (1 - \Phi_i)\rho_g\right] T_i}{S^*}$$

WHERE $S^* = \Sigma_1^i \; T_i^*$

Figure 3.20 Calculating ρ_s of a stratigraphic column.

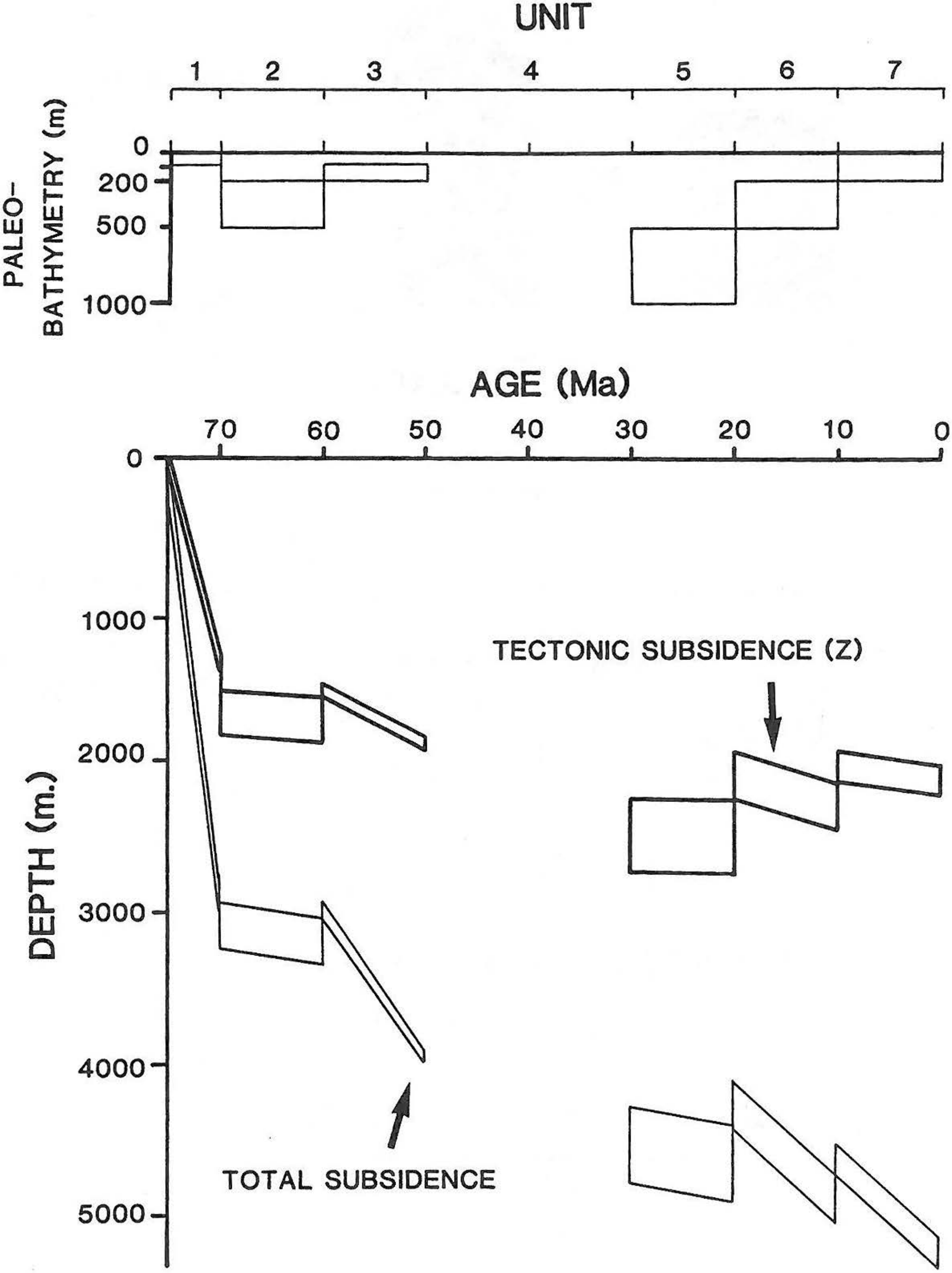


Figure 3.21 Tectonic subsidence versus time for the example.